Atmosphere, Climate & Environment
Information Programme, **aric**
Manchester Metropolitan University
Chester Street, Manchester M1 5GD
Tel: 0161 247 1590
Fax: 0161 247 6332
E-Mail: aric@mmu.ac.uk
Internet: **http://www.doc.mmu.ac.uk/aric/**

Global Climate Change Student Guide

A review of contemporary and prehistoric *global climate change*

Joe Buchdahl

1999

Contents

Page

Page

List of Figures

Page

List of Tables

Introduction

Climate is the long-term statistical expression of short-term weather. Climate can be defined qualitatively as "expected weather" (Bradley, 1985), or quantitatively by statistical expressions such as central tendencies and variances. The overall distribution of climatological parameters, bounded by weather extremes, defines the climatic variability. Changes in climate can be defined by the differences between average conditions at two separate times. Climate may vary in different ways and over different time scales. Variations may be periodic (and hence predictable), quasi-periodic or non-periodic (Hare, 1979).

This guide represents an up-to-date review of climate change. Throughout, the focus has essentially been on global climate change, although reference to regional scale climatic change has been made if and when necessary. On their own each chapter is a broadly self-contained discussion of a specific sub-issue of importance.

Chapters 1 and 2 review the climate system and the causes of climate change. These two chapters form the basis for understanding what climate change is, and how and why it occurs. Chapter 3 discusses the methods used to construct climatological time series from various instrumental data, and the reconstruction of palaeoclimates (past climates) from proxy data, whilst chapter 4 reviews the use of climate models in attempting to understand the climate system and climate change. Finally, chapters 5 and 6 are concerned with the topics of palaeoclimatology and contemporary climate change respectively.

The purpose of the guide is to serve as a first resort for undergraduates, applicable to first, second and third years in environmental sciences, geography and related disciplines. It is sufficient in depth and rigour for final year students but equally valuable as an information resource for those in their "pre-finals" years. It is not a complete guide but serves to review and illustrate the key factors of climate change over time and space. It contains an extensive list of references which the student is urged to consult.

The authors recognise that the study of global climate change is a rapidly developing discipline, and recent empirical evidence of climate change, and modelling evidence of the causes of climate change may have been omitted. In this sense, the guide represents a snapshot in time of the current understanding of climate change, which is open to future development.

The opportunity therefore exists for readers of this guide to comment upon its content, both regarding improvements to the existing material and new ideas for additional entries. Comments and responses regarding the Global Climate Change Student Guide should be sent to The Global Climate Change Information Programme at the address on the front cover.

Joe Buchdahl

1.1. Introduction

The key to understanding global climate change is to first understand what global climate is, and how it operates. This is the purpose of chapter 1.

The global climate system is a consequence of and a link between the atmosphere, oceans, the ice sheets (cryosphere), living organisms (biosphere) and the soils, sediments and rocks (geosphere). Only by consideration of the climate system in these terms is it possible to understand the flows and cycles of energy and matter in the atmosphere, an understanding which is required to investigate the causes (and effects) of climatic change.

Having stressed the interconnectedness of the elements that make up climate system, it might therefore seem inappropriate to divide a discussion on the climate system into separate sections, each dealing with a different component of the climate system. However, without this rationalisation, such a discussion would prove very difficult, in light of the huge complexity of the climate system. This discussion will therefore begin with the atmosphere and its energy budget or energy cycle, the balance of which ultimately controls the global climate. Following upon this, the other components of the climate system (the oceans, cryosphere, biosphere and geosphere) will be introduced, showing how each influence the atmospheric energy budget.

1.2. The Atmosphere

The atmosphere is a mixture of different gases and aerosols (suspended liquid and solid particles) collectively known as air which envelopes the Earth, forming an integrated environmental (climate) system with all the Earth's components. The atmosphere provides various functions, not least the ability to sustain life. Of prime interest for a discussion on climate change, however, is its ability to control the Earth's energy budget. To understand this process, it will be necessary to study in more detail the composition of the atmosphere.

1.2.1. The Composition of the Atmosphere

Let us consider first the atmospheric gases. Table 1.1 illustrates the average gaseous composition of dry air below 25km. Although traces of atmospheric gases have been detected well out into space, 99% of the mass of the atmosphere lies below about 25 to 30km altitude, whilst 50% is concentrated in the lowest 5km (less than the height of Mount Everest).

Table 1.1. *Average composition of the atmosphere below 25km*

Component	Chemical Abbreviation	Volume % (dry air)
Nitrogen	N_2	78.08
Oxygen	O_2	20.98
Argon‡	Ar	0.93
Carbon dioxide	CO_2	0.035
Neon‡	Ne	0.0018
Helium‡	He	0.0005
Hydrogen	H	0.00006
Krypton‡	Kr	0.0011
Xenon‡	Xe	0.00009
Methane	CH_4	0.0017
Ozone†	O_3	0.00006

† Strictly speaking, the concentration of ozone in the atmosphere is variable.
‡ Inert gases.

This gaseous mixture remains remarkably uniform in composition, and is the result of efficient biogeochemical recycling processes and turbulent mixing in the atmosphere. The two most abundant gases are nitrogen (78% by volume) and oxygen (21% by volume), and together they make up over 99% of the lower atmosphere. There is no evidence that the relative levels of these two gases are changing significantly over time (Kemp, 1994).

Despite their relative scarcity, the so-called greenhouse gases play an important role in the regulation of the atmosphere's energy budget.

1.2.1.1. Carbon dioxide

Carbon dioxide (CO_2), the most important of these minor gases, is involved in a complex global cycle (see section 6.4.1). It is released from the interior of the Earth via volcanic eruptions, and by respiration, soil processes, combustion of carbon compounds and oceanic evaporation. Conversely, it is dissolved in the oceans and consumed during plant photosynthesis. Currently, there are 359 parts per million by volume (ppmv) of CO_2 in the atmosphere (Schimel *et al.*, 1995), a concentration which is continuing to rise due to anthropogenic (man-made) emissions from the burning of fossil fuels and forests. The implications of this are discussed in chapter 6.

1.2.1.2. Methane

Methane (CH_4) is another greenhouse gas, and is produced primarily by anaerobic (oxygen-deficient) processes such as the cultivation of rice paddies or animal digestion. It is destroyed in the lower atmosphere (troposphere) by reactions with free hydroxyl radicals (OH):

$$CH_4 + OH \rightarrow CH_3 + H_2O$$

Like CO_2, its concentration in the atmosphere is increasing due to anthropogenic activities such as agricultural practices and landfills (Prather *et al*., 1995).

1.2.1.3. Nitrous oxide

Nitrous oxide (N_2O) is produced by both biological mechanisms in the oceans and soils, and by anthropogenic means including industrial combustion, vehicle exhausts, biomass burning and the use of chemical fertilisers. It is destroyed by photochemical reactions (involving sunlight) in the upper atmosphere (stratosphere) (Prather *et al*., 1995).

1.2.1.4. Ozone

Ozone (O_3) in the stratosphere filters out harmful ultra-violet radiation from the Sun, and so protects life on Earth. Recently, there have been fears about the destruction of the ozone layer, principally over the Antarctic (Farman *et al*., 1985; Hofmann *et al*., 1992), but increasingly now over the Arctic regions (Hofmann *et al*., 1991). The concentration of O_3 in the atmosphere is not uniform, unlike other trace gases, but varies according to altitude. O_3 is formed during a photochemical reaction (Chapmen, 1930) involving solar ultra-violet radiation, an oxygen molecule and an oxygen atom,

$$O_2 + O + M \rightarrow O_3 + M$$

where M represents the energy and momentum balance provided by collision with a third atom or molecule, for example oxides of nitrogen (NO_x). The destruction of O_3 involves the recombination with atomic oxygen, via the catalytic effect of agents such as OH radicals, NO_x and chlorine (Cl, ClO) radicals. The concentration of O_3 is determined by the finely balanced equilibrium of formation and natural destruction (Dotto & Schiff, 1978). Because the relative reaction rates of formation and destruction vary with temperature and pressure, so O_3 concentrations vary with altitude. Most of the ozone occurs in a layer between 15 to 35 km altitude (see Figure 1.1) where the relative reaction rates of formation and destruction rates are most conducive to O_3 formation. The present fear about ozone depletion is due to the increase in amount of agents (such as Cl) in the atmosphere that increase the destruction rate of O_3, so upsetting the delicate equilibrium that exists.

1.2.1.5. Halocarbons

Halocarbons are compounds containing carbon, halogens such as chlorine, bromine and fluorine, and sometimes hydrogen. They may be wholly anthropogenic such as the CFCs or they may have natural sources, such as some of the methylhalides (see section 6.4.4).

Chlorofluorocarbons (CFCs) are entirely anthropogenically produced by aerosol propellants, refrigerator coolants and air conditioners. They are made up of carbon, chlorine and fluorine molecules. CFCs are destroyed slowly by photochemical reactions in the upper atmosphere (stratosphere). CFCs were absent from the atmosphere before the 1930s, but over the last half century, their concentrations have steadily increased. Although their concentrations are measured in parts per trillion (by volume), they are seen as a significant threat to future global warming. They possess long atmospheric lifetimes measured in decades to centuries, and molecule for molecule, are thousands of time stronger as a greenhouse gas than CO_2 (IPCC, 1990a). Halons are similar anthropogenic species but contain bromine instead of chlorine.

Other halocarbons include the Hydrochlorofluorocarbons (HCFCs) and hydrofluorocarbons (HFCs). These are anthropogenic compounds currently being used as substitutes to replace CFCs which are being phased out under the terms of the Montreal Protocol, to protect the ozone layer.

Many of the other gases in the atmosphere are the noble or **inert gases**, including argon, neon, helium, krypton and xenon (see Table 1.1). Because of their chemical inertness, they play little or no part in the chemical and physical processes operating in the atmosphere.

1.2.1.6. Other Trace Gases

In addition to these gases, water vapour (H_2O) is a vital atmospheric constituent, averaging about 1% by volume, with significant variations across spatial and temporal scales. Its presence in the atmosphere forms part of the global hydrological cycle. Water vapour, being the most important natural greenhouse gas on account of its abundance, plays a crucial role in the regulation of the atmosphere's energy budget. Despite this, the total volume of water in the atmosphere is relatively small, and, if precipitated completely and evenly over the whole Earth, would yield only about 25mm or 1 inch of rainfall (Kemp, 1994). In reality, of course, rainfall distribution is highly uneven, due to the internal dynamic processes within the global climate system.

In addition to the gases in Table 1.1, there are other reactive gas species produced by cycles of sulphur (S), nitrogen (N_2) and chlorine (Cl) halogens. For a further discussion of these species, Wayne (1991) offers a useful text.

1.2.1.7. Aerosols

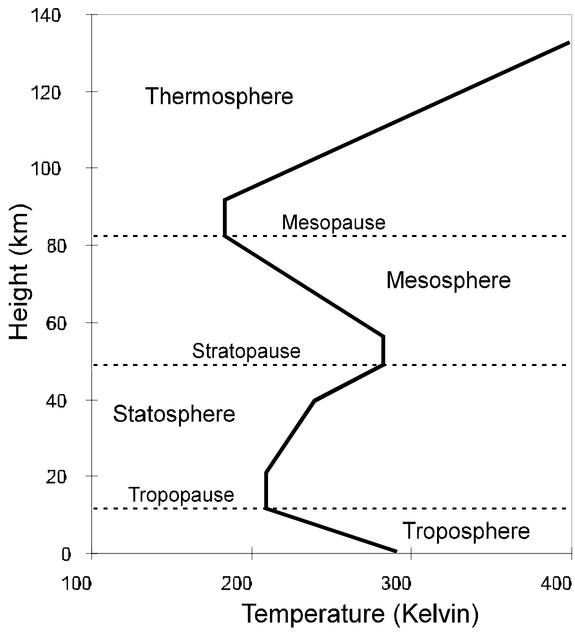
Variations in the abundance of the atmospheric greenhouse gases have the potential for changing the global climate. Variations in another group of species, namely the atmospheric aerosols, can also affect climate. Aerosols are solid or liquid particles dispersed in the air (Kemp, 1994), and include dust, soot, sea salt crystals, spores, bacteria, viruses and a plethora of other microscopic particles. Collectively, they are often regarded as air pollution, but many of the aerosols have a natural origin (Jonas *et al*., 1995). Although atmospheric turbidity (the abundance of aerosols) varies over short time scales, for example after a volcanic eruption (Sear *et al*., 1987; section 2.6.3), over the long term it maintains a fair degree of equilibrium, owing to the natural cleansing mechanisms of the Earth's climate system. Cleansing is never complete, however, and there always remains a background level of atmospheric aerosols which reflects the dynamic processes involved with aerosol input and aerosol removal. Natural sources of aerosols are probably 4 to 5 times larger than anthropogenic ones on a global scale (Barry & Chorley, 1992), but regional variations of anthropogenic emissions may change this ratio significantly in certain areas, particularly in the industrialised Northern Hemisphere.

1.2.2. The Vertical Structure of the Atmosphere

Most of the gaseous constituents are well mixed throughout the atmosphere. However, the atmosphere itself is not physically uniform but has significant variations in temperature and pressure with altitude. Figure 1.1. shows the structure of the atmosphere, in which a series of layers is defined by reversals of temperature. The lowest layer, often referred to as the lower atmosphere, is called the troposphere. It ranges in thickness from 8km at the poles to 16km over the equator, mainly as the result of the different energy budgets at these locations (section 1.2.5, Campbell, 1986; Lamb, 1982). Although variations do occur, the average decline in temperature with altitude (known as the lapse rate) is approximately 6.5^{o}C per kilometre. The troposphere contains up to 75% of the gaseous mass of the atmosphere, as well as nearly all of the water vapour and aerosols (Barry & Chorley, 1992), whilst 99% of the mass of the atmosphere lies within the lowest 30km.

Owing to the temperature structure of the troposphere, it is in this region of the atmosphere where most of the world's weather systems develop. These are partly driven by convective processes that are established as warm surface air (heated by the Earth's surface) expands and rises before it is cooled at higher levels in the troposphere.

Figure 1.1. *The Vertical Structure of the Atmosphere*



The tropopause (see Figure 1.1) marks the upper limit of the troposphere, above which temperatures remain constant before starting to rise again above about 20km. This temperature inversion prevents further convection of air, thus confining most of the world's weather to the troposphere.

The layer above the tropopause in which temperatures start to rise is known as the stratosphere. Throughout this layer, temperatures continue to rise to about an altitude of 50km, where the rarefied air may attain temperatures close to 0^{o}C. This rise in temperature is caused by the absorption of solar ultraviolet radiation by the ozone layer (see section 1.2.1). Such a temperature profile creates very stable conditions, and the stratosphere lacks the turbulence that is so prevalent in the troposphere.

The stratosphere is capped by the stratopause, another temperature inversion occurring at about 50km. Above this lies the mesosphere up to about 80km through which temperatures fall again to almost -100^{o}C. Above 80km temperatures rise continually (the thermosphere) to well beyond 1000^{o}C, although owing to the highly rarefied nature of the atmosphere at these heights, such values are not comparable to those of the troposphere or stratosphere.

1.2.3. Radiation Laws

The Earth's atmosphere has an important influence on the energy budget of the global climate system. This is determined by the thermodynamic processes involved in solar and terrestrial energy transfers.

The Earth's principal source of energy is the Sun, which produces electromagnetic radiation from nuclear fusion reactions involving hydrogen in its core. Radiation emitted from its surface has a temperature of approximately 5800 Kelvin (K). The radiation is emitted over a spectrum of wavelengths, with a specific quantity of energy for each wavelength, calculated using Planck's Law:

$$E_{\lambda} = a / [\lambda^{5} \{e^{(b/\lambda T)} - 1\}] \qquad \text{[Equation 1]}$$

where E_{λ} is the amount of energy ($Wm^{-2}\mu m^{-1}$) emitted at wavelength λ (μm) by a body at temperature T (K), with a and b as constants (Henderson-Sellers & Robinson, 1986). This assumes that the Sun is a perfectly radiating (black) body.

By differentiating Equation 1 (Planck's Law), it is possible to determine the wavelength of maximum radiation emission from the Sun:

$$\lambda = 2897 / T \qquad \text{[Equation 2]}$$

This is Wien's Law and for T = 5800K (the solar surface temperature) the wavelength of maximum energy is approximately 0.5μm. This represents radiation in the visible part of the spectrum.

By integrating Equation 1, one can determine the total energy emitted by the Sun, given by the Stefan-Boltzmann Law:

$$E_{Total} = \sigma T^{4} \qquad \text{[Equation 3]}$$

where σ is the Stefan-Boltzmann constant. Solving Equation 3 for the solar temperature of 5800K reveals a total energy output of about 64 million Wm^{-2}.

The solar radiation disperses uniformly in all directions. After travelling 93 million miles only a tiny fraction of the energy emitted by the Sun is intercepted by the Earth[1] (Critchfield, 1983). Therefore, the energy flux arriving at the top of the Earth's atmosphere is many orders of magnitude smaller than that leaving the Sun. The latest satellite measurements indicate a value of $1368Wm^{-2}$ for the energy received at the top of the atmosphere on a surface perpendicular to the solar beam. This is known as the solar constant.

Figure 1.2 shows the ideal solution to Planck's Law (Equation 1) for the spectrum of energy arriving at the top of the Earth's atmosphere. The high point of the curve represents the wavelength of greatest energy flux (0.5μm), as calculated by Wien's Law (Equation 2), whilst the area under the curve represents the total amount of energy received ($1368Wm^{-2}$), calculated by the Stefan-Boltzmann Law (Equation 3). 8% of the energy flux is in the ultraviolet part of the spectrum, whilst 39% is visible radiation.

Equations 1, 2 and 3 can be solved again for the Earth at a temperature of 255K (-18^{o}C), assuming the Earth to be a perfectly radiating (black) body. Since the Earth is much cooler than the Sun, its radiating energy is in the longer wavelength, invisible infrared part of the spectrum.

[1] The energy received is inversely proportional to the square of the solar distance (93 million miles or 150 million km)

Figure 1.2. *Spectral distribution of solar radiation reaching the Earth and terrestrial radiation leaving the Earth.*

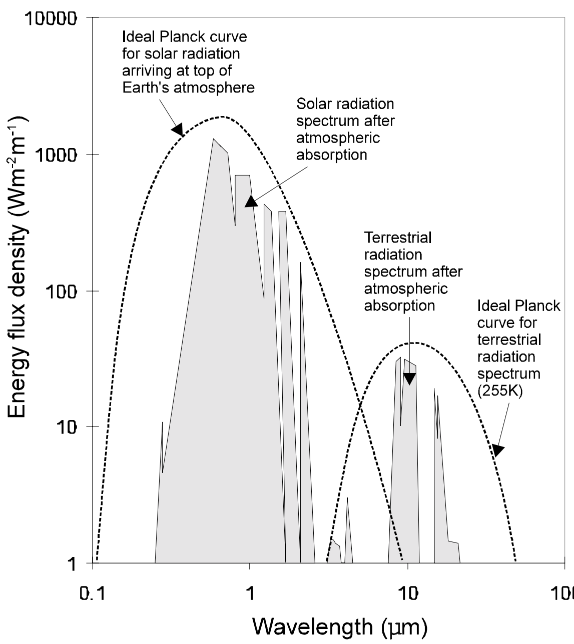


Figure 1.2 also shows a similar ideal Planck curve for the Earth at 255K. This would be the average global temperature for an Earth without an atmosphere, called the effective radiation temperature. It is the temperature at which the energy received by the Earth from the Sun balances the energy lost by the Earth back into space.

1.2.4. The Energy Budget of the Atmosphere

The Earth, however, does have an atmosphere (section 1.2.1) and this affects its energy balance. The average global temperature is, in fact, 288K or 15^{o}C, 33K warmer than the effective radiation temperature.

Although the Earth and Sun behave approximately as black bodies, this is not the case for the gases that make up the Earth's atmosphere. Certain atmospheric gases absorb radiation at some wavelengths but allow radiation at other wavelengths to pass through unimpeded.

Absorption of energy by a particular gas occurs when the frequency[2] of the electromagnetic radiation is similar to that of the molecular vibrational frequency of the gas in question. The atmosphere is mostly transparent (little absorption) in the visible part of the spectrum, but significant absorption of ultraviolet radiation (incoming short-wave solar radiation) by ozone, and infrared radiation (long-wave outgoing terrestrial radiation) by water vapour, carbon dioxide and other trace gases occurs. This is shown in Figure 1.2.

The absorption of terrestrial infrared radiation is particularly important to the energy budget of the Earth's atmosphere (Campbell, 1986). Such absorption by the trace gases heats the atmosphere, stimulating it to emit more long-wave radiation. Some of this is released into space (generally at higher, colder levels in the atmosphere) whilst most is re-radiated back to Earth. The net effect of this is that the Earth stores more energy near its surface than it would if there was no atmosphere, consequently the temperature is higher by about 33K.

This process is popularly known as the greenhouse effect. Glass in a greenhouse is transparent to solar radiation, but opaque to terrestrial infrared radiation. The glass acts like some of the atmospheric gases and absorbs the outgoing energy. Much of this energy is then re-emitted back into the greenhouse causing the temperature inside to rise. In reality, a greenhouse is warmer than its surroundings principally because of the shelter it offers rather than because of any radiative considerations. Nevertheless, the term has stuck, largely as a result of the media coverage.

Consequently, the gases in the atmosphere which absorb the outgoing infra-red radiation are known as greenhouse gases and include carbon dioxide, water vapour, nitrous oxide, methane and ozone. All the gases have molecules whose vibrational frequency lies in the infrared part of the spectrum. Despite the considerable absorption by these greenhouse gases, there is an atmospheric window through which terrestrial infrared radiation can pass (Kemp, 1994). This occurs at about 8 to 13μm, and its gradual closing is one of the effects of anthropogenic emissions of greenhouse gases (chapter 6).

As well as absorbing solar and terrestrial radiation, gases in the atmosphere, along with aerosols (see section 1.2.1), also scatter radiation. Of principal importance is the scattering of the incoming solar radiation, because this, too, can alter the overall energy budget of the atmosphere. Scattering occurs

[2] The frequency of electromagnetic radiation is related to its wavelength, λ by the equation, $c = f\lambda$, where f is the frequency of the radiation and c is the speed of light (300,000kms^{-1}).

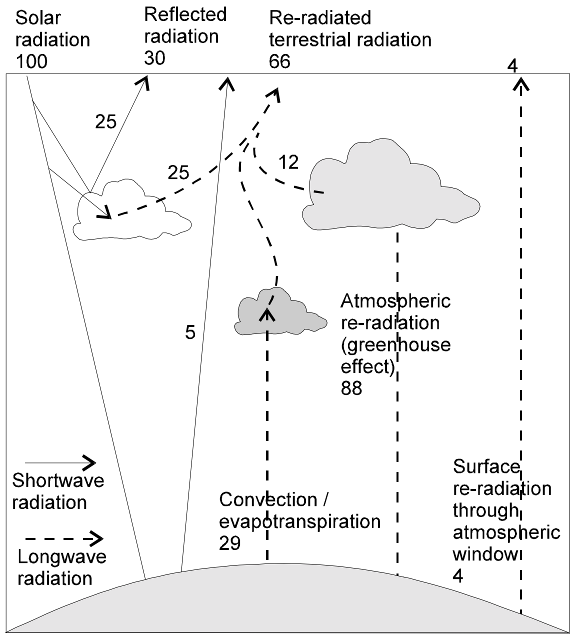
when a photon[3] impinges on an obstacle without being absorbed. Scattering changes only the direction of travel of that photon. Gas molecules, with small sizes relative to the wavelength of the incident radiation cause scattering in all directions, both forwards and backwards, known as Rayleigh scattering. Aerosols whose size is comparable to the incident radiation cause Mie scattering, which is mostly forward in direction. It can be seen that changes in atmospheric aerosol content could affect the energy budget (Shine *et al*., 1995), having implications for the state of the global climate. This is returned to in section 2.6.5.

Figure 1.3 summarises schematically the global energy transfers that have been discussed. Energy arriving at the top of the atmosphere starts an energy cascade involving numerous energy transformations. On entering the atmosphere, some of the solar (short-wave) radiation is absorbed by gases in the atmosphere (eg. ozone), some is scattered, some is absorbed by the Earth's surface and some is reflected directly back into space by either clouds or the surface itself. The amount of short-wave radiation reflected depends on a factor known as the albedo (or reflectivity). Albedo varies according to the surface. Ice and certain clouds have a high albedo (0.6 to 0.9) whilst the oceans generally have a low albedo (0.1) For the whole Earth this averages at about 0.30, meaning that 30% of the incoming solar radiation is reflected.

Of the terrestrial (long-wave) radiation re-emitted from the Earth's surface, most is re-absorbed by the greenhouse gases and only a little escapes directly through the atmospheric window. Long-wave radiation re-emitted from the atmosphere (greenhouse gases, clouds) is either returned to the Earth's surface or released into space. The net result of this greenhouse effect is to increase the amount of energy stored near the Earth's surface, with a consequent increase in temperature. There are also additional heat fluxes associated with evaporation and transpiration which balance the energy fluxes into and out of all parts of the Earth-atmosphere system.

[3] Electromagnetic radiation of all wavelengths is made up of photons, massless particles which, mathematically, behave more like waves.

Figure 1.3. *The Earth-Atmosphere Energy Balance*



1.2.5. Horizontal Energy Transfers

Figure 1.3 illustrates how the energy transfers of the Earth-atmosphere system are in equilibrium. On a global scale and over a time period of several days and more this assumption is valid, and is adequate for an understanding of the causes of climate change covered in chapter 2. However, the real world is more complex than this.

If energy fluxes are calculated for different areas around the globe, one finds that between about 40^{o}N and 35^{o}S the incoming solar radiation is greater than the outgoing terrestrial radiation. Elsewhere (i.e. nearer the poles), there is a net radiation deficit, that is, more radiation is lost than received (Trewartha & Horn, 1980; Figure 1.4).

To restore equilibrium to this balance a meridional interchange of heat exists from the tropics to the poles (Figure 1.5). If this energy transfer did not occur, the equator would be 14^{o}C warmer on average than now, whilst the North Pole would be 25^{o}C colder (Barry & Chorley, 1992). This latitudinal transfer of energy occurs in several ways, involving the movement of sensible heat (convection processes caused by heating, rising and dispersion of surface air), latent heat (evapotranspiration processes involving evaporation of water vapour from the oceans and transpiration from land plants) and ocean currents (section 1.3).

Figure 1.4. *Net Latitudinal Radiation Balance (after Barry & Chorley, 1992)*

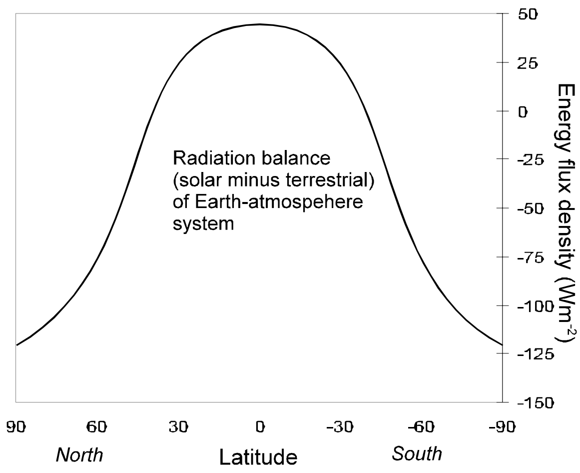
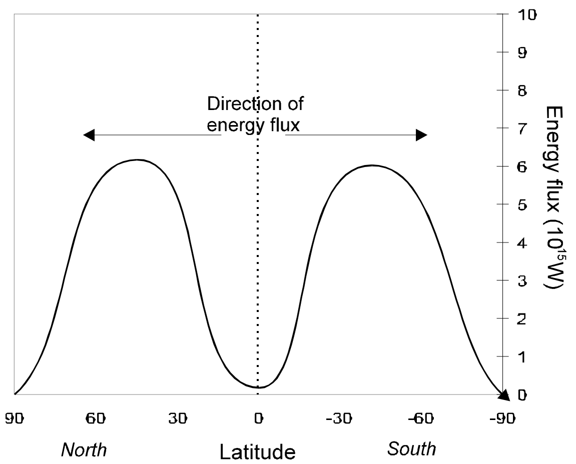


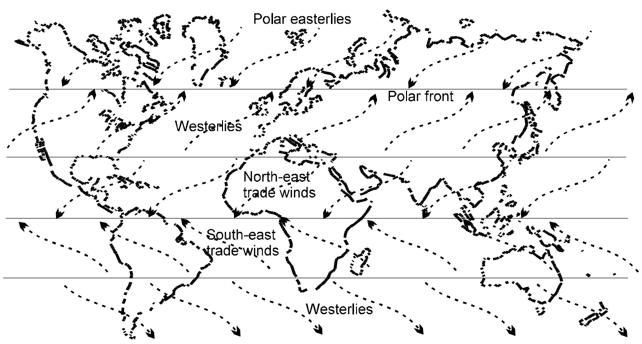
Figure 1.5. *Poleward Latitudinal Energy Flux*



As well as this movement of heat energy, there are other transfers which occur and must be balanced according to thermodynamic and physical principles. These include the transfer and balance of mass, momentum and moisture. The movement of heat involves the movement of air (sensible[4] heat), and moisture or water vapour (evapotranspiration). If a packet of air moves from the equator to the poles, this air must be replaced by colder air returning from the poles, having released its heat. In other words, the fluxes of air masses at specific locations around the Earth must be in equilibrium. By similar reasoning, both moisture and momentum fluxes must balance (Cubasch & Cess, 1990).

This picture is further complicated by the rotation of the Earth, which introduces a Coriolis Force on the moving atmosphere, and the axial tilt of the Earth, which affects the seasonal and latitudinal distribution of solar radiation. However, these phenomena are more usually covered in the study of meteorology, which is beyond the scope of this guide. Barry & Chorley (1992) provide a more detailed discussion for those interested. Figure 1.6 schematises a rudimentary Earth atmospheric circulation.

Figure 1.6. *Simple Atmospheric Circulation*



1.2.6. Summary

It is the fluxes principally of energy but also of moisture, momentum and mass, which determine the state of our climate. Factors which influence these on a global scale may be regarded as causes of global climate change. So far, however, only the fluxes in and out of and within the atmosphere have been considered. In the introduction, it was highlighted that the atmosphere forms just one, albeit, major component of the climate system. Before looking at the causes of global climate change then, it is worth devoting a little time to the other components of the climate system (the oceans, cryosphere, biosphere and geosphere), and how the fluxes of energy, moisture momentum and mass operate between them.

[4] Transfer of energy occurs in several ways: radiation, reflection, transmission and convection-advection. Most of the preceding discussion dealt with the radiation of energy (in the form of electro-magnetic waves) from both the Sun and the Earth. Energy may also be transferred by the movement of the agent which contains it. Air heated may rise (convection), owing to its lower density than cooler air, and disperse laterally (advection). Convection-advection of air transfers energy, this energy being called **sensible heat**. Oceans similarly transfer sensible heat.

1.3. Other Components of the Climate System

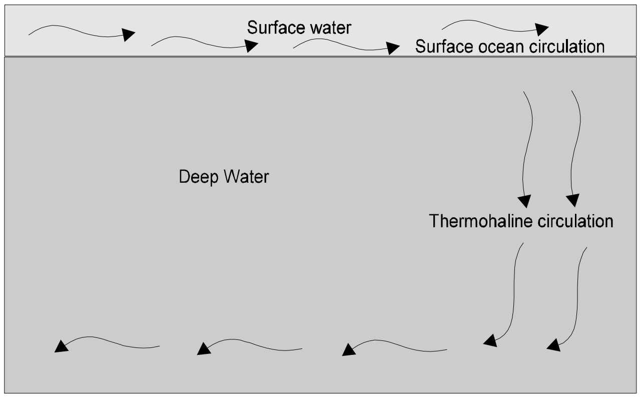
1.3.1. The Oceans

In section 1.2.5 it became clear that the atmosphere does not respond as an isolated system. Like the atmosphere, the thermodynamic state of the oceans is determined by the transfer of heat, momentum and moisture to and from the atmosphere. Ignoring for the moment the other components of the climate system, these fluxes within this coupled ocean-atmosphere system exist in equilibrium.

Momentum is transferred to the oceans by surface winds, mobilising the global surface ocean currents (Cubasch & Cess, 1990). Surface ocean currents assist in the latitudinal transfer of sensible heat in a similar fashion to the process occurring in the atmosphere. Warm water moves poleward whilst cold water returns towards the equator. Energy is also transferred via moisture. Water evaporating from the surface of the oceans stores latent heat which is subsequently released when the vapour condenses to form clouds and precipitation.

The significance of the ocean is that it stores a much greater quantity of energy than the atmosphere. This is on account of both its larger heat capacity (4.2 times that of the atmosphere) and its much greater density (1000 times that of air). The vertical structure of the ocean (Figure 1.7) can be divided into two layers which differ in the scale of their interaction with the overlying atmosphere. The lower layer comprises the cold deep water sphere, making up 80% of the oceans' volume. The upper layer, which has closest contact with the atmosphere, is the seasonal boundary layer, a mixed water sphere extending down only 100m in the tropics but several kilometres in polar regions. The seasonal boundary layer alone stores approximately 30 times as much heat as the atmosphere (Henderson-Sellers & Robinson, 1986). Thus for a given change in heat content of the ocean-atmosphere system, the temperature change in the atmosphere will be around 30 times greater than that in the ocean. Clearly then, small changes to the energy content of the oceans could have considerable effects on global climate.

Figure 1.7. *Vertical Structure and Circulation of the Oceans*



Energy exchanges also occur vertically within the oceans, between the mixed boundary layer and the deep water sphere (Figure 1.7). Sea salt remains in the water during the formation of sea ice in the polar regions, with the effect of increased salinity of the ocean. This cold, saline water is particularly dense and sinks, transporting with it a considerable quantity of energy. To maintain the equilibrium of water (mass) fluxes, a global thermohaline[5] circulation exists, which plays an important role in the regulation of the global climate. Broecker & Denton (1990) have proposed that changes in this thermohaline circulation influence climate changes over millennia time scales (chapter 5).

1.3.2. The Cryosphere

The cryosphere consists of those regions of the globe, both land and sea, covered by snow and ice. These include Antarctica, the Arctic Ocean, Greenland, Northern Canada, Northern Siberia and most of the high mountain ranges throughout the world, where sub-zero temperatures persist throughout the year. The cryosphere plays another important role in the regulation of the global climate system.

Snow and ice have a high albedo[6] (reflectivity), that is they reflect much of the solar radiation they receive.

[5] Driven by stratification of temperature and salinity.

[6] The Albedo of a surface is a measure of its reflectivity, i.e. how much solar radiation is reflected. A value of 1 defines a perfectly reflective surface. White surfaces such as snow and ice have high albedos of the order of 0.8 to 0.9. A value of 0 defines a perfectly absorbing surface. Dark surfaces such as forest and surface ocean have low albedos (0.1 to 0.3).

Some parts of the Antarctic reflect as much as 90% of the incoming solar radiation, compared to a global average of 31% (see section 1.2.4). Without the cryosphere, the global albedo would be considerably lower. More energy would be absorbed at the Earth's surface rather than reflected, and consequently the temperature of the atmosphere would be higher. Indeed, during the Cretaceous Period (120 to 65 million years ago) evidence suggests there was little or no snow and ice cover, even at the poles, and global temperatures were at least 8 to $10^{o}C$ warmer than today (Frakes, 1979; chapter 5).

The cryosphere also acts to decouple the atmosphere and oceans, reducing the transfer of moisture and momentum, so stabilising the energy transfers within the atmosphere (Henderson-Sellers & Robinson). The formation of sea ice in polar regions (section 1.3.1) can initiate global thermohaline circulation patterns in the oceans, which greatly influence the global climate system. Finally, the presence of the cryosphere itself markedly affects the volume of the oceans and global sea levels, changes to which can affect the energy budget of the climate system.

1.3.3. The Biosphere

Life may be found in almost any environment existing on Earth. Nevertheless, in a discussion on the climate system, it is convenient to regard the biosphere as a discrete component, like the atmosphere, oceans and cryosphere.

The biosphere, both on land and in the oceans, affects the albedo of the Earth's surface. Large areas of continental forest have relatively low albedos compared to barren regions such as deserts. The albedo of deciduous forests is about 0.15 to 0.18 whilst that of coniferous forests is 0.09 to 0.15 (Barry & Chorley, 1992). Tropical rainforest reflects even less energy, approximately 7 to 15% of that which it receives. In comparison, the albedo of a sandy desert is about 0.3. Clearly, the presence of the continental forests affect the energy budget of the climate system.

The biosphere also influences the fluxes of certain greenhouse gases such as carbon dioxide and methane. Plankton in the surface oceans utilise the dissolved carbon dioxide for photosynthesis. This establishes a flux of carbon dioxide, with the oceans effectively "sucking" down the gas from the atmosphere. On death, the plankton sink, transporting the carbon dioxide to the deep ocean. Such primary productivity reduces by at least four-fold the atmospheric concentration of carbon dioxide (Broecker, 1982), weakening significantly the Earth's natural greenhouse effect.

The biosphere also influences the amount of aerosols in the atmosphere. Millions of spores, viruses, bacteria, pollen and other minute organic species are transported into the atmosphere by winds, where they can scatter incoming solar radiation, and so influence the global energy budget (see section 1.2.4). Primary productivity in the oceans results in the emission of compounds known as dimethyl sulphides (DMSs). In the atmosphere these compounds oxidise to form sulphate aerosols called marine non-sea-salt (nss) sulphate (Charlson *et al*., 1987). These nss sulphates act as condensation nuclei for water vapour in the atmosphere, thus allowing the formation of clouds. Clouds have a highly complex effect on the energy budget of the climate system (see section 2.7). Thus changes in primary productivity in the oceans can affect, indirectly, the global climate system.

There are, of course, many other mechanisms and processes which couple the biosphere with the rest of the climate system, but the discussion has illustrated the major influences of the biosphere upon the global climate system.

1.3.4. The Geosphere

The fifth and final component of the global climate system is the geosphere, consisting of the soils, the sediments and rocks of the Earth's land masses, the continental and oceanic crust and ultimately the interior of the Earth itself. These parts of the geosphere each play a role in the regulation and variation of global climate, to a greater or lesser extent, over varying time scales.

Variations in global climate over tens of millions or even hundreds of millions of years are due to modulations within the interior of the Earth (Pickering & Owen, 1994; Raymo & Ruddiman, 1992; Ruddiman & Kutzbach, 1991). Changes in the shape of ocean basins and the size of continental mountain chains (driven by plate tectonic processes) may influence the energy transfers within and between the coupled components of the climate system.

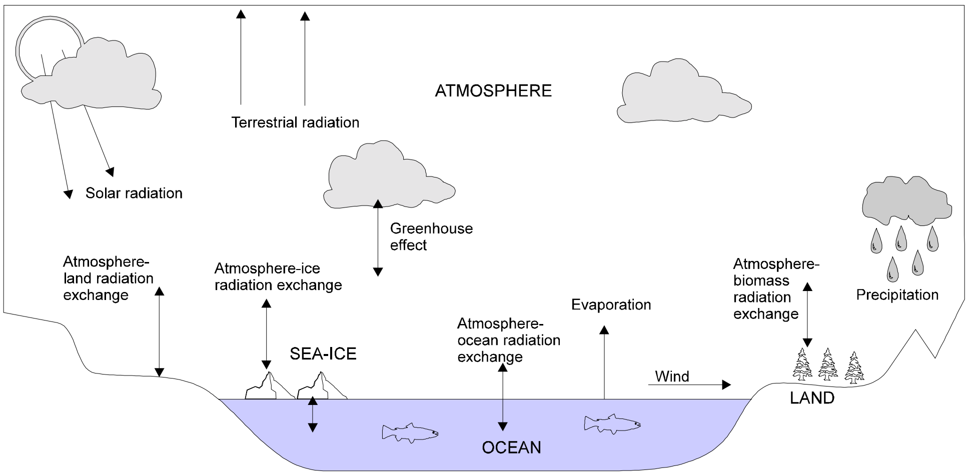
On much shorter time scales physical and chemical processes affect certain characteristics of the soil, such as moisture availability and water run-off, and the fluxes of greenhouse gases and aerosols into the atmosphere and oceans (Cubasch & Cess, 1990; McBean & McCarthy, 1990). Volcanism, although driven by the slow movement of the tectonic plates, occurs regularly on much shorter timescales. Volcanic eruptions replenish the carbon dioxide in the atmosphere, removed by the biosphere, and emit considerable quantities of dust and aerosols (see section 2.6.3). Volcanic activity can therefore affect the energy budget and regulation of the global climate system (Sear *et al.*, 1987).

1.4. Conclusion

The overall state of the of the global climate is determined by the balance of solar and terrestrial radiation budgets (see figure 1.3). How this energy balance is regulated depends upon the fluxes of energy, moisture, mass and momentum within global climate system, made up of its 5 components, the atmosphere, the oceans, the cryosphere, the biosphere and the geosphere. This is schematised in Figure 1.8.

Arguably there is a sixth component, an anthropogenic system, mankind. In the last 200 years, through increased utilisation of the world's resources, humans have begun to influence the global climate system, primarily by increasing the Earth's natural greenhouse effect. Chapter 6 reviews the subject of contemporary climate change. The next chapter examines some causes of climate change, with reference to the global climate system discussed in this chapter.

Figure 1.8. *The Global Climate System and its Energy Transfers*



2.1. Introduction

The global climate must be viewed as operating within a complex atmosphere/earth/ocean/ice/land system. Any change to this system, resulting in climate change, is produced by forcing agents - the causes of climate change. Such forcing agents may be either internal or external. External forcing mechanisms involve agents acting from outside the climate system. By contrast, internal mechanisms operate within the climate system itself. These are discussed separately in sections 2.5 and 2.6. In addition, forcing mechanisms may be non-radiative or radiative.

2.2. Non-Radiative Forcing

Any change in the climate must involve some form of energy redistribution within the global climate system. Nevertheless, forcing agents which do not directly affect the energy budget of the atmosphere (the balance between incoming solar radiation and outgoing terrestrial radiation (see Figure 1.3), are considered to be non-radiative mechanisms of global climate change. Such agents usually operate over vast time scales (10^7 to 10^9 years) and mainly include those which affect the climate through their influence over the geometry of the Earth's surface, such as location and size of mountain ranges and position of the ocean basins.

2.3. Radiative Forcing

A process which alters the energy balance of the Earth-atmosphere system (see Figure 1.3) is known as a radiative forcing mechanism (Shine *et al*., 1990). These may include variations in the Earth's orbit around the Sun, solar radiation, volcanic activity and atmospheric composition. Associating a particular cause with a particular change, however, is extremely difficult. The interlinked nature of the climate system ensures that there are feedbacks; a change in one component leads to a change in most, if not all, other components. The concept of feedback is discussed more thoroughly in section 2.7.

Before investigating some of the more important forcing mechanisms, both internal and external, there is one factor that needs elaborating: time scale.

2.4. Time Scale of Climatic Change

The importance of considering different time scales when investigating climate change has already been identified. Climate varies on all time scales (Mitchell, 1976), in response to random and periodic forcing factors. Across all time periods from a few years to hundreds of millions of years there is a white (background) noise of random variations of the climate, caused by internal processes and associated feedback mechanisms, often referred to as stochastic or random mechanisms (Goodess *et al*, 1992). Such randomness accounts for much of the climate variation, and owes its existence to the complex and chaotic behaviour of the climate system in responding to forcing (Lorenz, 1991; Nicolis & Nicolis, 1984; Palmer, 1989). An essential corollary of the existence of random processes is that a large proportion of climate variation cannot be predicted.

Of far more relevance are the periodic forcing factors, for by understanding their mechanisms and the impacts they have on the global climate, it is possible to predict future climate change. How the climate system responds to periodic forcing factors, however, is often not clear. If it is assumed that the climate system responds in a linear fashion to periodic forcing, variations in climate will exhibit similar periodicity. If, however, the response of the system to forcing is strongly non-linear, the periodicities in the response will not necessarily be identical to the periodicities in the forcing factor(s). Frequently, the climate responds in a fashion intermediate between the two.

There are many climate forcing factors spanning an enormous range of periodicities. The longest, 200 to 500 million years, involves the passage of our Solar System through the galaxy, and the variations in galactic dust (Williams, 1975a). These may be considered to be external forcing mechanisms (section 2.5.1). Other long time scale variations (10^6 to 10^8 years) include the non-radiative forcing mechanisms, such as continental drift, orogeny (mountain building) and isostasy (vertical movements in the Earth's crust affecting sea level) (Raymo & Ruddiman, 1992; Ruddiman & Kubasch, 1991). These are internal forcing mechanisms (sections 2.6.1 and 2.6.2). External changes in the amount of solar radiation (Wigley & Kelly, 1990; Eddy, 1976, 1977, 1982; section 2.5.3) and the Earth's orbit around the Sun (Milankovitch, 1941; Berger, 1978, 1984; section 2.5.2), and internal variations in volcanic activity

(Sear *et al*., 1987; section 2.6.3), ocean circulation (Broecker & Denton, 1990; section 2.6.4) and atmospheric composition (IPCC, 1990a, 1992, 1995; section 2.6.5), all occur over time scales from 1 year to 10^5 years. Additionally, there are numerous other internal feedback mechanisms (see section 2.7) which all contribute to the changing of the global climate. The actual climate state at any point in time represents an aggregate response to all cycles of variation superimposed on the background noise.

The response of the climate system to this combination of forcing factors itself depends upon the different response times of the various components of the system. The overall climatic response will then be determined by the interactions between the components. The atmosphere, surface snow and ice, and surface vegetation typically respond to climatic forcing over a period of hours to days. The surface ocean has a response time measured in years, whilst the deep ocean and mountain glaciers vary only over a period spanning hundreds of years (Henderson-Sellers & McGuffie, 1987). Large ice sheets advance and withdraw over thousands of years whilst parts of the geosphere (e.g. continental weathering of rocks) respond only to forcing periods lasting hundreds of thousands to millions of years.

The response of the climate system to episodes of forcing can be viewed as a form of resonance. When the time period of forcing matches most closely the response time of a particular system component, the climatic response will be greatest within that component. Milankovitch forcing (section 2.5.2), for example, with periods of tens of thousands of years will be manifest in the response of the ice sheets (section 5.3.1), and the overall response of the climate system will be dominated by changes within the cryosphere. In addition, longer response times of certain components of the climate system modulate, through feedback processes, the short term responses. The response of the deep ocean to short term forcing (e.g. enhanced greenhouse effect (section 2.6.5), solar variations (section 2.5.3)), for example, will tend to attenuate or smooth the response of the atmosphere.

Throughout the remainder of this chapter, it should be recognised that a range of time scales applies to climate forcing mechanisms, radiative and non-radiative, external and internal, and to the response of the different components of the climate system.

2.5. External Forcing Mechanisms

This section discusses some of the various external forcing mechanisms operating over time scales of 10 years to 10^9 years.

2.5.1. Galactic Variations

The orbit of the Solar System about the centre of the Galaxy has been considered as a possible external climate forcing mechanism (Huggett, 1991). During the course of a galactic year (now estimated at 303 million years), variations in the interstellar medium (Williams, 1975a) may influence the amount of solar radiation incident at the Earth’s surface, thus acting as a radiative forcing mechanism to induce climate change. Williams (1975a) also suggests that variations in gravitational torque induced by our Galaxy’s near neighbours, the Small and Large Magellanic Clouds, could have far-reaching consequences for the Earth’s climate.

Unfortunately, the enormous time scale associated with this forcing (and any hypothesised global climatic change) makes empirical confirmation of this premise exceedingly imprecise. Nevertheless, it is indeed possible that the ice age supercycles during the last 700 million years (Fischer, 1984) (see section 5.2.2) could be the result of such galactic forcing mechanisms.

2.5.2. Orbital Variations

In the mid-19th century, Croll (1867a, 1867b) proposed an astronomical theory linking the Pleistocene (2 Ma to 10 Ka) ice ages with periodic changes in the Earth’s orbit around the Sun. Croll’s ideas were later refined and elaborated by Milankovitch (1941). The Milankovitch theory is the name given to the astronomical theory of climate variations. Since these ideas were put forward, much evidence has been found to support the theory. A review of the mechanics of empirical climate reconstruction is given in chapter 3, whilst chapter 5 covers the global climate changes associated with the ice ages. In this section, the forcing mechanisms involved with the Milankovitch theory are discussed.

The original Milankovitch theory identifies three types of orbital variation which could act as climate forcing mechanisms, obliquity or tilt of the Earth’s axis, precession of the equinoxes and eccentricity of the Earth orbit around the Sun. Each variation has its specific time period.

2.5.2.1. Obliquity

Today the Earth is tilted on its rotational axis at an angle of 23.4^o relative to a perpendicular to the orbital plane of the Earth. Over a 41,000 year time period, this angle of inclination fluctuates between 22^o and 24.5^o, influencing the latitudinal distribution of solar radiation.

Obliquity does not influence the total amount of solar radiation received by the Earth, but affects the distribution of insolation in space and time. As obliquity increases, so does the amount of solar radiation received at high latitudes in summer, whilst insolation decreases in winter. Changes in obliquity have little effect at low latitudes, since the strength of the effect decrease towards the equator. Consequently, variations in the Earth's axial tilt affect the strength of the latitudinal temperature gradient. Increased tilt has the effect of raising the annual receipt of solar energy at high latitudes, with a consequent reduction in the latitudinal temperature gradient.

2.5.2.2. Eccentricity

The Earth's orbit around the Sun is not perfectly circular but follows an elliptical path (see Figure 2.1). A second orbital variation involves the strength of the ellipse, or eccentricity. This parameter, e, is determined by Equation 4, which compares the two focal lengths, x and y in Figure 2.1.

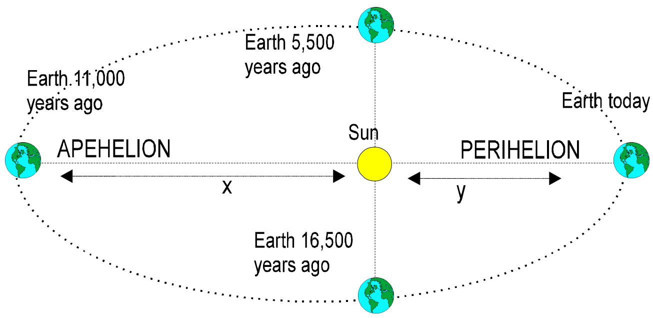
$$e = \{ (x^2 - y^2)^{½} \} / x \qquad \text{[Equation 4]}$$

When the orbit is circular, the lengths x and y are equal and e = 0. The Earth's orbit has been found to vary from being near circular (e = 0.005) to markedly elliptical (e = 0.06) with two primary periodicities of approximately 96,000 and 413,000 years (Berger, 1976). The current value of e is 0.018 (Henderson-Sellers & Robinson, 1986). Variations in eccentricity influence the total amount of solar radiation incident at the top of the Earth's atmosphere. With maximum eccentricity, differences in solar radiation receipt of about 30% may occur between perihelion and aphelion (Figure 2.1) (Goodess *et al*, 1992).

2.5.2.3. Precession

The third orbital variation is that of precession. The Sun lies at one of the focal points of the Earth's orbital ellipse. Due to the gravitational interaction of other planetary bodies in the solar system, primarily the Moon and the planet Jupiter, the perihelion (the point at which the Earth passes closest to the Sun) moves in space with a consequent shifting or precessing of the elliptical orbit. This phenomenon is known as the precession of the equinoxes, and effects the intensity of the seasons.

Figure 2.1 *Present and past orbital locations of the Earth during the Northern Hemisphere winter*



Precession has two components: an axial precession, in which the torque of the other planets exerted on the Earth's equatorial bulge causes the rotational axis to gyrate like a spinning top; an elliptical precession, in which the elliptical orbit of the Earth itself rotates about one focus. The net effect describes the precession of the equinoxes with a period of 22,000 years. This term is modulated by eccentricity which splits the precession into periods, 19,000 and 23,000 years (Crowley & North, 1991).

Like obliquity, precession does not affect the total amount of solar energy received by the Earth, but only its hemispheric distribution over time. If the perihelion occurs in mid-June i.e. when the Northern Hemisphere is tilted toward the Sun, then the receipt of summer solar radiation in Northern Hemisphere will increase. Conversely, if the perihelion occurs in December, the Northern Hemisphere will receive more solar radiation in winter (see Figure 2.1). It should be clear that the direction of changes in solar radiation receipt at the Earth's surface is opposite in each hemisphere.

2.5.2.4. Milankovitch cycles and ice ages

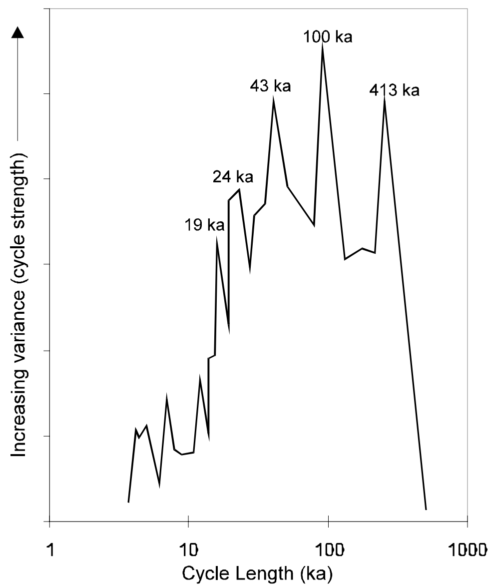
The three components of the orbital variations together effect both the total flux of incoming solar radiation and also the temporal and spatial distribution of that energy. These variations have the potential to influence the energy budget of the climate system (Milankovitch, 1941; Berger 1978), and can therefore be regarded as possible causes of climate change over a 10^4 to 10^5 year time scale. Being external to the climate system, they may be classified as external forcing mechanisms.

Milankovitch (1941) considered the changing seasonal (precession) and latitudinal (obliquity) patterns of incoming radiation to be critical factors in the growth of continental ice sheets and in the initiation of ice ages. He hypothesised that when axial tilt was small (large latitudinal temperature gradient), eccentricity was large and perihelion occurred during the Northern Hemisphere winter (warmer winters and colder summers), such a configuration would allow the persistence of accumulated snow throughout the summer months in the Northern Hemisphere. Additionally, the warmer winters and stronger atmospheric general circulation due to the increased temperature gradient would increase the amount of water vapour at the high latitudes available for snowfall.

For long-term proxy temperature data, spectral analysis[7], which permits the identification of cycles, has shown the existence of periodicities of 100,000, 43,000, 24,000 and 19,000 year (see Figure 2.2), all of which correspond closely with the theoretical Milankovitch cycles (Hays *et al*., 1976; Imbrie & Imbrie, 1979).

Nevertheless, verification of a causal link between the orbital forcing factors and the climatic response is far from being proved, and significant problems remain. Firstly, Figure 2.2 shows that the strongest signal in the observational data is the 100,000 year cycle. This would be the result of eccentricity variations in the Earth's orbit, which alone account for the smallest insolation changes. Secondly, it is not clear why changes in climate appear to be global. *A priori* reasoning indicates that the effects of precession would cause opposite responses in each hemisphere. In fact, climate change is synchronised between Southern and Northern Hemispheres, with a growth of ice sheets during glaciations occurring in the Arctic and Antarctic. It is now widely believed that the circulation of the oceans provides the forcing factor for synchronisation (Duplessy & Shackleton, 1985; Broecker, 1987; Broecker & Denton, 1990; Boyle & Keigwin, 1987). This is discussed further in section 2.6.4.

Figure 2.2. *Orbital periodicities identified through spectral analysis (after Hays et al., 1976)*



Most crucially of all, however, it seems that the orbital forcing mechanisms alone, could not account for the observed climatic variations over the past 2 million years (Hoyle, 1981). In order to explain the magnitude of the observed climatic changes, it seems necessary to invoke various feedback mechanisms. Indeed, Milankovitch himself had expected the direct effects of variations in insolation to be magnified by feedback processes, such as, at high latitudes, the ice albedo effect (section 2.7).

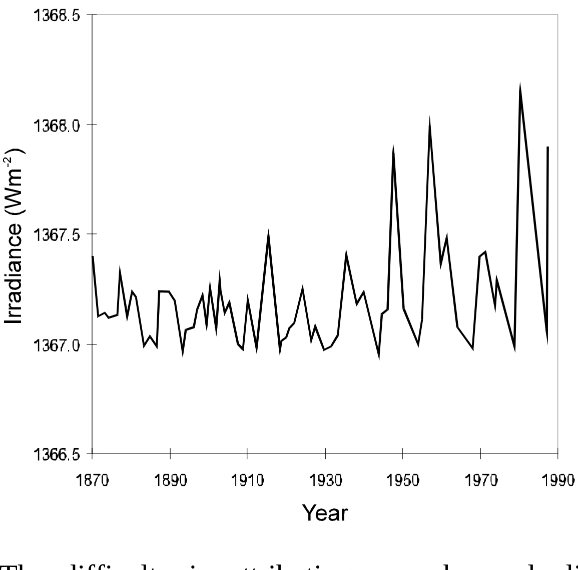
2.5.3. Solar Variations

Although solar variability has been considered, *a priori*, to be another external forcing factor, it remains a controversial mechanism of climate change, across all time scales. Despite many attempts to show statistical associations between various solar periodicities and global climate cycles, no realistic causal mechanism has been proposed to link the two phenomena.

[7] This mathematical technique calculates the strength of periodic variation across a range of time scales (see Figure 2.2).

The best known solar cycle is the variation in the number of sunspots over an 11 year period. Sunspot cycles are thought to be related to solar magnetic variations, and a double magnetic cycle (approximately 22 years) can also be identified (Foukal, 1990; Gough, 1990; Weiss, 1990). What is of interest to a climatologist is whether the sunspot cycles are accompanied by variations in solar irradiance - the solar constant - which, potentially, could force climate changes. The solar constant (approximately 1368Wm^{-2}) is a measure of the total solar energy flux integrated across all wavelengths of radiation. Two decades of satellite observations (e.g. Foukal, 1990) reveal that the solar constant varies over time scales of days to a decade, and there does appear to be a significant relationship with the sunspot number cycle. At times of high sunspot number, the value of the solar constant increases. Although sunspots are regions of cooler than average Sun surface temperature, their presence is accompanied by brighter (hotter) faculae which more than compensates for the increase in darker sunspot areas (Foukal, 1990; Kuhn *et al*., 1988). This relationship can be extended back over time using the long sunspot record. Solar irradiance changes thus calculated are reproduced in Figure 2.3.

Figure 2.3. *Variations in solar irradiance over the last 120 years (after Shine et al., 1990)*



The difficulty in attributing any observed climate change to these variations in solar irradiance is that the latter are small in magnitude - a change of much less than 1% over the course of the sunspot cycle. Wigley (1988) stressed that with such small variations in the solar constant, the global climatic response would be no more than a 0.03^{o}C temperature change. Nevertheless, many climatic records (e.g. indices of droughts, temperature and total atmospheric ozone) do appear, at least statistically, to display periodicity linked to one or both of the sunspot cycles (Mitchell *et al*., 1979; Newell *et al*., 1989). It should be clear, however, that a statistical association between solar variability and climate change does not prove cause and effect.

It is of course possible that the approximate 11-year cycle identified in many climate records is caused by some unknown internal oscillation and not by external solar forcing. It is conceivable that, simply by chance, the phase of the oscillation could coincide with the phase of the solar variability. More plausibly, an internal oscillation can become phase-locked to the solar cycles, thus augmenting the climatic response by a kind of feedback mechanism. For the time being, therefore, the link between the sunspot cycles and climate change must remain a speculative one.

However, there are other solar periodicities, with longer time scales that could be considered as climate forcing mechanisms. It has been suggested that the long-term variation in the amplitude of the sunspot cycles may have an influence on global climate (Eddy, 1982). Observations made with the naked eye reveal times when sunspot activity was very limited, including the Maunder Minimum (1654 to 1715) and the Spörer Minimum (1450 to 1534). These events occurred during the period known as the Little Ice Age (see section 5.3.2.4), and Eddy (1976, 1977) has hypothesised that the two may be causally linked. As with the sunspot cycles, however, the evidence is largely circumstantial. Other solar variations include cycles of sunspot cycle length (between about 9 and 13 years), changing solar diameter and the rate of change of solar diameter (Wigley & Kelly, 1990). Although some of these long-term variations may involve larger changes in solar output, this is again mere speculation.

Proxy records of solar irradiance changes are needed when even longer time scales are considered. A number of scientists (e.g. Hood & Jirikowic, 1990) have used records of ^{14}C in tree rings to investigate the relationships between potential solar forcing mechanisms and climate change. Changes in the output of energetic particles from the Sun (the solar wind) are believed to modulate the production of ^{14}C in the upper atmosphere. The magnetic properties of the solar wind change with the variation of sunspots, leading in turn to variations in the production of ^{14}C (Stuiver *et al*., 1991). The effect of the solar wind is such that high ^{14}C production is associated with periods of low sunspot number.

Relatively long and reliable ^{14}C records are now available. Spectral analysis has revealed a number of solar periodicities including a 2,400 year cycle, a 200 year cycle, a 80 to 90 year cycle and the shorter 11 and 22 year cycles. The ^{14}C records have also been correlated with a number a climate change indicators (Eddy, 1977), including glacial advance-retreat fluctuations and annual temperatures for England. Episodes of low ^{14}C production are associated with high sunspot activity and warmer climates. It is certainly feasible that the climatic variations of the Holocene (last 10,000 years since the end of the last ice age), and the shorter fluctuations associated with the Little Ice Age have been forced by the interacting millennia and century scale cycles of solar activity. However, conclusive evidence of a mechanism linking cause and effect is again missing. In addition, numerical modelling of Wigley & Kelly (1990) seems to indicate that solar irradiance changes would not be substantial enough to bring about the observed climatic changes without invoking additional internal feedback mechanisms.

2.6. Internal Forcing Mechanisms

This section discusses some of the various internal forcing mechanisms operating over time scales of 1 year to 10^8 years. They may be either radiative or non-radiative forcing mechanisms.

2.6.1. Orogeny

Orogeny is the name given to the tectonic process of mountain building and continental uplift. Such mechanisms operate only over tens or even hundreds of millions of years. The Earth's outer surface, a layer known as the lithosphere (made up of the crust and upper section of the mantle), is broken up into about 12 different plates which are constantly adjusting their positions relative to each other. Such movements are driven by the internal convective dynamics within the Earth's mantle. When plates collide, one may either be subducted beneath another, or both are pushed continually together, forcing upwards any continental land masses, to form long mountain ranges. The Himalayas formed when the Indian plate crashed into Asia about 20 to 30 million years ago.

There is now little doubt that the presence of mountain ranges on the Earth can dramatically influence global climate, and that orogenic uplift can act as a non-radiative (internal) forcing mechanism. North-south orientated mountain ranges in particular have the ability to influence global atmospheric circulation patterns, which usually maintain a more east-west trend on account of the Coriolis Force (Figure 1.6, section 1.2.5).

Ruddiman & Kutzbach (1991) have proposed that the uplift of the Tibetan Plateau, the Himalayas and the Sierra Nevada in the American south-west may have induced a global cooling during the last 40 million years (see section 5.2.2.3). Raymo & Ruddiman (1992) also suggest that increased uplift of these regions exposed more rock, thereby increasing the rate of physical and chemical weathering. During chemical weathering, carbon dioxide is extracted from the atmosphere to react with the decomposing rock minerals to form bicarbonates. These bicarbonates are soluble and can be transported via rivers and other fluvial channels, finally to be deposited on ocean floors as sediment. In essence, carbon dioxide is sequestered from the atmosphere, thereby decreasing the Earth's natural greenhouse effect, causing further cooling.

In view of this greenhouse feedback, mountain uplift seems to generate both non-radiative forcing (atmospheric circulation changes) and radiative forcing (greenhouse feedback). In such situations as described above, isolating a primary cause of climatic change from its secondary feedbacks, becomes ineffective. In section 2.7 the hypothesis will be investigated that climate change really results from a combination of impacts to different components (and sub-components) of the climate system, which cascade through the system.

Mountain uplift may also increase the land surface area covered by snow the year round. The subsequent increase in planetary albedo will reduce the amount of energy absorbed at the Earth's surface, initiating further cooling (Umbgrove, 1947). This is an example of the ice-albedo feedback effect, and is explained further in section 2.7.

2.6.2. Epeirogeny

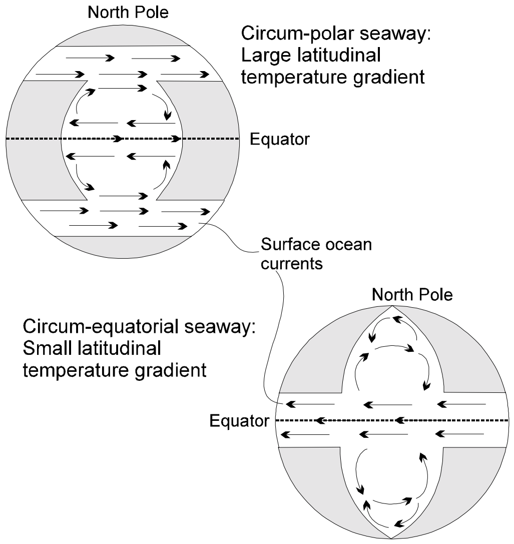
Epeirogeny is the term used to describe changes in the global disposition of land masses, and like orogenic processes, these changes are driven by internal plate tectonic movements. Because the internal dynamics of the Earth are slow, continents move about the globe at a rate of several centimetres per year. However, over tens or hundreds of millions of years, both the size and position of land area can change appreciably.

At times in Earth history, there have been super-continents in which all the continental plates were locked together in one area of the globe. The last of these occurred about 250 million years ago, and is named Pangea. Since that time, the continents have gradually moved apart, the most recent separation occurring between Europe and North America, during the last 60 to 70 million years. What is now the Pacific Ocean used once to be the vast expanse of water, called the Panthalassa Ocean, that surrounded Pangea.

A number of possible mechanisms which forced global climate to fluctuate between “greenhouse” and “icehouse” states have been explored (Crowell & Frakes, 1970; Beaty, 1978a, 1978b). First, as the continental area occupying high latitudes increases, as a result of continental drift, so the land area with permanent ice cover may expand, thus raising the planetary albedo, forcing (radiatively) a global cooling (the ice-albedo feedback). Second, the arrangement of continental land masses significantly affects the surface ocean circulation. Since ocean circulation is involved in the latitudinal heat transport regulating global climate (see section 1.2.5), so the wandering of land masses may force (non-radiatively) climate change over times scales involving tens or hundreds of millions of years.

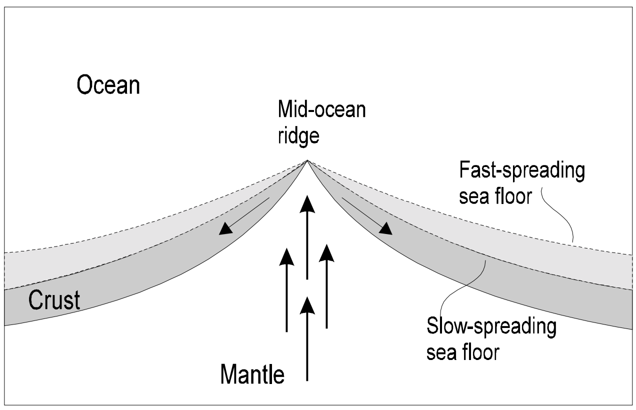
Such long term variations in ocean circulation as a result of continental drift, in addition to orogenic processes (see section 2.6.1), may have accounted for the return to a global “icehouse” that has taken place over the last 40 million years (Crowley & North, 1991; Frakes, 1979). Figure 2.4 postulates a particular scenario of hypothetical ocean circulation changes that may account for global climatic changes.
Associated with continental drift is the tectonic process of sea floor spreading. In the preceding section, it was explained how tectonic plates collide with one another and are consumed either by subduction or mountain building. New lithospheric plate material is formed at mid-ocean ridges, tectonic spreading centres, that mark the boundary between two diverging plates. These sea-floor regions, for example the Mid-Atlantic Ridge, release large amounts of energy and associated greenhouse gases. At times of enhanced tectonic activity and sea floor spreading, elevated levels of greenhouse gas emissions may initiate or augment a “greenhouse” world.

Figure 2.4. *Continental drift and ocean circulation changes*



As the newly formed plates diverge, they slowly begin to cool, and as the density of the exhumed rock increases, so the ocean crust begins to subside, as schematised in Figure 2.5. During times of increased tectonic activity, spreading rates are faster and the ocean crust has less time to cool and subside. The resulting ocean bathymetry is shallower that it otherwise would be and causes an (epeirogenic) rise in sea level.

Figure 2.5. *Formation of lithospheric plates*



During Cretaceous times (see section 5.2.2.2), mid-ocean ridges were indeed more active than they are today (Arthur *et al*., 1985; Davis & Solomon, 1981). Consequently, sea levels stood several hundred metres higher (due also to the absence of water-storing ice sheets), covering vast continental areas with shallow-level (epeiric) seas. Such a situation may have two important consequences. First, ocean circulation will be markedly affected, influencing global climate as illustrated above. Second, the large shallow seas, with relatively lower albedos than the land areas which they submerge, would be capable of storing considerably more energy, thus heating the Earth's surface.

2.6.3. Volcanic Activity

Explosive eruptions can inject large quantities of dust and gaseous material (such as sulphur dioxide) into the upper atmosphere (the stratosphere - see Figure 1.1, section 1.2.2), where sulphur dioxide is rapidly converted into sulphuric acid aerosols. Whereas volcanic pollution of the lower atmosphere is removed within days by the effects of rainfall and gravity, stratospheric pollution may remain there for several years, gradually spreading to cover much of the globe.

The volcanic pollution results in a substantial reduction in the direct solar beam, largely through scattering by the highly reflective sulphuric acid aerosols. This can amount to tens of percent. The reduction, is however, compensated for by an increase in diffuse radiation and by the absorption of outgoing terrestrial radiation (the greenhouse effect). Overall, there is a net reduction of 5 to 10% in energy received at the Earth's surface.

Clearly, this volcanic pollution affects the energy balance of the atmosphere whilst the dust and aerosols remain in the stratosphere. Observational and modelling studies (e.g. Kelly & Sear, 1984; Sear *et al*., 1987) of the likely effect of recent volcanic eruptions suggest that an individual eruption may cause a global cooling of up to 0.3^{o}C, with the effects lasting 1 to 2 years. Such a cooling event has been observed in the global temperature record in the aftermath of the eruption of Mount Pinatubo in June 1991. The climate forcing associated with individual eruptions is, however, relatively short-lived compared to the time needed to influence the heat storage of the oceans (Henderson-Sellers & Robinson, 1986). The temperature anomaly due to a single volcanic event is thus unlikely to persist or lead, through feedback effects, to significant long-term climatic changes.

Major eruptions have been relatively infrequent this century, so the long-term influence has been slight. The possibility that large eruptions might, during historical and prehistorical times, have occurred with greater frequency, generating long-term cooling, cannot, however, be dismissed. In order to investigate this possibility, long, complete and well-dated records of past volcanic activity are needed. One of the earliest and most comprehensive series is the Dust Veil Index (DVI) of Lamb (1970), which includes eruptions from 1500 to 1900. When combined with series of acidity measurements in ice cores (due to the presence of sulphuric acid aerosols), they can provide valuable indicators of past eruptions. Using these indicators, a statistical association between volcanic activity and global temperatures during the past millennia has been found (Hammer *et al*., 1980). Episodes of relatively high volcanic activity (1250 to 1500 and 1550 to 1700) occur within the period known as the Little Ice Age, whilst the Medieval Warm Period (1100 to 1250) can be linked with a period of lower activity.

Bryson (1989) has suggested a link between longer time scale volcanic variations and the climate fluctuations of the Holocene (last 10,000 years). However, whilst empirical information about temperature changes and volcanic eruptions remains limited, this, and other suggested associations discussed above, must again remain speculative.

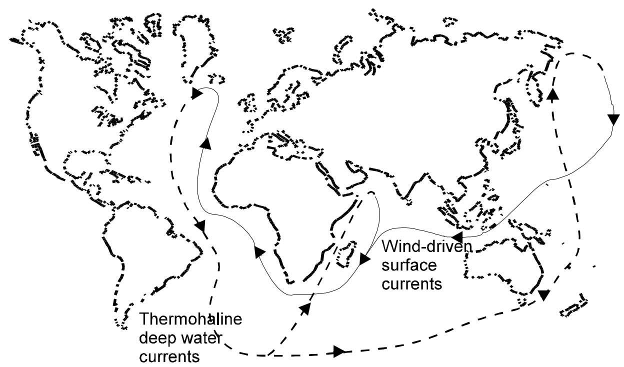
Volcanic activity has the ability to affect global climate on still longer time scales. Over periods of millions or even tens of millions of years, increased volcanic activity can emit enormous volumes of greenhouse gases, with the potential of substantial global warming (Pickering & Owen, 1994; Rampino & Volk, 1988). However, the global cooling effects of sulphur dioxide emissions (Officer & Drake, 1983) will act to counter the greenhouse warming, and the resultant climate changes remain uncertain. Much will depend upon the nature of volcanic activity. Basaltic outpourings release far less sulphur dioxide and ash, proportionally, than do the more explosive (silicic) eruptions.

2.6.4. Ocean Circulation

In section 1.3.1 it was identified that the oceans store an immense amount of heat energy, and consequently play a crucial role in the regulation of the global climate system. In order to explain the observed hemispheric synchroneity of glaciation, despite periods of directly opposed orbital forcing in the two hemispheres (see section 2.5.2), many researchers have looked to the oceans. Although, in this sense, changes in ocean circulation could be regarded as feedback resulting from orbital forcing, ocean circulation has traditionally been viewed as an internal forcing mechanism in its own right.

At present, northern maritime Europe is warmed by heat carried polewards by the Gulf Stream. When the warm water meets cold polar air in the North Atlantic, heat is released to the atmosphere and the water cools and sinks. This is assisted by the increases in salinity (and therefore density) that occur when sea ice forms in the Arctic regions (see sections 1.3.1 and 1.3.2). The bottom water so formed, called the North Atlantic Deep Water (NADW), flows southward through the western Atlantic, round Southern Africa and Australia, and then northwards into the Pacific Ocean. The North Atlantic is warmer than the North Pacific. The increased evaporation there therefore serves to increase salinity, relative to the North Pacific. This salinity gradient is thought to drive the global thermohaline ocean circulation. Such a picture of thermohaline circulation is schematised in Figure 2.6.

Figure 2.6. *The global thermohaline ocean circulation*



A number of the theories which have been put forward concerning the role of the oceans in the processes of climate change invoke changes in the rate of NADW production and other characteristics of the thermohaline circulation. Most attention has focused on the climatic transitions between glacial and interglacial episodes.

It has been suggested that during a glacial period, the formation of the NADW is much reduced or even totally shut down. At these times, the Arctic ice sheets extends much further south into the North Atlantic, pushing the position of the polar front southwards. Cooler sea surface temperatures reduce evaporation and therefore salinity, further precluding the initiation of a thermohaline circulation. The concomitant absence of the warm surface Gulf Stream could result in northern Europe being 6 to 8°C colder than during interglacial times (i.e. at present) (Broecker, 1987). The causes of the changes between the glacial and interglacial patterns of thermohaline circulation would then be seen as internal climate forcing mechanisms.

Indeed, Broecker (1987) has proposed that salinity changes between the North Atlantic and North Pacific may be so great that the entire global thermohaline circulation could be reversed. Such a theory of mode changes was developed in order to explain the rapid (<1,000 years) post-glacial climatic fluctuation of the Younger Dryas event about 11,000 years ago (see section 5.3.2.1), when the North Atlantic appeared to cool by several degrees (Dansgaard *et al*., 1989). Modelling appears to confirm the existence of at least two stable states of the thermohaline circulation (Manabe & Stouffer, 1988). Rapid transitions between these two states, and the corresponding climatic flips between glacial and interglacial periods, in response to internal forcing, would then be non-linear. Nevertheless, empirical evidence in support of mode changes is still inconclusive.

Broecker concedes, however, that the shutdown of the North Atlantic ‘heat conveyor belt system’ alone would not be sufficient to initiate global temperature changes and ice sheet development (Broecker & Denton, 1989). Other internal feedback mechanisms would need to be invoked, for example changes in the concentration of greenhouse gases and aerosol loading, together with reduced ocean heat transport and increased ocean alkalinity.

From the foregoing discussion on ocean circulation, one could conclude that such a mechanism of climate change should really be regarded as non-radiative (see also section 2.6.2, Figure 2.4), since what is at issue here is the transfer of energy within the ocean component of the climate system only. It is perhaps the resultant feedback processes identified in the preceding paragraph that allow most scientists to regard this mechanism as a radiative one.

2.6.5. Variations in Atmospheric Composition

The changing composition of the atmosphere, including its greenhouse gas and aerosol content, is a major internal forcing mechanism of climate change. As we have seen in section 1.2.4, the Earth's natural greenhouse effect (involving an increase in the downward energy flux) plays an important role in the regulation of the global climate. Obviously, then, changes in the atmospheric concentrations of greenhouse gases will modify the natural greenhouse effect, and consequently affect global climate.

Changes in the greenhouse gas content of the atmosphere can occur as a result of both natural and anthropogenic factors, the latter which has received considerable attention in the last 20 years (see chapter 5). Mankind, through the burning of fossil fuels, forest clearing and other industrial processes, has increased the amount of carbon dioxide and other greenhouse gases since the eighteenth century.

Natural changes in greenhouse gas concentrations can occur in numerous ways, most often in response to other primary forcing factors. In this sense, as with ocean circulation changes, such forcing should be more strictly regarded as secondary forcing or feedback. The role of feedback in global climate change is examined in the next section. Changes in atmospheric CO_2 and methane (CH_4) have been associated with transitions between glacial and interglacial episodes (Barnola *et al*., 1987; Neftel *et al*., 1988; Stauffer *et al*., 1988). Much of the empirical evidence suggests that these changes lag behind the climate signal, and must therefore act as feedback mechanisms to enhance climate change rather than as primary forcing mechanisms (Genthon *et al*., 1987; Lorius *et al*., 1990).

Changes in the atmospheric content of aerosols, again both natural and anthropogenic can act as climate forcing mechanisms, or more usually secondary feedback mechanisms. Increases in atmospheric turbidity (aerosol abundance) will affect the atmospheric energy budget by increasing the scattering of incoming solar radiation (see section 1.2.4). Atmospheric turbidity has been shown to be higher during glacial episodes than in interglacials (Legrand *et al*., 1988; Petit *et al*., 1990), with a consequent reduction in direct radiation reaching the Earth's surface. Such a situation will enhance the cooling associated with glacial periods.

2.7. Climate Feedback

The state of the global climate is one of general stability, engendered by a balance existing between the coupled components of the global climate system. The amount of incoming solar radiation is balanced by the amount of outgoing terrestrial radiation (section 1.2.3), so that the Earth neither continues indefinitely to heat up nor cool down. The Earth's climate is said to exist in equilibrium. When the climate system responds to radiative forcing (see section 2.3), this equilibrium is temporarily upset and a discrepancy between incoming and outgoing radiation exists. In an attempt to restore equilibrium, the global climate subsequently alters by either heating up or cooling down, depending on the direction of initial forcing.

Although the climate system is in balance, that balance is dynamic, ever-changing. The system is constantly adjusting to forcing perturbations and, as it adjusts, the climate alters. A change in any one part of the climate system will have much wider consequences as the initial effect cascades through the coupled components of the system. As the effect is transferred from one sub-component of the system to another, it will be modified in character or in scale. In some cases it will be amplified (positive feedback), in others, it may be reduced (negative feedback) (Cess & Potter, 1988). It is easiest to understand the concept of feedback by way of an example, the ice-albedo feedback.

Consider an Earth warmed as a result of increased radiative forcing, due, say, to changes in the orbital configuration of the Earth-Sun system. As the Earth's surface heats up, some of the ice at high latitudes begins to melt, exposing either bare ground or ocean, both of which have lower albedos (reflectivities) than ice. With a lower albedo, the exposed surfaces reflect less incident solar radiation, and the enhanced absorption causes further heating. Further rises in temperature initiate further melting of snow and ice, with further exposure of more energy-absorbent terrain. Thus a cyclic chain reaction of cause and effect is established, with each effect acting as the cause for the next step. This climatic phenomenon is called the ice-albedo feedback (Cubasch & Cess, 1990) and is an example of a positive feedback. The response to primary climatic forcing acts as a secondary forcing mechanism in the same direction as that of the initial forcing factor. Positive feedback augments the climatic response to forcing.

Negative feedback occurs when the response to primary climatic forcing acts as a secondary forcing mechanism in the opposite direction to that of the initial forcing factor. Negative feedback reduces the climate response to forcing. One example of a negative feedback due to increased radiative forcing would be cloud formation (Cubasch & Cess, 1990). As the Earth warms, so the rate of evaporation from the (warmer) oceans increases, supplying the atmosphere with more water vapour conducive to enhanced cloud formation. With greater global cloud coverage, more incident radiation is reflected, reducing radiative forcing and leading to a lowering of the global temperature.

This simple picture of the cloud feedback is, however, complicated by the fact that clouds also serve to trap terrestrial infrared radiation, augmenting the greenhouse effect, and thus act as a positive feedback, also, to increased radiative forcing. Numerical modelling has not been able to determine with any degree of certainty whether positive cloud feedbacks outweigh negative ones, or vice versa. Much depends upon the altitude of the cloud and the cloud type (Charlson *et al*., 1987; Wigley, 1989). High level clouds are expected to have a net positive feedback, with the effect of long-wave radiation absorption outweighing albedo effects. Clouds at high altitudes exist in colder air and tend to emit less radiation generating a stronger greenhouse effect. Low level clouds, on the other hand, probably have a net negative feedback effect.

There are many other feedback effects which have the potential to influence global climate in response to some initial radiative forcing. These may be found to operate within and between all components of the climate system. Any forcing mechanism which affects the amount of water vapour in the atmosphere will initiate a cloud feedback process. Water vapour, itself, is also a greenhouse gas and forcing perturbations will initiate a water vapour feedback (Cubasch & Cess, 1990; Gates *et al*., 1992)). Changes in ocean chemistry may occur as a result of primary climatic forcing. For example, warmer water stores less dissolved carbon dioxide, which then remains in the atmosphere to further enhance the greenhouse effect - a positive feedback. Circulation changes within the oceans will also introduce feedback processes affecting the transfer of heat, moisture and momentum (section 2.6.4). Changes in the surface vegetative cover, which has a marked effect on the Earth's albedo (see section 1.2.3), are likely to have feedback effects on the Earth's climate (Melillo *et al*., 1990).

2.8. Climate Sensitivity

The concept of feedback is related to the climate sensitivity or climate stability. It is useful to have a measure of the strength of various feedback processes which determine the ultimate response of the climate system to any change in radiative forcing. In general terms, an initial change in temperature due to a change in radiative forcing, $\Delta T_{forcing}$, is modified by the complex combination of feedback processes such that:

$$\Delta T_{final} = \Delta T_{forcing} + \Delta T_{feedback} \qquad \text{[Equation 5]}$$

where $\Delta T_{feedback}$ is the temperature change resulting from feedback and ΔT_{final} is the overall change in temperature between the initial and final equilibrium states. The degree to which feedback processes influence the final climatic response is a measure of the sensitivity of the climate system.

Equation 5 can be rewritten as follows:

$$\Delta T_{final} = f\Delta T_{forcing} \qquad \text{[Equation 6]}$$

where f is called the feedback factor (Henderson-Sellers & Robinson, 1986). When only one feedback mechanism is operative the solution to equation 6 is simple, assuming both f and $\Delta T_{forcing}$ are known. When more than one feedback is operative, matters become more complicated. For two feedbacks, the net effect is given by:

$$f = f_1f_2 / (f_1 + f_2 - f_1f_2) \qquad \text{[Equation 7]}$$

where f_1 and f_2 are the feedback factors of the two feedback processes (Henderson-Sellers & Robinson, 1986). Clearly, the feedback factors are neither additive nor multiplicative. A feedback operating alone with a factor of 2 would double the initial climatic response to forcing. If a second feedback with factor 1.5 acted with it, the overall feedback would be enhanced by a factor of 6. It can be seen, then, that a combination of feedback processes could dramatically affect the climate as it responds to only a small change in radiative forcing.

The climate's sensitivity can be mathematically determined in another way. From satellite observations, it has been shown that changes in global temperature are approximately proportional to changes in radiative forcing. If we assume an instantaneous[8] change in climate, from one equilibrium state to another, then:

$$\Delta Q = \lambda \Delta T \qquad \text{[Equation 8]}$$

where ΔQ is change in radiative forcing (expressed in terms of the net downward radiative flux at the top of the troposphere), ΔT is the change in global temperature, and λ is a measure of the climate sensitivity.

Based on equation 8, the climate sensitivity is usually expressed in terms of the temperature change associated with a specified change in radiative forcing, usually a doubling of the atmospheric carbon dioxide content. Thus, the carbon dioxide doubling temperature, ΔT_{2x}, is given by

$$\Delta T_{2x} = \Delta Q_{2x} / \lambda \qquad \text{[Equation 9]}$$

where ΔQ_{2x} is 4.2 Wm^{-2}. The magnitude (and sign) of ΔT_{2x} will depend on λ, the climate sensitivity, which is determined by the net affect of the climate feedback processes. Despite extensive climate modelling over the last 2 decades to understand the problem of contemporary global warming (see chapter 6), it is this parameter that is proving hard to define numerically.

As explained earlier, the idea of static equilibrium and instantaneous climatic response represents an unrealistic situation. To take into account the dynamic and transient nature of the climate's response to forcing, a more complicated equation linking ΔT and ΔQ is required if the evolution of the response with time is to be determined. In the case:

$$\Delta Q = \lambda \Delta T + C\delta\Delta T/\delta t \quad \text{[Equation 10]}$$

Here, the change in radiative forcing, ΔQ, is balanced by:
1) the change in the outgoing radiative flux at the tropopause caused by the response of the climate system including feedback; and
2) the energy stored in the system, $C\delta\Delta T/\delta t$, where C is the system's heat capacity and t is time.

The latter term in equation 10 simulates the time-dependent nature of the climate system's response. The main contributor to the system's heat capacity is the world's oceans. The heat capacity of water is large compared to that of air and is therefore able to store much more energy (see section 1.3.1). Additionally, the high heat capacity means that the oceans take time to heat up (or cool down) and hence slow the surface temperature response to any change in radiative forcing: the transient response will always be less than the equilibrium response.

Solution of Equation 10 leads to the definition of the response time of the climate system, τ, such that:

$$\tau = C / \lambda \qquad \text{[Equation 11]}$$

If the heat capacity of the climate system is large, the response time is large. Equally, if the climate sensitivity is small the response time is large. λ is also known as the radiative damping coefficient. Here, analogy is made between the response of the climate system and an oscillating spring. If a spring has a high damping coefficient, it will stop oscillating soon after it has been set in motion. Similarly, if the radiative damping coefficient is large, the climate will respond quickly and τ will be small.

2.9. Conclusion

Both climate forcing mechanisms and the response of the climate system operate over a variety of different time scales. Response to forcing may be linear, quasi-linear or non-linear. Non-linearity in climatic change is the result of a complex interaction of feedback processes. It should be appreciated that different primary forcing mechanisms will initiate different feedback processes. Primary feedback processes will give rise to secondary feedback processes. Some may be positive, some negative, but generally, climatic feedback acts in a direction which augments the initial climatic response to forcing. How much it does so, and how quickly, depends upon the sensitivity of the global climate to radiative forcing.

[8] As explained earlier, the climate does not respond instantaneously between two levels of static equilibrium but responds transiently (i.e. over time) and dynamically.

3.1. Introduction

To both understand the present climate and to predict future climate change, it is necessary to have both theory (chapters 1 and 2) and empirical observation. Any study of climate change involves the construction (or reconstruction) of time series of climate data. How these climate data vary across time provides a measure (either quantitative or qualitative) of climate change. Types of climate data include temperature, precipitation (rainfall), wind, humidity, evapotranspiration, pressure and solar irradiance.

During the most recent history, scientists have been able to construct climate time series from empirically observed instrumental data. Although the longest of these is a temperature record from central England beginning in the 17th century (Manley, 1974), the period traditionally associated with instrumental records extends back only to the middle of the 19th century. Analysis of instrumental records is the subject of contemporary climate change, and is the focus of chapter 6. For periods prior to the recording of instrumental data, climate changes have to be reconstructed from indirect or proxy sources of information. This is the discipline of palaeoclimatology, and is dealt with in chapter 5. In this chapter, the empirical methods of both instrumental and proxy observation are reviewed.

3.2. Climate Construction from Instrumental Data

Contemporary climate change may be studied by constructing records of values (daily, monthly and annual) which have been obtained with standard equipment. The instruments must be properly installed in suitable places, carefully maintained and conscientiously observed. The instruments should be exposed in such a manner to ensure that a representative or homogeneous measurement of the climate in question is made (Oldeman, 1987). The concept of homogeneity is discussed further in section 3.2.2. It is not possible to measure the climate *per se*, but only the individual elements of climate. A climate element is any one of the various properties or conditions of the atmosphere which together specify the physical state of the climate at a given place, for a particular period of time (Linacre, 1992). The most commonly measured element is temperature.

3.2.1. Measurement of Climate Elements

Sections 3.2.1.1 to 3.2.1.4 aim to provide only an overview of the standard practises observed in the collection of climate data. For a fuller account, the reader is referred to Linacre (1992).

3.2.1.1. Measurement of Temperature

Many surface air temperature records extend back to the middle part of the last century. The measurement of the surface air temperature is essentially the same now as it was then, using a mercury-in-glass thermometer, which can be calibrated accurately and used down to -39^{o}C, the freezing point of mercury. For lower temperatures, mercury is usually substituted by alcohol. Maximum and minimum temperatures measured during specified time periods, usually 24 hours, provide useful information for the construction and analysis of temperature time series. Analysis involves the calculation of averages and variances of the data and the identification, using various statistical techniques, of periodic variations, persistence and trends in the time series (Mitchell *et al.*, 1966; Barry & Perry, 1973). This is discussed further in section 3.2.3.

Observations of temperature from surface oceans are also collected in order to construct time series. In recent decades, much effort has also been directed towards the measurement of temperature at different levels in the atmosphere. There are now two methods of measuring temperatures at different altitudes: the conventional radiosonde network; and the microwave-sounding unit (MSU) on the TIROS-N series of satellites. The conventional network extends back to 1958 (Angell, 1988) and the MSU data to 1979 (Spencer & Christy, 1990).

Temperature is a valuable climate element in climate observation because it directly provides a measure of the energy of the system under inspection (see Equation 3, section 1.2.3). For example, a global average temperature[9] reveals information about the energy content of the Earth-atmosphere system. A higher temperature would indicate a larger energy content. Changes in temperature indicate changes in the energy balance, the causes of which were

[9] Global temperature time series are constructed using a representative network of measuring stations across the Earth, including data on both land surface air temperatures and sea surface temperatures. Analysis of the time series is performed as with any other time series.

discussed in chapters 1 and 2. Variations in temperature are also subject to less variability than other elements such as rainfall and wind. In addition, statistical analysis (see section 3.2.3) of temperature time series is often less complex than that associated with other series. Perhaps most importantly of all, our own perception of the state of the climate are intimately linked to temperature.

3.2.1.2. Measurement of Rainfall

Rainfall is measured most simply by noting periodically how much has been collected in an exposed vessel since the time of the last observation. Care must be taken to avoid underestimating rainfall due to evaporation of the collected water and the effects of wind (Folland, 1988; Mueller & Kidder, 1972). Time series can be constructed and analysis performed in a similar manner to those of temperature.

The measurement of global rainfall offers an indirect or qualitative assessment of the energy of the Earth-atmosphere system. Increased heat storage will increase the rate of evaporation from the oceans (due to higher surface temperatures). In turn, the enhanced levels of water vapour in the atmosphere will intensify global precipitation. Rainfall is, however, subject to significant temporal and spatial variability, and the occurrence of extremes, and consequently, analysis of time series is more complex (Huff, 1970).

3.2.1.3. Measurement of Humidity

The amount of water vapour in the air can be described in at least 5 ways, in terms of:

1) the water-vapour pressure;
2) the relative humidity;
3) the absolute humidity
4) the mixing ratio
5) the dewpoint.

A full account of these definitions may be found in Linacre (1992). The standard instrument for measuring humidity is a psychrometer. This is a pair of identical vertical thermometers, one of which has the bulb kept wet by means of a muslin moistened by a wick dipped in water. Evaporation from the wetted bulb lowers its temperature below the air temperature (measured by the dry bulb thermometer). The difference between the two measured values is used to calculate the air's water-vapour pressure, from which the other indices of humidity can be determined.

3.2.1.4. Measurement of Wind

Wind is usually measured by a cup anemometer which rotates about a vertical axis perpendicular to the direction of the wind. The exposure of wind instruments is important (Johnson & Linacre, 1978); any obstruction close by will affect measurements. Wind direction is also measured by means of a vane, accurately balanced about a truly vertical axis, so that it does not settle in any particular direction during calm conditions.

3.2.2. Homogeneity

Non-climatic influences - inhomogeneities - can and do affect climatic observations. Any analyst using instrumental climate data must first assess the quality of the observations. A numerical series representing the variations of a climatological element is called homogeneous if the variations are caused only by fluctuations in weather and climate (Conrad & Pollak, 1962). Leaving aside the misrecording of data, the most important causes of inhomogeneity are:

1) changes in instrument, exposure and measuring technique (for example, when more technologically advanced equipment is introduced);

2) changes in station location (i.e. when equipment is moved to a new site);

3) changes in observation times and methods used to calculate daily averages; and

4) changes in the station environment, particularly urbanisation (for example, the growth of a city around a pre-existing meteorological station).

When assessing the homogeneity of a climate record, there are three major sources of information: the variations evident in the record itself; the station history; and nearby station data. Visual examination and statistical analysis of the station record may reveal evidence of systematic changes or unusual behaviour which suggest inhomogeneity. For example, there may be a step-change in the mean, indicating a change in station location. A steady trend may indicate a progressive change in the station environment, such as urbanisation. An extreme value may be due to a typing error.

Often these inhomogeneities may be difficult to detect and other evidence is needed to confirm their presence. One source of evidence is the station history, referred to as metadata. The station history should include details of any changes of location of the station, changes in instrumentation or changes in the timing and nature of observation. Very often, though, actual correction factors to observation data containing known inhomogeneities will be difficult to calculate, and in these cases, the record may have to be rejected.

The third approach to homogenisation involves empirical comparisons between stations close to each other. Over time scales of interest in climate change studies, nearby stations (i.e. within 10km of each other) should be subject to similar changes in monthly, seasonal and annual climate. The only differences should be random. Any sign of systematic behaviour in the differences (e.g. a trend or step-change) would suggest the presence of inhomogeneities.

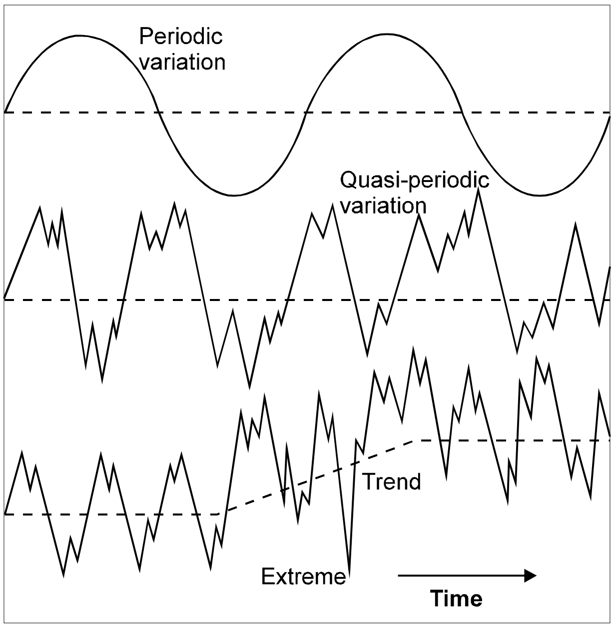
In light of the foregoing discussion on homogeneity, careful attention has to be paid to eliminating sources of non-climatic error when constructing large-scale record, such as the global surface air temperature time series (Jones, 1988 & Farmer *et al*., 1989). This, and similar records including sea surface temperatures, rely on the collection of millions of individual observations from a huge network made up of thousands of climate stations. A number of these have been reviewed by Jones *et al*., (1991). The effects of urbanisation (the artificial warming associated with the growth of towns and cities around monitoring sites) were considered to be the greatest source of inhomogeneity, but even this, it was concluded, accounts for at most a 0.05^{o}C warming (or 10% of the observed warming) over the last 100 years.

Conrad & Pollak (1962), Folland *et al*. (1990), and Jones *et al*. (1991) provide useful references investigating the problems of homogeneity and data reliability of instrumental records of climate data.

3.2.3. Statistical Analysis of Instrumental Records

Once climate data has been collected and corrected for inhomogeneities, it will need to be analysed. The aim of any statistical analysis is to identify systematic behaviour in a data set and hence improve understanding of the processes at work to compliment the theory. Statistical analysis is a search for a signal in the data that can be distinguished from the background noise (see discussion in section 2.4). In climate change research that signal will be a periodic variation, a quasi-periodic variation, a trend, persistence or extreme events in the climate element under analysis (Figure 3.1).

Figure 3.1. *Typical climatic variations (after Hare, 1979)*



Before undertaking a statistical analysis of a climate record, a number of questions about the task in hand should be asked.

1. *What is the purpose of the analysis?*

In its simplest form, the statistical analysis may be:

a) descriptive; or
b) investigative (Mitchell *et al*., 1966).

Descriptive analyses set out solely to document particular aspects of the variations present in the data set (the signal). Indices calculated will include the mean and variance (or standard deviation). The occurrence of extreme events, cycles and trends will also be noted (Gibbs *et al*., 1978). Significance testing is crucial to this category of analysis. Significance testing establishes whether or not the variation under consideration is different from what one would expect to arise in a random time series.

Investigative analyses set out to test a pre-defined hypothesis. The hypothesis should *a priori* have a sound physical basis. “Does the time series contain an El Niño cycle?” would be an example of a hypothesis that could be investigated.

2. *What is the most appropriate data set to use?*

Any data set used for a statistical analysis should be:

a) representative of the relevant physical processes;

b) sufficient in quantity to support the statistical method(s) to be used; and

c) accurate and reliable (homogeneous).

To investigate the impact of El Niño[10] on drought in eastern Australia, it is necessary to firstly identify a representative indicator of El Niño, such as sea surface temperatures in the SE Pacific. Secondly, a reliable indicator of drought in eastern Australia is required, for example, rainfall. The data set would need to be of sufficient duration to permit the testing for a relationship on the time scale under consideration, i.e. does El Niño cause drought in eastern Australia? Since El Niño has a quasi-periodicity of 2 to 5 years, then a data set of length at least 7 to 10 times this duration (i.e. up to 50 years) is required to have confidence in the statistical methods. To investigate longer-term trends, the data requirement becomes more stringent.

3. *What is the most appropriate technique to use and how should it be applied?*

Often it will be clear as to which statistical method of analysis is required. However, its application may be less straightforward. The nature of the data may determine whether or not a particular technique is valid (or, at least the way in which the technique is applied). For example, if the data are not normally distributed[11] then this may invalidate assumptions on which the technique is based. What ever technique is used, it goes without saying that testing statistical significance must be a critical concern.

Barry & Perry (1973) offer a detailed introduction of the mathematical aspects of statistical analysis, with many useful examples. Other useful references are provided by Gani (1975) and Godske (1966). Before concluding this section, however, a couple of points need illustrating. Firstly, much of what has been said about statistical analysis of instrumental records applies equally well to the study of palaeoclimatology, and to the reconstruction of past climates from proxy data (discussed in the next section). Secondly, the statistical analysis of climate data serves to compliment and support theories developed to explain the causes (and effects) of climate change. Statistical associations do not prove cause and effect for they are solely based upon the laws of probability. When analysing and interpreting climate data in the effort to aid understanding of the causes of climate change, it is necessary to bare this in mind.

3.3. Palaeoclimate Reconstruction from Proxy Data

Climate varies over different time scales, from years to hundreds of millions of years, and each periodicity is a manifestation of separate forcing mechanisms (section 2.4). In addition, different components of the climate system change and respond to forcing factors at different rates; in order to understand the role such components play in the evolution of climate it is necessary to have a record considerably longer than the time it takes for them to undergo significant changes (Bradley, 1985).

Palaeoclimatology is the study of climate and climate change prior to the period of instrumental measurements. A geological chronology of Earth history is provided in the appendix, which will provide a useful time frame for the material discussed both in the remainder of this chapter and in chapter 5.

Instrumental records span only a tiny fraction ($<10^{-7}$) of the Earth’s climatic history and so provide a

[10] El Niño is a quasi-periodic variation lasting 2 to 5 years during which an anomalously warm body of surface water forms and spreads throughout the SE Pacific. It is associated with the Southern Oscillation of pressure anomalies occurring in the south Pacific region, and the whole phenomenon is termed the El Niño Southern Oscillation or ENSO. ENSO has a significant influence on global and regional climates.

[11] A discussion of statistical methods and associated terminology in a climatological context may be found in Barry & Perry (1973).

inadequate perspective on climatic variation and the evolution of the climate today and in the future. A longer perspective on climate variability can be obtained by the study of natural phenomena which are climate-dependent. Such phenomena provide a proxy record of the climate.

Many natural systems are dependent on climate, and from these it may be possible to derive palaeoclimatic information from them. By definition, such proxy records of climate all contain a climatic signal, but that signal may be weak and embedded in a great deal of random (climatic) background noise. In essence, the proxy material has acted as a filter, transforming climate conditions in the past into a relatively permanent record. Deciphering that record is often a complex business.

Table 3.1. *Principle sources of proxy data for palaeoclimatic reconstructions*

Glaciological (Ice Cores)
- Oxygen isotopes
- Physical properties
- Trace element & microparticle concentrations

Geological
- A. Sediments
  - 1. Marine (ocean sediment cores)
    - i) Organic sediments (planktonic & benthic fossils)
      - Oxygen isotopes
      - Faunal & floral abundances
      - Morphological variations
    - ii) Inorganic sediments
      - Mineralogical composition & surface texture
      - Distribution of terrigenous material
      - Ice-rafted debris
      - Geochemistry
  - 2. Terrestrial
    - Periglacial features
    - Glacial deposits & erosional features
    - Glacio-eustatic features (shorelines)
    - Aeolian deposits (sand dunes)
    - Lacustrine deposits/varves (lakes)
- B. Sedimentary Rocks
  - Facies analysis
  - Fossil/microfossil analysis
  - Mineral analysis
  - Isotope geochemistry

Biological
- Tree rings (width, density, isotope analysis)
- Pollen (species, abundances)
- Insects

Historical
- meteorological records
- parameteorological records (environmental indicators)
- phenological records (biological indicators)

The major types of proxy climatic data available are listed in Table 3.1. Each proxy material differs according to: a) its spatial coverage; b) the period to which it pertains; and c) its ability to resolve events accurately in time (Bradley, 1985). Some proxy records, for example ocean floor sediments, reveal information about long periods of climatic change and evolution (10^7 years), with a low-frequency resolution (10^3 years). Others, such as tree rings are useful only during the last 10,000 years at most, but offer a high frequency (annual) resolution. The choice of proxy record (as with the choice of instrumental record) very much depends on what physical mechanism is under review. As noted, climate responds to different forcing mechanisms over different time scales, and proxy materials will contain necessary climatic information on these to a greater or lesser extent, depending on the three factors mentioned.

Other factors that have to be considered when using proxy records to reconstruct palaeoclimates include the continuity of the record and the accuracy to which it can be dated (Bradley, 1985). Ocean sediments may provide continuous records for over 1 million years (Ma) but typically they are hard to date using existing techniques. Usually there is an uncertainty of +/- 5% of the record's true age. Ice cores are easier to date but may miss layers from certain periods due to melting and wind erosion. Glacial deposits are highly episodic in nature, providing evidence only of discrete events in the past. Different proxy systems also have different levels of inertia with respect to climate, such that some systems may vary exactly in phase with climate whereas others lag behind by as much as several centuries (e.g. Bryson & Wendland, 1967).

Like climate construction from instrumental records, palaeoclimate reconstruction may be considered to proceed through a number of stages (Hecht *et al.*, 1979). The first stage is that of proxy data collection, followed by initial analysis and measurement. This results in primary data. The next stage involves the calibration of the data with modern climate records. In this, the uniformitarian principle is assumed, whereby contemporary climatic variations form a modern analogue for palaeoclimatic changes. It is important to be aware, however, of the possibility that palaeo-environmental conditions may not have modern analogues (Sachs *et al.*, 1977). The calibration may be only qualitative, involving subjective assessment, or it may be highly quantitative. The secondary data provide a record of past climatic variation. The third stage is the statistical analysis of this secondary data. The palaeoclimatic record is now statistically described and interpreted, providing a set of tertiary data.
Table 3.1 is obviously not exhaustative. Bradley (1985) offers an excellent review of the various proxy

methods and techniques employed to reconstruct Quaternary[12] palaeoclimatic change, whilst Frakes (1979) provides a useful commentary on the evidence for pre-Quaternary climates spanning most of geologic time. In the following sections, some of the more widely used proxy techniques will be reviewed. In all of the accounts, attention should be paid to the issues of reliability, dating, interpretation and meaning for all forms of climate reconstruction.

3.3.1. Historical Records

Historical records have been used to reconstruct climates dating back several thousand years (i.e. for most of the Holocene). Historical proxy data can be grouped into three major categories (see Table 3.1). First, there are observations of weather phenomena *per se*, for example the frequency and timing of frosts or the occurrence of snowfall. Secondly, there are records of weather-dependent natural or environmental phenomena, termed parameteorological phenomena, such as droughts and floods. Finally, there are phenological records of weather-dependent biological phenomena, such as the flowering of trees, or the migration of birds.

Major sources of historical palaeoclimate information include: ancient inscriptions; annals and chronicles; government records; estate records; maritime and commercial records; diaries and correspondence; scientific or quasi-scientific writings; and fragmented early instrumental records.

There are a number of major difficulties in using this kind of information. First, it is necessary to determine exactly what the author meant in describing the particular event. How severe was the “severe” frost? What precisely does the term drought refer to? Content analysis - a standard historical technique - has been used to assess, in quantitative terms, the meaning of key climatological phrases in historical accounts (Baron, 1982). This approach involves assessment of the frequency of use made of certain words or phrases by a particular author. Nevertheless, the subjectivity of any personal account has to be carefully considered. Very often, the record was not kept for the benefit of the future reader, but to serve some independent purpose. During much of the dynastic era in China, for example, records of droughts and floods would be kept in order to gain tax exemptions at times of climatic adversity (Yao, 1943).

Secondly, the reliability of the account has to be assessed. It is necessary to determine whether or not the author had first-hand evidence of the meteorological events (Wigley, 1978; Ingram *et al*., 1981). Thirdly, it is necessary to date and interpret the information accurately. The representativeness of the account has to be assessed. Was the event a localised occurrence or can its spatial extent be defined by reference to other sources of information? What was the duration of the event? a day? a month? a year?

Finally, the data must, as with all proxy records, be calibrated against recent observations and cross-referenced with instrumental data (e.g. Bergthorsson, 1969) This might be achieved by a construction of indices (e.g. the number of reports of frost per winter) which can be statistically related to analogous information derived from instrumental records.

3.3.2. Ice Cores

As snow and ice accumulates on polar and alpine ice caps and sheets, it lays down a record of the environmental conditions at the time of its formation. Information concerning these conditions can be extracted from ice and snow that has survived the summer melt by physical and chemical means. When melting does occur, the refreezing of meltwater can provide a measure of the summer conditions.

Palaeoclimate information has been obtained from ice cores by three main approaches (see Table 3.1). These involve the analysis of: a) stable isotopes of water; b) dissolved and particulate matter in the firn[13] and ice; and c) the physical characteristics of the firn and ice, and of air bubbles trapped in the ice. Each approach has also provided a means of dating the ice at particular depths in the ice core.

[12] See appendix.

[13] Firn is the term given to snow that has been converted to ice by compaction from overlying layers of subsequent snowfall.

3.3.2.1. Stable isotope analysis

The basis for palaeoclimatic interpretations of variations in the stable isotope[14] content of water molecules is that the vapour pressure of $H_2{}^{16}O$ is higher than that of $H_2{}^{18}O$. Evaporation from a water body thus results in a vapour which is poorer in ^{18}O than the initial water; conversely, the remaining water is enriched in ^{18}O. During condensation, the lower vapour pressure of the $H_2{}^{18}O$ ensures that it passes more readily into the liquid state than water vapour made up of the lighter oxygen isotope (Dansgaard, 1961). During the poleward transportation of water vapour, such isotope fractionation continues this preferential removal of the heavier isotope, leaving the water vapour increasingly depleted in $H_2{}^{18}O$. Because condensation is the result of cooling, the greater the fall in temperature, the lower the heavy isotope concentration will be. Isotope concentration in the condensate can thus be considered as a function of the temperature at which condensation occurs. Water from polar snow will thus be found to be most depleted in $H_2{}^{18}O$.

This temperature dependency allows the oxygen isotope content of a ice core to provide a proxy climate record. The relative proportions of ^{16}O and ^{18}O in an ice core are expressed in terms of departures, $\delta^{18}O$, from the Standard Mean Ocean Water (SMOW) standard (Craig, 1961), such that:

$$\delta^{18}O = \frac{(^{18}O/^{16}O)_{sample} - (^{18}O/^{16}O)_{SMOW}}{(^{18}O/^{16}O)_{SMOW}} \times 10^{3}\ ‰$$

[Equation 12]

All measurements are made using a mass spectrometer and results are normally accurate to within 0.1‰ (parts per mille). A $\delta^{18}O$ value of -10‰ indicates a sample with an $^{18}O/^{16}O$ ratio 1% or 10‰ less than SMOW. For most palaeoclimate reconstructions, typical values for $\delta^{18}O$ obtained from ice cores range between -10 and -60‰ (Morgan, 1982).

Similar palaeoclimatic studies can be carried out using isotopes of hydrogen (^1H and ^2H (Deuterium)), but these are rarer in nature and the laboratory techniques involved are more complex.

[14] There are 3 stable isotopes of oxygen, ^{16}O, ^{17}O and ^{18}O in the proportions 99.76%, 0.04% and 0.2% respectively. 16, 17 and 18 refer to the number of nucleons contained within the O atom. An atom of each isotope contains 8 protons. It is the number of neutrons (8, 9, or 10) which varies. Consequently, ^{18}O is heavier than ^{17}O which is heavier than ^{16}O.

3.3.2.2. Physical and chemical characteristics of ice cores

The occurrence of melt features in the upper layers of ice cores are of particular palaeoclimatic significance. Such features include horizontal ice lenses and vertical ice glands which have resulted from the refreezing of percolating water (Langway, 1970; Koerner, 1977a). They can be identified by their deficiency in air bubbles. The relative frequency of melt phenomena may be interpreted as an index of maximum summer temperatures or of summer warmth in general (Koerner, 1977b). Other physical features of ices cores which offer information to the palaeoclimatologist include variations in crystal size, air bubble fabric and crystallographic axis orientation (Langway, 1970).

Another important component of ice cores which is of palaeoclimatic significance is the atmospheric gas content, as the air pores are closed off during the densification of firn to ice (Raynaud & Lorius, 1973). Considerable research effort has been devoted to the analysis of carbon dioxide concentrations of air bubbles trapped in ice cores. It will be seen in chapter 5 that variations in atmospheric carbon dioxide may have played an important role in the glacial-interglacial climatic variations during the Quaternary.

Finally, variations of particulate matter, particularly calcium, aluminium, silicon and certain atmospheric aerosols can also be used as proxy palaeoclimatic indicators (Koerner, 1977b; Petit *et al*., 1981).

3.3.2.3. Dating ice cores

One of the biggest problems in any ice core study is determining the age-depth relationship. Many different approaches have been used and it is now clear that fairly accurate time scales can be developed for the last 10,000 years. Prior to that, there is increasing uncertainty about ice age. The problem lies with the fact that the age-depth is highly exponential, and ice flow models (e.g. Dansgaard & Johnson, 1969) are often needed to determine the ages of the deepest sections of ice cores. For example, the upper 1000m of a core may represent 50,000 years, whilst the next 50m may span another 100,000 year time period, due to the severe compaction, deformation and flow of the ice sheet in question.

Radio isotope dating[15], using ^{210}Pb (lead) (Crozaz & Langway, 1966), ^{32}Si (silicon), ^{39}Ar (argon) (Oeschger *et al*., 1977) and ^{14}C (carbon) (Paterson *et al*., 1977) have all been used with varying degrees of success, over different time scales, to determine the age of ice cores.

Certain components of ice cores may reveal quite distinct seasonal variations which enable annual layers to be identified, providing accurate time scales for the last few thousand years. Such seasonal variations may be found in $\delta^{18}O$ values, trace elements and microparticles (Hammer *et al*., 1978).

Where characteristic layers of known ages can be detected, these provide valuable chronostratigraphic markers against which other dating methods can be verified. So-called reference horizons have resulted from major explosive volcanic eruptions. These inject large quantities of dust and gases (principally sulphur dioxide) into the atmosphere, where they are globally dispersed. The gases are converted into aerosols (principally of sulphuric acid) before being washed out in precipitation. Hence, after major eruptions, the acidity of snowfall increases significantly above background levels (Hammer, 1977). By identifying highly acidic layers (using electrical conductivity) resulting from eruptions of known age, an excellent means of checking seasonally based chronologies is available.

3.3.3. Dendroclimatology

The study of the annual growth of trees and the consequent assembling of long, continuous chronologies for use in dating wood is called dendrochronology. The study of the relationships between annual tree growth and climate is called dendroclimatology. Dendroclimatology offers a high resolution (annual) form of palaeoclimate reconstruction for most of the Holocene.

The annual growth of a tree is the net result of many complex and interrelated biochemical processes. Trees interact directly with the microenvironment of the leaf and the root surfaces. The fact that there exists a relationship between these extremely localised conditions and larger scale climatic parameters offers the potential for extracting some measure of the overall influence of climate on growth from year to year. Growth may be affected by many aspects of the microclimate: sunshine, precipitation, temperature, wind speed and humidity (Bradley, 1985; Fritts, 1976). Besides these, there are other non-climatic factors that may exert an influence, such as competition, defoliators and soil nutrient characteristics.

There are several subfields of dendroclimatology associated with the processing and interpretation of different tree-growth variables. Such variables include tree-ring width (the most commonly exploited information source, e.g. Briffa & Schweingruber, 1992), densitometric parameters (Schweingruber *et al*., 1978) and chemical or isotopic variables (e.g. Epstein *et al*., 1976).

A cross section of most temperate forest tree trunks[16] will reveal an alternation of lighter and darker bands, each of which is usually continuous around the tree circumference. These are seasonal growth increments produced by meristematic tissues in the cambium of the tree. Each seasonal increment consists of a couplet of earlywood (a light growth band from the early part of the growing season) and denser latewood (a dark band produced towards the end of the growing season), and collectively they make up the tree ring. The mean width of the tree ring is a function of many variables, including the tree species, tree age, soil nutrient availability, and a whole host of climatic factors (Bradley, 1985). The problem facing the dendroclimatologist is to extract whatever climatic signal[17] is available in the tree-ring data from the remaining background “noise” (Fritts, 1976).

Whenever tree growth is limited directly or indirectly by some climate variable, and that limitation can be quantified and dated, dendroclimatology can be used to reconstruct some information about past environmental conditions. Only for trees growing

[15] Isotopes of certain elements, rather than being stable (like H and O) are radioactive and unstable, and decay by emitting nuclear particles (α or β particles). The rate at which they do this is invariable so that a given quantity of a radioactive isotope will decay in a known interval of time; this is the basis of radio-isotope dating methods.

[16] Not all trees are suitable for dendroclimatology. Many trees in tropical areas where growth is not seasonal do not reliably reflect varying conditions and are unsuitable as proxy indicators of palaeoclimates.

[17] The climate signal may be regarded as the response due to an identified forcing factor, as opposed to inherent random climatic variations (the noise)

near the extremities of their ecological amplitude[18], where they may be subject to considerable climatic stresses, is it likely that climate will be a limiting factor (Fritts, 1971). Commonly two types of climatic stress are recognised, moisture stress and temperature stress. Trees growing in semi-arid regions are frequently limited by the availability of water, and dendroclimatic indicators primarily reflect this variable. Trees growing near the latitudinal or altitudinal treeline are mainly under growth limitations imposed by temperature; hence dendroclimatic indicators in such trees contain strong temperature signals.

Furthermore, climatic conditions prior to the growth period may precondition physiological processes within the tree and hence strongly influence subsequent growth (Bradley, 1985). Consequently, strong serial correlation or autocorrelation may establish itself in the tree-ring record. A specific tree ring will contain information not just about the climate conditions of the growth year but information about the months and years preceding it.

Several assumptions underlie the production of quantitative dendroclimatic reconstructions. First, the physical and biological processes which link toady's environment with today's variations in tree growth must have been in operation in the past. This is the principle of uniformitarianism. Second, the climate conditions which produce anomalies in tree-growth patterns in the past must have their analogue during the calibration period. Third, climate is continuous over areas adjacent to the domain of the tree-ring network, enabling the development of a statistical transfer function relating growth in the network to climate variability inside and outside of it. Finally, it is assumed that the systematic relationship between climate as a limiting factor and the biological response can be approximated as a linear mathematical expression. Fritts (1976) provides a more exhaustive review of the assumptions involved in the use of dendroclimatology.

The general approach taken in dendroclimatic reconstruction is:

1) to collect (sample) data from a set of trees (within a tree population) which have been selected on the basis that climate (e.g. temperature, moisture) should be a limiting factor (Bradley, 1985);

2) to assemble the data into a composite site chronology by cross-dating the individual series after the removal of age effects by standardization[19]. This master chronology increases the (climate) signal to (non-climate) noise ratio (Fritts, 1971);

3) to build up a network of site chronologies for a region;

4) to identify statistical relationships between the chronology times series and the instrumental climate data for the recent period - the calibration period (Fritts, 1976);

5) to use these relationships to reconstruct climatic information from the earlier period covered by the tree-ring data, and;

6) finally, to test, or verify, the resulting reconstruction against independent data.

Bradley (1985) gives a full account of the methods (1 to 6 above) of palaeoclimate reconstruction from tree-ring analysis. This approach may be applied to all the climate-dependent tree-growth variables, specifically tree-ring width, but also wood density and isotopic measurements. The latewood of a tree ring is much denser than the earlywood and interannual variations contain a strong climatic signal (Schweingruber *et al*., 1978). Density variations are particularly valuable in dendroclimatology because they to not change significantly with tree age, and the process of standardisation (removal of growth function) can therefore be avoided.

The use of isotopic measurements in dendroclimatology also avoids the need for a standardisation process. The basic premise of isotope dendroclimatology is that since $^{18}O/^{16}O$ and D/H (deuterium/hydrogen) variations in meteoric (atmospheric) waters are a function of temperature (see also section 3.3.2.1), tree growth which records such isotope variations should preserve a record of past temperature fluctuations (Epstein *et al*., 1976). Unfortunately, isotope fractionation effects within the tree, which are themselves temperature dependent,

[18] The range of habitats that a tree can grow and reproduce within is termed the ecological amplitude.

[19] Younger trees generally produce wider rings, and this confounding influence of age on growth must be removed before analysis. Standardisation is achieved by fitting a mathematical function to the data which characterizes the low frequency variations in the data. Division of the original series by the resulting set of values then removes the growth trend.

will create problems associated with this technique (Libby, 1972).

3.3.4. Ocean Sediments

Billions of tonnes of sediment accumulate in the ocean basins every year, and this may be indicative of climate conditions near the ocean surface or on the adjacent continents. Sediments are composed of both biogenic (organic) and terrigenous (inorganic) materials. The biogenic component includes the remnants of planktonic (surface ocean-dwelling) and benthic (deep-water- or sea floor-dwelling) organisms which provide a record of past climate and oceanic circulation. Such records may reveal information about past surface water temperatures, salinity, dissolved oxygen and nutrient availability. By contrast, the nature and abundance of terrigenous materials provides information about continental humidity-aridity variations, and the intensities and directions of winds. Ocean sediment records have been used to reconstruct palaeoclimate changes over a range of time scales, from thousands of years to millions and even tens of millions of years in the past.

3.3.4.1. Palaeoclimatic reconstruction from biogenic material

Biogenic sea floor sediments are called oozes, and are usually either calcareous or siliceous in nature. Calcareous oozes consist mainly of the carbonate tests (skeletons) of millions of marine organisms, whilst siliceous oozes are made up of their silicate counterparts. For palaeoclimatic purposes, the most important materials are the tests of foraminifera (calcareous zooplankton), coccoliths (calcareous algae), radiolarians and silicoflagellates (siliceous zooplankton), and diatoms (siliceous algae).

Palaeoclimate reconstruction from the study of calcareous and siliceous tests has resulted from basically three types of analysis:

a) the oxygen isotope composition of calcium carbonate;

b) the relative abundance of warm- and cold-water species;

c) the morphological variations in particular species resulting from environmental factors.

Most work has concentrated on the study of the foraminifera, in particular oxygen isotopic analyses (e.g. Emiliani, 1977, 1978; Shackleton, 1977; Duplessy & Shackleton, 1985; Shackleton, 1988).

If calcium carbonate (of a marine organism) is crystallised slowly in water, ^{18}O is slightly concentrated in the precipitate relative to that remaining in the water. This fractionation process is temperature dependent, with the concentrating effect diminishing as temperature increases. When the organism dies, the test[20] sinks to the sea bed and is laid down, with millions of other tests, as sea floor sediment (calcareous ooze), thus preserving a temperature signal (in the form of an oxygen isotopic ratio) from a time when the organism lived. If a record of oxygen isotope ratios is built up from cores of ocean sediment, and the cores can be accurately dated[21], this will provide a method of palaeoclimate reconstruction (Urey, 1948).

As for isotope ratios from ice cores, the oxygen isotopic composition of a sample is generally expressed as a departure, $\delta^{18}O$, from the $^{18}O/^{16}O$ ratio of an arbitrary standard, $^{18}O/^{16}O_{SMOW}$ (see equation 12, section 3.3.2.1). The fractionation effect is much smaller than that which occurs during evaporation/condensation of water, and typically, $\delta^{18}O$ values are no more than a few parts per mille ($^{o}/_{oo}$) above or below the SMOW isotopic ratio.

Empirical studies relating the isotopic composition of calcium carbonate deposited by marine organisms to the temperature at the time of deposition have demonstrated the following relationship:

$$T = 16.9 - 4.2\,(\delta_c - \delta_w) + 0.13\,(\delta_c - \delta_w)^2$$

[Equation 13]

where T is the water temperature (^{o}C), δ_c is departure from SMOW of the carbonate sample and δ_w is the departure from SMOW of the water in which the sample precipitated (Shackleton, 1974). For modern analyses, δ_w can be measured directly in ocean water samples; in fossil samples, however, the isotopic composition of sea water is unknown and cannot be assumed to have been the same as it is today. In

[20] The test is the external shell of the organism.

[21] Standard techniques used to date oceanic sediment cores include palaeomagnetic analysis and radio-isotope studies, such as radiocarbon and uranium series dating methods (see Bradley (1985)).

particular, during glacial times, sea water was isotopically heavier (i.e. enriched in ^{18}O) compared to today; large quantities of isotopically lighter water were land-locked as huge ice sheet formations. Thus, the expected increase in δ_c due to colder sea surface temperatures during glacial times, is complicated by the increase in δ_w at these times.

By analysing isotopic records of deep water organisms, it is possible to resolve how much of the increase in δ_c for surface organisms was due to decreases in surface temperature and how much due to continental ice sheet formation. It is expected that bottom water temperatures ($\approx$ -1°C to 2°C) have changed very little since glacial times (the last glacial maximum being 18,000 Ka) and increases in δ_c for deep water organisms would reflect only changes in the isotopic composition of the glacial ocean. On this basis, Duplessy (1978) has concluded that 70% of the changes in the isotopic composition of surface dwelling organisms was due to changes in the isotopic composition of the oceans, and only 30% due to temperature variations.

Unfortunately, changes in the isotopic composition of the ocean reservoirs are not the only complications affecting a simple temperature interpretation of δ_c variations. The assumption that marine organisms precipitate calcium carbonate from sea water in equilibrium is sometimes invalidated. Certain vital effects of marine organisms, such as the incorporation of metabolically produced carbon dioxide, may cause a departure from the thermodynamic equilibrium of carbonate precipitation (Urey, 1947). However, by careful selection of species either with no vital effects or where the vital effects may be quantified, this problem can be avoided.

In addition to stable isotope analyses, the reconstruction of palaeoclimates can also be achieved by studying the relative abundances of species, or species assemblages (e.g. Williams & Johnson, 1975), and their morphological variations (e.g. Kennett, 1976). In the case of the latter, test coiling directions (either right-coiling (dextral) or left-coiling (sinistral)) often reveal useful proxy information about palaeo-temperatures of the oceans. Other variations include differences in test size, shape and surface structure.

3.3.4.2. Palaeoclimatic reconstruction from terrigenous material

Weathering and erosion processes in different climatic zones on the continental land masses may produce characteristic inorganic products. When these are carried to the oceans (by wind, rivers or floating ice) and deposited on the ocean floor, they convey information about the climate of their origin or transportation route, at the time of deposition (e.g. Kolla *et al*., 1979).

Terrestrial detritus dilutes the relatively constant influx of calcium carbonate; the “purity” of a calcareous ooze shows an inverse relationship with the influx of terrestrial material. Because terrestrial influx is related to climatic factors, the “purity” of calcareous sediments therefore provides a palaeoclimatic indicator (e.g. Hays & Perruzza, 1972). Hence times of high carbonate abundance indicate low terrestrial influx, i.e. low rates of continental weathering. Conversely, carbonate minima correspond to increased levels of continental weathering.

3.3.5. Terrestrial Sediments

The range of non-marine sediment studies providing relevant palaeoclimatic information is vast. Aeolian, glacial, lacustrine and fluvial deposits are, to a greater degree, a function of climate, though it is often difficult to distinguish specific causes of climatic change. Erosional features such as ancient lacustrine or marine shorelines, or glacial striae also reveal a palaeoclimatic signal. A number of these are discussed in the following sub-sections.

3.3.5.1. Periglacial features

Periglacial features are morphological features which are associated with continuous (permafrost) or discontinuous (diurnal or seasonal freezing) periods of sub-zero temperatures. Such features on which palaeoclimatic inferences can be based include: fossil ice wedges; pingos; sorted polygons; stone stripes; and periglacial involutions. Detailed descriptions of these features may be found in Washburn (1979). Unfortunately, palaeoclimate reconstructions based on such phenomena are subject to a fair degree of uncertainty. First, the occurrence of periglacial activity during the past can only indicate an upper limit on palaeotemperatures, not a lower one (Williams, R.B.G., 1975). Second, periglacial features are generally difficult to date accurately; dating of the sediments with which they are associated provides only a maximum age estimate.

3.3.5.2. Glacier fluctuations

Glacier fluctuations result from changes in the mass balance of glaciers; if snow accumulation at the source outweighs glacier ablation, glacier thickening and the forward advance of the glacier snout occurs; conversely, if the rate of ablation exceeds snow accumulation, the glacier thins and recedes. Such glacial movements will always lag behind changes in climatic factors, and different glaciers have different response times to mass balance variations (Nye, 1965). Additional complexity in interpreting glacial movements in terms of climate change exists because there are many combinations of climatic conditions which might correspond to specific mass balance fluctuations. Temperature, precipitation (snowfall) and wind speed are three factors that need to be considered (Karlén, 1980).

A record of glacial front movements is generally derived from moraines (piles of sediments carried by advancing glaciers and deposited when they retreat). Periods of glacial recession, and the magnitude of recession, are naturally much harder to identify. Additionally, repeated glacial movements may destroy evidence from earlier advances, thus limiting the period open to palaeoclimatic reconstruction. Dating glacial movements is also prone to considerable error (Matthews, 1980). Radiocarbon dates on organic material in soils which have developed on moraines provide only a minimum age for glacial advance, since a considerable time lag may exist between moraine deposition and soil formation. Lichenometry (lichens) and tephrochronology (lava flows) may sometimes be used to assist the dating of glacial events (Porter, 1979), but again, the reliability is restricted.

3.3.5.3. Lake-level fluctuations

In regions where surface water discharge (via rivers and other waterways) is restricted to inland basins instead of the oceans, changes in the hydrological balance may provide evidence for past climatic fluctuations. In these land-locked basins, water loss is almost entirely due to evaporation. During times of positive water budgets (wetter climates), lakes may develop and expand over large areas; during times of negative water budgets (drier climates), lake levels drop and the aerial expanses recede. Such palaeoclimatic studies are particularly useful in arid or semi-arid areas.

Many factors will influence the hydrological balance of a lake (Bradley, 1985). Factors affecting the rates of evaporation include temperature, cloudiness, wind speed, humidity, lake water depth and salinity. Factors influencing the rate of water runoff include ground temperature, vegetation cover, soil type, precipitation frequency, intensity and type (i.e. rain, snow etc.), slope gradients and stream sizes and numbers.

Episodes of lake growth may be identified by abandoned wave-cut shorelines, beach deposits, perched river deltas and exposed lacustrine sediments (e.g. Bowler, 1976). Episodes of lake retreat may be identified in lake sediment cores or by palaeosols and evaporites developed on exposed lake bed. As well as stratigraphy, microfossil analysis and geochemistry may be used to decipher lake level history (Bradbury *et al.*, 1981).

3.3.6. Pollen Analysis

Pollen grains and spores form the basis of another important aspect of palaeoclimate reconstruction, generally referred to as pollen analysis, or palynology. Where pollen and spores have accumulated over time, a record of the past vegetation of an area may be preserved. Often, changes in the vegetation of an area may be due to changes of climate. Interpreting past vegetation through pollen analysis may therefore offer a form of palaeoclimatic reconstruction.

Pollen grains and spores are extremely resistant to decay and are produced in huge quantities which are distributed widely from their source. A particular genus or species of plant may possess unique morphological characteristics to aid the reconstruction of past vegetation assemblages.

Differences in pollen productivity and dispersion rates pose a significant problem for palaeoclimatic reconstruction because the relative abundances of pollen grains in a deposit cannot be directly interpreted in terms of species abundance in the study area. For example, a vegetation community composed of 15% x, 35% y and 50 % z may be represented in a deposit by approximately equal amounts of pollen from x, y and z, because of the inter-species differences in pollen productivity and dispersion. To resolve this, it is necessary to have some function that relates pollen abundance and spatial distribution to species frequency. The various approaches to this have employed the uniformitarian principle i.e. the “present is the key to the past” (e.g. Davis, 1963), although the evidence that this assumption is valid is by no means conclusive.

Pollen being an aeolian (wind-blown) sediment will accumulate on any undisturbed surface. Sediments containing fossil pollen have been taken from peat bogs, lake beds, alluvial deposits, ocean bottoms and ice cores (Jacobson & Bradshaw, 1981). Where pollen has been deposited in water, one must be aware of the non-climatic effects that cause variations in pollen type and abundances. These include differential settling, turbulence mixing and the effects of burrowing organisms (Davis, 1973).

Unfortunately, the difficulties associated with pollen analysis have meant that most palaeoclimate reconstructions have proceeded in a qualitative way only - the climate was wetter/drier or warmer/colder. Sometimes it may be possible to quantify palaeoclimatic variations by the use, not of total pollen assemblages, but of individual indicator species, plants which may not be abundant but which are thought to be limited by specific climatic conditions (e.g. holly (*Ilex*), ivy (*Hedera*) and mistletoe (*Viscum*) (Iversen, 1944)).

3.3.7. Sedimentary Rocks

Many of the techniques used for palaeoclimatic reconstruction discussed in the preceding sections have only a limited time scale open to their period of study. Most ice cores are restricted to the last million years, whilst tree ring analysis can only provide proxy climate information for at best the last 10,000 years. Ocean sediments provide some of the longest proxy records available, and offer a window on palaeoclimates dating back to the age of the dinosaurs, 100 million years ago. Most older sediments, however, will have been subducted beneath overriding tectonic plates as the continents continue to drift about the Earth. To reconstruct climates older than this, therefore, one needs to look elsewhere for the evidence.

Sediments laid down on the ocean floor become progressively buried by subsequent debris transported from continental interiors. Deeply buried sediments are subjected to considerable pressures from the overlying layers, and after tens to hundreds of millions of years, the sediments are gradually lithified, forming sedimentary rocks. If, through tectonic movements, these sedimentary rocks are uplifted and exposed, scientists may study them, as they do other forms of evidence, to reconstruct past climates.

Numerous techniques of analysing sedimentary rocks are used for palaeoclimate reconstruction. Principally, rock type provides valuable insights into past climates, for rock composition reveals evidence of the climate at the time of sediment deposition. However, depositional climatic regimes vary not only due to actual climatic changes but also due to continental movements. The Carboniferous limestones and coals (evidence of warm, humid climates) of Northern England (300Ma), for example, were laid down at a time when Britain was located near the equator, whilst large scale glaciation was occurring in the high latitudes of the Southern Hemisphere (section 5.2.2.2).

The study of rock type is geologically known as facies analysis. Facies analysis investigates how the rock type changes over time, and therefore provides a potential tool for investigating past climatic change. A sedimentary formation consisting of a shale layer (fine-grained mudstone) interbedded between two sandstone layers (coarse-grained), for example, provides evidence of a changing sea level, potentially linked to climatic change (caused either by epeirogeny (section 2.6.2) or ice formation (section 2.5.2.4)). Sandstones are deposited in coastal zones where the water is shallow, whilst mudstones (shales) are deposited in deeper water of the continental shelf region. A change in the rock type in the vertical cross section must therefore reflect a change in sea level and associated coastline movements.

Other principal marker rock types include evaporites (lithified salt deposits and evidence of dry arid climates), coals (lithified organic matter and evidence of warm, humid climates), phosphates and cherts (lithified siliceous and phosphate material and evidence of ocean upwelling due to active surface trade winds) and reef limestone (lithified coral reef and evidence of warm surface ocean conditions).

As well as facies analysis, other techniques, including analysis of sedimentation rates, sediment grain morphology and chemical composition provide information on the climatic conditions prevailing at the time of parent rock weathering. In addition, some of the methods used to reconstruct past climate discussed in earlier sections may be equally applied to sedimentary rocks. For example, the type and distribution of marine and continental fossils within fossil-bearing rocks (principally limestones and mudstones, but occasionally sandstones) are valuable palaeoclimate indicators. Microfossil type, abundance and morphology may also be studied, and palaeotemperatures derived from their oxygen isotope analysis (section 3.3.4.1).

3.4. Conclusion

Chapter 3 has reviewed the various methods and techniques used both to construct contemporary climate from instrumental records, and to reconstruct palaeoclimates from proxy data sources. Whenever an indicator, whether instrumental or proxy, is chosen to represent some aspect of the climate, one must be sure that there exists a physical basis for the choice of that indicator i.e. variations in the record of that indicator truly reflect variations in that aspect of the climate one is attempting to measure.
In all climatic and palaeoclimatic analyses, it should be the aim of the investigator to maximise the (climate) signal to (non-climatic) noise ratio, without compromising the validity of the data.

In chapter 5 the use of the proxy methods for palaeoclimate reconstruction across geologic time are discussed. Particular attention is paid to the climates and climate changes of the Quaternary. First, however, chapter 4 investigates another tool used to assist the scientist in understanding climate change: climate modelling.

4.1. Introduction

Climate models attempt to simulate the behaviour of the climate system. The ultimate objective is to understand the key physical, chemical and biological processes which govern climate. Through understanding the climate system, it is possible to: obtain a clearer picture of past climates by comparison with empirical observation, and; predict future climate change. Models can be used to simulate climate on a variety of spatial and temporal scales. Sometimes one may wish to study regional climates; at other times global-scale climate models, which simulate the climate of the entire planet, will be desired.

Henderson-Sellers & Robinson (1986), Henderson-Sellers & McGuffie (1987) and Schneider (1992) provide detailed introductory discussions of the methods and techniques involved in climate modelling.

There are three major sets of processes which must be considered when constructing a climate model:

1) **radiative** - the transfer of radiation through the climate system (e.g. absorption, reflection);

2) **dynamic** - the horizontal and vertical transfer of energy (e.g. advection, convection, diffusion);

3) **surface process** - inclusion of processes involving land/ocean/ice, and the effects of albedo, emissivity and surface-atmosphere energy exchanges.

These are the processes fundamental to the behaviour of the climate system that were discussed in chapter 1.

The basic laws and other relationships necessary to model the climate system are expressed as a series of equations. These equations may be empirical in derivation based on relationships observed in the real world, they may be primitive equations which represent theoretical relationships between variables, or they may be a combination of the two. Solving the equations is usually achieved by finite difference methods; it is therefore important to consider the model **resolution**, in both time and space i.e. the time step of the model and the horizontal/vertical scales.

4.2. Simplifying the Climate System

All models must simplify what is a very complex climate system. This is in part due to the limited understanding that exists of the climate system, and partly the result of computational restraints. Simplification may be achieved in terms of spatial dimensionality, space and time resolution, or through parameterisation of the processes that are simulated.

The simplest models are zero order in spatial dimension. The state of the climate system is defined by a single global average. Other models (see section 4.4) include an ever increasing dimensional complexity, from 1-D, 2-D and finally to 3-D models. Whatever the spatial dimension of a model, further simplification takes place in terms of spatial resolution. There will be a limited number of, for example, latitude bands in a 1-D model, and a limited number of gridpoints in a 2-D model. The time resolution of climate models varies substantially, from minutes to years depending on the nature of the models and the problem under investigation.

In order to preserve computational stability, spatial and temporal resolution have to be linked. This can pose serious problems when systems with different equilibrium time scales have to interact as a very different resolution in space and time may be needed.

Parameterisation involves the inclusion of a process as a simplified (sometimes semi-empirical) function rather than an explicit calculation from first principles. Subgridscale phenomena such as thunderstorms, for example, have to be parameterised as it is not possible to deal with these explicitly. Other processes may be parameterised to reduce the amount of computation required.

Certain processes may be omitted from the model if their contribution is negligible on the time scale of interest. For example, there is no need to consider the role of deep ocean circulation whilst modelling changes over time scales of years to decades. Some models may handle radiative transfers in great detail but neglect or parameterise horizontal energy transport. Other models may provide a 3-D representation but contain much less detailed radiative transfer information.

Given their stage of development, and the limitations imposed by incomplete understanding of the climate system and computational constraints, climate models cannot yet be considered as predictive tools of future climate change. They can, however, offer a valuable window on the workings of the climate system, and of the processes that have influenced both past and present climate.

4.3. Modelling the Climatic Response

The ultimate purpose of a model is to identify the likely response of the climate system to a change in any of the parameters and processes which control the state of the system. The climate response occurs in order to restore equilibrium within the climate system. For example, the climate system may be perturbed by the radiative forcing associated with an increase in carbon dioxide (a greenhouse gas) in the atmosphere. The aim of the model is then to assess how the climate system will respond to this perturbation, in an attempt to restore equilibrium.

The nature of the model can be one of two modes. In equilibrium mode, no account is taken of the energy storage processes which control the evolution of the climate response with time. It is assumed that the climate response occurs instantaneously following the system perturbation. The inclusion of energy storage processes allows the model to be run in transient mode, simulating the development of the climate response with time. Generally, both equilibrium and transient models will be run twice, once in a control run with no forcing, then over the test run including forcing and perturbation of the climate system. (For a further discussion of the concepts of equilibrium and transient climate responses, the reader may wish to refer back to section 2.7.)

The climate sensitivity and the role of feedback are critical parameters whatever the model formulation. In the most complex models, the climate sensitivity will be calculated explicitly through simulations of the processes involved. In simpler models this factor is parameterised by reference to the range of values suggested by the more complex models. This approach, whereby more sophisticated models are nested in less complex models, is common in the field of climate modelling.

4.4. The Climate Models

It is often convenient to regard climate models as belonging to one of four main categories:

1) energy balance models (EBMs);
2) one dimensional radiative-convective models (RCMs);
3) two-dimensional statistical-dynamical models (SDMs);
4) three-dimensional general circulation models (GCMs).

These models increase in complexity, from first to last, in the degree to which they simulate the particular processes and in their temporal and spatial resolution. The simplest models permit little interaction between the primary processes, radiation, dynamics and surface processes, whereas the most complex models are fully interactive.

It is not always necessary or beneficial to the analysis, however, to invariably choose the more sophisticated models. The choice of model depends upon the nature of the analysis For example, the simpler models, unlike the 3-D GCMs, may be run many times in sensitivity studies which test the influence of modelling assumptions. For simulation experiments, which require complex modelling of the physical, chemical and biological processes inherent in the climate system, more sophisticated models may indeed be more appropriate. Computational cost is always an important factor to consider when choosing a climate model.

4.4.1. Energy Balance Models

Energy balance models (EBMs) simulate the two most fundamental processes governing the state of the climate:

a) the global radiation balance (i.e. between incoming solar and outgoing terrestrial radiation) (see section 1.4), and;

b) the latitudinal (equator-to-pole) energy transfer (see section 1.5).

EBMs are usually 0-D or 1-D in form. In 0-D EBMs, the Earth is considered as a single point in space, and in this case only the first process listed above is modelled. In 1-D models, the dimension included is latitude. Temperature for each latitude band is calculated using the appropriate latitudinal value for the various climatic parameters (e.g. albedo, energy flux etc.). Latitudinal energy transfer is usually estimated from a linear empirical relationship based on the difference between the latitudinal temperature and the global average temperature. Other factors that may be included in the model equation are the time-dependent energy storage and the energy flux into the deep ocean.

4.4.2. Radiative-Convective Models

Radiative-convective models (RCMs) are 1-D or 2-D, with height the dimension that is invariably present. RCMs simulate in detail the transfer of energy through the depth of the atmosphere, including:

a) the radiative transformations that occur as energy is absorbed, emitted and scattered, and;

b) the role of convection, energy transfer via vertical atmospheric motion, in maintaining stability.

2-D RCMs also simulate horizontally-averaged energy transfers.

RCMs contain detailed information about the radiation streams or energy cascades (see section 1.4) - the fluxes of terrestrial and solar radiation - that occur throughout the depth of the atmosphere. By considering parameters such as surface albedo, cloud amount and atmospheric turbidity, the heating rates of a number of atmospheric layers are calculated, based on the imbalance between the net radiation at the top and bottom of each layer. If the calculated vertical temperature profile (lapse rate) exceeds some stability criterion (critical lapse rate), convection is assumed to take place (i.e. the vertical mixing of air) until the stability criterion is no longer breached. This process is called the convective adjustment. (A fuller explanation of convective processes and lapse rates in the atmosphere is given in Barry & Chorley, 1992)

As with EBMs, time-dependent calculations (energy storage and fluxes into the deep ocean) may be made, depending on the nature of the analysis. RCMs are most useful in studying forcing perturbations which have their origin within the atmosphere, such as the effects of volcanic pollution.

4.4.3. Statistical-Dynamical Models

Statistical-dynamical models (SDMs) are generally 2-D in form, with usually one horizontal and one vertical dimension, though variants with two horizontal dimensions have been developed. Standard SDMs combine the horizontal energy transfer modelled by EBMs with the radiative-convective approach of RCMs. However, the equator-to-pole energy transfer is simulated in a more sophisticated manner, based on theoretical and empirical relationships of the cellular flow between latitudes.

Parameters such as wind speed and wind direction are modelled by statistical relations whilst the laws of motion are used to obtain a measure of energy diffusion as in an EBM. Hence the description statistical-dynamical. They are particular useful in investigations of the role of horizontal energy transfer, and the processes which disturb that transfer directly.

4.4.4. General Circulation Models

General circulation models (GCMs) represent the most sophisticated attempt to simulate the climate system. The 3-D model formulation is based on the fundamental laws of physics:

a) conservation of energy;
b) conservation of momentum;
c) conservation of mass, and;
d) the Ideal Gas Law[22].

These are the same physical laws which formed the discussion of the climate system in theoretical terms in chapter 1. A series of primitive equations describing these laws are solved, resulting in an estimate of the wind field, which is expressed as a function of temperature. Processes such as cloud formation are also simulated.

To compute the basic atmospheric variables at each gridpoint requires the storage, retrieval, recalculation and re-storage of 10^5 figures at every time-step. Since the models contain thousands of grid points[23], GCMs are computationally expensive. However, being 3-D, they can provide a reasonably accurate representation of the planetary climate, and unlike simpler models, can simulate global and continental scale processes (e.g. the effects of mountain ranges on atmospheric circulation) in detail. Nevertheless, most GCMs are not able to simulate synoptic (regional)

[22] The Ideal Gas Law is given by: $PV = nRT$, where P is the pressure of the gas, V is the volume of the gas, T is the temperature of the gas, n is the number of moles of the gas (with 1 mole containing 6.2×10^{23} atoms), and R is Joules constant (8.2 Joules mol^{-1} $Kelvin^{-1}$).

[23] Typical global scale GCMs will be made up of a 5^o latitude by 5^o longitude horizontal grid, and a third dimension of 6 to 15 vertical layers.

meteorological phenomena such as tropical storms, which play an important part in the latitudinal transfer of energy and momentum. The spatial resolution of GCMs is also limited in the vertical dimension. Consequently, many boundary layer processes must be parameterised.

To date, most GCMs only model the atmospheric component of the climate system, and have therefore been equilibrium in nature (e.g. Mitchell *et al*., 1990). Recently, a new breed of transient coupled atmosphere-ocean GCMs has evolved, in an attempt to simulate more accurately the climate system (e.g. Bretherton *et al*., 1990; Gates *et al*., 1992). Such coupled models create major computational difficulties because the atmosphere and ocean components respond over vastly different time scales.

In summary, GCMs can be considered to simulate reasonably accurately the global and continental-scale climate, but confidence is lacking in the regional detail.

4.5. Confidence and Validation

Although climate models should aid understanding in the processes which control and perturb the climate, the confidence placed in such models should always be questioned. Critically, it should be remembered that all climate models represent a simplification of the climate system, a system which indeed may ultimately prove to be too complex to model.

Given that many of the processes which are modelled occur over time scales so long that it is impossible to test model results against real-world observations, it is also arguable that climate modelling is, in some respects. philosophically suspect. Model performance can be tested through the simulations of shorter time scale processes but short-term performance may not necessarily reflect long-range accuracy.

Climate models must therefore be used with care and their results interpreted with due caution. Margins of uncertainty must be attached to any model projection. Uncertainty margins can be derived by the comparison of the results of different model experiments or through sensitivity studies, in which key assumptions are altered to determine the role they play in influencing the final climatic response.

Validation of climate models (testing against real-world data) provides the only objective test of model performance. As far as GCMs are concerned, validation exercises have revealed a number of deficiencies in their simulations of present-day conditions (e.g. Gates *et al*., 1990):

a) modelled stratospheric temperatures tend to be lower than equivalent instrumental observations;

b) modelled mid-latitude westerlies tend to be too strong; easterlies are too weak;

c) modelled sub-polar low pressure systems in winter tend to be too deep and displaced east, and;

d) day-to-day variability is lower than in the real world.

Finally, it has been observed that some models suffer from climate drift. The background climate shifts as the simulation proceeds, despite the absence of any climate forcing.

4.6. Conclusion

Through much of the history of climate modelling, division between modelling and observational studies has hampered the development of both sides. The coupling of these theoretical and empirical disciplines, to test both model accuracy and understanding gained from analysis of observational data, has only recently been addressed. Much of the discrepancy between the climate model and real world is the result of this division. Such a divide must be bridged if accurate forecasts of future climates are to be produced.

5.1. Introduction

Chapter 5 is a review of the major episodes of climate change that have occurred through Earth history, prior to the advent of the use of instrumental data. Although the contemporary climate change record spans only a tiny fraction (10^{-7}) of Earth history, in light of its current and future implications to the global society, it forms the basis of a separate chapter (chapter 6).

This chapter begins with a discussion of climate changes over the longest time scales (10^6 to 10^8 years), those associated with continental drift, followed by an investigation into the Quaternary glacials (last 2 Ma), assumed to be driven by orbital forcing mechanisms (see chapter 2). Finally, century and millennia time scales of climate change over the last 10,000 years will be reviewed. For all sections, the evidence for climate change will be drawn from both empirical investigations (see chapter 3) and modelling studies (see chapter 4).

5.2. Pre-Quaternary Climates

The pre-Quaternary spans 99.95% of Earth History (see appendix). Nevertheless, knowledge of pre-Quaternary climates is significantly poorer than that of the last 2 million years. Going further back in time, more and more evidence for past climate change will have been removed by subsequent climatic episodes.

5.2.1. Precambrian Climates

The Precambrian comprises 85% of Earth history, yet very little can be said about palaeoclimates from these ancient times, and what is known is not known with any degree of confidence. In passing, it may be noted that there is evidence for two major periods of glaciation, one at 2.7 to 2.3 billion years (Ga) (Frakes, 1979), the other more recently at 0.9 to 0.6Ga (Frakes, 1979; Williams, 1975b). The former of these occurred at a time when the current tectonic regime of continental drift was in early growth. There appears to be no evidence for glaciation at other times during the Precambrian, a feature that has puzzled palaeoclimatologists, since it is generally assumed that the Sun was considerably fainter at that time. For the interested reader, Frakes (1979) and Crowley & North (1991) offer reliable accounts of Precambrian climates.

5.2.2. Phanerozoic Climates

The **Phanerozoic** covers the last 570 Ma of Earth history, made up of three **Eras**, the Palaeozoic (570 to 225Ma), Mesozoic (225 to 65Ma) and Cenozoic (65Ma to present) (see appendix). Each Era is divided up into different Periods, and each Period is divided into a number of Epochs. The most recent Period of the geologic record is the Quaternary (considered in section 5.3), comprising the Pleistocene (2Ma to 10Ka) (see section 5.3.1) and Holocene (10Ka to present) (see section 5.3.2). In this section, only pre-Quaternary climates will be considered.

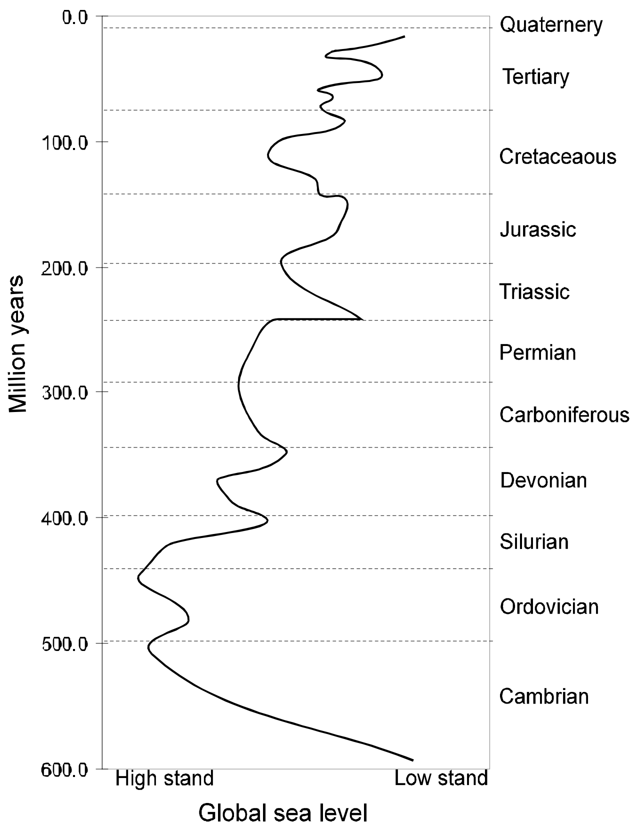
The most likely candidates for long term climate forcing include those associated with continental drift, orogeny (section 2.6.1) and epeirogeny (section 2.6.2). The growth of mountain ranges may affect atmospheric circulation patterns; movement of land masses into high latitude region may initiate strong ice-albedo feedbacks; variations in the rate of sea-floor spreading may alter ocean bathymetry and global carbon dioxide emissions.

The Phanerozoic has witnessed the evolution of a major tectonic cycle, involving the coming together of land masses to form a single super-continent known as Pangea (about 220Ma), followed by its disintegration, resulting in the configuration of continents that exist today. Associated with this continental drift were variations in tectonic activity and ocean-floor spreading, with subsequent fluctuations in ocean bathymetry and global sea level (Vail *et al*., 1977). Figure 5.1 shows how global sea level has varied during the course of the Phanerozoic.

5.2.2.1. Palaeozoic Climates

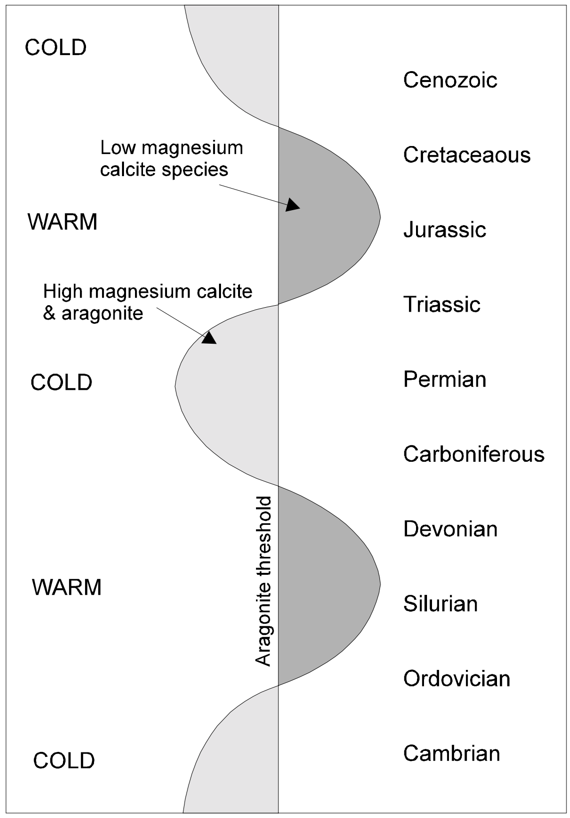
During the early Palaeozoic, about 530Ma, the Northern Hemisphere was entirely oceanic north of about 30°N palaeolatitude. Most of today's Southern Hemisphere continents (India, Antarctica, Australia, Africa, South America) had formed into a super-continent known as Gondwanaland. Sea levels were at or near an all time high, perhaps reflecting increased volcanic activity/oceanic ridge expansion after the break-up of a postulated late Precambrian supercontinent (Bond *et al*., 1984).

Figure 5.1. *Global sea level variations during the Phanerozoic (after Hallam, 1984)*



In response to increased tectonic activity, atmospheric carbon dioxide may also have been significantly higher during the early Palaeozoic. Geochemical evidence that links changes in preservation of various species of carbonate minerals with the partial pressure of carbon dioxide in surface waters (and hence the atmosphere) supports this conclusion. Wilkinson & Given (1986) have estimated that the atmospheric concentration of carbon dioxide may have been 10 times higher during the early Palaeozoic than at present. In general, variations in the preservation of marine carbonate species support the sea level/CO_2 connection assumed for the Phanerozoic (Sandberg, 1983), and Fischer (1982) has correlated these variations (see Figure 5.2) to conjectured climate (greenhouse-icehouse) supercycles. During low sea stands and lower atmospheric CO_2 concentration, high-magnesium carbonate and aragonite[24] shells are preferentially precipitated from ocean water. Conversely, calcite shells are more common during high sea stands and accompanying higher partial pressure of CO_2 (pCO_2).

Figure 5.2. *Greenhouse-icehouse supercycles (after Fischer, 1982)*



With such elevated levels of atmospheric CO_2 one would expect a considerable warming associated with greenhouse radiative forcing. Indeed, it seems that the early Palaeozoic was dominated by a maritime[25] climate associated with the expansion of the seaways in the tropics. It is therefore surprising to find evidence (Frakes, 1979) of global ice growth during the Ordovician Period (about 440Ma), at a time when sea levels were at there highest (Figure 5.1), with presumably CO_2 concentrations elevated as well. Considerable ice sheet expansion occurred over northern Africa, at that time situated in the vicinity of the South Pole[26].

[24] Aragonite and calcite are two crystallographic forms of calcium carbonate ($CaCO_3$)

[25] Global climate may be designated as being either maritime or continental. In the latter, sea levels are lower exposing larger areas of land masses which experience larger ranges in temperature due to their smaller heat capacity (relative to the oceans). Maritime climates, conversely, are normally associated with global transgressions (high sea stands), covering much of the global land mass, thereby reducing temperature extremes. Maritime climates are typically warmer, on average, than continental ones.

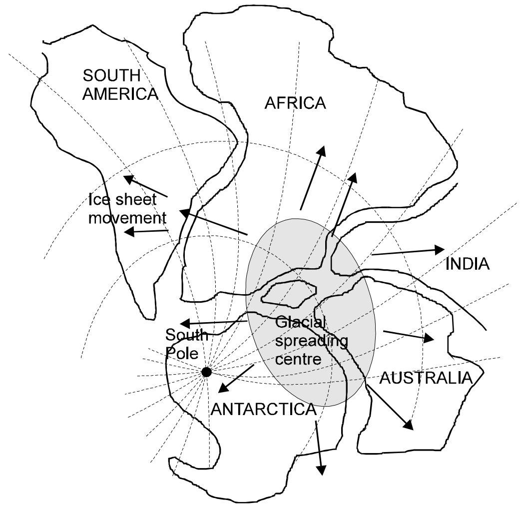
[26] Past latitudinal positions of continents can be reconstructed from palaeomagnetic information contained within the minerals (principally

Energy balance models (EBMs) have gone some way to resolve this paradox. Simulations in which the South Pole is located in coastal regions generate low (below freezing) summer temperatures due to the reduced seasonal temperature cycle associated with the coastal environment (Crowley *et al*., 1987). The influence of enhanced greenhouse forcing due to elevated pCO_2 affects only annual temperature, rather than the seasonal cycle. In addition, the continental flooding associated with the early Palaeozoic marine transgression would increase the water surface area, thus reducing further the seasonality, and hence summer temperatures, of the polar region. Geological evidence (Caputo & Crowell, 1985) appears to support the hypothesis of a coastally situated South Pole.

For most of the Silurian and Devonian Periods (410 to 345Ma), there is no evidence for glaciation (Caputo & Crowell, 1985). Although the palaeogeography for the Devonian is uncertain, a South Pole located in central Africa would increase seasonality and summer temperatures, thereby preventing the formation of ice sheets (Crowley *et al*., 1987). Expansion of land plants also occurred for the first time during the Devonian. Newly vegetated areas would decrease surface albedo (Posey & Clapp, 1964) allowing increased absorption of short-wave radiation. The intensification of the hydrological cycle would also influence climate (Shukla & Mintz, 1982).

A second, and larger ice age began towards the end of the Palaeozoic during the Carboniferous Period (about 305Ma), and continued well into the Permian. Powell & Veevers (1987) have postulated that orogenic uplift in Australia and South America triggered the climatic event. The aerial extent of glaciation covered much of Gondwanaland, and glacial striations (erosional markings due to the movement of glaciers over bedrock) dating from this time have been found in Antarctica, Australia, Africa, India and South America (see Figure 5.3). Again, it appears that the period of major glaciation was associated with a South Pole located at the edge of the Gondwanan supercontinent (Crowley *et al*., 1987). Such a combination of modelling research and empirical reconstruction appears to favour the hypothesis that climate change was driven by tectonic movements. Additionally, sea levels were at that time falling (Figure 5.1), associated with a slowing down of sea-floor spreading, and presumably a lowering of pCO_2 (and greenhouse forcing), as all the continents began to converge.

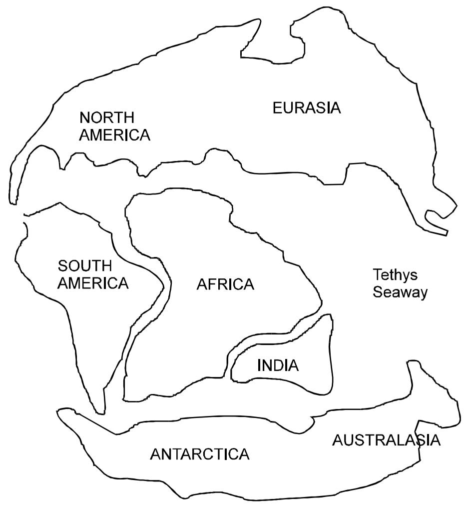
Figure 5.3. *Carboniferous glaciation on Gondwanaland (after Sullivan, 1974)*



5.2.2.2. Mesozoic Climates

By the early Mesozoic (Triassic Period), the final suturing of all continents to form Pangea (see Figure 5.4) was complete (about 220Ma). The combination of a gigantic landmass and lower global sea level (associated with reduced rates of tectonic movement) should have resulted in extremely continental climates, with associated aridity. This conjecture is consistent with geological data (Parrish *et al*., 1982); there is abundant evidence of extensive red beds (aeolian deposits) and evaporites (salt deposits) throughout the Permian and Triassic, which form in arid environments.

Figure 5.4. *Pangea supercontinent*



rich in iron) of certain rocks, which are sensitive to the Earth's magnetic field.

Nevertheless, the assumption that the climate of the early Mesozoic was hot and arid on a global scale has been questioned by various modelling studies (Crowley *et al*., 1989; Kutzbach & Gallimore, 1989). These simulations have suggested that with such extreme continentality and increased seasonality prevalent, considerable amounts of sea ice would be expected at high latitudes. Even with enhanced anticyclonic ocean gyres, transporting heat into high latitude regions, one cannot escape the fact that polar continental areas should have experienced sub zero temperatures at some time during the annual cycle. Despite the lack of empirical glacial evidence, the premise that the early Mesozoic was a time of arid warmth must remain ambiguous.

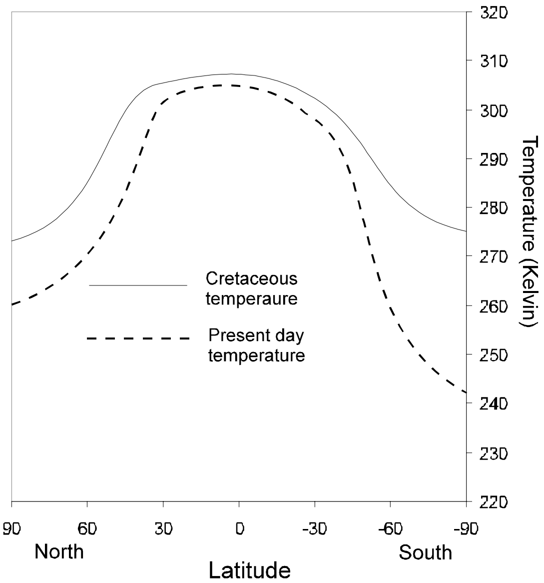
From geological evidence (e.g. Lloyd, 1982; Savin, 1977) it appears that the postulated warmth of the Triassic and Jurassic Periods continued into the Cretaceous (136 to 65Ma). During the Jurassic and Cretaceous, global sea level rose again (Figure 5.1), presumably associated with increased sea-floor spreading as Pangea began to break up.

The elevated sea level may have created a near transglobal equatorial seaway (called the Tethys Sea), flooding large parts of western Europe, North Africa and North America, bringing considerable warmth and moisture to low latitude regions. There is also considerable geological evidence for warmer temperatures in higher latitudes during the mid-Cretaceous (e.g. Barron, 1983). From oxygen isotope records, Savin (1977) has indicated that deep water temperatures at 100Ma may have been as high as 20^{o}C. Figure 5.5 illustrates the comparison between present day surface temperatures and those during the Cretaceous estimated empirically.

As for the early Mesozoic, however, modelling studies do not conclusively predict high latitude ice-free environments during the Cretaceous. Indeed, the hypothesis that increased high latitude warmth is due to changes in land-sea distribution appears to be only partially supported by climate simulations (Barron *et al*, 1981; Berner *et al*, 1983; Schneider *et al*., 1985). Simulated climate changes for seasonal experiments do not result in temperature patterns compatible with geologic data; temperatures are still too low in high latitudes.

Various explanations have been proposed for the discrepancy, and subsequently incorporated into further modelling studies. Two of these include ocean circulation changes and the role of CO_2. Of these, only an elevated atmospheric CO_2 concentration could come close to reconciling the models with geological evidence. High levels of CO_2 do not altogether seem unreasonable, considering the high global sea level and ensuing break-up of Pangea (presumably due to increased tectonic activity) (Berner *et al*., 1983). In addition to increased outgassing of CO_2, the reduced continental area (due to the global marine transgression) would result in a decreased rate of weathering of silicates and removal of CO_2 from the atmosphere.

Figure 5.5. *Comparison of Cretaceous temperatures with the present (after Barron & Washington, 1985)*



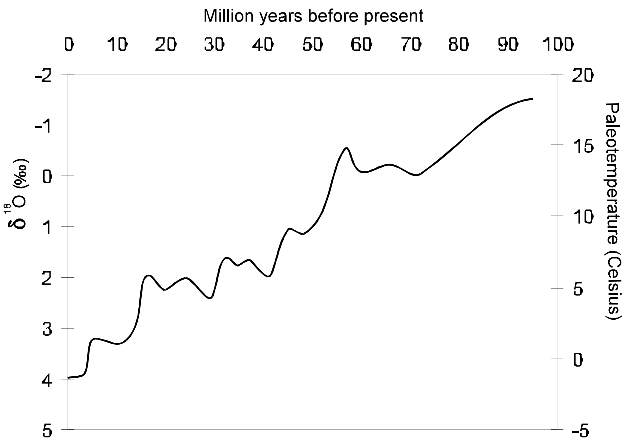
Unfortunately, there is little reliable evidence to support the CO_2 model. Kominz (1984) has estimated that average rates of the velocities of major tectonic plates were higher in the late Cretaceous and ocean ridge volumes were greater. In addition, Cretaceous sea beds were dominated by calcite minerals (Sandberg, 1983), implying higher aqueous, and consequently atmospheric, CO_2 concentrations (see also Figure 5.2). Nevertheless, further intercomparison of different CO_2 proxy records is necessary if full evaluation of the CO_2/Cretaceous climate paradigm is to be evaluated.

5.2.2.3. Cenozoic Climates

The Cenozoic Era is divided into the Tertiary (65 to 2Ma) and Quaternary (2Ma to present) periods. The climate of the Quaternary is reviewed in section 5.3. In this section, the climatic deterioration and global cooling of the Tertiary Period is investigated. The ensuing discussion will serve as a useful introduction to the ice age climates of the Quaternary and their orbital forcing mechanisms.

Although late Cretaceous temperatures were cooler than the mid-Cretaceous thermal maximum (120 to 90Ma), values remained relatively high into the early Cenozoic. In particular, evidence from oxygen isotope records (Douglas & Woodruff, 1981) revealed that deep sea ocean temperatures were at least 10°C to 15°C warmer than they are today (see Figure 5.6). Early Cenozoic sea surface temperatures around Antarctica were also considerably warmer than today (Shackleton & Kennet, 1975) .

Figure 5.6. *Deep ocean temperatures during the last 100 million yeras (after Douglas & Woodruff, 1981)*



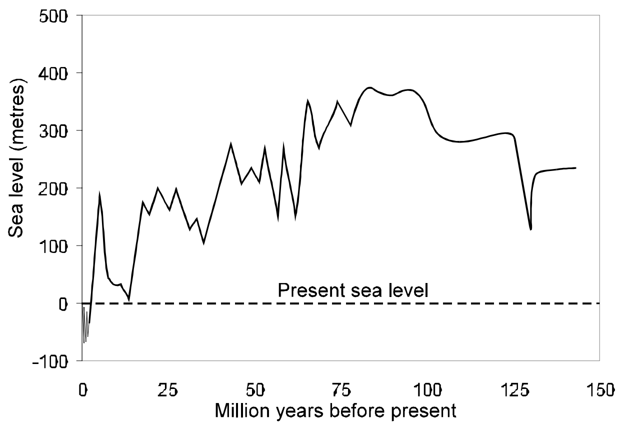
The early Eocene (55 to 50Ma) was the warmest period during the Cenozoic. Various climatic indices suggest that tropical conditions extended 10° to 15° of latitude poleward of their present limits. Eocene tropical assemblages of foraminifera and coccoliths have been found in North Atlantic sediments (Haq *et al*., 1977). Vertebrate fossils of alligators and flying lemurs have been found from a site on Ellesmere Island, west of Greenland (Dawson *et al*., 1976).

It has been suggested that the early Eocene warming may have resulted from an increase in atmospheric CO_2, due to a significant reorganisation in tectonic plate motion, as North America separated from the Eurasian plate (Berner *et al*., 1983). Unfortunately, little proxy evidence exists to test this premise.

During the late Eocene and Oligocene Epochs (40 to 25Ma) a transition occurred between the warm periods of the early Cenozoic and the cold periods of the later Cenozoic. Antarctic glaciation may have been initiated at about this time. Evidence of ice-rafted debris in the Southern Ocean has been dated at 34Ma (Hayes & Frakes, 1975). Changes in equatorial Pacific planktonic and benthonic $\delta^{18}O$ records suggest that considerable and rapid continental ice formation was taking place at this time (Keigwin, 1980).

Another significant cooling transition occurred during the Miocene (15 to 10Ma). A dramatic increase in the $\delta^{18}O$ record between 14 and 15Ma has been interpreted both as the rapid growth of the Antarctic ice sheet (Shackleton & Kennet, 1975) and a deep water cooling event (4 to 5°C) (Moore *et al*., 1987). Both the Oligocene and Miocene cooling events reveal themselves in the sea level record in Figure 5.7 (Haq *et al*., 1987). The rapidity of these falls in sea level precludes the possibility that changes in ocean bathymetry were the cause. Rather, increases in continental ice volume must be proposed.

Figure 5.7. *Global sea levels during the last 150 million years (after Haq et al., 1987)*



The exact timing of the onset of mid-latitude Northern Hemisphere Glaciation is uncertain, but some oxygen isotope records suggest a date towards the end of the Pliocene (3 to 2Ma). Variations in planktonic abundances indicate that large changes in sea surface temperature were occurring prior to 2.4Ma (Raymo *et al*., 1986) whilst ice-rafted debris from the Norwegian Sea has been dated at 2.8 to 2.6Ma (Jansen *et al*., 1988).

Northern Hemisphere glaciation proceeded throughout the Pleistocene (Quaternary Period) and continues today. Much scientific work has been devoted to the analysis of the Pleistocene glaciations; this is reviewed in the next section. In the meantime, however, one issue remains unresolved. What was the cause of the Cenozoic climatic deterioration?

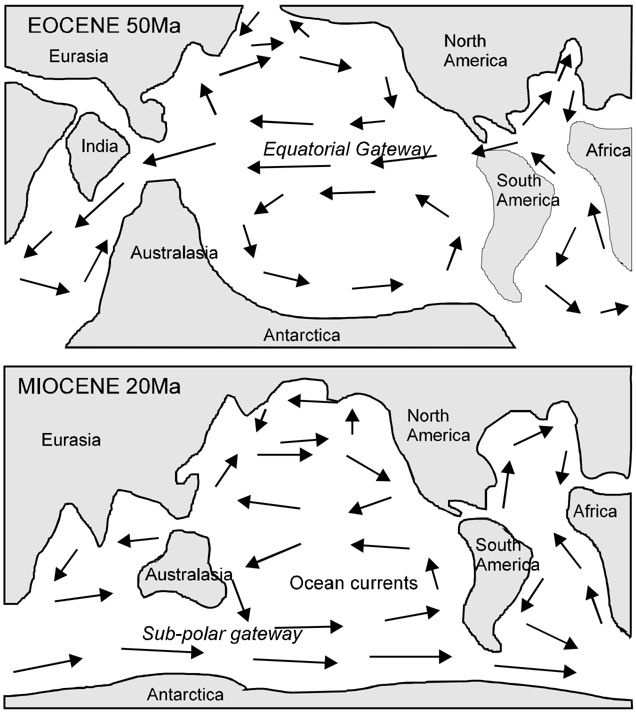
A number of hypotheses have been proposed to account for the Cenozoic cooling. Changes in land-sea distribution have been extensively modelled using GCMs. However, the increase in high latitude land masses (with initiation of ice-albedo feedbacks) and the decrease in continentality (with accompanying decreases in seasonality and summer temperatures) do not seem to account for all of the temperature changes (Barron, 1985).

Variations in ocean circulation due to changes in continental positions have also been considered as a causal mechanism for Cenozoic climate change. Indeed, geologists have long been interested in the effects of ocean gateways (between two continental land masses) on past climates. Experiments with ocean circulation models (Seidov, 1896) also indicate that changing continental position can have a significant effect on ocean heat transport.

At the beginning of the Cenozoic a circum-equatorial seaway existed (the Tethys Sea) that spanned the entire Earth. Equatorial currents flowed unimpeded several times around the globe before diverting north and south. Waters in this equatorial seaway would have experienced considerable heating, resulting in a more even distribution of warmth across the latitudes. As the continents continued to separate the Tethys Sea gradually closed, cutting off the circum-equatorial currents. Latitudinal heat transport was thus reduced. During the late Eocene and Oligocene, the Gondwanan continents fully separated, and by 25 to 30Ma, the opening of the Drake Passage between Antarctica and South America allowed the formation of a circum-polar ocean flow (Figure 5.8). With heat transport effectively cut off from the high southern latitudes, the latitudinal temperature gradient increased and glaciation in the Southern Hemisphere ensued.

Changes in continental topography during the Cenozoic has been proposed as another cause of the long-term climatic deterioration. Specifically, model simulations suggest that uplift of the Colorado and Tibetan Plateaux initiated winter cooling of North America, northern Europe, northern Asia and the Arctic Ocean, as a result of atmospheric circulation changes (Ruddiman and Kutzbach, 1989).

Figure 5.8. *The formation of a circum-polar ocean circulation*



Accompanying the global escalation in orogenic uplift (during a period which witnessed the building of the Alpine and Himalayan mountain chains) was an increase in the rate of weathering of silicate rocks (Raymo *et al*., 1988). CO_2 is removed from the atmosphere during this geochemical process (see section 2.6.1). Cenozoic mountain building may therefore have indirectly reduced the greenhouse forcing of the Earth-atmosphere system, enhancing global cooling. It was recognised in earlier sections that atmospheric CO_2 concentration was probably several times greater than at present, but little reliable proxy data exist to constrain the timing and magnitude of the postulated fall in pCO_2.

In summary, it is likely that a combination of processes - changes in land-sea distribution, ocean heat transport, orography and CO_2 - are involved in the long-term evolution of the Cenozoic climate, and probably the climates of the earlier Phanerozoic. These processes operate over time scales involving tens or even hundreds of millions of years. It is evident that model simulations are sometimes in disagreement with the proxy climatic records, whilst the validity of such records may often be questioned. If a thorough and unequivocal reconstruction pre-Quaternary climates is to be achieved, these problems will need to be addressed by future researchers in palaeoclimatology.

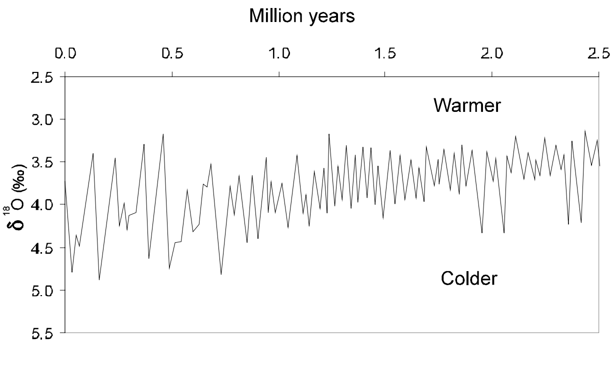
5.3. Quaternary Climates

Proxy reconstructions for the climate of the Quaternary Period are considerably more abundant and reliable than for earlier periods. The Quaternary spans the last 2Ma of Earth history and is separated into two Epochs, the Pleistocene (2Ma to 10Ka) and the Holocene (10Ka to present). In the preceding section, evidence that the Earth had entered a long-term climatic decline, was reviewed. For at least the last 10Ma, and probably much longer, the Earth has been in the grip of a long-term ice age. This ice age has continued during the Quaternary, and is in evidence today, as ice caps remain at both poles. Nevertheless, within the Quaternary, global climate has fluctuated between times of relative warmth and frigidity. This section will review the evidence for these fluctuations, and discuss their probable causal mechanisms.

5.3.1. Pleistocene Glacials and Interglacials

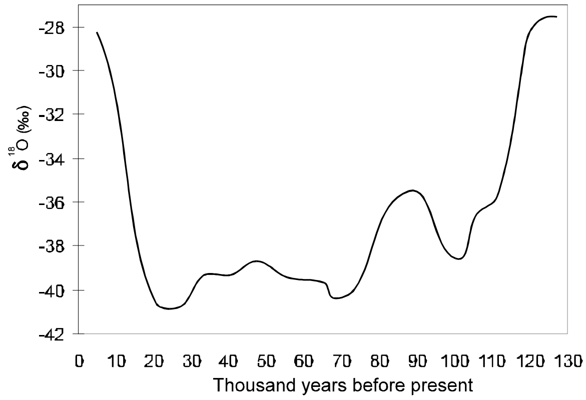
Numerous proxy records have been used to reconstruct Pleistocene climate variations. Before the widespread use of deep-sea sediment cores, it was known that there had been a number of fluctuations of Pleistocene glaciers. Windblown loess deposits have also been used to demonstrate climate change on the continents (Kukla, 1970). However, it was the advent of oxygen isotope analysis of nannofossils in deep-sea cores which really marked the breakthrough in Pleistocene climate reconstruction. Figure 5.9 shows a record of $\delta^{18}O$ fluctuations for the last 2.5Ma. Within it, distinct cycles are evident that demonstrate changes both in ocean temperature and global ice volume (recall section 3.3.4.1).

Figure 5.9. *Ocean $\delta^{18}O$ (temperature) record for the last 2.5 million years (after Raymo et al., 1990)*



Analysis of the $\delta^{18}O$ record indicates two basis climate states, one glacial and one interglacial. Evidence for these bistable climate state is further provided by oxygen isotope analysis of numerous ice cores. Figure 5.10 shows a $\delta^{18}O$ profile along the Camp Century (Greenland) ice core (Dansgaard *et al.*, 1984) for the last 130,000 years. The record clearly reveals the last major interglacial period at about 120 thousand years (Ka) and the ensuing glaciation.

Figure 5.10. *Camp Century ice core $\delta^{18}O$ (temperature) record for the last 130 thousand years, smoothed curve (after Dansgaard et al., 1984)*

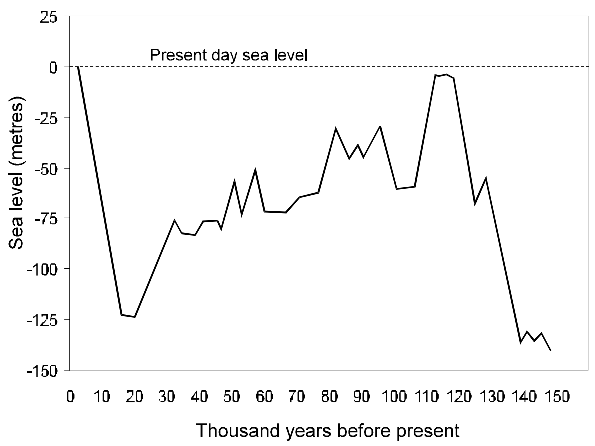


Sea level estimates from isotope analysis reconstructions (Shackleton, 1988) bear a striking resemblance to palaeo-temperature and ice volume curves (see Figure 5.11). The crucial issue to researchers was to determine the cause of such pronounced variations in the climate. A mechanism was needed which could force changes in climate over periods of tens to hundreds of thousands of years. Today, it is generally accepted that the glacial-interglacial transitions of the Pleistocene Epoch are driven by variations in the Earth's orbit around the Sun.

5.3.1.1. Orbital Variations

The hypothesis that gravitational effects of planetary bodies can cause orbital perturbations that periodically vary the geographic distribution of incoming solar radiation, and so act as an external forcing mechanism on the Earth's climate, was discussed in section 2.5.2. These Milankovitch Cycles include the Precession of the Earth (with periodicities of 19 and 23Ka), the Obliquity or tilt of the Earth's axis (41Ka) and the Eccentricity of the orbital ellipse (100 and 413Ka).

Figure 5.11. *Global sea level during the last 150 thousand years (after Shackleton, 1988)*



In a landmark paper, Hays *et al*. (1976) clearly demonstrated the existence of spectral peaks in the proxy ice volume record from a sediment core from the Indian Ocean that matched the significant Milankovitch periodicities (Figure 2.2). Also apparent in the record was the fairly consistent phase relationship between insolation, sea surface temperature and ice volume; each preceded the next by about 2 to 4Ka. Somewhat more puzzling, however, was the evidence that the 100Ka contribution to the total variance of the record far exceeded that expected *a priori* from a simple linear relationship between insolation and ice volume. Of the three orbital periodicities, eccentricity has the least potential as a climate forcing mechanism. Nevertheless, its phase-locking with tilt and precession indicates a primary orbital influence on ice volume fluctuations (Imbrie *et al*., 1984).

To complicate matters still further, the strength of the 100Ka cycle has not been constant throughout the Pleistocene Epoch. Distinct cycles of 100Ka duration are only present for the last 700Ka. Before this time, fluctuations with a periodicity of about 40Ka (obliquity) seem to dominate (see Figure 5.9).

During the last two decades, a number of modelling studies have attempted to explain the relationship between astronomical forcing and climate change, and to reproduce some of the wealth of geological evidence that supports the Milankovitch hypothesis. Most of these studies have been motivated by the recognition that the amount of insolation perturbation associated with the 100Ka cycle (eccentricity) is insufficient to cause a climate change of ice-age magnitude. Two different modelling approaches are evident.

The first considers that ice volume changes are primarily **driven** by orbital forcing. Imbrie *et al*. (1984) demonstrated that at the 23Ka and 41Ka periods, ice volume (as measured by $\delta^{18}O$) responds linearly to orbital forcing. At 100Ka the effect is non-linear. 100Ka power in the frequency spectrum (see Figure 2.2) is generated by transmission of the 19Ka and 23Ka frequencies (precession) through a non-linear system (Wigley, 1976). The non-linearity of the climatic response is most evident at the terminations of glacial episodes, with a rapid transition to the interglacial state (see Figures 5.10 and 5.11). To explain such rapid transitions, internal climatic feedback loops must be inferred.

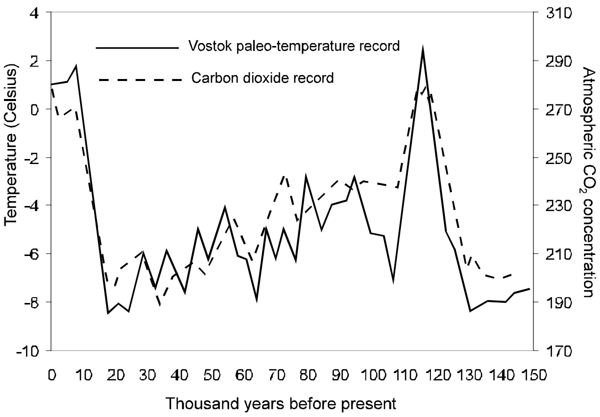
An entirely different approach to modelling the 100Ka climate variability involves the hypothesis that glacial-interglacial fluctuations are the consequence of non-linear internal interactions in a highly complex system (Nicolis, 1984; Saltzman, 1985). In this situation, ice volume fluctuations are **modulated**, rather than driven, by orbital forcing. Quasi-periodic fluctuations within various components of the climate system (e.g. variations in CO_2, ocean circulation, surface- and deep-water temperatures) are phase-locked to the changes in insolation due to the 100Ka cycle.

Whether climate change is driven or modulated by orbital forcing is as yet unresolved. Nevertheless, it does appear that the astronomical signal in some way interacts with "internally" generated climate change. Indeed, it is clear *a priori* that insolation changes associated with orbital variations are alone, not sufficient to account for the Pleistocene glacial-interglacial transitions (e.g. Hoyle, 1981).

5.3.1.2. CO_2 Feedbacks

Variations in atmospheric CO_2 (measured from air bubbles trapped in ice cores) have generated perhaps the most significant interest regarding the internal mechanisms proposed to explain Pleistocene climate change. The recognition that CO_2 variations appeared to correlate extremely well with temperature over the last 160Ka (Barnola *et al*., 1987; Figure 5.12) suggested that changes in greenhouse forcing was playing a part in Pleistocene climate regulation. During the last major glacial-interglacial transition (at approximately 14Ka), when global temperatures increased by about 4 to 5^{o}C, atmospheric CO_2 concentration rose 80ppm, from 200ppm to 280ppm (see Figure 5.12).

Figure 5.12. *Vostok CO_2 (after Barnola et al., 1987) and temperature (after Jouzel et al., 1987) records for the last 160 thousand years*



Hypotheses as to the causes of CO_2 changes, and the phase relationships of CO_2, ice volume and temperature, have passed through many stages over the last decade or so. All of them involve a redistribution of carbon between its major global reservoirs - atmosphere, ocean, land biosphere and oceanic sediments. (The carbon cycle is fully reviewed in chapter 6, in the context of contemporary climate change.) Most importantly, the time scale of carbon redistribution must match that of the CO_2 variations; tectonic processes, which may have accounted for CO_2 variations throughout the Phanerozoic (see section 5.2) can therefore be discounted.

Over time scales of millennia, the ocean can be considered to be the most important carbon reservoir, in terms of its potential to influence CO_2 changes. Dissolved inorganic carbon in the ocean, or "total CO_2" (ΣCO_2) includes HCO_3^- (carbonate ion) (90%), CO_3^{2-} (bicarbonate ion) (9%) and dissolved CO_2 gas ($CO_{2\,(aq)}$) (1%). These agents exist in equilibrium, according to Equation 14.

$$CO_{2\,(aq)} + CO_3^{2-} + H_2O \rightarrow 2HCO_3^-$$

[Equation 14]

Atmospheric CO_2 is directly regulated by the chemical composition of the surface sea water, according to Equation 15.

$$CO_2 + H_2O \rightarrow H_2CO_3\ \{CO_{2\,(aq)}\}$$

[Equation 15]

Equations 14 and 15 combine to form Equation 16, the governing equation for carbonate chemistry in the ocean system.

$$CO_3^{2-} + H_2CO_3 \rightarrow 2HCO_3^-$$

[Equation 16]

Equation 15 can be simplified to:

$$C_w = sC_a \qquad \text{[Equation 17]}$$

where C_w is the CO_2 concentration (or partial pressure) in water, C_a is the CO_2 concentration (or partial pressure) in air and s is the solubility (equilibrium) constant for Equation 15.

There are two ways in which the concentration of atmospheric CO_2 (C_a) can be altered: varying s or varying C_w. The solubility of CO_2 increases with decreasing temperature, thus decreasing C_a for constant C_w. One could postulate that changes in surface ocean temperature associated with glacial-interglacial transitions could indirectly influence the strength of greenhouse forcing; cooler surface oceans during a glacial episode would lower CO_2 levels in the atmosphere, thereby initiating a positive feedback. Unfortunately, evidence from the last major glacial-interglacial transition, at about 14Ka, does not support this hypothesis. The magnitude of the global sea surface temperature change (inferred from oxygen isotope records of sediments cores) would only account for about 25% of the change in CO_2 (Broecker, 1982). In addition, changes in salinity due to changes in ice volume affect CO_2 concentrations in an opposite direction to the effect of temperature fluctuations. Consequently, another mechanism to explain the nature of the CO_2 variations is needed.

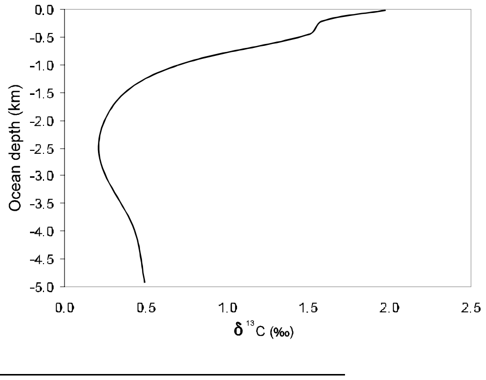
A 1% increase in ΣCO_2 in the surface ocean increases C_w, and consequently C_a by about 10%. Minor changes in the total dissolved inorganic carbon may therefore have considerable effect on atmospheric CO_2. ΣCO_2 in the surface ocean is smaller than for the entire ocean (Broecker & Peng, 1982). Surface-dwelling plankton (which make up 90% of ocean biota) extract carbon for both photosynthesis and carbonate shell formation. On death, these organisms sink to the sea floor and the carbon is oxidised, returning it to the deep ocean water. Consequently, a

ΣCO_2 gradient between surface and deep ocean exists (Figure 5.13a).

Figure 5.13a. *Ocean ΣCO_2 profile*

If the biological productivity of the surface ocean could vary, this would have a direct effect on CO_2 in the atmosphere. To explain reduced glacial CO_2 levels in the atmosphere, Broecker (1982) suggested that transfer of nutrients from eroding continental shelves, exposed by the marine regression (low sea stand), would raise the productivity level of the ocean, with the increased biomass in the surface waters acting as a "biological pump" to draw down atmospheric CO_2. Broecker (1974) argued that since marine plankton preferentially use ^{12}C for photosynthesis, leaving surface water enriched in ^{13}C (see Figure 5.13b), past productivity levels in the ocean could be monitored by measuring the differences in $\delta^{13}C$ between the surface ($\delta^{13}C_{planktonic}$) and deep ($\delta^{13}C_{benthic}$) waters[27]. Increased surface productivity would increase the difference between surface and deep $\delta^{13}C$, $\Delta\delta^{13}C$ (as measured from the carbonate tests of planktonic and benthic organisms respectively).

Figure 5.13b. *Ocean $\delta^{13}C$ profile*



Unfortunately, changes in $\Delta\delta^{13}C$ over the last 160Ka do not correspond well with variations in atmospheric CO_2. In addition, deposition and erosion of shelf sediments is a slow process and depends upon major changes in sea level. Such a response time is too long to account for the more rapid fluctuations in CO_2. Finally, evidence of nutrient variations, in particular dissolved phosphate, does not seem to reveal significant changes between glacial and interglacial nutrient content of the oceans (Boyle & Keigwin, 1987).

A number of other factors have been considered which may influence atmospheric CO_2 concentrations. Two of the more significant hypotheses are detailed here.

1) *Changes in the organic carbon/carbonate ratio of biogenic material falling out of the upper water column* (Broecker & Peng, 1982). Increased upwelling in equatorial oceans, due to enhanced glacial trade winds[28], would raise biogenic productivity. Carbonate-secreting plankton are less common in more productive regions of the ocean, at the expense of siliceous plankton. Thus, enhanced upwelling might cause a shift in the dominant phytoplankton group, which in turn would cause a change in the organic carbon/carbonate rain ratio. With less carbonate (CO_3^{2-}) falling out of the upper water column, its concentration would increase, thereby raising the alkalinity of the surface waters[29]. Atmospheric CO_2 is very sensitive to changes in alkalinity (Sarmiento *et al*., 1988); an increase in ocean alkalinity would cause a decrease in pCO_2.

2) *Enhanced utilisation of nutrients in high-latitude surface waters* (Wenk & Siegenthaler, 1985; Toggweiler & Sarmiento, 1985). At present, not all nutrients are consumed in high-latitude oceans because vertical ocean circulation rates are fast enough to prevent equilibrium between phytoplankton fertility and nutrient levels. During the last glacial episode, production rates of the North Atlantic Deep Water (NADW) decreased by as much as 50% (Boyle & Keigwin, 1987), allowing a more efficient nutrient utilisation to take place. Consequently, the glacial North Atlantic was nutrient-depleted whilst deep ocean water became nutrient-enriched (Figure 5.14). The subsequent reduction in surface ΣCO_2 would

[27] The δ notation used here is the same as that for oxygen isotope analysis.

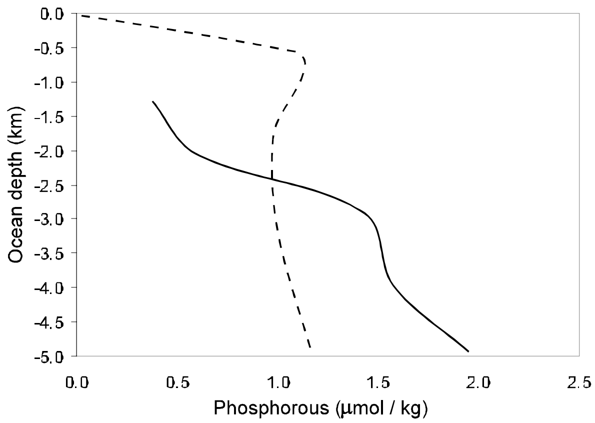
[28] The premise here is that a stronger equator to pole temperature gradient during glacial times would enhance the trade winds.

[29] Ocean alkalinity is given by: $[HCO_3^- + CO_3^{2-}]$.

lower pCO_2, thereby reducing the atmospheric concentration of CO_2. Model calculations show that a 50% reduction in high latitude vertical ocean circulation would cause a 50ppm decrease in atmospheric CO_2, almost two-thirds of the observed CO_2 change at the last glacial-interglacial transition (Sarmiento & Toggweiler, 1984; Siegenthaler & Wenk, 1984). Possible explanations for the slow down in vertical thermohaline circulation include the larger extent of sea ice, and colder sea surface temperatures (see section 2.6.4).

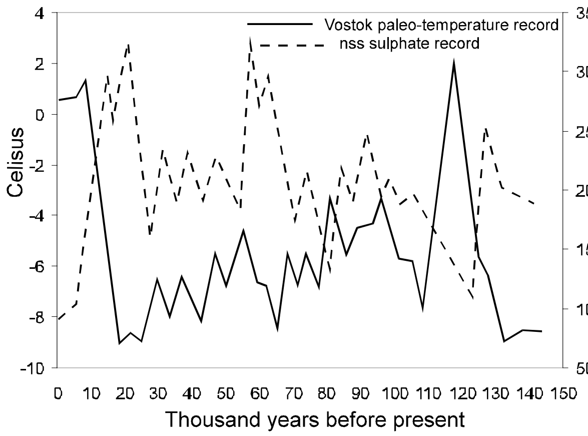
Again, both these suppositions have their problems. On current evidence, it seems implausible that ocean circulation changes in the small area of the North Atlantic Ocean could account for the entire glacial-interglacial difference in pCO_2. Changes in ocean circulation in the Southern Ocean around Antarctica would need to be invoked, but there is no clear evidence for reduced vertical exchange rates during the last glacial period. In addition, enhanced utilisation of surface nutrients, and subsequent increase in high-latitude biogenic productivity, predicts an increase in the $\delta^{13}C_{planktonic}$. The available isotope data seem not to confirm this expectation.

Figure 5.14. *Ocean phosphorous profile: glacial and present (after Boyle & Keigwin, 1987)*



Despite the obvious difficulties in linking changes in ocean productivity (and ocean circulation and ocean chemistry) with variations in atmospheric CO_2, other evidence appears to suggest a definite relationship. Most significantly, an index of non-sea salt (nss) sulphate from an ice core at Vostok, Antarctica (Legrand *et al.*, 1988) correlates well with the ice core deuterium/temperature record (Figure 5.15). Nss sulphates are produced by marine phytoplankton (section 1.3.3); an index of nss sulphate is therefore a measure of surface ocean primary productivity. High levels of nss sulphate coincide with cooler temperatures. Hence, it appears that ocean productivity (and presumably atmospheric CO_2) is in some way related to glacial-interglacial changes in climate.

Figure 5.15. *Vostok ice core temperature, smoothed curve (after Jouzel et al., 1987) and non-sea-salt sulphate (after Legrand et al., 1988) records for the last 150 thousand years*



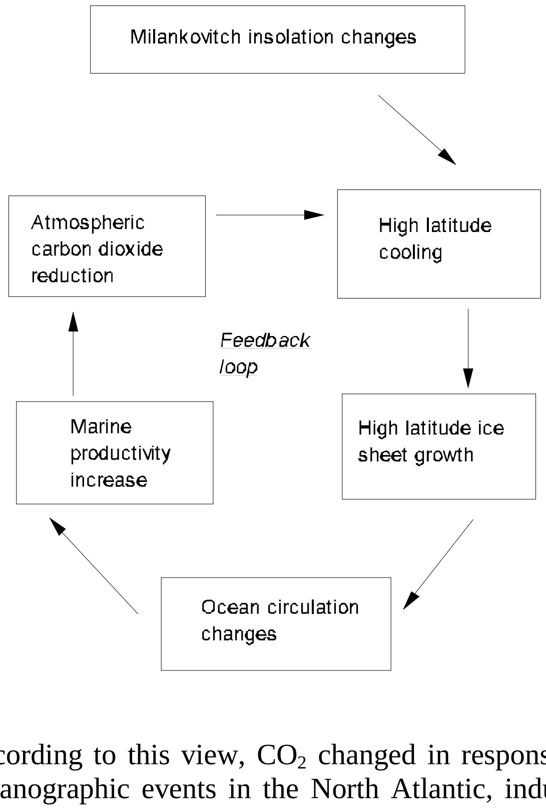
Nss sulphates form Dimethyl sulphide (DMS) aerosols in the atmosphere (section 1.3.3). Consequently, increases in glacial ocean productivity may initiate another feedback mechanism, in addition to the productivity-CO_2 connection. DMS represents the primary source of cloud condensation nuclei (CCN) in the marine atmosphere (Bates *et al.*, 1987). Greater ocean productivity would thereby increase cloud cover, increasing the planetary albedo, and initiating further global cooling. Nevertheless, the magnitude of this feedback is uncertain. Indeed, increased cloud cover may enhance greenhouse warming of the lower atmosphere (section 2.7) and hence the reality of this mechanism must, for the while, remain in doubt.

5.3.1.3. Coupled Internally-Externally Driven Climate Change

Despite the obvious relationship between atmospheric CO_2 and global temperature over the course of glacial-interglacial cycles (Figure 5.12), the precise role of CO_2 in climate regulation remains uncertain. Specifically, it is not clear whether changes in CO_2 were the effect of fluctuations in climate, which then acted as a feedback factor, or whether they were the primary cause of temperature variations, albeit driven by Milankovitch-type insolation variations. Clearly, changes in ocean circulation, productivity and chemistry all played a significant role in the events, particularly during glacial-interglacial transitions. Nevertheless, existing data from ice core does not allow the resolution of the phase relationship between CO_2 changes and temperature variations (Siegenthaler, 1988). The cause-and-effect relationships operating within the internal dynamics of the climate system over such time scales of millennia and tens of millennia must therefore remain unsettled.

Figure 5.16 schematises a possible scenario of events associated with the last glacial (Würm) episode. Changes in insolation at high latitudes of the Northern Hemisphere, steered by Milankovitch-type variations, acted as a global pacemaker, driving or regulating the complex non-linear internal variations within the climate system. Such internal variations may include the growth and decay of ice sheets, changes in vertical ocean circulation and productivity, and CO_2 feedback.

Figure 5.16. *Postulated climate change at the last interglacial-glacial transition 120 thousand years ago*



According to this view, CO_2 changed in response to oceanographic events in the North Atlantic, induced by the Northern Hemisphere cooling (due to external insolation changes) and ice growth (due to ice-albedo feedback) at the beginning of the last glacial episode, and then acted as a climatic feedback. The time delay between climate and CO_2 may have been rather small; the chemistry of the ocean has been shown to adjust itself within a few centuries to changes in circulation or productivity (Siegenthaler & Wenk, 1984). Clearly, however, more higher resolution proxy data and improved modelling studies are required if the full history of the Würm glacial, and indeed the glacial-interglacial cycles of the Pleistocene, is to be unravelled.

5.3.2. Holocene Climates

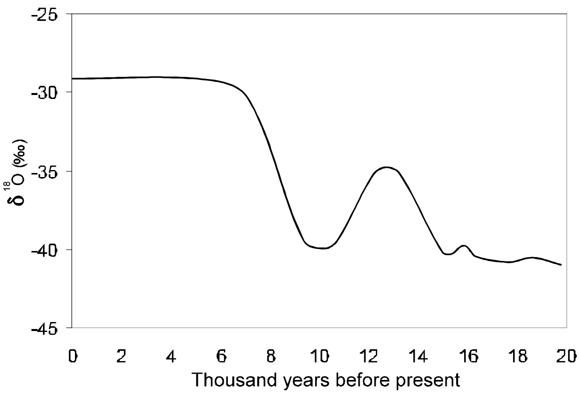
The last glacial maximum occurred at 18Ka. By 14Ka, the ice sheets of the Northern Hemisphere had already began retreating, and by 6Ka, global climate reached a Holocene thermal maximum. The Holocene Epoch is thus delineated as a period of relative warmth, and is considered to be an example of an interglacial episode that occurs in conjunction with glaciations throughout the Quaternary. The penultimate interglacial occurred at about 120Ka (Figure 5.12).

A number of text books discuss in great detail the climatic events of the last 10,000 years. Lamb (1982) provides a particularly useful commentary, whilst Bradley & Jones (1992) discuss the evidence for climatic variations since A.D. 1500. In this section, I shall be concerned with century to millennia time scales, and in this respect, I will review only the most significant climate changes which serve to illustrate the relevant climate forcing mechanisms.

5.3.2.1. The Younger Dryas Event

Although deglaciation had been taking place for at least 4,000 years, a rapid deterioration (cooling) in climate occurred at about 10 to 11Ka, roughly marking the boundary between Holocene and Pleistocene Epochs. This event is known as the Younger Dryas Cooling. The North Atlantic polar front readvanced far southward to approximately 45^oN (only 5 or 10^o north of the glacial maximum position), and cooling was especially strong in the circum-subpolar basin (Broecker *et al*., 1988). Figure 5.17 shows a record of carbon and oxygen isotopes of surface waters that provide evidence for North Atlantic cooling (Boyle & Keigwin, 1987). In particular, increases in ocean productivity (higher $\delta^{13}C$) coincide with cooler temperatures in Greenland (higher $\delta^{18}O$).

Figure 5.17. *Camp Century (Greenland) $\delta^{18}O$ record for the last 20 thousand years, smoothed curve (after Boyle & Keigwin, 1987)*



The Younger Dryas cooling event may have been caused by meltwater-induced changes in the atmosphere-ocean circulation. During the early stages of deglaciation, much of the meltwater from the Laurentide (North American) Ice Sheet emptied primarily into the Gulf of Mexico. By about 11Ka, the ice margin had retreated sufficiently to open up drainage into the St. Lawrence Seaway, near Newfoundland (Broecker *et al.*, 1988). The subsequent outflow of meltwater could have created a low-salinity lens in the subpolar North Atlantic. Such low-salinity water would not be dense enough to sink through the pycnocline (salinity density gradient at about 100m depth). The lack of overturn might reduce or temporarily shut down production of NADW. Geochemical evidence (Boyle & Keigwin, 1987) indeed supports this scenario.

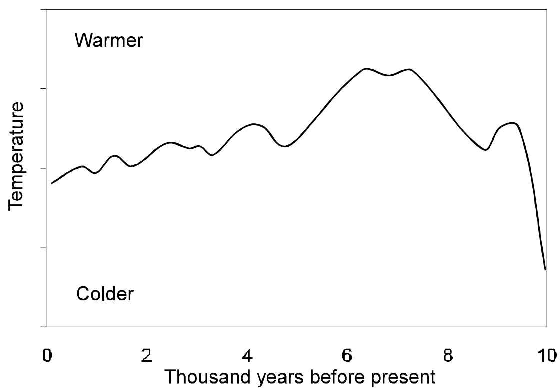
Subsequent drawdown of CO_2 from the atmosphere would then reduce greenhouse forcing of the climate. Unfortunately, evidence of CO_2 reduction during the Younger Dryas (Stauffer *et al.*, 1985) remains inconclusive.

Since the production of NADW results in an export of cold deep water from the North Atlantic basin and an import of warm South Atlantic surface waters (Figure 2.6), a reduction in NADW production rates might be expected to result in decreased flow of South Equatorial Current across the equator (today known as the Gulf Stream), cooling waters north of the equator and warming them to the south. Such a reduction in northward heat transport would enhance the cooling of the North Atlantic region. Indeed, Broecker (1987) has proposed that such a mechanism of circulation mode interchanges may be a primary factor in the regulation of millennia scale climate change (section 2.6.4).

5.3.2.2. Mid-Holocene Thermal Maximum

Although remnants of the Laurentide Ice Sheet did not disappear until about 7 Ka, the early to mid-Holocene (4,500 to 10,000 years) has often been considered to have been warmer than the last 4,500 years. A thermal maximum occurred at about 6 to 7 Ka (Figure 5.18). Conclusions about the mid Holocene warmth are based on several lines of evidence - latitudinal displacements of vegetation zones (Ritchie *et al.*, 1983) and vertical displacements of mountain glaciers (Porter & Orombelli, 1985).

Figure 5.18. *Holocene temperature optimum (after Mörner & Wallin, 1977)*



Quantitative estimates of mid-Holocene warmth (COHMAP, 1988) suggest that the Earth was perhaps 1 or 2°C warmer than today. Most of this warmth may primarily represent seasonal (summer) warmth rather than year-round warmth. Accompanying the higher global temperatures were significant changes in precipitation patterns, most noticeably in the monsoon belt of Africa and Asia. Reconstructions from palaeo-lake levels and latitudinal vegetation shifts (Ritchie & Haynes, 1987) suggest that these regions were considerably wetter than they were during the arid conditions of the last glacial maximum (18Ka), when moisture availability from cooler Northern Hemisphere sub-tropical oceans was reduced (Street-Perrott & Perrott, 1990).

5.3.2.3. Late Holocene Neoglacial Fluctuations

Records for the last 4,500 years generally indicate that temperatures were lower than the Holocene thermal maximum. A general cooling, known as the Iron Age neoglaciation, occurred between 2,500 and 4,500 years ago. A moderate climate amelioration followed near the dawn of the Roman Empire, before a return to cooler climates during the second half of the first millennia A.D. (the Dark Ages). Thence followed the Medieval optimum (1100 to 1300 A.D.), in which European temperatures reached some of the warmest levels for the last 4,000 years.

5.3.2.4. The Little Ice Age

Beginning about 1450 A.D. there was a marked return to colder conditions. This interval is often called the Little Ice Age, a term used to describe an epoch of renewed glacial advance. Although many regions of the world experience cooling during the period 1450 to 1890 A.D., its use has been criticised because it could not conclusively be considered an event of global significance (Bradley & Jones, 1992). Nevertheless, within the framework of Holocene climate fluctuations, its terminology may be justified. Since this period overlaps with the advent of instrumental measurement of climate indices, considerably greater confidence can be attributed to the proxy reconstructions (e.g. tree rings, ice cores, periglacial features) through the use of calibration techniques.

There is considerable evidence that the Little Ice Age consisted of two main cold stages of about a century's length (Bradley & Jones, 1992). These occurred in the seventeenth an nineteenth centuries, with relative warmth arising in the sixteenth and eighteenth centuries. Glaciers advanced in Europe, Asia and North America, whilst sea ice in the North Atlantic expanded with detrimental effects for the colonies of Greenland and Iceland (Lamb, 1982).

5.3.2.5. Holocene Climate Forcing Mechanisms

The causes of centennial to millennial scale climate fluctuations are not well understood. Three forcing mechanisms have been considered: volcanism (section 2.6.3), solar variability (section 2.5.3) and internal ocean circulation dynamics (section 2.6.4).

Detailed studies of volcanic events of the last 100 years indicate that significant hemispheric and global cooling occurs a number of months after large eruption (Sear *et al.*, 1987). Such cooling has been associated with the increase in atmospheric aerosol content, due to the emissions of large quantities of dust and sulphur dioxide during eruptions (section 2.6.3). Less certain is whether longer periods (centuries or millennia) of enhanced volcanism may have a climatic connection.

Porter (1986) demonstrated a significant correlation between glacial advances of the last 1,400 years and the frequency of volcanic eruptions as estimated by changes in acid content[30] of a Greenland ice core. However, changes in oceanic productivity can also vary the release of DMS to the atmosphere (section 5.3.1.2), which after conversion to sulphate can be deposited in the ice cores (Charlson *et al.*, 1987). Thus, while individual spikes in the acidity record probably represent specific volcanic eruption, the background acidity level in the ice core could represent a response to climate change rather than a cause of climate change.

It is also necessary to examine from a modelling viewpoint how episodic volcanic impulses are transmitted into the lower frequency part of the spectrum; observations from both the instrumental record and ice cores indicate that the volcanic signal and the detectable temperature effect are manifested in only the first couple of years after an eruption (Legrand & Delams, 1987). One possible explanation involves some type of ice-albedo feedback (Robock, 1978), in which short-term cooling events affects sea ice cover. Due to the longer time response of sea ice, the effect may then be felt over a period of decades.

Another mechanism which may have forced Holocene climate changes is solar variability. A number of solar periodicities were reviewed in section 2.5.3, including variations in sunspot number, solar diameter and total solar output. Eddy (1976) noted that a well-known minimum of sunspot activity at the end of the seventeenth century (the Maunder Minimum) coincided with one of the coldest periods of the Little Ice Age. Proxy records of ^{14}C and ^{10}Be (Beryllium) isotopes suggest a link between solar output changes and Holocene climate variations (Eddy, 1977). However, considering the uncertainty associated with a solar-climate link, due partially to the somewhat

[30] Acidity changes are assumed to be due to atmospheric sulphate changes from volcanic eruptions.

improper use of statistical analysis, it would seem unwise at this stage to investigate further any causal mechanism of century-millennia climate change involving solar variability.

The third mechanism invoked to explain climatic changes during the Holocene involves ocean circulation. Because the surface mixed layer and deep ocean have different response times, essentially high frequency random fluctuations in climatic forcing may generate responses at lower frequencies (Gaffin *et al.*, 1986). Such internal responses are typically non-linear in nature. Although capable of self-generation through such non-linear dynamics, ocean circulation changes may often be driven or regulated by external forcing perturbations. In this sense, changes in ocean circulation could be regarded as a primary climatic feedback.

Mode changes in ocean circulation have been considered (Broecker, 1987) as forcing agents of climatic change during the last glacial-interglacial transition, and in particular the Younger Dryas Cooling (section 5.3.2.1). Boyle and Keigwin (1987) also suggest that the neoglaciation between 4,500 and 2,500 years ago was associated with decreased NADW production. Quinn & Neal (1992) have proposed that the quasi-periodic nature of El Niño events may account for some decadal scale climate change during the last 500 years. Nevertheless, improved proxy records will be required if the reality of these internal oscillations and their climatic response is to be determined.

5.4. Conclusion

Throughout this chapter, some of the evidence for episodes of climate change during the course of Earth History has been reviewed, with particular emphasis on the last 2 million years. For each era, the time scale of climate change was explicitly defined and the relevant climatic forcing mechanisms evaluated.

The longest time scales of climate change (10^7 to 10^8 years) involve the shifting of the positions of continents and associated mountain building and ocean bathymetry variations. Such mechanisms can account for the supercycles of greenhouse and icehouse worlds during the Phanerozoic (last 550Ma), and the Cenozoic (last 65Ma) climatic deterioration.

Over time scales of 10^4 to 10^5 years, the so-called Milankovitch orbital periodicities, which vary the amounts of insolation at the Earth's surface, are invoked to explain the glacial-interglacial transitions that have been well-documented for the Pleistocene Epoch (2Ma to 10Ka). Such external forcing acts as a pacemaker to many non-linear perturbations within the climate system, including CO_2 feedbacks, ocean circulation changes and ice-albedo effects, which are required to account for the magnitude of glacial-interglacial climate change.

In the last 10,000 years, climate has fluctuated over decadal to millennial time scales. Volcanism, solar variability and ocean circulation changes have all been suggested, to a varying degree of certainty, as causes of these shorter episodes of climate change.

The state of the climate at any one moment is determined by a combination of all these forcing factors that affect the radiation budget of the global climate system.

6.1. Introduction

The last chapter examined episodes of past (palaeo) climate change reconstructed from proxy indicators. Since the late seventeenth century, and more commonly since the nineteenth century, instrumental records (section 3.2) have been kept that document more recent climate changes. Typically, this is labelled the study of contemporary climate change and it will be the focus of this last chapter. In particular, it will review the evidence for anthropogenic warming during the twentieth century, and possible future climate change. The numerous scientific assessments from the Intergovernmental Panel on Climate Change (e.g. IPCC, 1990a, 1990b, 1992, 1995) provide excellent reference material for those wishing to study contemporary climate change in more detail.

6.2. The Greenhouse Effect

Today a consensus amongst scientists is growing that mankind, through the emissions of greenhouse gases (section 1.2.1), is enhancing the Earth's natural greenhouse effect (section 1.2.4). Before investigating the evidence for anthropogenic climate change and global warming, it is useful to review the nature of the Earth's greenhouse effect.

Solar energy in the form of visible and ultraviolet radiation is partially absorbed by the atmosphere (25%) and Earth's surface (45%), and partially reflected by the atmosphere (25%) and Earth's surface (5%) (Figure 1.3). Energy absorbed at the Earth's surface is re-radiated back to the atmosphere as infrared radiation, longer in wavelength and lower in intensity, since the Earth is much cooler than the Sun (section 1.2.3).

Certain gases in the atmosphere are transparent to the incoming short-wave solar radiation but trap (absorb) the outgoing long-wave terrestrial radiation. This increases the kinetic energy of the gas molecules causing the temperature of the atmosphere, and subsequently the Earth's surface, to rise. Most absorption of infrared radiation takes place in the lower atmosphere, the troposphere (section 1.2.2). This warming phenomenon (±33^{o}C) is known as the natural greenhouse effect since the atmosphere is acting like the glass in a greenhouse, allowing short-wave radiation through but trapping much of the long-wave radiation trying to escape. The absorbing gases are called greenhouse gases. Natural greenhouse gases include water vapour, carbon dioxide, methane, nitrous oxide and ozone. A more detailed narrative of the greenhouse effect may be found in section 1.2.

6.3. The Enhanced Greenhouse Effect

From the end of the last glacial episode about 7 to 10Ka up to the mid-eighteenth century, the levels of greenhouse gases in the atmosphere remained fairly constant. Since the Industrial Revolution, concentrations of most of the major greenhouse gases have increased, to a greater or lesser extent. The effect of this has been to increase the greenhouse radiative forcing of the climate. With more greenhouse gases in the atmosphere, more outgoing terrestrial radiation is trapped in the lower atmosphere, leading, presumably, to increases in surface temperature.

The debate as to whether this anthropogenic pollution of the atmosphere has increased global surface temperatures is ongoing. The following sections of this chapter (sections 6.4 to 6.8) offer a review of the evidence for anthropogenic global warming. I will begin with a discussion of the sources and sinks of the various greenhouse gases (section 6.4), before investigating how changes in their concentrations over the last 200 years have affected the radiative forcing of the global climate (section 6.5). In section 6.7 observed climate changes are reviewed, whilst section 6.8 discusses the question of detection [of anthropogenic warming].

6.4. Sources, Sinks and Concentrations of Greenhouse Gases

Just as energy, moisture and momentum are transferred throughout the climate system (section 1.2.5), greenhouse gases are cycled through the various components that make up the global biogeochemical system.

As for all systems and subsystems in equilibrium, the fluxes of the different greenhouse gases must balance. For the atmospheric component, this means that the sources of greenhouse gases must equal the sinks of greenhouse gases. Similarly, in/out fluxes for the other components of the system will balance.
The global biogeochemical system, like the climate system, is in a dynamic equilibrium, in which fluxes

of greenhouse gases respond to forcing perturbations. If a source of a particular greenhouse gas into a particular component of the system increases, the concentration of that greenhouse gas in that component will rise in order to re-establish equilibrium, by increasing the sink outlet of the gas. Over time, the biogeochemical system continually adjusts to many forcing perturbations involving many mechanisms over a wide range of time scales.

For example, an increase in plate tectonic activity will enhance the outgassing of carbon dioxide, with a consequent rise in the concentration of the gas in the atmosphere, over time scales involving tens of millions of years (section 2.6.2). Similarly, changes in ocean temperature, productivity and circulation can affect the removal of CO_2 from the atmosphere over centuries to millennia (section 5.3.1.2). Clearly, since greenhouse gases affect the radiative balance of the climate system, their global biogeochemical cycles are intimately linked to the causes of climate change.

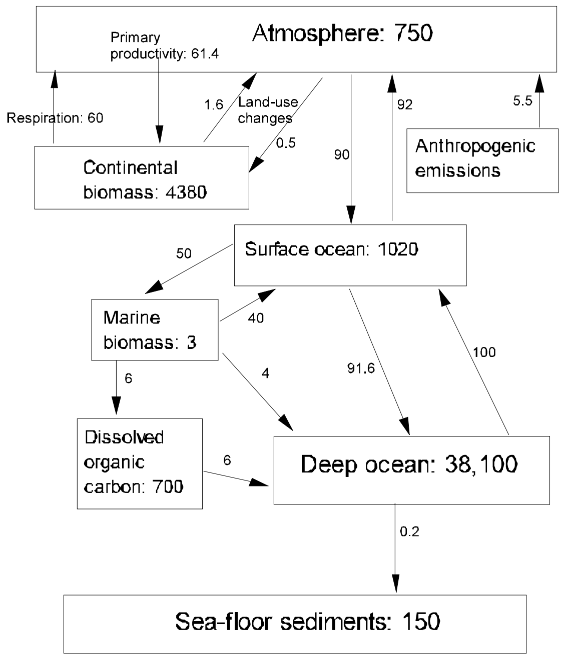
Recently, the impact of man's emissions of greenhouse gases into the atmosphere has introduced another climate forcing perturbation to the biogeochemical system. Sources of greenhouse gases to the atmosphere are now outstripping the sinks. Consequently, atmospheric concentrations of greenhouse gases over the last two hundred years have risen. In the following sub-sections, the sources (both natural and anthropogenic) and sinks of the greenhouse gases are reviewed in the context of contemporary climate change. Particular emphasis is placed upon CO_2 and its role in the global carbon cycle.

6.4.1. Carbon Dioxide and the Carbon Cycle

Carbon in the form of CO_2, carbonates and organic compounds is cycled between the main reservoirs of its biogeochemical system: the atmosphere, ocean, land and marine biota, and over geologic time scales, the sediments and rocks (Butcher *et al*., 1992). This is schematised in Figure 6.1.

The global carbon cycle, like other biogeochemical cycles, exists in dynamic equilibrium. Today, the atmosphere stores approximately 750 billion tonnes (Gt) of carbon (in the form of CO_2). The deep ocean represents an enormous store of carbon, with over 38,000Gt, whilst the surface ocean contains roughly 1,000Gt (Schimel *et al*., 1995).

Figure 6.1. *The Global Carbon Cycle (after Schimel et al., 1995)*



6.4.1.1. Sources of Atmospheric CO_2

Sources of atmospheric CO_2 can today be divided into two groups: natural and anthropogenic. Natural sources include the respiration of animals (60Gt per annum[31]) and the surface ocean (90Gt per annum) (Schimel *et al*., 1995). Anthropogenic sources include the combustion of fossil fuels (power stations and transport) and cement production (5.5Gt per annum[32])and land-use changes (mainly deforestation) (1.6Gt per annum)

6.4.1.2. Sinks of Atmospheric CO_2

The surface ocean also acts as a natural sink for atmospheric CO_2, with an annual removal flux of 92Gt carbon. The interaction of CO_2 between atmosphere and surface ocean was more fully addressed in section 5.3.1.2 (Equations 14 to 17). The

[31] Figures quoted here from IPCC (1995) are the best estimates for present day annual carbon fluxes

[32] Estimates for anthropogenic sources of CO_2 are usually accompanied with ranges of uncertainty. For example, IPCC (1995) estimate emissions from fossil fuels and cement production to be 5.5Gt ± 0.5Gt.

other major natural sink is the primary productivity of land vegetation (photosynthesis), which sequesters 61.4Gt carbon every year (Schimel *et al.*, 1995). The regrowth of Northern Hemisphere forests represents the only major anthropogenic sink of atmospheric CO_2, although enhanced fertilisation effects due to elevated CO_2 concentrations and other climatic feedbacks have also been considered.

6.4.1.3. Carbon Cycle Disequilibrium

It has conventionally been assumed that, prior to the Industrial Revolution (≈1765), the carbon fluxes into and out of the atmosphere were in equilibrium. Since that time, anthropogenic emissions of CO_2 have added a further source of atmospheric carbon. In response to this forcing perturbation, the disturbed equilibrium is currently in a process of readjustment, and consequently the atmospheric concentration of CO_2 has risen. A simple mass balance calculation from flux values in Figure 6.1 reveals a current net imbalance of 3.2Gt of carbon per annum entering the atmosphere.

Equation 18 demonstrates the nature of this disequilibrium that has existed since 1765.

$$d\Delta M/dt = I + D_n - F - X \quad \text{[Equation 18]}$$

where $\Delta M = M - M_0$ (atmospheric CO_2 mass change from pre-industrial, M_0, to present, M);
t= time (years);
I = CO_2 emissions from fossil fuel burning and cement production (source);
D_n = CO_2 emissions from land-use changes (source);
F = net oceanic uptake of CO_2 (sink); and
X = the net terrestrial uptake of CO_2 (sink).

Estimates of present day values for the terms in Equation 18 have been made by the IPCC (1990a, 1992, 1995). The most recent estimates, together with their associated ranges of uncertainty, are detailed in Table 6.1.

Inserting these estimates into Equation 18 reveals a present day figure of 3.2Gt/a for $d\Delta M/dt$, equivalent to that calculated from flux values in Figure 6.1 above.

6.4.1.4. Increases in Atmospheric CO_2 Concentration

The escalating rate of anthropogenic CO_2 emissions during the last 200 years has maintained the disequilibrium between atmospheric carbon fluxes. Figure 6.2 shows how the rate of atmospheric CO_2 accumulation (d M/dt) has increased since about 1840. Accompanying this mass accumulation has been a steady and exponential rise in the atmospheric CO_2 concentration (Figure 6.3). The CO_2 record prior to the late 1950s has been reconstructed from air bubbles trapped in ice cores (section 3.3.2.2). Since then, instrumental monitoring stations have measured concentrations directly.

Table 6.1. *Sources and sinks of atmospheric CO_2*

Net Atmospheric CO_2 Sources		
Source	Sources (Gt/a)	Uncertainty range (Gt/a)
Fossil fuel /cements (I)	5.5	±0.5
Land-use change (D_n)	1.6	±1.0
Total	7.1	±1.1
Net Atmospheric CO_2 Sinks		
Sink	Uptake (Gt/a)	Uncertainty range (Gt/a)
Oceanic uptake (F)	2.0	±0.8
Land biomass (X)	1.9†	±1.6
Total	3.9	±1.8

† Land biomass sink includes uptake by Northern Hemisphere forest regrowth (0.5 ± 0.5 Gt/a) and additional terrestrial sinks including CO_2 fertilisation, nitrogen fertilisation and climatic effects (1.4 ± 1.5Gt/a).

Figure 6.2. *Carbon emissions and atmospheric accumulation of CO_2 since 1860 (after Schimel et al., 1995)*

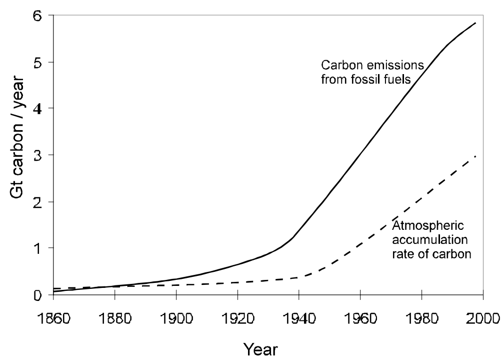
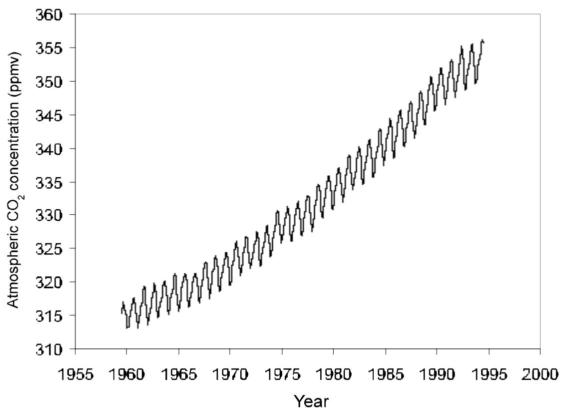


Figure 6.3. *Increase in atmospheric CO_2 since 1750 (after IPCC, 1990)*

Since pre-industrial times, atmospheric CO_2 concentration has increased by about 27% from about 290ppm to 359ppm (IPCC, 1995).

Figure 6.4 shows the recent record of CO_2 concentrations measured at the Mauna Loa station in Hawaii, since 1958. As well as the obvious upward trend, the record displays a distinct seasonal cycle. Every spring and early summer, the rate of global photosynthesis dramatically increases due to the enhanced growth of Northern Hemisphere terrestrial biomass, in response to elevated levels of insolation. Consequently, atmospheric CO_2 concentrations fall several ppm, before recovering in the autumn and winter seasons.

Figure 6.4. *Recent trend and seasonal variation of atmospheric CO_2 (after Schimel et al., 1995)*



6.4.1.5. Restoring Carbon Cycle Equilibrium

As for the climate system, the rate at which the carbon cycle, and more specifically the atmospheric reservoir of CO_2, adjusts to a new equilibrium in response to a forcing perturbation depends upon both the time scale of the forcing and the response period of the relevant components within the system.

The atmospheric adjustment time of CO_2 depends on the different time constants of the various carbon reservoirs (Watson *et al*., 1990). A relatively rapid adjustment (years) takes place between the atmosphere, the surface oceans and the terrestrial biosphere when anthropogenic CO_2 is added to the atmosphere. However, the long-term response of the atmospheric concentration of CO_2 to anthropogenic emissions depends primarily on the processes that control the rate of storage of CO_2 in the deep ocean, which have characteristics time scales of several decades to centuries (Watson *et al*., 1992).

6.4.2. Methane

Methane (CH_4) is a chemically and radiatively active trace gas that is produced from a wide variety of anaerobic (oxygen deficient) processes and is primarily removed by reaction with hydroxyl radicals (OH) in the atmosphere. (See also section 1.2.1.)

6.4.2.1. Sources of Atmospheric Methane

Anaerobic production of methane today occurs both naturally and anthropogenically. Natural sources include natural wetlands (115Mt/a[33]), termites (20Mt/a), and oceans (10Mt/a). Anthropogenic sources include the mining and burning of fossil fuels (100Mt/a), Enteric fermentation in ruminant animals (85Mt/a), rice paddies (60Mt/a), biomass burning (40Mt/a), landfills (40Mt/a) and other wastes (50Mt/a). When all minor sources are included, the total present annual emissions of methane are 535Mt/a (±125Mt).

6.4.2.2. Sinks of Atmospheric Methane

The major sink for atmospheric methane is the chemical reaction with hydroxyl radicals (OH) in the troposphere (section 1.2.1). Current removal estimates (Prather *et al*., 1995) are 445Mt/a. This natural process is, however, affected by the reaction of OH

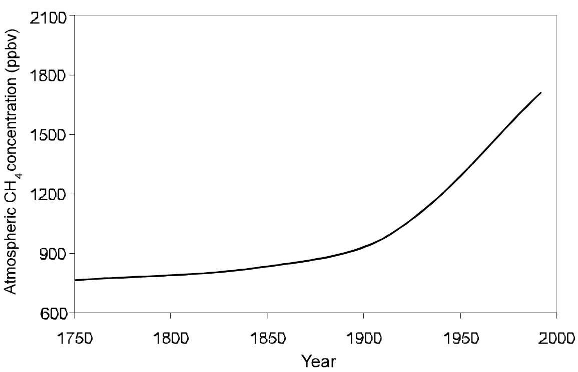
[33] As for CO_2 emissions, figures quoted in the text are best estimates for current annual methane emissions (Prather *et al*., 1995).

with other anthropogenic emissions, principally carbon monoxide (CO) and hydrocarbons from motor vehicles (Watson *et al*, 1990). Other sinks of atmospheric methane include stratospheric removal (again by reaction with OH) (40Mt/a) and consumption by microbial communities in the upper soils (30Mt/a). Total methane sinks are 515Mt/a (±85Mt/a)

6.4.2.3. Increases in Atmospheric Methane Concentration

The recent addition of anthropogenic sources of methane, as for CO_2, has disturbed the pre-industrial atmospheric flux equilibrium. Currently, the amount of methane emitted into the atmosphere is greater than the total removed, and observed atmospheric increase of methane is currently 37Mt/a (±2.5Mt/a). Consequently, atmospheric methane concentrations are increasing. Figure 6.5 shows how methane has increased in the atmosphere during the last two hundred years. A similar exponential growth to that seen in the CO_2 record is clearly noticeable.

Figure 6.5. *Increase in atmospheric methane since 1750 (after IPCC, 1990)*



Average global atmospheric methane concentrations have increased from about 730ppbv to 1,700ppbv, representing a 130% increase (Machida *et al.*, 1994). A hemispheric gradient exists, with concentrations in the Northern Hemisphere being about 6% higher than those south of the equator, since most of the principal (land-based) sources are within the Northern Hemisphere (Nakazawa *et al*., 1993).

6.4.3. Nitrous Oxide

Nitrous oxide (N_2O) is an important, long-lived greenhouse gas that is emitted predominantly by biological sources in soils and water, and removed in the upper atmosphere via photochemical reactions. Unfortunately, estimates of sources and sinks are not well quantified in global terms.

6.4.3.1. Sources of Atmospheric Nitrous Oxide

Tropical soils are probably the single most important source of N_2O to the atmosphere (Prather *et al.*, 1995), and intensification of tropical agriculture is likely to increase this (Matson & Vitousek, 1990). The current IPCC (1995) estimate for N_2O emissions from tropical soils is 4Mt/a (nitrogen), 75% from wet forests soils and 25% from dry savannahs. Temperate soil sources (including forest soils and grasslands) have been estimated at about 2Mt/a. The Earth's oceans are significant sources of N_2O. Emissions may be larger than previously estimated (Law & Owen, 1990), accounting for 3Mt/a.

Anthropogenic sources may account for about 40% of total N_2O emissions (5.7Mt/a). These include cultivated soils, biomass burning and industrial sources (e.g. nylon production). Total source emissions (natural and anthropogenic) are currently estimated at 14.7Mt/a (± 3.5Mt/a).

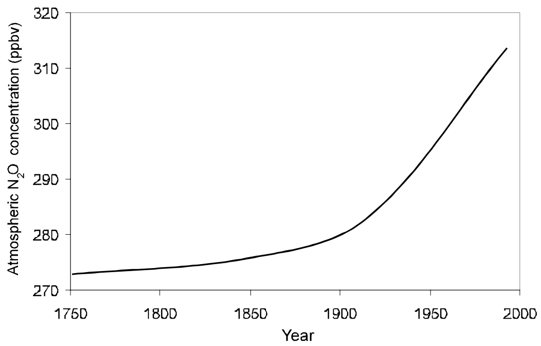
6.4.3.2. Sinks of Atmospheric Nitrous Oxide

The major sinks for N_2O are stratospheric photo-dissociation and photo-oxidation, estimated at 12.3Mt/a (±3.5Mt/a). Consumption by soils may also be a small sink but, to date, this contribution has not been quantitatively evaluated.

6.4.3.3. Increases in Atmospheric Nitrous Oxide Concentration

The observed increase in atmospheric N_2O concentrations at the present time implies that an excess of 3.9Mt/a is emitted to the atmosphere. Since pre-industrial times N_2O concentrations have increased by approximately 13% (Machida *et al.*, 1994), and now stand at about 310ppbv (Figure 6.6).

Figure 6.6. *Increase in atmospheric nitrous oxide since 1750 (after IPCC, 1990)*



6.4.4. Halocarbons

Halocarbon species include the chlorofluorocarbons (CFCs), the hydrochlorofluorocarbons (HCFCs), methylhalides, carbon tetrachloride (CCl_4), carbon tetrafluoride (CF_4)), and the halons (bromide species). They are all considered to be powerful greenhouse gases (section 6.5) because they strongly absorb terrestrial infrared radiation and reside in the atmosphere for many decades. All species containing chlorine and bromine play a role in lower stratospheric ozone depletion, and hence climate cooling since ozone is also a greenhouse gas (section 6.4.5), and this tends to offset their ability to cause surface warming. Emissions of the major chlorine- and bromine-containing halocarbons are now largely controlled by the requirements of the Montreal Protocol and its subsequent amendments.

6.4.4.1. Sources of Halocarbons

The CFCs and HCFCs are wholly anthropogenic and do not exist naturally. They have been widely used as propellants in aerosols, as blowing agents in foam manufacture, in air conditioning units and refrigerants (CDIAC, 1991, 1993). The Montreal Protocol has recently cut emissions of many species by over 90%. Methylhalides are primarily produced in the oceans, usually associated with algal growth (Moore & Tokarczyk, 1993), although a significant fraction may come from biomass burning. Annual emissions of halocarbons may be found in IPCC (1995).

6.4.4.2. Sinks of Halocarbons

Fully halogenated halocarbons are destroyed primarily by photodissociation and photo-oxidation in the stratosphere (section 1.2.1.5), but because of their relative inertness, remain in the atmosphere for long periods of time (section 6.4.7). Hydrogen-bearing halocarbons, such as the methylhalides and HCFCs are removed from the troposphere mainly by reaction with OH, but also NO_3. Some of these gases are, to a greater or less extent, removed by the ocean, presumably in hydrolysis reactions (Butler *et al*., 1991). Atmospheric adjustment times are shorter than for the fully halogenated species (section 6.4.7).

6.4.4.3. Increases in Atmospheric Halocarbon Concentrations

There are many species of halocarbons, and it would be time consuming to detail the increases in atmospheric concentration of these species. Concentrations of halocarbons are typically measured in pptv. Those with solely anthropogenic sources have increased from zero concentration since they were first manufactured in the 1930s. There is now evidence (Kaye *et al*., 1994) that the growth rates of species covered by the Montreal Protocol are slowing significantly. Current atmospheric concentrations for CFC-12 are shown in Table 6.2 (section 6.4.8).

6.4.5. Ozone

Ozone (O_3) is found throughout the atmosphere, with the majority (about 90%) in the stratosphere and the remainder in the troposphere. In both regions it is continually formed and destroyed through photochemical processes (section 1.2.1.4), involving carbon monoxide (CO), methane, non-methane hydrocarbons and nitrogen oxides (NO_x). Ozone is radiatively important in both the ultraviolet and infrared parts of the spectrum, and the radiative forcing due to ozone changes depends on whether the ozone is in the stratosphere or the troposphere (see section 6.5).

Decreases in stratospheric ozone have occurred since the 1970s. The most obvious feature is the annual appearance of the Antarctic ozone hole in September and October every year. The October average total ozone values over Antarctica are now 50-70% lower than those observed in the 1960s. Ozone loss is greatest at altitudes between 14 and 24km and is caused by photodissociation reactions involving chlorine and bromine released from halocarbons.

Observations show that free tropospheric ozone has increased above many locations in the Northern Hemisphere over the last 30 years (Prather *et al*., 1995), although changes in concentration are highly spatially variable, both regionally and vertically, making assessment of long-term trends difficult.

6.4.6. Other Trace Gases

Other trace gases such as the nitrogen oxides (NO_x), carbon monoxide (CO) and the volatile organic compounds (VOCs) have little direct radiative impact on the atmosphere, but they can indirectly influence the chemistry and concentrations of certain greenhouse gases, particularly ozone. The major sources of these compounds are technological (fossil fuel combustion in power stations and transport) and biomass burning. For a more detailed discussion see Prather *et al*. (1995).

6.4.7. Atmospheric Adjustment Times of Greenhouse Gases

In section 6.4.1.5, the atmospheric adjustment time of CO_2 was discussed. How quickly the additional influx of anthropogenic CO_2 is removed from the atmosphere depends upon the time constants of the removal processes. Similarly, for other greenhouse gases, the adjustment time describes the rate at which the compound is removed from the atmosphere by chemical processes or by irreversible uptake by the land or ocean.

The size of the adjustment time for different greenhouse gases considerably influences the rate at which any forcing perturbation in emissions is removed, and thus controls the time evolution of the increasing atmospheric concentrations of greenhouse gases. CO_2, the major greenhouse gas increasing in atmospheric concentration, has an adjustment time of 50 to 200 years. Such a long turnover period has serious implications for the radiative forcing of the climate system and future global climate change (section 6.9). Methane has an adjustment time of 12 to 17 years whilst for nitrous oxide it is 120 years (IPCC, 1995). Lifetimes for the halocarbons are variable.

6.4.8. Summary

With the exception of stratospheric ozone, all greenhouse gases discussed throughout section 6.4 have increased in concentration in the atmosphere over the last 200 years since the beginning of the Industrial Revolution. Pre-industrial and current concentrations of the key greenhouse gases, and their recent rates of increase (IPCC, 1995), are displayed in Table 6.2.

Table 6.2. *Pre-industrial and 1992 atmospheric concentrations of greenhouse gases*

	CO_2	CH_4	N_2O	CFC-12
Pre-industrial concentration	280 ppmv	700 ppbv	275ppbv	zero
Concentration in 1992	355 ppmv	1,715 ppbv	310ppbv	503pptv
Rate of change of concentration during 1980s	1.5 ppmv/a	13 ppbv/a	0.75 ppbv/a	20 pptv/a
Atmospheric adjustment time	50-200 yrs	12-17 yrs	120 yrs	102 yrs

6.5. Radiative Forcing of Greenhouse Gases.

Accompanying the increasing concentrations of greenhouse gases in the atmosphere has been an increase in greenhouse radiative forcing (enhanced greenhouse effect) due to the enhanced absorption of terrestrial infrared radiation. This radiative forcing can be quantified from the increases in greenhouse gases since the beginning of the Industrial Revolution.

6.5.1. Factors Affecting Greenhouse Radiative Forcing

A number of basic factors affect the behaviour of different greenhouse gases as forcing agents within the climate system (Shine *et al*., 1990). First, the absorption strength and wavelength of the absorption in the thermal infrared are of fundamental importance in dictating whether a molecule can be an important greenhouse forcing agent; this effect is modified by the overlap between the absorption bands and those of other gases present in the atmosphere. For example, the natural quantities of CO_2 are so large (compared to other trace gases) that the atmosphere is very opaque over short distance at the centre of its 15μm absorption band. The addition of a small amount of gas capable of absorbing at this wavelength has negligible effect on the net radiative flux at the tropopause. Other greenhouse gases with absorption bands in the more transparent regions of the infrared spectrum, in particular between 10 and 12μm, will have a far greater radiative forcing effect.

Second, the atmospheric residence time (lifetime) of a greenhouse gas can greatly influence its potential as a radiative forcing agent. Gases that remain in the atmosphere for considerable periods of time, before being removed via their sinks, will have a greater forcing potential over longer time horizons.

Third, the existing quantities of a greenhouse gas in the atmosphere can dictate the effect that additional molecules of that gas can have. For gases such as halocarbons, where the naturally occurring concentrations are zero or very small, their forcing is close to linear for present-day concentrations. Gases such as methane and nitrous oxide are present in such quantities that significant absorption is already occurring, and it is found that their forcing is approximately proportional to the square root of their concentration. For carbon dioxide, parts of the spectrum are already so opaque that additional molecules are almost ineffective; the forcing is found to be only logarithmic in concentration.

As well as the direct effects on radiative forcing, many greenhouse gases also have indirect radiative effects on the climate through their interactions with atmospheric chemical processes. For example, the oxidation of methane in the atmosphere leads to additional production of CO_2. Certain halocarbons such as the CFCs significantly affect the distribution of ozone, another greenhouse gas, in the atmosphere. The hydroxyl (OH) radical, itself not a greenhouse gas, is extremely important in the troposphere as a chemical scavenger. Reactions with OH largely control the atmospheric lifetime, and, therefore, the concentrations of a number of greenhouse gases, in particular methane and many of the halocarbons.

6.5.2. Greenhouse Warming Potentials

Although there are a number of ways of measuring and contrasting the radiative forcing potential of different greenhouse gases, the Global Warming Potential (GWP) is perhaps the most useful, particularly as a policy instrument. GWPs take account of the various factors influencing the radiative forcing potential of greenhouse gases. Such measures combine the calculations of the absorption strength of a molecule with assessments of its atmospheric lifetime; it can also include the indirect greenhouse effects due to chemical changes in the atmosphere caused by the gas. A number of GWPs are listed in Table 6.3 (IPCC, 1995).

Table 6.3. *Global Warming Potentials of the major greenhouse gases*

Trace Gas	Global Warming Potential (relative to CO_2)		
	Integration Time Horizon, Years		
	20	100	500
CO_2	1	1	1
CH_4 (incl. indirect)	62	24.5	7.5
N_2O	290	320	180
CFC-12	7900	8500	4200
HCFC-22	4300	1700	520

6.5.3. ΔF-ΔC relationships

The change in net radiative flux (Wm^{-2}) at the tropopause, ΔF, associated with a particular greenhouse gas, is usually expressed as some function of the change in atmospheric concentration of that gas. Direct-effect ΔF-ΔC relationships are calculated using detailed radiative-convective models (chapter 4). The form of the ΔF-ΔC relationship depends primarily on the existing gas concentration, as explained in section 6.5.1. For low/moderate/high concentrations, the form is well approximated by a linear/square-root/logarithmic dependence of ΔF on concentration. For example, the ΔF-ΔC relationship for CO_2 is given by:

$$\Delta F = 6.3 \ln (C/C_0) \qquad \text{[Equation 19]}$$

where C_0 is the initial CO_2 concentration, C is the final concentration and ln is the natural logarithm (Wigley, 1987; Hansen *et al*., 1988). This relationship is valid for concentrations up to 1000ppmv. Alternatively, the ΔF-ΔC relationship for CFC-12 is given by:

$$\Delta F = 0.22 (X - X_0) \qquad \text{[Equation 20]}$$

where X_0 is the initial CFC-12 concentration and X is the final concentration (Hansen *et al.*, 1988). The relationship holds for X less than 2ppbv (2000pptv).

Nevertheless, it should be appreciated that such relationships are empirical in nature and are therefore subject to uncertainties. First, there are uncertainties in the basic spectroscopic data for many gases. Second, uncertainties arise through details in the radiative-convective modelling. Third, the assumptions used to model ΔF-ΔC relationships are subject to uncertainties, for example the assumed vertical profile of concentration, temperature and moisture changes, and the indirect effects on radiative forcing due to chemical interactions.

6.5.4. Greenhouse Radiative Forcing 1765 to 1990

From the modelled ΔF-ΔC relationships, the increase in radiative forcing due to the enhanced greenhouse gas concentrations can be calculated. Instrumental records exist for the most recent decades, whilst proxy data are used to calculate greenhouse gas concentrations earlier in the study period. Table 6.4 gives the concentration changes for five of the commonly known greenhouse gases, whilst Table 6.5 details their contribution to radiative forcing for a number of time intervals. Here, 1765 has been regarded as the onset of the Industrial Revolution.

Table 6.4. *Changes in atmospheric concentration of the major greenhouse gases since 1750*

Year	CO_2 (ppmv)	CH_4 (ppbv)	N_2O (ppbv)	CFC-11 (pptv)	CFC-12 (pptv)
1765	279	790	275	0	0
1900	296	974	292	0	0
1960	316	1272	297	18	30
1970	325	1421	299	70	121
1980	337	1569	303	158	273
1992	355	1714	311	270	504

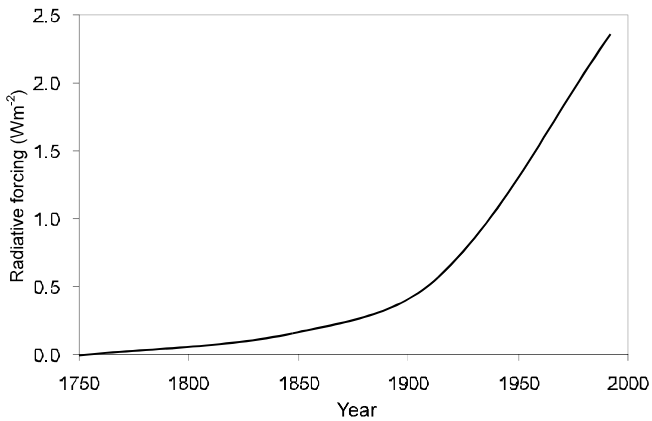
Table 6.5. *Change in radiative forcing (Wm^{-2}) due to concentration changes in greenhouse gases*

Time period	CO_2	CH_4	N_2O	CFC-11	CFC-12	Total†
1765-1900	0.37	0.1	0.027	0.0	0.0	0.53
1765-1960	0.79	0.24	0.045	0.004	0.008	1.17
1765-1970	0.96	0.30	0.054	0.014	0.034	1.48
1765-1980	1.20	0.36	0.068	0.035	0.076	1.91
1765-1990	1.50	0.42	0.10	0.062	0.14	2.45

† Direct radiative forcing greenhouse gases (i.e. excludes radiative effects of ozone loss).

Changes in CO_2 concentration over the last two centuries have contributed most to the greenhouse radiative forcing. Over the period 1765 to 1990, CO_2 forcing has accounted for 61% of the total enhanced greenhouse forcing. Nevertheless, other greenhouse gases, in particular the halocarbons, are now accounting for an increasing proportion of the total greenhouse forcing, due to their relatively larger GWPs. The total increase in radiative forcing since 1765 is shown in Figure 6.7.

Figure 6.7. *Increase in radiative forcing since 1750 (after IPCC, 1990)*



The total increase in direct radiative greenhouse forcing is approximately $2.5Wm^{-2}$ (Table 6.5). This value should be compared to the solar constant, $1368Wm^{-2}$, the amount of total solar radiation intercepted by the Earth. The average amount of radiation arriving at the top of the troposphere is about 270Wm^{-2}. This figure is lower than the solar constant, for it takes into account the latitudinal and temporal variations in insolation.

6.5.5. Radiative Forcing of Ozone

The ΔF-ΔC relationship for atmospheric ozone is more complex than for other trace gases because of its marked vertical variations in absorption and concentration. Changes in ozone can cause greenhouse forcing by influencing both solar and infrared radiation (section 1.2.4). The net change in radiative forcing is strongly dependent on the vertical distribution of ozone concentration changes, and is particularly sensitive to variations around the tropopause.

Decreases in stratospheric ozone, principally over the Antarctic at altitudes between about 14 and 24km, have been occurring since the 1970s due to the anthropogenic release of CFCs and halons. These changes in ozone substantially perturb both solar and long-wave radiation (WMO, 1991). While the solar effects due to ozone loss are determined by the total column ozone amounts, the long-wave effects are determined both by the amount and its vertical location (Lacis *et al*., 1990). In general, stratospheric ozone loss will tend to increase the solar forcing, resulting in surface-troposphere warming, whilst decreasing greenhouse forcing, with consequent surface-troposphere cooling (Isaksen *et al*., 1992; WMO, 1991). Most models unambiguously demonstrate that changes in greenhouse forcing are dominant (Wang *et al*., 1993; Schwarzkopf & Ramaswamy, 1993). Hansen *et al*. (1993) compute a global mean forcing of -0.2 ±0.1Wm^{-2} between 1970 and 1990. Such a value represents a significant offset to the positive greenhouse forcing from changes in halocarbons over the same period, estimated at 0.22Wm^{-2} (Shine *et al*., 1990).

The build up of tropospheric ozone due to chemical reactions involving precursors produced in various industrial processes (section 6.4.5) has potentially important consequences for radiative forcing. Hauglustine *et al*. (1994) used a 2-D radiative-convective model to estimate that changes in tropospheric ozone since pre-industrial times have contributed a global mean forcing of 0.55Wm^{-2}. This agrees well with other studies, both modelling (WMO, 1994) and observational (Marenco *et al*., 1994). Nevertheless, the increases in tropospheric ozone will be highly regional and so will the positive greenhouse forcing associated with them.

6.6. Atmospheric Aerosols

Atmospheric aerosol particles are conventionally defined as those particles suspended in air having diameters in the region of 0.001 to 10 μm. They are formed by the reaction of gases in the atmosphere, or by the dispersal of material at the surface. Although making up only 1 part in 10^9 of the mass of the atmosphere, they have the potential to significantly influence the short-wave radiative transfer. The recent addition of anthropogenic aerosols to the atmosphere has introduced a negative change in radiative forcing which partially offsets the positive greenhouse forcing discussed in the last section.

6.6.1. Sources and Sinks of Aerosols

Atmospheric aerosol particles may be emitted as particles (primary sources) or formed in the atmosphere from gaseous precursors (secondary sources). Table 6.6 summarises the estimated recent annual emissions into the troposphere or stratosphere from the major sources of atmospheric aerosol, both natural and anthropogenic, primary and secondary. These include sulphates from the oxidation of sulphur-containing gases, nitrates from gaseous nitrogen species, organic materials from biomass combustion and oxidation of VOCs[34], soot from combustion, and mineral dust from aeolian (wind-blown) processes (Andreae, 1994). It should be noted that each aerosol flux estimate is a best guess, and uncertainty ranges of ±100% are not untypical.

Removal of aerosol mass is mainly achieved by transfer to the Earth's surface or by volatilisation (Jonas *et al*., 1995). Such transfer is brought about by precipitation (wet deposition) and by direct uptake at the surface (dry deposition). The efficiency of both these deposition processes, and hence the time spent in the atmosphere by an aerosol particle, is a complex function of the aerosol's physical and chemical characteristics (e.g. particle size), and the time and location of its release.

34 Volatile organic compounds.

For fine sulphate aerosols (0.01 to 0.1μm) released into or formed near the Earth's surface, an average lifetime is typically of the order of several days (Chamberlain, 1991). This time scale is dominated mainly by the frequency of recurrence or precipitation. Conversely, particles transported into or formed in the upper troposphere are likely to remain there for weeks or months because of the less efficient precipitation scavenging (Balkanski *et al*., 1993). Stratospheric aerosols, formed as a consequence of large volcanic eruptions (section 2.6.3) can remain there for up to one or two years.

Table 6.6. *Recent global annual emissions estimates from major aerosols, Mt (after Andreae, 1994)*

Source	Flux Estimate	Particle Size†
Natural		
Primary		
Mineral aerosol	1500	mainly coarse
Sea salt	1300	coarse
Volcanic dust	33	coarse
Organic aerosols	50	coarse
Secondary		
Sulphates from biogenic gases	90	fine
Sulphates from volcanic SO_2	12	fine
Organic aerosols from VOCs	55	fine
Nitrates from NO_x	22	mainly coarse
Total	**3062**	
Anthropogenic		
Primary		
Industrial dust	100	coarse & fine
Soot	10	mainly fine
Biomass burning	80	fine
Secondary		
Sulphates from SO_2	140	fine
Organic aerosols from VOCs	10	fine
Nitrates from NO_x	40	mainly coarse
Total	**380**	

† coarse: <1μm diameter; fine: >1μm diameter

Owing to the short lifetime of aerosol particles in the atmosphere, particularly in the troposphere, and the non-uniform distribution of sources, their geographical distribution is highly non-uniform. As a consequence, the relative importance of the numerous sources shown in Table 6.6 varies considerably over the globe. In certain areas of the Northern Hemisphere, for example Europe and North America, industrial sources are relatively much more important.

6.6.2. Radiative Forcing by Aerosols

Atmospheric aerosols influence climate in two ways, directly through the reflection and absorption of solar radiation (in both the troposphere and stratosphere), and indirectly through modifying the optical properties and lifetimes of clouds (mainly in the troposphere) (Shine *et al*., 1995).

Estimation of aerosol radiative forcing is more complex and hence more uncertain than radiative forcing due to the well-mixed greenhouse gases (see section 6.5) for several reasons. First, both the direct and indirect radiative effect of aerosol particles are strongly dependent on the particle size and chemical composition and cannot be related to mass source strengths in a simple manner, For this reason, anthropogenic aerosols contribute between 10 and 20% of the atmospheric mass burden, yet 50% to the global mean aerosol optical depth[35]. Second, the indirect radiative effects of aerosols depend on complex processes involving aerosol particles and the nucleation and growth of cloud droplets. Third, most aerosols have short lifetimes (days to weeks) and therefore their spatial distribution is highly inhomogeneous and strongly correlated with their sources.

6.6.2.1. Direct radiative forcing

Aerosol particles in the 0.1 to 1.0μm diameter range have the highest efficiency per unit mass for optical interactions with incoming solar radiation due to the similarity of particle size and radiation wavelength. Sulphate aerosols and organic matter (Table 6.6) are therefore most effective at scattering and absorbing the short-wave radiation, with the effect of negative radiative forcing on the climate system (see sections 1.2.4 and 2.6.5). The negative radiative forcing due to

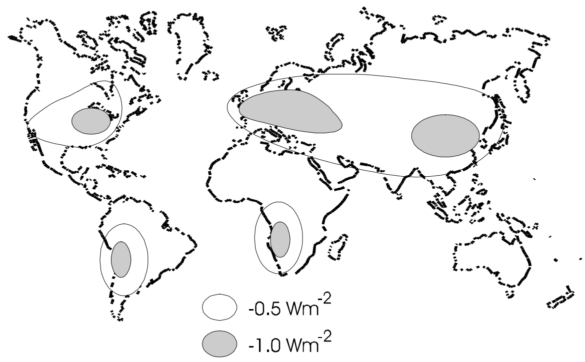
[35] The optical depth is a measure of the scattering and absorption of shortwave radiation due to a column of air (see also Shine *et al*., 1995).

large injections of sulphate aerosol precursors into the stratosphere from volcanic eruptions is well known (Sear *et al*., 1987).

The major contributions to the anthropogenic component of the aerosol optical depth arise from sulphates produced from sulphur dioxide released during fossil fuel combustion and from organics released by biomass burning. To date, estimations of global negative forcing associated with anthropogenic aerosols are based on a number of modelling studies, due primarily to a lack of observational data.

Using a aerosol radiative-convective model, Charlson *et al*. (1992) obtained a figure of -0.6 Wm^{-2} for direct global radiative forcing due to anthropogenic sulphate alone, whilst Kiehl & Brigleb (1993) calculated a value of -0.3Wm^{-2} for the Earth, and -0.43Wm^{-2} for the Northern Hemisphere. Figure 6.8 shows the geographic distribution of annual mean direct radiative forcing (Wm^{-2}) from anthropogenic sulphate aerosols. The regionality of the forcing, due to localised emission sources and the short lifetimes of tropospheric sulphate aerosols, is clearly noticeable.

Figure 6.8. *Global aerosol forcing (after Shine et al., 1995)*



Other model-calculated values for direct global radiative forcing due to sulphate aerosols include -0.25Wm^{-2} (Hansen at al., 1993) and -0.9Wm^{-2} (Taylor & Penner, 1994). Such a large sensitivity of the results of these simulations clearly indicate the need for more observational data on the chemical and physical properties of aerosols as well as more refined sulphate distribution calculations.

The effects of aerosols emitted as a result of biomass burning has received much less attention and global estimates of radiative forcing due to this source are subject to considerable uncertainty. The direct global-mean radiative forcing since pre-industrial times may lie in the range -0.05 to -0.6Wm^{-2} (Shine *et al*., 1995).

6.6.2.2. Indirect Radiative Forcing

The global energy balance is sensitive to cloud albedo, and most particularly towards marine stratus (low-level) clouds which cover about 25% of the Earth. Cloud albedo is itself sensitive to changes in the cloud droplet number concentration. This droplet number depends, in a complex manner, on the concentration of cloud condensation nuclei (CCN), which, in turn, depends on aerosol concentration. Through this indirect effect, the negative radiative forcing caused by anthropogenic aerosol emissions might be further increased.

Particles of sizes around 0.1μm diameter composed of water soluble substances are highly effective as CCN. This includes both sulphate aerosols and organic aerosols from biomass burning. Direct observations of the impact of CCN on cloud albedos have been reported (Coakley *et al*., 1987; Kim & Cess, 1993). Estimates from modelling studies of the indirect radiative effects of aerosols vary widely. Jones *et al*. (1994) calculated the indirect global-mean effect since pre-industrial times to be about -1.3Wm^{-2}, whilst Kaufman & Chou (1993) found a forcing of -0.45Wm^{-2}. Despite the level of uncertainty, indirect negative radiative forcing is believed to be comparable to the direct forcing.

6.6.3. Total Anthropogenic Radiative Forcing: Greenhouse Gases and Aerosols

Table 6.7 shows the estimate (based on IPCC, 1995) of the radiative forcing due to both the anthropogenic increase of atmospheric greenhouse gases and aerosols since pre-industrial times. Each estimate of forcing is accompanied by an uncertainty range to express the level of confidence associated with the "best guess" value. Direct greenhouse forcing has been estimated with a fairly high level of confidence, whilst levels of aerosol forcing and indirect ozone forcing are poorly constrained. An estimate of the net global-mean radiative forcing due to all anthropogenic activity since pre-industrial times is not presented, as the usefulness of combining estimates of global-mean radiative forcings of different signs, and resulting from different temporal and spatial patterns, is not currently understood. Nevertheless, it is qualitatively possible to recognise that much of the positive radiative forcing due to increases in greenhouse gases has been partially offset by the negative forcing associated with increases in atmospheric aerosols.

Table 6.7. *Estimates of globally averaged radiative forcing (Wm^{-2}) due to increases in atmospheric greenhouse gas and aerosol concentrations*

	Radiative forcing Wm^{-2}	Uncertainty range Wm^{-2}
Direct greenhouse forcing	2.45	±0.4
Stratospheric ozone forcing‡	-0.15	±0.075
Tropospheric ozone forcing‡	0.4	±0.25
Direct aerosol forcing	-0.9	±0.6
Indirect aerosol forcing	uncertain (negative)	±0.75
Solar forcing	0.3	±0.2

‡Indirect greenhouse forcing

6.7. Observed Climate Variations

Sections 6.4 to 6.6 discussed the increases in concentration of atmospheric greenhouse gases and aerosols due to anthropogenic pollution since the Industrial Revolution, and the accompanying global changes in radiative forcing. There is now considerable model-based evidence that indicates a causal relationship between the net positive global-mean radiative forcing and the observed global temperature increase during the twentieth century. Before reviewing this evidence (section 6.8), temperature, and other contemporary climatic changes which have occurred, will be considered by examining the modern instrumental records.

6.7.1. Surface Temperatures Variations

Early attempts to produce a global surface temperature record from instrumental data were confined to land surface air temperatures (e.g. Jones, 1988; Hansen & Lebedeff, 1987, 1988). All analyses indicate that during the last decade globally averaged land temperatures have been higher than in any decade in the past 140 years. Over the whole period, a global temperature increase of 0.45°C/100 years has been observed (Jones, 1988). Since the interpretation of the rise in temperature is a key issue for global warming, so the accuracy of the data needs careful consideration. A number of problems may have affected the land temperature record:

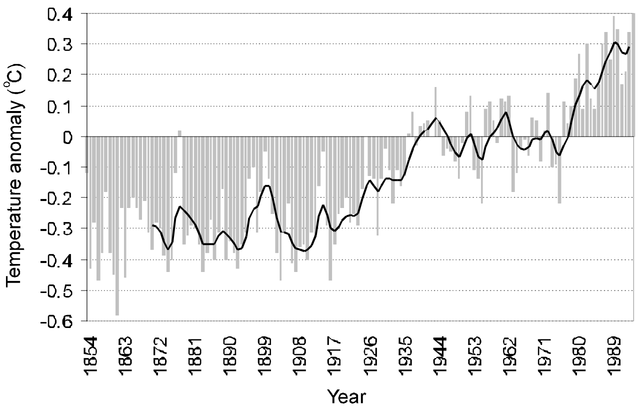
1) spatial coverage of the data is incomplete and varies considerably;
2) changes have occurred in the observing schedules and practices;
3) changes have occurred in the exposure of thermometers;
4) recording stations have changed their locations;
5) changes in the environment, especially urbanisation, have taken place around many recording stations.

These problems of inhomogeneity are discussed more fully in section 3.2.2.

The oceans comprise about 70% of the global surface. Obviously, a compilation of global temperature variations must include ocean surface temperatures. Farmer *et al*. (1989) have created historical analyses of global sea surface temperatures (SSTs) which are derived mostly from observations taken by commercial ships.

The combined land air and sea surface temperature is shown in Figure 6.9 (DoE, 1994). A linear trend fitted between 1860 and 1993 indicates temperature increases of 0.47°C, 0.53°C and 0.50°C for the Northern Hemisphere, Southern Hemisphere and globe respectively (Folland *et al*., 1990). Associated uncertainties are of the order of ±30% (±0.15°C), when all sources of inhomogeneity are considered.

Figure 6.9. *Annual deviation of global mean land-air and surface ocean temperatures relative to 1951-1980 mean (DoE, 1994)*



Combined land and ocean temperatures have increased rather differently in the two hemispheres. A rapid increase in the Northern Hemisphere temperature during the 1920s and 1930s contrasts with a more gradual increase in the Southern

Hemisphere. Both hemispheres had relatively stable temperatures from the 1940s to the 1970s, although there is some evidence of cooling in the Northern Hemisphere during the 1960s. Since the 1960s in the Southern Hemisphere but after 1975 in the Northern Hemisphere, temperatures have risen sharply.

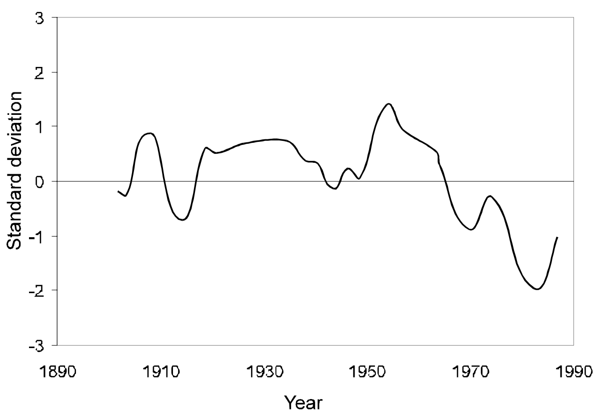
Whilst globally-averaged records offer a means of assessing climate change, it is important to recognise that they represent an over-simplification. Significant latitudinal and regional differences in the extent and timing of warming exist (Folland *et al*., 1990). In addition, winter temperatures and night-time minimums may have risen more than summer temperatures and day-time maximums.

6.7.2. Precipitation Variations

With globally increasing temperatures, increases in global precipitation would be expected *a priori*, due to the greater rates of evaporation of sea surface water. Unfortunately, no reliable estimates of evaporation increase exist. One problem is the effect of varying wind speed on evaporation rates, which may or may not be related to increases in temperature.

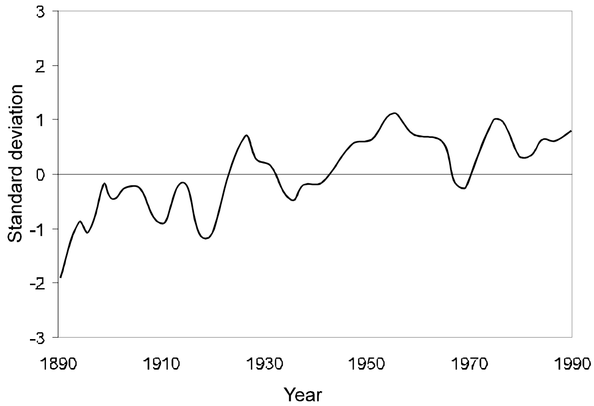
Several large-scale analyses of precipitation changes in both hemispheres have been carried out (Bradley *et al*., 1987; Diaz *et al*., 1989; Vinnikov *et al*., 1990). These have demonstrated that during the last few decades precipitation has tended to increase in the mid-latitudes, but decrease in the Northern Hemisphere subtropics and generally increase throughout the Southern Hemisphere.

Figure 6.10a. *Standardised deviation of precipitation in the Sahel region, N. Africa, smoothed curve (after Folland et al., 1990)*



Two of the more striking precipitation changes have occurred in the African Sahel (6.10a) and the former Soviet Union nations (6.10b). The dramatic drying up of sub-Saharan Africa has been linked to changes in ocean circulation (Street-Perrott & Perrott, 1990) and tropical Atlantic sea surface temperatures (Folland *et al*., 1991). Nevertheless, the accuracy of other precipitation records should be treated with caution. Precipitation is more difficult to monitor than temperature due to its greater temporal and spatial variability (see section 3.2.1.2). Other uncertainties in the data set may be due to the collection efficiency of raingauges.

Figure 6.10b. *Standardised deviation of precipitation in the old USSR nations, smoothed curve (after Folland et al., 1990)*



6.7.3. Other Climatic Variations

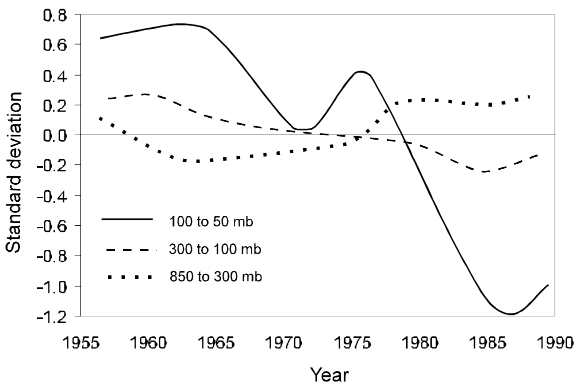
A number of other climatic indices reveal changes during the twentieth century and are discussed in the following sub-sections.

6.7.3.1. Tropospheric and Stratospheric Temperatures Changes

Tropospheric and stratospheric temperatures are central to the problem of greenhouse warming because GCMs predict that temperature changes with enhanced concentrations of greenhouse gases will have a characteristic profile in these layers, with warming in the mid-troposphere and cooling in the much of the stratosphere. The cooler stratospheric temperatures would be an expected consequence of the increased trapping of terrestrial radiation in the troposphere.

Layer mean temperatures from a set of 63 radiosonde stations covering most of the globe have been derived by Angell (1988). Layer mean temperatures from this network have been globally integrated. Figure 6.11 shows that, over the globe as a whole, mid-tropospheric (850-300mb) temperatures have increased between the 1970s and 1990s, in parallel with surface temperature. In the upper troposphere (300-100mb) there has been a steady decline in temperature of about 0.4^{o}C since the 1960s. Such a finding is in general disagreement with model simulations that show warming at these levels in response to greenhouse forcing. Temperatures in the lower stratosphere (100-50mb) show the greatest change, especially since 1980. It is mostly attributed to changes over and around Antarctica, probably related, in part, to the decrease in springtime stratospheric ozone.

Figure 6.11. *Standardised deviation of air temperature in the troposphere and lower stratosphere, smoothed curves (after Folland et al., 1990)*



6.7.3.2. Variations in the Cryosphere

Variations in cryospheric variables, such as snow, ice and glacial extent occur in response to changes in temperature, sunshine amount, precipitation, and for sea-ice, changes in wind-stress. Since 1966 Northern Hemisphere snow cover maps have been produced using National (U.S) Oceanographic and Atmospheric Administration (NOAA) satellite imagery. Consistent with the surface and tropospheric temperature measurements is the decrease in snow cover and extent around 1980. Variations in sea ice extent have also been reported (Mysak & Manak, 1989; Gloerson and Campbell, 1988) but little long term trend is noticeable. Nevertheless, considerable interest in retreating sea ice was generated in early 1995, when a large iceberg broke off the Larsen Ice Shelf on the Antarctic continent. At the same time, sea ice formerly blocking the Gustav Channel between the Antarctic peninsula and James Ross Island, completely disintegrated. In view of the rapidity at which they took place, such events have been viewed as a signal of greenhouse warming.

Measurements of glacial ice volume and mass balance provide further valuable information about climatic changes, but they are considerably scarcer. Glacial advance and retreat is influenced by temperature, precipitation and cloudiness, and depending on the size of the glacier, their movements tend to lag behind climatic variations. A substantial, but not continuous, recession of mountain glaciers has taken place almost everywhere since the latter half of the nineteenth century (Grove, 1988). The rate of recession appears to have been generally greatest between about 1920 and 1960.

6.7.3.3. Variations in Atmospheric Circulation

The atmospheric circulation is the main control behind regional changes in wind, temperature, precipitation, moisture and other climatic variables. Variations in many of these are quite strongly related through large-scale features of the atmospheric circulation. Changes in a number of circulation features, including El Niño Southern Oscillation (ENSO) events in the south-east Pacific, mid-latitude Northern Hemisphere westerlies and the location and intensity of the Aleutian low pressure system in the North Pacific, may all be related to the more general global warming of the twentieth century (Folland *et al*., 1990).

6.7.3.4. Cloudiness

Increased global cloudiness, as for increased global evaporation and precipitation, would be an expected consequence of higher global temperatures. Annual mean cloudiness has been found to have increased over Europe (6% / 80 years), Australia (8% / 80 years), the Indian sub-continent (7% / 50 years) and North America (10% / 90 years) (Henderson-Sellers, 1986, 1989). There is some suggestion that these values may represent over-estimates, due to a change in recording practices during the late 1940s and early 1950s, and the obscuring effects of smoke, haze, dust and fog (Karl & Steurer, 1990). When cloudiness over the oceans is also considered (Warren, 1988), it is not possible to be confident that average global cloudiness has really increased.

Other climatic variables that can be considered include sub-surface ocean temperatures and salinity variations, atmospheric moisture, and climate variability and extremes, including droughts, floods and cyclones. A discussion of their variations is provided in IPCC (1990).

6.8. Detection of Anthropogenic Global Warming

Sections 6.4. to 6.6. reviewed the anthropogenic increase in atmospheric concentrations of greenhouse gases, and the associated increase in greenhouse radiative forcing. Section 6.7 examined the observed climate changes, principally in surface temperature, that have taken place during the last 100 to 140 years. In this section, the evidence for a causal link between the anthropogenic increase in greenhouse radiative forcing and the observed global warming will be reviewed.

The word "detection" by climate scientists has been used to refer to the identification of the significant change in climate during the twentieth century and its association with the anthropogenically enhanced greenhouse effect. Until 1995 most reviews (MacCracken & Luther, 1985; Bolin *et al*., 1986; Wigley & Barnett, 1990) had concluded that the enhanced greenhouse effect has not yet been detected unequivocally in the observational record. However, they have also noted that the global-mean temperature change over the past 100 years is consistent with the greenhouse hypothesis. Since then, the most recent scientific report from the Intergovernmental Panel on Climate Change (IPCC, 1995) has proposed that the balance of [modelling] evidence suggests a discernible human influence on the global climate does exist.

Traditionally, the difficulty in greenhouse detection has arisen because there are numerous other causes of climatic variability (see chapter 2), and some of these [forcing mechanisms] may be operating on time scales (10^1 to 10^2 years) comparable to that of the anthropogenic greenhouse forcing. In addition, Wigley & Raper (1990) have shown that the inherent variability (random fluctuations) of the climate system can produce warming or cooling trends of up to 0.3^{o}C per century. The detection problem can be conveniently described in terms of "signal" and "noise" (Maddan & Ramanathan, 1980). The signal here is the time-dependent climatic response to greenhouse forcing, whilst the noise is any non-greenhouse climate variation, either periodic, quasi-periodic or random.

6.8.1. Greenhouse Modelling versus Observation

Global-mean temperature has increased by around 0.3 to 0.6^{o}C over the past 100 years (section 6.7.1). At the same time, greenhouse gas concentrations, and atmospheric aerosol loadings have increased substantially (sections 6.4 and 6.6). To assess whether the two are associated requires the use of model simulations of the likely climatic effects of the changing atmospheric composition, and the comparison of the results with observations. Three important detection experiments will be discussed here.

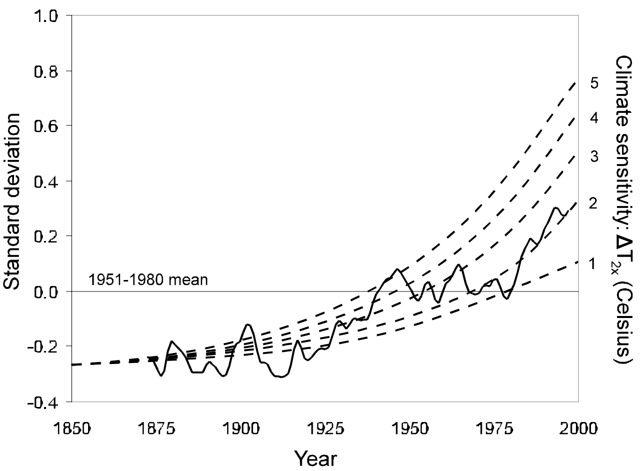
Wigley & Barnett (1990) used an energy balance climate model (incorporating upwelling and diffusion within the oceans to account for their radiative damping effect). The model was forced from 1765 to 1990 using only the changes in greenhouse gas concentrations, and the response could be varied by changing the value of the climate sensitivity[36]. In this way, climatic feedbacks (see section 6.9.2), not explicitly modelled, can be incorporated into the model.

The model results were qualitatively consistent with the observations on the century time scale (Figure 6.12). On shorter time scales, the model failed to reproduce the inter-decadal variability of the instrumental record. Indeed, this caveat has often been used as an argument against the greenhouse hypothesis altogether. However, Wigley & Barnett (1990) point out that such variability represents the background noise against which the greenhouse signal has to be detected. Significantly, the observational record seemed to lie at the low climate sensitivity end of the output range of that predicted by GCMs (1.5 to 4.5^{o}C). However, the situation becomes more complex if other forcing mechanisms, in addition to the enhanced greenhouse effect, are invoked. If the net century time scale effect of non-greenhouse

[36] The climate sensitivity is here defined as the equilibrium global-mean temperature change for a CO_2 doubling (ΔT_{2x}). Outputs from GCMs indicate that ΔT_{2x} lies in the range 1.5 to 4.5^{o}C.

factors (e.g. solar variability, volcanism) involved a warming, the climate sensitivity would be less than 1^{o}C. If their combined effect were a cooling, the sensitivity could be larger than 4^{o}C.

Figure 6.12. *Comparison of observed global surface temperature changes with model outputs (after Wigley & Barnett, 1990)*



One possible explanation for the decadal time scale discrepancies between the model and observed data is that some other forcing mechanism has been operating which has either offset or reinforced the general warming trend at different times. Using another energy balance model, Kelly & Wigley (1992) considered solar variability as a possible candidate. The model was run with a series of sensitivities spanning the accepted range of uncertainty in order to identify the best fit between modelled and observed temperature. Two sets of forcing histories (determined by a 1-D radiative-convective model) were considered, one involving only the effect of enhanced greenhouse gas concentrations (the IPCC 1990 forcing record), the other including also the negative radiative effect of aerosol loading and stratospheric ozone depletion (the IPCC 1992 forcing record).

Table 6.8 summarises the results of Kelly & Wigley (1992). The model explicitly calculated the best fit CO_2 doubling temperature (climate sensitivity) and the amount of explained variance in the observational record by each forcing history. As well as IPCC 1990 and 1992 forcing histories, different solar variables (sunspot number, length on sunspot cycle, solar diameter and rate of change of solar diameter) were considered, and combined with the greenhouse forcing.

Table 6.8. *Summary results of the Wigley/Kelly model*

Forcing		Doubling Temp. (^{o}C)	Explained variance (%)		
Green-house	Solar		Green-house	Solar	Total
IPCC 90/92	None	1.8/3.8	46.1/49.2		46.1/49.2
IPCC 90/92	Number	1.2/2.9	39.7/46.1	10.7/6.4	50.4/52.5
IPCC 90/92	Length	0.9/1.9	30.4/33.8	22.6/19.8	53.0/53.6
IPCC 90/92	Diameter	1.1/2.6	37.1/43.6	16.4/10.4	53.5/54.0
IPCC 90/92	Gradient	1.8/3.6	46.1/49.0	8.9/8.0	55.0/57.0

Table 6.8 demonstrates that a considerable difference in the best-fit climate sensitivities exists between the two IPCC forcing histories. If greenhouse forcing alone is considered the CO_2 doubling temperature is at the low end of the range predicted by GCMs, and is in agreement with Wigley & Barnett (1990). If the negative forcing of aerosol loading and ozone depletion is also included, the climate sensitivity is much greater. This result is intuitively correct, since the radiative effects of aerosols and ozone loss would offset the warming due to increased greenhouse gases.

When solar variability is included into the model, the explained variance of the observational record is greater than for greenhouse forcing alone. This is true for all of the solar variables considered here. The forcing combination that explains the most variance in the observational record (57%) includes the effects of greenhouse gases, aerosols and ozone depletion, and the rate of change of solar diameter. The latter, it seems, is accounting for much of the inter-decadal variability in the instrumental record.

There now seems to exist a plausible explanation for why the warming has not been regular over the past 100 years, but Kelly & Wigley (1992) point out a number of caveats. Most importantly, the explained variance varies little over a range of estimates, about the best-fit value, for the CO_2 doubling temperature. Thus, it remains difficult to define a precise value for the climate sensitivity. Ultimately, knowledge of the climate sensitivity provides the key to projecting the future impact of greenhouse gas emissions (section 6.9). Whilst uncertainty remains (see section 2.8), forecasting future climates will remain an imprecise science.

In addition, other possible mechanisms of climate forcing have not been considered. These could include the effects of volcanism (section 2.6.3), ocean circulation (section 2.6.4) or even the inherent random variability of the climate system. However, as other potential mechanisms are included into the model, the uncertainty in the climate sensitivity rises. If sufficient data existed to define the natural level of climate variability prior to human interference, this estimate could be used to attach error margins to the analysis of the greenhouse signal without explicit consideration of the source of the background noise. Unfortunately, such data do not exit since the period of the instrumental record contains this greenhouse signal.

The most recent attempt to simulate the observational temperature record used a GCM which modelled the transient[37] response to anthropogenic radiative forcing (Hadley Centre, 1995). Being more computationally complex than energy-balance models (see chapter 4), the Hadley Centre model could simulate both global-mean temperature changes and inter-regional differences. The model was run, incorporating the forcing histories of both greenhouse gases and aerosols.

For the first time, a GCM was able to replicate in broad terms the slow rise in global temperature since the middle of the last century. If greenhouse gases alone were influencing climate, one would expect global temperatures to have risen by some 0.6 to 1.3°C over the last 100 years (IPCC, 1990a). By taking into account the anthropogenic sulphate aerosols, the Hadley Centre model simulated a rise in temperature close to the observed 0.5°C.

What makes the experiment so interesting was that the GCM contained simulations of the atmosphere, oceans, ice and vegetation, and can therefore be considered to be a much better representation of the climate system than earlier GCMs, containing only the atmospheric component. Such a model significantly increased the confidence in scientists' assertion that current global warming is due to an anthropogenically enhanced greenhouse effect, albeit muted by the radiative effects of atmospheric aerosol loading. Indeed, it was in response to the Hadley Centre model, and similar ones to it more recently, that the Intergovernmental Panel on Climate Change (IPCC, 1995) indicated that the balance of [modelling] evidence suggests a discernible human influence on the global climate does exist. Nevertheless, in view of the reservations highlighted by Kelly & Wigley (1992) concerning the uncertainty of the climate sensitivity, it remains arguable whether the cause (enhanced greenhouse forcing) and effect (global warming) have been linked unequivocally.

[37] The distinction between transient and equilibrium climate change is fully discussed in section 2.8.

6.8.2. Attribution and the Fingerprint Method

In the last section, it was recognised that although a good case for linking the enhanced greenhouse effect to global warming could be proposed, a high degree of confidence could not be attached to such a cause-and-effect relationship from studies of a single variable (surface temperature). Linking cause and effect is referred to as attribution. Confidence in the attribution is increased as model predictions of changes in various components of the climate system are borne out by the observed data in more and more detail. This method is known as the fingerprint approach; namely, identification of an observed signal that has a structure unique to the predicted enhanced greenhouse effect (Madden & Ramanathan, 1980). It was noted that the use of global-mean surface temperature as a fingerprint variable does not permit attribution.

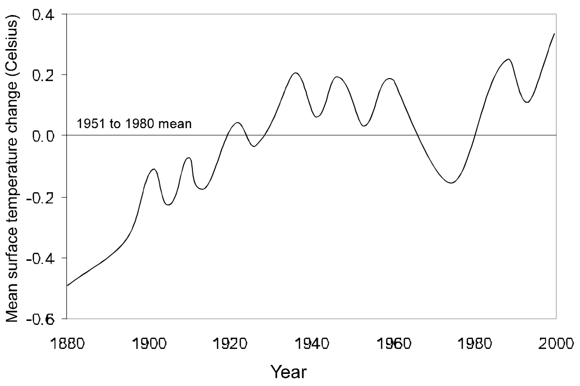
The fingerprint method is essentially a form of model validation (see section 4.5), where the perturbation experiment that is being used to test the models is the currently uncontrolled emissions of greenhouse gases into the atmosphere. In addition to global-mean surface temperature, there are a number of other fingerprint variables that can be used. In choosing them, three issues must be considered:

1) the signal-to-noise ratio - should be maximised;
2) uncertainties in both the predicted signal and the noise - these should be minimised, and;
3) the availability of suitable observational data.

6.8.2.1. Latitudinal Surface Temperatures

Most simulations suggest that the warming north of 50°N in the winter half of the year should be enhanced due to ice-albedo feedback mechanisms (Manabe & Stouffer, 1980; Ingram *et al*., 1989). Over the last 100 years, high northern latitudes have warmed slightly more than the global mean during winter, but since the 1920s there is little noticeable trend (see Figure 6.13). More importantly, however, high-latitude warming does not necessarily represent a unique fingerprint of greenhouse warming. Ice-albedo feedbacks could be equally invoked for other causes of climate change.

Figure 6.13. *Standardised deviation of surface air temperature 50^o - 90^{o}N, smoothed curve (after Wigley & Barnett, 1990)*



6.8.2.2. Tropospheric Warming and Stratospheric Cooling

All equilibrium model simulations show a warming in the mid-troposphere and cooling in the stratosphere. It has been suggested that this contrast in trends between the troposphere and stratosphere might prove a useful detection fingerprint (Karoly, 1987, 1989). In addition, high signal-to-noise ratios have been obtained (Barnett & Schlesinger, 1987) for free tropospheric temperatures. However, stratospheric cooling may not solely be attributed to greenhouse forcing, and can arise due to volcanic pollution injections and ozone depletion. Model simulations are also inconsistent with recent observations (Angell, 1988, and section 6.7.3.1), in particular at the level at which warming reverses to cooling. Perhaps most significantly of all, though, is the lack of reliable instrumental data. The instrumental record for upper atmosphere temperatures extends back only to the late 1950s with the use of radiosonde, whilst satellite-based data is even younger.

6.8.2.3. Global-Mean Precipitation Increase

GCM models suggest an increase in global-mean precipitation, as one might expect from the associated increase in atmospheric temperature. However, because the spatial variability of rainfall is much greater than for surface temperature, regional and local details of changes are highly uncertain (IPCC, 1990a). Instrumental data from which long-term changes in precipitation can be determined are only available over land areas (Bradley *et al*., 1987; IPCC, 1990a), and this data suffers from the problems of incomplete coverage and inhomogeneity. With the likelihood that the precipitation signal-to-noise ratio is low, any meaningful comparison between observation and model, and therefore attribution, is precluded. In addition, since global precipitation is partially dependent upon global temperature, rises in global precipitation may be expected as a consequence of other causes of climate change.

6.8.2.4. Sea Level Rise

Increasing greenhouse gases are expected to cause a rise in the global-mean sea level, due partly to oceanic thermal expansion and partly to the melting of land-based ice masses. However, as for global precipitation, such a variable is not wholly independent of global temperature. Thus, whilst both thermal expansion and glacial melting are consistent with global warming, neither provides any independent information about the cause of the warming.

6.8.2.5. Multivariate Fingerprints

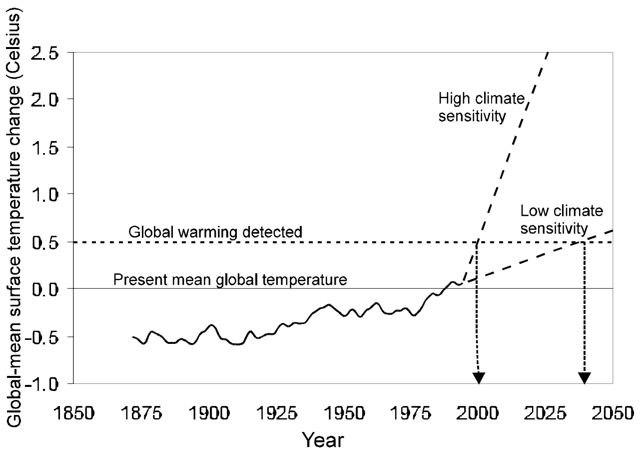
Although most univariate detection methods of global climate change due to an enhanced greenhouse effect have their limitations, it is likely that multivariate fingerprint methods, which involve the simultaneous use of several time series, will facilitate attribution. In its most general form one might consider the time evolution of a 3-D spatial field, comparing model results with observations. This may be achieved by comparing changes in mean values and variances, or correlating spatial patterns between simulation and observation (Wigley & Barnett, 1990).

6.8.3. When Will Attribution Occur?

The fact that the cause (the enhanced greenhouse effect) and effect (global warming) have not unambiguously been linked leads to the question: when is this [attribution] likely to occur? Detection is not a simple yes/no issue. Rather, it involves the gradual accumulation of observational evidence in support of model predictions. The scientist's task is to reduce the uncertainties associated with the understanding of the climate system. Thus, while the ultimate objective of climate research is detection, it is really the climate sensitivity (section 2.8) one is interested in. A better understanding of this key parameter will not only allow one to argue with increased confidence that anthropogenic global warming is (or is not) occurring, but it will support prediction of future climate change, both anthropogenic and natural.

Figure 6.14 illustrates how the uncertainty in the climate sensitivity puts constraints on the timing of the detection issue. GCMs indicate that the climate sensitivity (the CO_2 doubling temperature) lies in the region 1.5 to 4.5^{o}C. A high climate sensitivity will allow detection[38] before 2010. Conversely, if it is low, detection may not occur for another 50 years.

Figure 6.14. *Timing of detection of anthropogenic global warming (after Wigley & Barnett, 1990)*



6.9. Future Climate Change

The issue of climate change detection discussed in the last section introduces the wider concern of future climate change. Although predictions of climate change thousand, millions and even hundreds of millions of years from now could be made based on our knowledge of longer term climate changes (see chapters 2 and 5), I will be concerned in this section only with the current decadal to century time scale.

[38] Arbitrarily, detection may be agreed to mean at least a 1^{o}C warming since the late nineteenth century (i.e. another 0.5^{o}C) (Wigley & Barnett, 1990).

Clearly, prediction of climate change over the next 100 to 150 years is based solely on model simulations. Understandably, the vast majority of modelling has concentrated on the effects of continued anthropogenic pollution of the atmosphere by greenhouse gases, and to a lesser extent, atmospheric aerosols. The main concern, at present, is to determine how much the Earth will warm in the near future.

6.9.1. GCM Climate Simulations

During the last decade or so, a number of complex GCMs have attempted to simulate future anthropogenic climate change. Earlier models (e.g. Manabe & Stouffer, 1980; Hansen *et al*., 1984; Mitchell *et al*., 1989; see also IPCC, 1990a) studied equilibrium climate change associated between two, presumably stable, states of climate. The results of these are discussed in the first IPCC report (1990a). Associated with a doubling of pre-industrial atmospheric CO_2, the following conclusions have been made:

a) a global average warming at or near the Earth's surface of between 1.5 and 4.5^{o}C, with a "best guess" of 2.5^{o}C, will occur;

b) the stratosphere will experience a significant cooling;

c) surface warming will be greater at high latitudes in winter, but less during the summer;

d) global precipitation will increase by 3 to 15%;

e) year-round increases in precipitation in high-latitude regions are expected, whilst some tropical areas may experience small decreases.

More recent time-dependent GCMs (e.g.. Washington & Meehle, 1989; Stouffer *et al*., 1989; Cubasch *et al*., 1990; see also IPCC, 1992) which couple the atmospheric and oceanic components of the climate system together, provide more reliable estimates of greenhouse-gas-induced climate change. For steadily increasing greenhouse forcing, the global rise in temperature is typically less than the equilibrium rise corresponding to an instantaneous forcing. Significant results indicate:

a) a global average warming of 0.3°C per decade, assuming non-interventionist greenhouse gas emission scenarios (see section 6.9.2);

b) a natural variability of about 0.3°C in global surface air temperature on decadal time scales;

c) regional patterns of temperature and precipitation change similar to equilibrium experiments, although warming is reduced in high latitude oceans where deep water is formed.

The ability to model the time-dependent nature of the climate system more adequately has allowed scientists to investigate the damping effects of the oceans on climate change (section 2.4). Because the response time of the oceans, in particular the deep ocean, is much longer than for the free atmosphere, they have a regulating or delaying effect on the warming associated with enhanced greenhouse forcing. In addition, the transient GCMs have allowed increased attention to focus on the critical role of feedback processes (section 2.7) in determining the climate's response to forcing perturbations.

6.9.2. Greenhouse Feedbacks

GCMs have estimated the CO_2 doubling temperature change in the absence of feedback processes to be approximately 1.2°C. The existence of feedback loops within the climate system results in a climate sensitivity of 1.5°C to 4.5°C, and makes an otherwise linear association between radiative forcing and global temperature distinctly non-linear. Three of the most important direct climatic feedbacks to greenhouse forcing include the water vapour feedback, the cloud feedback and the ice-albedo feedback. (See also section 2.7).

6.9.2.1. Water Vapour Feedback

The importance of water vapour feedback in climate change has long been recognised (Manabe & Wetherald, 1967). The concentration of water vapour in the atmosphere increases rapidly with rising temperature (about 6%/°C); this is the basis for the strong positive water vapour feedback seen in current climate models (whereby increases in temperature produce increases in atmospheric water vapour which in turn enhance the greenhouse forcing leading to further warming). All current GCMs simulate a strong positive water vapour feedback (Cess *et al*., 1990) and are in agreement with observational data (Raval & Ramanathan, 1989).

6.9.2.2. Cloud Feedback

Cloud feedback is the term used to encompass effects of changes in cloud and their associated radiative properties on a change of climate, and has been identified as a major source of uncertainty in climate models (Cubasch & Cess, 1990). This feedback mechanism incorporates both changes in cloud distribution (both horizontal and vertical) and changes in cloud radiative properties (cloud optical depth and cloud droplet distribution) (Wigley, 1989; Charlson *et al*., 1987); these are not mutually independent. Although clouds contribute to the greenhouse warming of the climate system by absorbing more outgoing infrared radiation (positive feedback), they also produce a cooling through the reflection and reduction in absorption of solar radiation (negative feedback) (Cubasch & Cess, 1990). It is generally assumed that low clouds become more reflective as temperatures increase, thereby introducing a negative feedback, whilst the feedback from high clouds depends upon their height and coverage and could be of either sign (Gates *et al*., 1992).

6.9.2.3. Ice-Albedo Feedback

The conventional explanation of the amplification of global warming by snow and ice feedback is that a warmer Earth will have less snow cover, resulting in a lower global albedo and consequent absorption of more radiation, which in turn causes a further warming of the climate. Most GCMs (Cess *et al*., 1990, 1991) have simulated this positive surface albedo feedback, but significant uncertainties exist over the size of the effect, particularly with sea-ice (Ingram *et al*., 1989).

6.9.2.4. Greenhouse Gas Feedbacks

In addition to these climatic feedbacks, there are other non-climatic feedbacks which may enhance or diminish the increase in atmospheric greenhouse gas accumulation. Potentially, there are numerous feedback processes which may act on the carbon cycle, thus affecting the transfer of carbon dioxide between the various components (section 6.4.1). Carbon storage in the oceans will be affected by both changes in water temperature and ocean circulation resulting from global warming. Warmer water stores less CO_2 (section 5.3.1.2), and future atmospheric CO_2 increase may be amplified by something like 5 to 10% by this effect (Lashof, 1989). A slow down in vertical overturning within the oceans will reduce the uptake of atmospheric CO_2. In addition, reduced nutrient recycling would limit primary biological productivity in the surface oceans, thereby increasing further the surface water partial pressure of CO_2, and consequently atmospheric CO_2 (section 5.3.1.2).

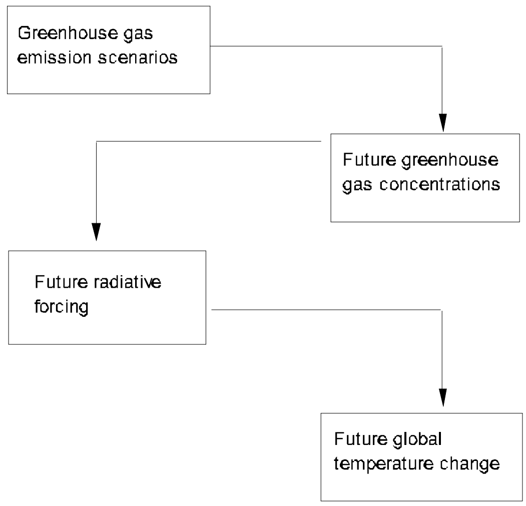
Carbon cycle feedbacks involving terrestrial carbon storage may include carbon dioxide fertilisation of plant photosynthesis (a negative feedback) (Strain & Cure, 1985; Rotmans & den Elzen, 1993); eutrophication involving nitrogen and phosphate nutrient fertilisation (a negative feedback) (Melillo *et al*., 1993); temperature feedbacks on the length of growing season and photosynthesis/respiration rates (both positive and negative feedbacks) (Lashof, 1989; Melillo *et al*., 1993); and changes in the geographical distribution of vegetation (positive or negative feedbacks(?)) (Watson *et al*., 1990). The methane gas cycle may also experience feedback effects. Recent attention has been focused on northern wetlands and permafrost, where increases in temperature and soil moisture would result in significant increases in methane release into the atmosphere (Kvenvolden, 1988).

Despite the increased ability of transient GCMs to model climate feedback processes, there is no compelling evidence to warrant changing the equilibrium sensitivity to doubled CO_2 from the range of 1.5 to 4.5^{o}C as estimated by IPCC, 1990a.

6.9.3. Climate in the 21st Century

Although GCMs provide the most detailed simulations of future climate change, computational constraints preclude their use in sensitivity studies that allow one to investigate the potential effects of real-world future greenhouse gas emissions. By using simpler climate models, it is possible to estimate future changes in the global-mean surface air temperature induced by different scenarios of radiative forcing. A number of these scientific scenarios have been identified by the IPCC (1990a, 1992). The radiative forcing scenarios themselves are determined by policy scenarios which estimate the time evolution of greenhouse gas concentrations in response to various emission scenarios. Figure 6.15 illustrates the sequence of modelling studies that has been undertaken by the IPCC.

Figure 6.15. *The IPCC modelling process (after IPCC, 1990)*



For the 1990 IPCC Scientific Assessment, four future greenhouse gas emission scenarios were developed and used to produce estimates of the resulting rate and magnitude of climate change, represented by the global average surface temperature, up to 2100. These were updated to six emissions-climate response scenarios in the 1992 report. Calculations for the 1992 report were performed using a 1-D climate model and a series of gas cycle models that relate emissions to concentration changes (STUGE, Wigley *et al*., 1991). Climate feedbacks were not explicitly modelled but their effects are accounted for by varying the value of the climate sensitivity, as determined by GCMs.

The assumptions used to formulate the IPCC 1992 emissions scenarios (IS92a to f) may be found in IPCC (1992 & 1995), whilst Figure 6.16 shows the time evolution of CO_2 emissions, the main greenhouse gas, for each scenario. IS92a and IS92b are revised versions of the IPCC 1990 "business-as-usual" scenario (SA90) which assumed no climate-interventionist policy during the next century. IS92c and IS92d provide example of possible "best-case" scenarios" whilst IS92e and IS92f represent possible "worst-case" scenarios. Similar time evolutions exist for the other greenhouse gases. Figure 6.17 illustrates the greenhouse radiative forcing associated with each of the emission scenarios, calculated from the estimated increase in greenhouse gas concentrations. Figure 6.18 then shows the global-mean temperature change estimates for each case, assuming the "best-guess" value for the climate sensitivity (2.5°C) as modelled by GCMs. The simulated temperature increases across the range climate sensitivities are shown for the IS92a scenario in Figure 6.19.

Figure 6.16. *Projected CO_2 emissions to 2100 for the IPCC scenarios (after IPCC, 1992)*

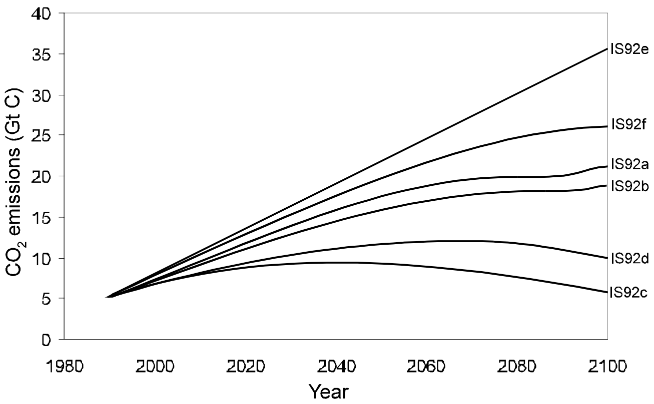
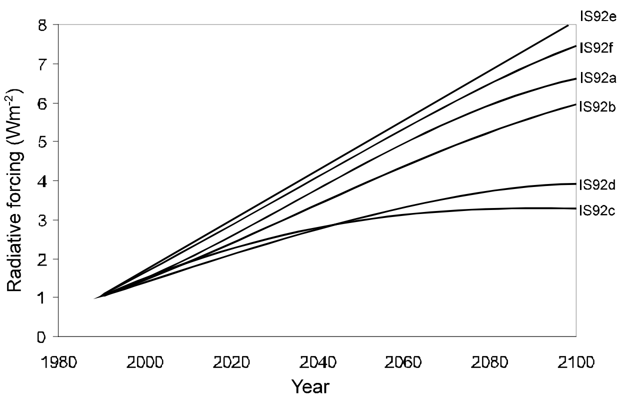


Figure 6.17. *Projected direct radiative forcing due to greenhouse gas increases for the IPCC emission scenarios (after IPCC, 1992)*



Using the "best-estimate" sensitivity, these scenarios give a range of warming from 1.5 to 3.5°C by the year 2100 (Figure 6.18). Under "business as usual" conditions, global-mean surface temperature is expected to rise by between 2 and 4°C over the next hundred years, depending on the climate sensitivity (Figure 6.19). Clearly, even the most optimistic projections of greenhouse gas accumulation seem unable to prevent a significant change in global climate during the next century. Under a "worst-case" scenario with a high climate sensitivity, global-mean surface temperature could rise by 6°C by 2100. Significantly, however, these projections are based solely on changes in greenhouse gases; no account has been taken of the negative forcing influence of stratospheric ozone loss or sulphate aerosols.

Figure 6.18. *Projected global mean surface temperature increases for the IPCC emission scenarios (after IPCC, 1992)*

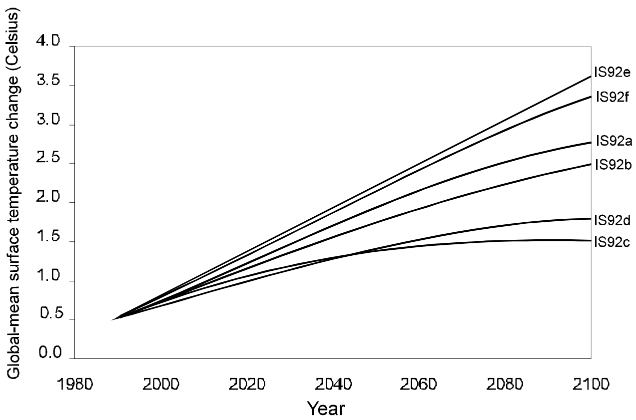
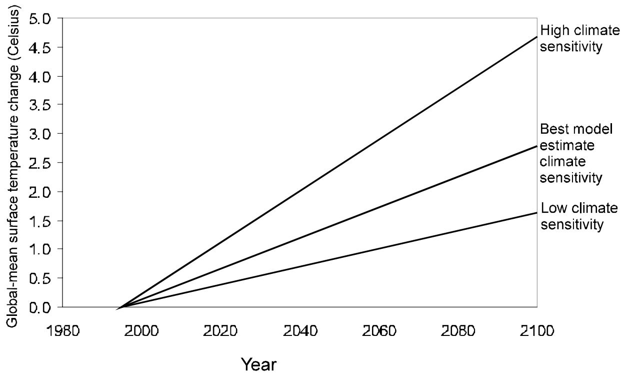
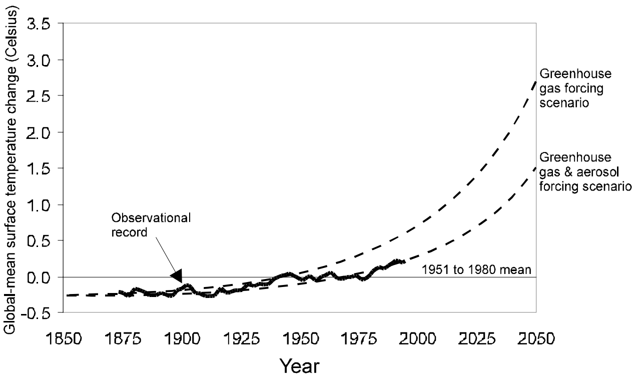


Figure 6.19. *Projected global mean surface temperature increase for the IS92a emission scenario with varying climate sensitivity (after IPCC, 1992)*



Using "business-as-usual" projections of greenhouse gas emissions and sulphate aerosols, the Hadley Centre used a transient GCM to simulate the climatic effects of both positive greenhouse forcing and negative aerosol and stratospheric ozone forcing (Hadley Centre, 1995). As well as replicating the twentieth century global temperature increase (section 6.8), the model also refined future projections of climate change. Figure 6.20 illustrates the modelled past and future global-mean temperature changes, from 1860 to 2050. The negative forcing of sulphate aerosols, despite its regionality, unarguably reduces the projected temperature increase beyond the present. Under "business-as-usual" conditions, and with a "best-estimate" climate sensitivity, the global-mean surface temperature is projected to rise by approximately 0.2°C per decade. This can be compared to the IPCC (1992) prediction of a 0.3°C rise per decade.

Figure 6.20. *Projected global mean surface temperature increase to 2050 for the Hadley Centre model (1995)*



6.10. Impacts of Future Climate Change

If the climate system responds to enhanced greenhouse forcing in a manner that has been projected by climate models, the Earth is likely to experience a number of detrimental impacts. Obviously, much depends upon the magnitude of the climate sensitivity, a variable as yet poorly constrained by the present understanding of the climate system. In addition, climate change, and therefore climate impacts, are poorly defined at the regional level. Nevertheless, it is possible to make generalisations of the likely consequences of future anthropogenic climate change, and to attach levels of certainty to these. The IPCC Climate Change Impacts Assessment (IPCC, 1990b) provides an excellent review in this context.

As the climate system shifts to a new equilibrium, each of its components and sub-components will respond to the complex interaction of feedback loops. Climate impacts can therefore be expected to occur throughout the system. Some of the major impact areas identified (IPCC, 1990b) include agriculture, forestry, natural ecosystems, hydrology and water resources, human settlements and health, and oceans and coastal zones. Global-mean surface temperature increase, regional temperature increases, precipitation increases and decreases, soil moisture availability, climatic variability and the occurrence of extreme events such as hurricanes are all likely to influence the nature of these impacts. Climatic impacts, whether detrimental or benign, will also have to be viewed in the context of a rapidly expanding global population.

6.10.1. Agriculture

There is no overall consensus as to whether anthropogenic climate change will increase or decrease global agricultural potential. Negative impacts could be experienced at the regional level as a result of changes in weather, moisture availability (Parry, 1990), pests associated with climate change (EPA, 1989) and changes in ground level ozone associated with pollutants. Areas of existing vulnerability, including SE Asia, the Sahel in N. Africa and parts of South America, may experience the most severe effects (UNEP, 1989). Parts of the high and mid latitudes may see increases in potential productivity as the length of the growing season expands, but this is likely to be confined largely to the Northern Hemisphere (Parry & Carter, 1988; Parry *et al*., 1989).

On balance, global food production may be maintained at or near present day levels (Rosenzweig *et al*., 1994); however, the cost of achieving this is unclear, especially in view of a rapidly growing world population.

6.10.2. Forestry

The rotation period of forests is long and trees will find it increasingly difficult to adjust to the swifter changes of climatic regimes (Solomon, 1986; Solomon & West, 1986). Actual impacts will depend upon the physiological adaptability of trees and their host-parasite relationships. Tree species close to their biological limits (in terms of temperature and moisture) will be most sensitive to climate change. Areas experiencing warmer, drier climates may also see an increase in the incidence of forest-fires (Frosberg, 1989). As well as the impact to forest ecology, climate change will also augment social stresses of forest-dependent populations, and consequent anthropogenic damages to forests can be expected. Such increased and non-sustainable forest use will put further pressure on forest investments, conservation and management (Winget, 1987).

6.10.3. Natural Terrestrial Ecosystems

Most climate models agree that a reduction in the equator to pole temperature gradient, at least on land, will be experienced in a greenhouse world. Consequently, climatic zones could shift several hundred kilometres towards the poles during the next century (Emmanuel *et al*., 1985), although the rate of transition is uncertain. As flora and fauna would lag behind these climatic shifts, they would be subject to considerable stresses. Many species may be unable to adapt to different climatic and meteorological regimes, and will become extinct.

The rate of projected climate changes is critical in determining the nature of impacts on terrestrial ecosystems and their ability to adapt. Even a projected temperature increase of 0.2^{o}C per decade (Hadley Centre, 1995) is at least twice the maximum rate of increase that most ecosystems could withstand (IPCC, 1990b). The most susceptible ecological systems will include montane, alpine, polar, island and coastal communities where climatic changes will add to existing stresses.

6.10.4. Hydrology and Water Resources

Relatively small climatic variations can create large problems for water resources, especially in arid and semi-arid regions. However, the climatic impacts on hydrology at the regional scale are uncertain, and will be influenced by a complex mix of temperature, precipitation, evaporation, soil moisture availability and run off changes. For many marginal areas, such as the Sahelian region in Africa, the rates of present desertification (Nicholson, 1989; Demarée & Nicolis, 1990) are likely to increase, having significant implications for agriculture, water storage, distribution and power generation. Other areas, such as NW Europe, may experience increases in precipitation, with resultant increases in the risk of flooding, as water tables rise.

Patterns of water demand will need to alter in response to changes in water supply (Fiering & Rogers, 1989). This will be true for areas at risk from both drought and flooding, and effective water management strategies will be required.

6.10.5. Oceans and Coastal Zones

Global warming will accelerate sea-level rises due to thermal expansion of seawater and the melting of land-based ice sheets. If current GCMs projections of a 4cm per decade rise are reasonable, sea level may stand at least half a metre higher by the end of the next century. This will seriously threaten many low-lying islands and coastal zones, rendering some countries uninhabitable (Lewis, 1988). Other impacts include the increased risk of coastal flooding (Kana *et al*., 1984; Titus *et al*., 1987) and salination of fresh groundwater supplies, exacerbated by an increasing occurrence of droughts or storms. Rapid sea-level rise would damage the coastal ecology, threatening important fisheries.

Changes in ocean circulation are also projected to occur as a result of changing global climate. This will not only influence marine ecosystems (due to changes in upwelling zones), but will dramatically affect regional meteorological patterns (in response to the shifting heat balance). Palaeoclimatic studies (e.g. Boyle & Keigwin, 1987; Boyle, 1988) have revealed that rapid circulation changes, within only a few decades, have occurred in the past.

6.10.6. Human Settlements and Health

The most vulnerable populations to the impacts of climate change will be those already under pressure from social, economic and existing climate stresses. These will include those in developing countries, in the lower income groups, residents of coastal lowlands and islands, populations in semi-arid grasslands, and the urban poor (Tickell, 1989). Increased exposure to natural hazards, such as coastal or river flooding, drought, landslides, storms and hurricanes, will prove detrimental to those most at risk. In the industrialised world many of the largest cities are situated in low-lying regions, and rises in sea level will threaten large land areas of immense economic wealth.

Major health impacts can be expected as changes in precipitation and temperature will affect the patterns of vector-borne and viral diseases. Incidences of malaria, for example, could spread to higher latitudes outside pre-existing tropical bounds (Martens *et al*., 1995). Such impacts could initiate large migrations of people, causing substantial disruptions of settlement patterns and social instability.

6.11. Responses to Anthropogenic Climate Change

Section 6.9 examined the possible future scenarios of anthropogenic climate change, dependent on the rate at which mankind continues to alter the composition of the atmosphere and affect the radiative balance of the climate system. Even the IPCC's most optimistic emissions projections do not preclude the possibility that severe impacts of climate change (discussed in section 6.10) will be experienced during the twenty-first century. Although detection of greenhouse-forced climate change has not been detected with absolute certainty (section 6.8), the time has now come to assess just what is required to avert a potentially unprecedented destabilisation of the climate system's present equilibrium.

6.11.1. Stabilising Greenhouse Gas Concentrations

Table 6.9 illustrates the reductions in emissions of the most important greenhouse gases that would be required now to stabilise atmospheric concentrations at present day levels (IPCC, 1990a).

Table 6.9. *Reductions in greenhouse gas emissions required to stabilise atmospheric concentrations*

Greenhouse Gas	Reduction Required
Carbon Dioxide	60%
Methane	15 - 20%
Nitrous Oxide	70 - 80%

The **Montreal Protocol** and subsequent amendments have already significantly reduced many of the halocarbon species, particularly the CFCs, by at least 90% in developed nations, to halt the destruction of stratospheric ozone. Paradoxically, as the ozone layer slowly repairs itself, its greenhouse forcing will increase.

Far more importantly, however, are the other greenhouse gases, most noticeably carbon dioxide, the major greenhouse gas that is increasing anthropogenically, and the by-product of the world's energy production. Clearly, a 60% cut in carbon dioxide emissions, either now or over the next few years, would be almost impossible to achieve. Even the most optimistic IPCC emissions scenario (IS92c) projects a rise in carbon emissions by 2025, with only a gradual decline by the year 2100. In view of this, it is worth examining what action is being taken by the global community to safeguard against the impacts of anthropogenic climate change.

6.11.2. The Framework Convention on Climate Change

In response to the reports from the IPCC, the United Nations Intergovernmental Negotiating Committee for a Framework Convention on Climate Change (INC/FCCC) was established, and adopted by over 150 countries in 1992 as a **blueprint** for **precautionary action**. The main objective of the FCCC (Article 2) was to achieve a stabilisation of greenhouse gas concentrations in the atmosphere at a level that would prevent dangerous anthropogenic interference with the climate system (UNEP, 1992). In order to achieve this, the party nations were committed to drawing up national programmes to return emissions of greenhouse gases to 1990 levels by the year 2000.

The United Kingdom signed the Convention in 1992 and ratified it in December 1993. The then Government was obligated to: formulate and implement a national programme to reduce greenhouse gas emissions, in particular carbon dioxide; provide assistance to developing countries and countries with their economies in transition; support research into climate change, and; promote public education and awareness concerning climate change.

6.11.3. The Kyoto Protocol

An historic agreement to cut emissions of the main greenhouse gases which are attributed to global warming was agreed in December 1997 in Kyoto, Japan, at the third Conference of Parties to the Framework Convention. Industrial nations agreed to reduce their collective emissions of greenhouse gases by 5.2% from 1990 levels by the period 2008 to 2012.

Crucially, the Kyoto Protocol commits developed countries to make legally binding reductions in their greenhouse gas emissions. The six gases include carbon dioxide, methane, nitrous oxide, HFCs, perfluorocarbons (PFCs) and sulphur hexafluoride.

The Kyoto Protocol was endorsed by 160 countries. It will become legally binding provided at least 55 countries sign up to it, including developed nations responsible for at least 55% of emissions from the industrialised world.

Significantly, the global cut of 5.2% is to be achieved by differential reductions for individual nations. The European Union, Switzerland and the majority of Central and Eastern European nations will deliver reductions of 8%; the US will cut emissions by 7%; and Japan, Hungary, Canada and Poland by 6%. New Zealand, Russia and the Ukraine are required to stabilise their emissions, whilst Australia, Iceland and Norway are permitted to increase slightly, although at a reduced rate to "business as usual" scenarios.

Included in the Protocol is the provision for emission reduction trading amongst developed countries, and a "Clean Development Mechanism" that allows developed countries to credit projects aimed at reducing emissions in developing nations towards their own reduction targets. Carbon sinks are also included within the Protocol. They may be used to offset increases in emissions elsewhere, provided that they result from "direct human-induced land-use change and forestry activities".

6.11.4. The UK Programme

The first UK national strategy, in response to the Framework Convention, aimed to stabilise CO_2 emissions at 1990 levels by 2000. As a consequence of the Kyoto Protocol, the UK is now committed to a 12% reduction in greenhouse gas emissions by 2008 to 2012 (as part of a Europe-wide 8% reduction) from 1990 levels. Within the latest UK Programme for greenhouse gas emission reduction, the current Government has proposed that a full 20% reduction in CO_2 emissions can be achieved by 2010. The emission of CO_2 over the next 10 to 15 years will be a function of the development of the UK economy, the demand for goods and services, population increase and the strategies and policies adopted by government. The UK Programme focuses on the need to reduce CO_2 emissions, the main greenhouse gas, although this does not detract from measures formulated to reduce other greenhouse gases (DoE, 1994). A range of measures aimed at all sectors of the population has been identified including legislation, economic instruments, regulation, information and education.

The emissions of CO_2 from various end user categories (Table 6.10) in the UK in 1990 totalled 158 million tonnes (MtC). Latest energy projections indicate that CO_2 emissions will be 4-8% lower by the beginning of the 21st century than those of 1990 (Dti, 1995).

Table 6.10. *Emissions of CO_2 by final energy user in the UK in 1990 (DoE, 1994).*

End User	Mt C
Industry	48
Domestic	42
Road Transport	33
Commercial	13
Public Service	10
Other Transport	5
Other Emissions	4
Agriculture	2
Exports	1
TOTAL	158

The volume of CO_2 emitted is directly related to the type of fuel used and the amount of energy consumed, thus any reduction in CO_2 emissions will be dependent on 1) using less carbon intensive fuels, 2) improving the production and delivery of energy, and 3) utilising energy more efficiently.

6.11.4.1. Energy Demand

Four energy demand sectors have been identified in the UK Programme as areas where emission reduction measures can be effectively implemented.

1. Domestic energy consumption produced 42 MtC in 1990, equivalent to 27% of emissions. A reduction in emissions from this sector should be achieved with measures aimed at improving energy efficiency. They include: 8% VAT on domestic fuel (introduced in April 1994); financial assistance from the Energy Saving Trust and advice from local energy advice centres (LEACs); public awareness campaigns; and efficiency standards for energy use in buildings.

2. Energy consumption in the business sector was responsible for 48 MtC or 30% of CO_2 emissions in

1990. During the 1990s, the aim has been to reduce emissions through the implementation of the *Making a Corporate Commitment Campaign* (involving over 1600 organisations), and promoting the use of the Energy Efficiency Office's advisory services (via the Best Practice Programme).

3. Energy consumption in the public sector resulted in 9.6 MtC in 1990. The aim has been to reduce energy consumption in the public sector by 20%.

4. Transport accounted for 24% of the UK's CO_2 emissions in 1990, and this is forecast to rise to 26% by 2000. Emissions from this sector can be reduced through improved fuel efficiency of vehicles, increased fuel duties and increased use of public transport.

6.11.4.2. Energy Supply

Electricity generation from fossil fuels is a major source of carbon emissions, accounting for over 30% of CO_2 emissions in 1990. A reduction in emissions attributable to electricity generation will be achieved through the development of new and renewable energy sources, including the increased use of combined heat and power schemes (CHP), and wind, solar and hydroelectric technologies through the Non Fossil Fuel Obligation (NFFO).

6.11.5. Evaluation of the FCCC, the Kyoto Protocol and the UK Programme

The UK is now committed, through the Kyoto Protocol, to a 12% reduction in greenhouse gas emissions by 2008 to 2012 from 1990 levels. The UK Programme of CO_2 emission reduction furthermore aims to reduce CO_2 emissions by 20% by 2010. Recent energy projections (Dti, 1995) show that the UK is confident of meeting the original commitment of the Framework Convention for a stabilisation in CO_2 emissions by 2000, and may exceed it by between 6 and 13 MtC below 1990 levels.

Despite the Kyoto Protocol however, emission reductions of the magnitude shown in Table 6.9 have not yet established themselves on the global or national political agenda. Yet it seems that without more stringent targets for emission reductions, the world will be unable to avert a climate change more rapid than anything that has occurred during geologic time.

Despite current scientific uncertainties regarding the causal link between increased greenhouse gases and a rise in global temperature, it seems probable that current FCCC commitments are at odds with its own underlying objective, to prevent dangerous anthropogenic interference with the climate system in a time-frame sufficient to allow natural ecosystems to adapt.

Although the FCCC aims to be precautionary in nature, just what is required by global policy to prevent dangerous anthropogenic interference with the climate system is far from clear. Comparison of temperature changes observed in the natural analogue, as indicated above, provide a starting framework. Nevertheless, the whole issue of the precautionary principle is at present, poorly defined (O'Riordan & Jordan, 1995). Most usually, it is interpreted to mean the implementation of action to prevent interference with the global climate in spite of scientific uncertainty (Article 3.3, UNEP, 1992). Included within this definition are the ideas of cost-effectiveness and responsibility.

Assigning a monetary (and indeed a moral) value to the impacts of global warming is fraught with the difficulties of ethical subjectivism (Buchdahl, 1997). Turner (1995), and O'Riordan & Jordan (1995) have offered some sort of framework, distinguishing between the weak and strong precautionary standpoints of traditional economists and contemporary environmental analysts respectively. Such plurality of opinion depends to a large extent on differing views on the vulnerability or resilience of the global climate. All parties call for scientific evidence to back up their respective arguments. Ironically, as much of this chapter has illustrated, the greatest scientific uncertainty in climate change prediction is this climate sensitivity or vulnerability to the change in radiative forcing associated with a build up of greenhouse gases. Although the essence of the precautionary principle is its call for preventative action, in spite of existing scientific uncertainty, it paradoxically seems that the very success of the FCCC rests on finding a better understanding of the climate system.

The issue of responsibility for global warming may prove to be even more of a contentious issue as nations try to agree on strategies to protect the global climate. Ultimately, it is not clear how the burden of responsibility should be shared amongst the parties of the FCCC. Article 4.2a of the FCCC recognised the different economic starting points of nations to be considered when drawing up national CO_2 reduction strategies. However, should developed nations, such as the UK, be required to shoulder more of the burden of CO_2 emission reduction, in light of their substantially greater energy expenditure, present, past and future? Equally, can developing nations reasonably be expected to stabilise or reduce their greenhouse gas emissions to the detriment of their fledgling economies?

Many developing nations regard the developed world as the cause of the global warming problem. Before developing nations attempt to reduce emissions, developed countries, they argue, should implement their own effective reduction strategies. Some see the concept of joint implementation (UNEP, 1992) as a workable solution, but here again, poorer states view this as ineffective at mitigating the threat of climate change.

Perhaps what is really at issue here, and is often conveniently overlooked, is that developing nations, whilst being only minor contributors to the global warming problem at present, are most likely to suffer the severest impacts resulting from any future climate change.

To date, the instruments for achieving policy goals have been solely advisory, as nations have shied away from using legislative measures to implement greenhouse gas emission reductions. This piecemeal approach to climate change abatement could shift the burden of responsibility on to those less able to manage the task (Buchdahl *et al*., 1995). Clearly, the issue for a nation state is providing the maximum environmental protection commensurate with political acceptability. In the case of the UK the level of acceptable carbon abatement has been defined through a comprehensive consultation process which identified the target reductions, the instruments to achieve these targets and the sectors in which instruments could be deployed with the greatest cost effectiveness and maximum efficiency.

6.12. Conclusion

This final chapter has comprehensively reviewed the evidence for contemporary anthropogenic climate change. The changing composition of the atmosphere (involving both greenhouse gases and aerosols) was discussed and its potential influence on greenhouse forcing and climate change examined. Despite the lack of unequivocal attribution for the enhanced greenhouse hypothesis, the evidence that mankind is changing the Earth's climate is beginning to mount.

If climate model projections prove to be even moderately accurate, global temperatures by the end of the next century will be higher than at any time during the last 120,000 years (see chapter 5). With such unprecedented climate change, impacts to all parts of the climate system are likely to be substantial. Failure to introduce some form of global greenhouse gas emission reduction strategy will merely extend the time frame of anthropogenic global warming that humanity may already be witnessing.

Epilogue

Global climate change results from a combination of periodic external and internal forcing mechanisms, and a complex series of interactive feedbacks within the climate system itself. These climate changes occur over a whole range of time scales from a few years to hundreds of millions of years.

The traditional view of climate change has been one of cause and effect, where the climate system responds, generally in a linear fashion to climatic forcing. Only now is humanity beginning to realise that this view is too simple. In recent years climate models have re-emphasised the complex nature of the climate system and our limited understanding of its behaviour. The very nature of the feedback processes highlights the non-linearity of the climatic system.

External forcing sets the pace of climate change; but the internal (non-linear) dynamics of the climate system modulate the final response. This is true for all time scales. Galactic variations impose external variations in insolation, whilst tectonic movements through mantle convection regulate the climate changes over hundreds of millions of years. Milankovitch orbital variations act as a pacemaker to the internal variations of ocean circulation and atmospheric composition.

Today we may be witnessing one of the most profound climatic changes in the Earth's history. Certainly, larger changes in global climate have occurred in the past, but over much longer time periods. The danger facing the global society today is that anthropogenic global warming may be too fast to allow humans, and other species, to adapt to its detrimental impacts. In addition, through enhanced greenhouse forcing, we may be pushing the climate system towards a bifurcation point, where climatic responses may become highly non-linear through complex feedback processes, driving the system to a completely different, and most probably, inhospitable state for humankind.

The challenge for scientists is to understand the climate system, and ultimately predict changes in global climate. To this end, greater collaboration is required between modellers, empiricists and policy makers. Ultimately, the climate system may be too complex to simulate reliably, and the study of global climate change will remain an imprecise science.

In light of this, the precautionary approach to mitigating the threats of anthropogenic global climate change must be fully recognised and adopted by the international community. In addition, increased emphasis will need to focus on the study and modelling of the impacts of future global warming. Greater integration between scientific and policy scenarios will be beneficial to the management and control of future impacts to society. Greater emphasis on impact scenarios at the regional level is also needed, if society is truly to "think globally" and "act locally". Indeed, the challenge for society as a whole is to respond to current dangers regarding global warming, and ultimately to "manage" the climate system in a sustainable and responsible manner.

Appendix: The Geological Time Scale

	ERA	PERIOD	AGE (Ma)	EPOCH	MAJOR GEOLOGICAL EVENTS
PHANEROZOIC	CENOZOIC	Quaternary	0.01	Holocene	
			2	Pleistocene	
		Tertiary	5	Pliocene	
			26	Miocene	Himalayan mountain building
			37	Oligocene	
			53	Eocene	Alpine mountain building
			65	Palaeocene	
	MESOZOIC	Cretaceous	136		Atlantic opens
		Jurassic	190		
		Triassic	225		Pangea breaks apart
	PALAEOZOIC	Permian	280		
		Carboniferous	345		Continents converge to form Pangea
		Devonin	395		
		Silurian	430		
		Ordovocian	500		
		Cambrian	570		Formation of Gondwanaland
	PRECAMBRIAN	Proterozoic	2300		
		Archean			

References

Andreae, M.O., 1994. Climatic effects of changing atmospheric aerosol levels. In: *World Survey of Climatology, Vol. XVI: Future Climates of the World*, Henderson-Sellers (ed.), Elsevier, Amsterdam.

Angell, J.K., 1988. Variations and trends in tropospheric and stratospheric global temperatures, 1958-87. *J. Climate*, **1**, pp. 1296-1313.

Arthur, M.A., Dean, W.E. & Schlanger, S.O., 1985. Variations in the Cretaceous related to climate volcanism, and changes in atmospheric CO_2. In: *The Carbon Cycle and Atmospheric CO_2: Natural Variations Archean to Present.* Sundquist, E.T. & Broecker, W.S (eds.). Geophys. Mon., **32**, Amer. Geophys. Union, Washington D.C., pp. 504-529.

Balkanski, Y.J., Jacob, D.J., Gardner, G.M., Graustein, W.C. & Turekian. K.K., 1993. Transport and residence times of continental aerosols inferred from a global three-dimensional simulation of ^{210}Pb. *J. Geophys. Res.*, **98**, pp. 20573-20586.

Barnett, T.P. & Schlesinger, M.E., 1987. Detecting changes in global climate induced by greenhouse gases. *J. Geophys. Res.*, **92**, pp. 14772-14780.

Barnola, J.M., Raynaud, D., Korotkevich, Y.S. & Lorius, C., 1987. Vostok ice core provides 160,000 year record of atmospheric CO_2, *Nature*, **329**, pp. 408-414.

Baron, W.R., 1982. The reconstruction of eighteenth century temperature records through the use of content analysis. *Cliamtic Change*, **4**, pp. 385-398.

Barron, E.J., 1983. A warm, equable Cretaceous: The nature of the problem. *Earth Sci. Rev.*, **19**, pp. 305-338.

Barron, E.J., 1985. Explanations of the Tertiary global cooling trend. *Palaeogeog., Palaeoclim., Palaeoecol.*, **50**, pp. 45-61.

Barron, E.J., Thompson, S.L. & Schneider, S.H., 1981. An ice-free Cretaceous? Results from climate model simulation. *Science*, **212**, pp. 501-508.

Barron, E.J. & Washington, 1985. Warm Cretaceous climates: High atmospheric CO_2 as a plausible mechanism. In: *The Carbon Cycle and Atmospheric CO_2: Natural Variations Archean to Present.* Sundquist, E.T. & Broecker, W.S (eds.). Geophys. Mon., **32**, Amer. Geophys. Union, Washington D.C., pp.546-553.

Barry, R.G. & Perry, A.H., 1973. Statistical Methods. In: *Synoptic Climatology: Methods & Applications*. Methuen, London.

Barry, R.G. & Chorley, R.J., 1992. *Atmosphere, Weather & Climate*, 6th edn. Routledge, London. 392pp.

Bates, T.S., Charlson, R.J. & Gammon, R.H., 1987. Evidence for the climate role of marine biogenic sulphur. *Nature*, **329**, pp. 319-321.

Beaty, C., 1978a. The causes of glaciation. *Amer. J. Sci.*, **66**, pp. 452-459.

Beaty, C., 1978b. Ice ages and continental drift. *New Scientist*, **80**, pp. 776-777.

Berger, A.L., 1976. Obliquity and precession for the last 5,000,000 years. *Astron. Astrophys.*, **51**, pp. 127-135.

Berger, A.L., 1978. Long-term variations of caloric insolation resulting from the Earth's orbital elements. *Quat. Res.*, **9**, pp. 139-167.

Berger, A.L., 1984. Accuracy and frequency stability of the Earth's orbital elements during the Quaternary. In: *Milankovitch & Climate*, Berger, A.L., Imbrie, J., Hays, J., Kukla, G. & Saltzman, B. (eds.). D. Reidal, Dordrecht, Netherlands, pp. 3-39.

Bergthorsson, P., 1969. An estimate of drift ice and temperature in Iceland in 1000 years. *Jökull*, **19**, pp. 94-101.

Berner, R.A., Lasaga, A.C. & Garrels, R.M., 1983. The carbonate-silicate geochemical cycle and its effect on atmospheric carbon dioxide over the last 100 million years. *Amer. J. Sci.*, **283**, pp. 641-683.

Bolin, B, Döös, B.R., Jäger, J. & Warrick, R.A. (eds.), 1986. *The Greenhouse Effect, Climate Change, and Ecosystems*. SCOPE Vol. 29, John Wiley & Sons Ltd., Chicester, 539pp.

Bond, G.C., Nickeson, P.A. & Kominz, M.A., 1984. Breakup of a supercontinent between 625Ma and 555Ma: New evidence and implications for continental histories. *Earth Plan. Sci. Lett.*, **70**, pp. 325-345.

Bowler, J.M., 1976. Aridity in Australia: age origins and expression in aeolian landforms and sediments. *Earth Sci. Rev*. **12**, pp. 279-310.

Boyle, E.A., 1988. Vertical oceanic nutrient fractionation and glacial/interglacial CO_2 cycles. *Nature*, **331**, pp. 55-58.

Boyle, E.A. & Keigwin, L.D., 1987. North Atlantic thermohaline circulation during the past 20,000 years linked to high-latitude surface temperatures. *Nature*, **330**, pp. 35-40.

Bradbury, J.P., Leyden, B., Salgado-Labouriau, M., Lewis, W.M., Schubert, C., Binford, M.W., Frey, D.G., Whitehead, D.R. & Weibezahn, F.H., 1981. Late Quaternary environmental history of lake Valencia, Venezuela. *Science*, **214**, pp. 1299-1305.

Bradley, R.S., 1985. *Quaternary Palaeoclimatology: Methods of Palaeoclimatic Reconstruction*. Unwin Hyman, London, 472pp.

Bradley, R.S. & Jones, P.D., 1992. *Climate Since A.D. 1500*. Routledge, London, 679pp.

Bradley, R.S., Diaz, H.F., Eischeid, J.K., Jones, P.D., Kelly, P.M. & Goodess, C.M., 1987. Precipitation fluctuations over Northern Hemisphere land areas since the mid-19th century. *Science*, **237**, pp. 171-175.

Bretherton, F.P., Bryan, K. & Woods, J.D., 1990. Time-dependent greenhouse-gas-induced climate change. In: *Climate Change: The IPCC Scientific Assessment*, Intergovernmental Panel on Climate Change (IPCC), Houghton, J.T., Jenkins, G.J. & Ephraums, J.J. (eds.). Cambridge University Press, Cambridge, pp. 173-193.

Briffa, K.R. & Schweingruber, F.H., 1992. Recent dendroclimatic evidence of northern and central European summer temperatures. In: *Climate Since A.D.1500*. Routledge, London, pp. 366-392.

Broecker, W.S., 1974. *Chemical Oceanography*. Harcourt Brace, Jovanovitch, New York.

Broecker, W.S., 1982. Ocean chemsitry during glacial time. *Geochim. Cosmochim. Acta.*, **46**, pp. 1689-1705.

Broecker, W.S., 1987. Unpleasant surprises in the greenhouse? *Nature*, **328**, pp.123-126.

Broecker, W.S. & Peng, T.-H., 1982. *Tracers in the Sea*. Eldigio Press, Palisades, New York.

Broecker, W.S. & Denton, G.H., 1989. The role of ocean-atmosphere reorganisations in glacial cycles. *Geochimica et Cosmochimica Acta*, **53(10)**, pp. 2465-2502.

Broecker, W.S. & Denton, G.H., 1990. What drives glacial cycles? *Scientific American*, **262**, pp. 42-50.

Broecker, W.S. *et al*, 1988. The chronology of the last deglaciation: Implications to the cause of the Younger Dryas event. *Palaeoceanog.*, **3**, pp. 1-19.

Bryson, R.A., 1989. Late Quaternary volcanic modulation of Milankovitch climate forcing. *Theoretical & Applied Climatology*, **39(3)**, pp. 115-125.

Bryson, R.A. & Wendland, W.M., 1967. Tentative climate patterns for some late-glacial and post-glacial episodes in central North America. In: *Life, land and water*, Mayer-Oakes, W.J. (ed.). University of Manitoba Press, Winnipeg, pp. 271-298.

Buchdahl, J.M., 1997. Valuing Climate Change. In: Proceedings of the 8th Global Warming Conference, New York.

Buchdahl, J.M., Longhurst, J.W.S. & Rowe, E.J., 1995. An evaluation of the UK Programme for CO_2 reduction. In: *Air Pollution III Volume 2: Pollution Control and Monitoring*, Baldasano, J.M., Brebbia, C.A., Power, H. & Zannetti, P (eds.). Computational Mechanics Publications, Southampton.

Butcher, S.S., Charlson, R.J., Orians, G.H., Wolfe, G.V. (eds.), 1992. *Global Biogeochemical Cycles*. Academic Press Ltd., London.

Butler, J.H., Elkins, J.W., Thompson, T.M., Hall, B.D., Swanson, T.H. & Koropalov, V., 1991. Oceanic consumption of CH_3CCl_3: implications for tropospheric OH. *J. Geophys. Res.*, **96**, pp. 22347-22355.

Campbell, I.M., 1986. *Energy and the atmosphere: A physical-chemical approach*, 2nd edn. John Wiley & Son Ltd., Chichester, England, 337pp.

Caputo, M.V. & Crowell, J.C., 1985. Migration of glacial centres across Gondwana during the Palaeozoic Era. *Geol. Soc. Amer. Bull.*, **96**, pp. 1020-1036.

Carbon Dioxide Information Analysis Centre (CDIAC), 1991, 1993. *Trends `91, `93: A Compendium of Data on Global Change*. Oak Ridge National Laboratory, USA.

Cess, R.D. & Potter, G.L., 1988. A methodology for understanding and intercomparing atmospheric climate feedback processes in general circulation models. *J. Geophys, Res.*, **93**, pp. 8305-8314.

Cess, R.D., Potter, G.L., Blanchet, J.P., Boer, G.J., Del Genio, A.D., Déqué, M., Dymnikov, V., Galin, V., Gates, W.L., Ghan, S.J., Kiehl, J.T., Lacis, A.A., Le Treut, H., Li, Z.-X., Liang, X.-Z., McAvaney, B.J., Meleshko, V.P., Mitchell, J.F.B., Morcrette, J.-J., Randall, D.A., Rikus., L., Roeckner, E., Royer, J.F., Schlese, U., Sheinin, D.A., Slingo, A., Sokolov, A.P., Taylor, K.E., Washington, W.M., Wetherald, R.T., Yagai, I. & Zhang, M.-H.,1990. Intercomparison and interpretation of climate feedback processes in nineteen atmospheric general circulation models. *J. Geophys. Res.*, **95**, pp. 16601-16615.

Cess, R.D., Potter, G.L., Zhang, M.-H., Blanchet, J.P., Chalita, S., Colman, R., Dazlich, D.A., Del Genio, A.D., Dymnikov, V., Galin, V., Jerret, D., Keup, E., Lacis, A.A., Le Treut, H., Liang, X.-Z., Mahouf, J.-F., McAvaney, B.J., Meleshko, V.P., Mitchell, J.F.B., Morcrette, J.-J., Norris, P.M., Randall, D.A., Rikus., L., Roeckner, E., Royer, J.F., Schlese, U., Sheinin, D.A., Slingo, A., Sokolov, A.P., Taylor, K.E., Washington, W.M., Wetherald, R.T. & Yagai, I., 1991. Interpretation of snow-climate feedback as produced by seventeen general circulation models. *Science*, **253**, pp. 888-892.

Chamberlain, A.C., 1991. *Radiative Aerosols*. Cambridge Univeristy Press, Cambridge, 255pp.

Chapman, S., 1930. A theory of upper atmospheric ozone. *Quar. J. Royal Met. Soc.*, **3**, pp. 103.

Charlson, R.J., Lovelock, J.E., Andreae, M.O. & Warren, S.G., 1987. Oceanic phytoplankton, atmsopheric sulphur, cloud albedo and climate. *Nature*, **326**, pp. 655-661.

Charlson, R.J., Schwartz, S.E., Hales, J.M., Cess, R.D., Coakley, J.A., Hansen, J.E. & Hofmann, D.J., 1992. Climate forcing by anthropogenic aerosols. *Science*, **255**, pp. 422-430.

Coakley, J.A., Bernstein, R.L. & Durkee, P.A., 1987. Effects of ship-stack effluents on cloud reflectivities. *Science*, **237**, pp. 1020-1022.

COHMAP Members, 1988. Climatic change of the last 18,000 years: Observations and model simulations. *Science*, **241**, pp. 1043-1052.

Conrad, V. & Pollak, L.D., 1962. *Methods in Climatology*. Harvard University Press, Harvard.

Craig, H., 1961. Standard for reporting concentrations of deuterium and oxygen-18 in natural waters. *Science*, **133**, pp. 1833-1834.

Critchfield, H.J., 1983. *General Climatology*. Englewood Cliffs: Prentice-Hall.

Croll, J., 1867a. On the excentricity of the Earth's orbit, and its physical relations to the glacial epoch. *Philosophical Magazine*, **33**, pp. 119-131.

Croll, J., 1867b. On the change in the obliquity of the eliptic, its influence on the climate of the polar regions and on the level of the sea. *Philosophical Magazine*, **33**, pp. 426-445.

Crowell, J.C. & Frakes, L.A., 1970. Phanerozoic glaciation and the causes of ice ages. *Amer. J. Sci.*, **268**, pp. 193-224.

Crowley, T.J. & North, G.R., 1991. *Palaeoclimatology*. Oxford Monographs on Geology and Geophysics No. 18. Oxford University Press, New York, 339pp.

Crowley, T.J., Mengel, J.G. & Short, D.A., 1987. Gondwanaland's seasonal cycle. *Nature*, **329**, pp. 803-807.

Crowley, T.J., Hyde, W.T. & Short, D.A., 1989. Seasonal cycle variations on the supercontinent of Pangea. *Geology*, **17**, pp. 457-460.

Crozaz, G. & Langway, C.C. Jr., 1966. Dating Greenland firn-ice cores with ^{210}Pb. *Earth Planet. Sci. Lett.*, **1**, pp. 194-196.

Cubasch, U. & Cess, R.D., 1990. Processes and modelling. In: *Climate Change: The IPCC Scientific Assessment*, Intergovernmental Panel on Climate Change (IPCC), Houghton, J.T., Jenkins, G.J. & Ephraums, J.J. (eds.). Cambridge University Press, Cambridge, pp. 69-91.

Cubasch, U., Maier-Reimer, E., Santer, B., Mikolajewicz, U., Roeckner, E. & Boettinger, M., 1990. The response of a global coupled O-AGCM to CO_2 doubling. MPI report, MPI fuer Meteorologie, Hamburg, Germany.

Dansgaard, W. 1961. The isotopic composition of natural waters with special reference to the Greenland Ice Cap. *Meddr Grønland*, **165**, pp. 1-120.

Dansgaard, W. & Johnsen, S.J., 1969. A flow model and a time scale for the ice core from Camp Century. *J. Glaciol.*, **8**, pp. 215-223.

Dansgaard, W., Johnsen, S.J., Clausen, B., Dahl-Jensen, D., Gundestrup, N., Hammer, C.U. & Oeschger, H., 1984. North Atlantic climatic oscillations revealed by deep Greenland ice cores. In: *Climate Processes and Climate Sensitivity*. Hansen, J.E. & Takahashi, T. (eds.). Geophys. Mono., **29**, Amer. Geophys. Union, Washington D.C., pp. 288-298.

Dansgaard, W., White, J.W.C. & Johnsen, S.J., 1989. The abrupt termination of the Younger Dryas climatic event. *Nature*, **339**, pp. 532-534.

Davis, M.B., 1963. On the theory of pollen analysis. *Amer. J. Sci.*, **261**, pp. 899-912.

Davis, M.B., 1973. Redeposition of pollen grains in lake sediment. *Limnol. Oceanogr.*, **18**, pp. 44-52.

Davis, D.D. & Solomon, S.C., 1981. Variations in the velocities of the major plates since the late Cretaceous. *Technophys.*, **74**, pp. 189-208.

Dawson, M.R., West, R.M., Langston, W. & Hutchinson, J.H., 1976. Palaeogene terrestrial vertebrates: Northernmost occurrence, Ellsemere Island, Canada. *Science*, **192**, pp. 781-782.

Demarée, G. & Nicolis, C., 1990. Onset of Sahelian drought viewed as a fluctuation-induced transition. *Q.J.R. Meteorol. Soc.*, **116**, pp. 221-238.

Department of Environment (DoE), 1994. *Climate Change: the UK Programme.* HMSO, London.

Department of Trade and Industry (Dti), 1992. *Energy-Related Carbon Emissions in Possible Future Scenarios for the United Kingdom*, Energy Paper 59. HMSO, London

Department of Trade and Industry (Dti), 1995. *Energy Use and Energy-Related Emissions of Carbon Dioxide in UK, 1995-2020*, Energy Paper 65. HMSO, London.

Diaz, H.F., Bradley, R.S. & Eischeid, J.K., 1989. Precipitation fluctuations over global land areas since the late 1800s. *J. Geophys. Res.*, **94**, pp. 1195-1210.

Dotto, L. & Schiff, H., 1978. *The Ozone War.* Garden City, New York: Doubleday.

Douglas, R.G. & Woodruff, F., 1981. Deep sea benthic foraminifera. In: *The Sea*, **vol. 7**. Emiliani (ed.). Wiley-Interscience, New-York, pp. 1233-1327.

Duplessy, J.C., 1978. Isotope studies. In: *Climate Change*, Gribben, J. (ed.). Cambridge Univeristy Press, Cambridge, pp. 46-67.

Duplessy, J.C. & Shackleton, N.J., 1985. Response of global deep water circulation to Earth's climatic change 135,000 - 107,000 years ago. *Nature*, **316**, pp. 500-506.

Eddy, J.A., 1976. The Mauder Minimum. *Science*, **192**, pp. 1189-1202.

Eddy, J.A., 1977. Climate and the changing Sun. *Climatic Change*, **1(2)**, pp. 173-190.

Eddy, J.A., 1982. The solar constant and surface temperature. In: *Interpretation of Climate and Photochemical Models, Ozone and Temperature Measurements*, Peck, R.A. & Hummel, J.R. (eds.). American Institute of Physics, New York.

Emmanuel, W.R., Shugart, H.H. & Stevenson, M.P., 1985. Climate change and the broad-scale distribution of terrestrial ecosystem complexes. *Climate Change*, **7**, pp. 29-43.

Emiliani, C., 1977. Oxygen isotope analysis of the size fraction betwen 62 and 250 micrometres in Caribbean cores P6304-8 and P6304-9. *Science*, **198**, pp. 1255-1256

Emiliani, C., 1978. The cause of the Ice Ages. *Earth Plant. Sci. Lett.*, **37**, pp. 349-352.

Environmental Protection Agency (EPA), 1989. *The Potential Effects of Global Climate Change on the United States*. Report to Congress.

Epstein, S., Yapp, C.J. & Hall, J.H., 1976. The determination of the D/H ratio of non-exchangeable hydrogen in cellulose extracted from aquatic and land plants. *Earth Planet. Sci. Lett.*, **30**, pp. 241-251.

Farman, J.C., Gardiner, B.G. & Shanklin, J.D., 1985. Large losses of total ozone in Antarctica reveal seasonal ClO_x/NO_x interaction. *Nature*, **315**, pp. 207-210.

Farmer, G., Wigley, T.M.L., Jones, P.D. & Salmon, M., 1989. Documenting and explaining recent global-mean temperature changes. Climatic Research Unit, Norwich, Final Report to NERC, UK, Contract GR3/6565.

Fiering, M. & Rogers, P., 1989. *Climate Change & Water Resources Planning Under Uncertainty*. US Army Institute for Water Resources. Ft. Belvoir, 105pp.

Fischer, A.G., 1982. Long term climate oscillations recorded in stratigraphy. In: *Climate In Earth History*. Natl. Acad. Press, Washington D.C., pp. 97-104.

Fischer, A.G., 1984. The two Phanerozoic supercycles. In: *Catastrophes and Earth History: the new uniformitarianism*, Berggren, W.A. & Van Couvering, J.A. (eds.). Princeton University Press, Princeton, New Jersey, pp. 129-150.

Folland, C.K., 1988. Numerical models of the rainguage exposure problem, field experiments and an improved collector design. *Quar. Royal Met. Soc.*, **114**, pp. 1488-1516.

Folland, C.K., Karl, T.R. & Vinnikov, K.Ya., 1990. Observed climate variations and change. In: *Climate Change: The IPCC Scientific Assessment*, Intergovernmental Panel on Climate Change (IPCC), Houghton, J.T., Jenkins, G.J. & Ephraums, J.J. (eds.). Cambridge University Press, Cambridge, pp. 195-238.

Folland, C.K., Owen, J., Ward, M.N. & Colman, A., 1991. Prediction of seasonal rainfall in the Sahel region using empirical and dynamical methods. *J. Forecasting*, **10**, pp. 21-56.

Foukal, P.V., 1990. The variable Sun. *Scientific American*, **262(2)**, pp. 34-41.

Frakes, L.A., 1979. *Climates Throughout Geologic Time*. Elsevier, Amsterdam.

Fritts, H.C., 1971. Dendroclimatology and dendroecology. *Quat. Res.*, **1**, pp. 419-449.

Fritts, H.C., 1976. *Tree rings and climate*. Academic Press, London.

Frosberg, M.A., 1989. Climate change and forest fires. In: *Coping with Climate Change - Proceedings of the Second North American Conference on Preparing for Climate Change: A Cooperative Approach*, pp. 292-296, Topping Jr., J.C. (ed.). The Climate Institute, Washington D.C.

Gaffin, S.R., Hoffert, M.I. & Volk, T., 1986. Nonlinear coupling between surface temperature and ocean upwelling as an agent in historical climate variations. *J. Geophys. Res.*, **91**, pp. 3944-3950.

Gani, J., 1975. The use of statistics in climatological research. *Search*, **6**, pp. 504-508.

Gates, W.L., Rowntree, P.R. & Zeng, Q.-C., 1990. Validation of climate models. In: *Climate Change: The IPCC Scientific Assessment*, Intergovernmental Panel on Climate Change (IPCC), Houghton, J.T., Jenkins, G.J. & Ephraums, J.J. (eds.). Cambridge University Press, Cambridge, pp. 93-130.

Gates, W.L., Mitchell, J.F.B., Boer, G.J., Cubasch, U. & Meleshko, V.P., 1992. Climate modelling, climate prediction and model validation. In: *Climate Change 1992: The supplementary report to the IPCC Scientific Assessment*, Intergovernmental Panel on Climate Change (IPCC), Houghton, J.T., Callander, B.A. & Varney, S.K. (eds.). Cambridge University Press, Cambridge, pp. 97-134.

Genthon, C., Barnola, J.M., Raynaud, D., Lorius, C., Jouzel, J., Barkov, N.I., Korotkevich, Y.S. & Kotlyakov, V.M., 1987. Vostock ice core: climatic response to CO_2 and orbital forcing changes over the last climatic cycle. *Nature*, **329**, pp. 414-418.

Gibbs, W.J., Maher, J.V. & Coughlan, M.J., 1978. Climatic variability and extremes. In: *Climatic Change and Variability*, Pittock, A.B., Frakes, L.A., Jensun, D., Peterson, J.A. & Zillman, J.W., (eds.). Cambridge Univesity Press, pp. 135-150.

Gloersen, P. & Campbell, W.J., 1988. Variations in the Arctic, Antarctic, and global sea ice covers during 1979-1987 as observed with the Nimbus 7 Scanning Multichannel Microwave Radiometer. *J. Geophys. Res.*, **93**, pp. 10666-10674.

Godske, C.L., 1966. Methods of statistics and some applications to climatology. In: *Statistical Analysis and Prognosis in Meteorology*, World Meteorological Organisation (WMO), 1966, Tech. Note 71, pp. 9-86.

Goodess, C.M., Palutikof, J.P. & Davies, T.D., 1992. *The nautre and causes of climate change*. Belhaven Press, London. 248pp.

Gough, D.O., 1990. On Possible origins of relatively short- term changes in the solar structure. *Phil. Trans. Royal Soc. London, A*, **330**, pp. 627-640

Grove, J.M., 1988. *The Little Ice Age*. Methuen, London, 498pp.

Hadley Centre, 1995. *Modelling Climate Change 1860-2050*. Hadley Centre for Climate Prediction & Research, 12pp.

Hallam, A., 1984. Pre-Quaternary sea-level changes. *Ann. Rev. Earth Plan. Sci.* **12**, pp. 205-243.

Hammer, C.U., 1977. Past volcanism reveled by Greenland Ice Sheet impurities. *Nature*, **270**, pp. 482-486.

Hammer, C.U., Clausen, H.B., Dansgaard, W., Gundestrup, N, Johnsen, S.J. & Reeh, N., 1978. Dating of Greenland ice cores by flow models, isotopes, volcanic debris and continental dust. *J. Glaciol.*, **20**, pp. 3-26.

Hammer, C.U., Clausen, H.B. & Dansgaard, W., 1980. Greenland ice sheet evidence of post-glacial volcanism and its climatic impact. *Nature*, **288**, pp. 230-235.

Hansen, J.E. & Lebedeff, S., 1988. Global trends of measured surface air temperature. *J. Geophys. Res.*, **92**, pp. 13345-13372.

Hansen, J.E., Lacis, A, Rind, D., Russel, L., Stone, P., Fung, I., Ruedy, R. & Lerner, J., 1984. Climate sensitivity analysis of feedback mechanisms. In: *Climate processes and climate sensitivity*, (Hansen, J.E. & Takahashi, T.), Geophys. Monogr. Ser., **29**, AGU, Washington D.C., pp. 130-163.

Hansen, J.E., Fung, I., Lacis, A., Rind, D., Lebedeff, S., Ruedy, R. & Russell, G., 1988. Global climate changes as forecast by Goddard Institute for Space Studies (GISS) Three-Dimensional Model. *J. Geophys. Res.*, **93**, pp. 9341-9364.

Hansen, J.E., Sato, M., Lacis, A. & Ruedy, R., 1993. Climatic impact of ozone change. In: *The Impact on Climate of Ozone Change & Aerosols, Background Material from the Joint Workshop of IPCC Working Group I and the International Ozone Assessment Panel*, Hamburg, May, 1993.

Haq, B.U., Premoli-Silva, I. & Lohmann, G.P., 1977. Calcareous planktonic palaeobiogeographic evidence for major climatic fluctuations in the Early Cenozoic Atlantic Ocean. *J. Geophys. Res.*, **8**, pp. 427-431.

Haq, B.U., Hardenbol, J., & Vail, P.R., 1987. Chronology of fluctuating sea levels since the Triassic. *Science*, **235**, pp. 1156-1167.

Hare, F.K., 1979. Climatic variation and variability: empirical evidence from meteorological and other sources. In: *Proceedings of the World Climate Conference*, World Meteorological Organisation Publication No. 537. WMO, Geneva, pp. 51-87.

Hauglustaine, D.A., Granier, C, Brasseur, G.P. & Megie, G., 1994. The importance of atmospheric chemistry in the calculation of radiative forcing on the climate system. *J. Geophys. Res.*, **99**, pp. 1173-1186.

Hayes, D.E. & Frakes, L.A., 1975. General synthesis. In: *Initail Report of the Deep Sea Drilling Project*, **vol. 28**, US Govt. Print. Office, Washington D.C., pp. 919-942.

Hays, J.D. & Perruzza, A., 1972. The significance of calcium carbonate oscillations in eastern equitorial Atlantic deep-sea sediments for the end of the Holocene warm interval. *Quat. Res.*, **2**, pp. 335-362.

Hays, J.D., Imbrie, J. & Shackleton, N.J., 1976. Variations in the Earth's orbit: Pacemaker of the ice ages. *Science*, **194**, pp. 1121-1132.

Hecht, A.D., Barrry, R.G., Fritts, H.C., Imbrie, J., Kutzbach, J., Mitchell, J.M. & Savin, S.M., 1979. Palaeoclimatic research: status and opportunities. *Quat. Res.*, **12**, pp. 6-17

Henderson-Sellers, A., 1986. Cloud changes in a warmer Europe. *Clim. Change*, **8**, pp. 25-52.

Henderson-Sellers, A., 1989. North American total cloud amount variations this century. *Glob. & Planet. Change*, **1**, pp. 175-194.

Henderson-Sellers, A. & Robinson, P.J., 1986. *Contemporary Climatology*. Longman Group UK Ltd., Harlow, England. 439pp.

Henderson-Sellers, A. & McGuffie, K., 1987. *A Climate Modelling Primer*. Wiley, Chichester.

Hofmann, D.J., Oltmans, S.J., Harris, J.M., Solomon, S., Deshler, T. & Johnson, B.J., 1992. Observation and possible causes of new ozone depletion in Antarctica in 1991. *Nature*, **359**, pp. 283-287.

Hofmann, D.J., Deshler, T.L., Aimedieu, P., Matthews, W.A., Johnston, P.V., Kondo, Y., Sheldon, W.R., Byrne, G.J. & Benbrook, J.R., 1989. Stratospheric clouds and ozone depletion in the Arctic during January 1989. *Nature*, **340**, pp. 117-121.

Hood, L.L. & Jirikowic, J.L., 1990. Recurring variations of probable solar origin in the atmospheric $\delta^{14}C$ time record. *Geophys. Res. Lett.*, **17(1)**, pp. 85-88.

Hoyle, F., 1981. *Ice*. Hutchinson, London.

Huff, F.A., 1970. Sampling errors in measurement of mean precipitation. *J. App. Met.*, **9**, pp. 35-44.

Huggett, R.J., 1991. *Climate, Earth Processes and Earth History*. Springer-Verlag. 281pp.

Imbrie, J. & Imbrie, K.P., 1979. *Ice Ages: Solving the Mystery*. Macmillan Press Ltd., London, 224pp.

Ingram, M.J., Underhill, D.J. & Farmer, G., 1981. The use of documentary sources for the study of past climates. In: *Climate and History*, Wigley, T.M.L, Ingram, M.J. & Farmer, G. (eds.). Cambridge University Press, Cambridge, pp. 180-213.

Ingram, W.J., Wilson, C.A. & Mitchell, J.F.B., 1989. Modelling climate change: an assessment of sea ice and surface albedo feedbacks. *J Geophys. Res.*, **94**, pp. 8609-8622.

Intergovernmental Panel on Climate Change (IPCC), 1990a. *Climate Change: The IPCC Scientific*

Assessment, Houghton, J.T., Jenkins, G.J. & Ephraums, J.J. (eds.). Cambridge University Press, Cambridge, 365pp.

Intergovernmental Panel on Climate Change (IPCC), 1990b. Tegart, W.J.McG, Sheldon, G.W. & Griffiths, D.C. (eds.). *Climate Change: The IPCC Impacts Assessment*. Australian Government Publishing Service, Canberra.

Intergovernmental Panel on Climate Change (IPCC), 1992. Houghton, J.T., Callander, B.A. & Varney, S.K. (eds.). *Climate Change 1992: The supplementary report to the IPCC Scientific Assessment*, Cambridge University Press, Cambridge, 200pp.

Intergovernmental Panel on Climate Change (IPCC), 1995. Houghton, J.T., Meira Filho, L.G., Bruce, J., Lee, H., Callander, B.A., Haites, E., Harris, N. & Maskell, K. (eds.). *Climate Change 1994: Radiative Forcing of Climate Change, and An Evaluation of the IPCC IS92 Emission Scenarios*, Cambridge University Press, Cambridge, 339pp.

Intergovernmental Panel on Climate Change (IPCC), 1996. Houghton, J.T., Meira Filho, L.G., Callander, B.A., Harris, N., Kattenberg, A. & Maskell, K. (eds.). *Climate Change 1995: The Science of Climate Change*, Cambridge University Press, Cambridge, 572 pp.

Isaksen, I.S.A., Ramaswamy, V., Rodhe, H. & Wigley, T.M.L., 1992. Radiative forcing of climate. In: *Climate Change 1992: The supplementary report to the IPCC Scientific Assessment*, Intergovernmental Panel on Climate Change (IPCC), Houghton, J.T., Callander, B.A. & Varney, S.K. (eds.). Cambridge University Press, Cambridge, pp. 47-67.

Iversen, J., 1944. *Viscum, Hedera* and *Ilex* as climatic indicators. A contribution to the study of past-glacial temperature climate. *Geol. Fören. Stockholm Förhandl.*, **66**, pp. 463-483.

Jacobson, G. & Bradshaw, R., 1981. The selection of sites for palaeovegetational studies. *Quat. Res*., **16**, pp. 80-96.

Jansen, E., Bleil, U., Henrich, R., Kringstad, L. & Slettemark, B., 1988. Palaeoenvironmental changes in the Norwegian Sea and the northeast Atlantic during the last 2.8 m.y.: Deep Sea Drilling Project/Ocean Drilling Project sites 610, 642, 643 and 644. *Palaeoceanog.*, **3**, pp. 563-581.

Johnson, G.T. & Linacre, E.T., 1978. Estimation of winds at particular places. In: *Proceedings of International Clean Air Conference, Brisbane*, White, E.T., Hetherington, P. & Thiele, B.R. (eds.). Ann Arbor Scientific, pp. 751-762.

Jonas, P.R., Charlson, R.J. & Rodhe, H., 1995. Aerosols. In: *Climate Change 1994: Radiative Forcing of Climate Change, and An Evaluation of the IPCC IS92 Emission Scenarios*, Intergovernmental Panel on Climate Change, Houghton, J.T., Meira Filho, L.G., Bruce, J., Lee, H., Callander, B.A., Haites, E., Harris, N. & Maskell, K. (eds.). Cambridge University Press, Cambridge, pp. 127-162.

Jones, P.D., 1988. Hemespheric surface air tempertaure variations: recent trends and an update to 1987. *J. Climate*, **1**, pp. 654-660.

Jones, P.D., Wigley, T.M.L. & Farmer, G., 1991. Marine and land temperature data sets: A comparison and a look at recent trends. In: *Greenhosue-Gas-Induced Climate Change: A Critical Appraisal of Simulations and Observations*, Schlesinger, M.E. (ed.). Elsevier, Amsterdam, pp. 153-172.

Jones, A., Roberts, D.L. & Slingo, A., 1994. A climate model study of the indirect radiative forcing by anthropogenic sulphate aerosols. *Nature*, **370**, pp. 450-453.

Jouzel, J. *et al.*, 1987. Vostok ice core: A continuous isotope temperature record over the last climatic cycle (160,000 years). *Nature*, **329**, pp. 403-408.

Kana, T.W., Michel, J., Hayes, M.O. & Jensen, J.R., 1984. The physical impact of sea level rise in the area of Charlston, South Carolina. In: *Greenhouse Effect and Sea Level Rise*, Barth, M.C. & Titus, J.G. (eds.). Van Nostrand Reinhold Co., New York, NY, Chap. 4, pp. 105-150.

Karl, T.R. & Steurer, P.M., 1990. Increased cloudiness in the United States during the first half of the twentieth century: Fact or fiction. *Geophys. Res. Lett.*

Karlén, W., 1980. Reconstruction of past climatic conditions from studies of glacier-front variations. *World Met. Org. Bull.*, **29**, pp. 100-104.

Karoly, D.J., 1987. Southern Hemisphere temperature trends: A possible greenhouse gas effect? *Geophys. Res. Lett.*, **14**, pp. 1139-1141.

Karoly, D.J., 1989. Northern Hemisphere temperature trends: A possible greenhouse gas effect? *Geophys. Res. Lett.*, **16**, pp. 465-468.

Kaufman, Y.J. & Chou, M.D., 1993. Model simulations of the competing climatic effects of SO_2 and CO_2. *J. Climate*, **6**, pp. 1241-1252.

Kaye, J., Penkett, S. & Ormond, F., 1994. Report on *Concentrations, Lifetimes and Trends of CFCs, Halons and Related Species*, NASA Reference Publication 1339, NASA Office of Mission to Planet Earth, Science Division, Washington D.C.

Keigwin, L.D., 1980. Palaeoceanographic change in the Pacific at the Eocene-Oligocene boundary. *Nature*, **287**, pp. 722-725.

Kelly, P.M. & Sear, C.B., 1984. Climatic impact of explosive volcanic eruptions. *Nature*, **311**, pp. 740-743.

Kelly, P.M. & Wigley, T.M.L., 1992. Solar cycle length, greenhouse forcing, and global climate. *Nature*, **360**, pp. 328-330.

Kemp, D.D., 1994. *Global environmental issues: a climatological approach*, 2nd ed. Routledge, London, 224pp.

Kennett, J.P., 1976. Phenotypic variation in some recent and late Cenozoic planktonic foraminifera. In: *Foraminifera*, **Vol. 2**, Hedley, R.H. & Adams, C.G. (eds.), Academic Press, New York, pp. 111-170.

Kiehl, J.T. & Brigleb, B.P., 1993. The relative role of sulphate aerosols and greenhouse gases in climate forcing. *Science*, **260**, pp. 311-314.

Kim, Y. & Cess, R.D., 1993. Effect of anthropogenic sulphate aerosols on low level cloud albedo over oceans. *J. Geophys. Res.*, **98**, pp. 14883-14885.

Koerner, R.M., 1977a. Devon Island Ice Cap: core stratigraphy and palaeoclimate. *Science*, **196**, pp. 15-18.

Koerner, R.M., 1977b. Distribution of microparticles in a 299m core through the Devon Island ice cap, North West Territories, Canada. In: *Symposium on Isotopes and Impurities in Snow and Ice*, International Association of Scientifc Hydrology Publication No. 118. IASH, Washington D.C., pp. 371-376.

Kolla, V., Biscaye, P.E. & Hanley, A.F., 1979. Distribution of quartz in late Quaternary Atlantic sediments in relation to climate. *Quat. Res.*, **11**, pp. 261-277.

Kuhn, J.R., Libbrecht, K.G. & Dicke, R.H., 1988. The surface temperature of the Sun and changes in the solar constant. *Science*, **242**, pp. 908-910.

Kukla, G.J., 1970. Correlation between loesses and deep-sea sediments. *Geologiska Föreningeas i Stockholm Förhandlingar*, **92**, pp. 148-180.

Kutzbach, J.E. & Gallimore, R.G., 1989. Pangean climates: Megamonsoons of the megacontinent. *J. Geophys. Res.*, **94**, pp. 3341-3357.

Kvenvolden, K.A., 1988. Methane hydrates and global climate. *Global Biogeochem. Cycles*, **2**, pp. 221-230.

Lacis, A.A., Wuebbles, D.J. & Logan, J.A., 1990. Radiative forcing of climate by changes in the vertical distribution of ozone. *J. Geophys. Res.*, **95**, pp. 9971-9981.

Lamb, H.H., 1970. Volcanic dust in the atmosphere; with a chronology and assessment of its meteorological significance. *Phil. Trans. Royal Soc. A*, **266**, pp. 452-533.

Lamb, H.H., 1982. *Climate, History and the Modern World*. Methuen & Co., Ltd., London.

Langway, C.C., Jr., 1970. *Stratigraphic analysis of a deep ice core from Greenland*, Geological Society of America Special Paper 125. Geological Society of America, Boulder, Colorado.

Lashof, D.A., 1989. The dynamic greenhouse: Feedback processes that may influence future concentrations of atmospheric trace gases and climate change. *Climatic Change*, **14**, pp. 213-242.

Law, C.S. & Owen, N.J.P., 1990. Significant flux of atmospheric nitrous oxide from the north-west Indain Ocean. *Nature*, **346**, pp. 826-828.

Legrand, M. & Delmas, R.J., 1987. A 220-year continuous record of volcanic H_2SO_4 in the Antarctic ice sheet. *Nature*, **327**, pp. 671-676.

Legrand, M.R., Delmas, R.J. & Charlson, R.J., 1988. Climate forcing implications from Vostok ice-core sulphate data. *Nature*, **334**, pp. 418-420.

Lewis, J., 1988. The Implications of Sea Level Rise for Island and Low Lying Countries. Report for Commonwealth Secretariat.

Libby, L.M., 1972. Multiple thermometry in palaeoclimatic and historic climate. *J. Geophys. Res.*, **77**, pp. 4310-4317.

Linacre, E., 1992. *Cliamte data and resources: a reference and guide*. Rouledge, London, 366pp.

Lloyd, C.R., 1982. The mid-Cretaceous earth: Palaeogeography, ocean circulation and temperature, atmospheric circulation. *J. Geol.*, **90**, pp. 393-413.

Lorenz, E.N., 1991. Chaos, spontaneous climatic variations and detection of the greenhouse effect. In: *Greenhosue-Gas-Induced Climate Change: A Critical Appraisal of Simulations and Observations*, Schlesinger, M.E. (ed.). Elsevier, Amsterdam, pp. 445-453.

Lorius, C., Jouzel, J., Raynaud, D., Hansen, J. & Le Treut, H., 1990. The ice core record: climate sensitivity and future greenhouse warming. *Nature*, **347**, pp. 139-145.

MacCracken, M.C. & Luther, F.M. (eds.), 1985. *Detecting the Climate Effects of Increasing Carbon Dioxide*. US Department of Energy, Carbon Dioxide Research Division, Washington D.C., 198pp.

Machida, T., Nakazawa, T., Tanaks, M., Fujii, Y., Aoki, S. & Watanabe, O., 1994. Atmospheric methane and nitrous oxide concentrations during the last 250 years deduced from H15 Ice Core, Antarctica. *Proceedeings International Symposium on Global Cycles of Atmsopheric Greenhouse gases*, Sendai, Japan, 7-1- March, 1994, pp. 113-116.

Madden, R.A. & Ramanathan, V., 1980. Detecting climate change due to increasing carbon dioxide. *Science*, **209**, pp. 763-768.

Manabe, S. & Wetherald, R.T., 1967. Thermal equilibrium of the atmosphere with a given distribution of relative humidity. *J. Atmos. Science*, **24**, pp. 241-259.

Manabe, S. & Stouffer, R.J., 1980. Sensitivity of a global climate model to an increase of CO_2

concentration in the atmosphere. *J. Geophys. Res.*, **85**, pp. 5529-5554.
Manabe, S. & Stouffer, R.J., 1988. Two stable equilibria of a coupled ocean-atmosphere model. *J. Climate*, **1(9)**, pp. 841-866.
Manley, G., 1974. Central England temperatures: monthly means 1659 to 1973. *Quar. J. Royal Met. Soc.*, **100**, pp. 389-405.
Marenco, A., Gouget, H., Nédelec, P., Pagés, J-P. & Karcher, F., 1994. Evidence of a long-term increase in tropospheric ozone from Pic du Midi data series - consequences: positive radiative forcing. *J. Geophys. Res.*, **99**, pp. 16617-16632.
Martens, W. *et al.*, 1995. *Environmental Helath Perspectives*, **103**, pp. 458-464.
Matson, P. & Vitousek, P., 1990. Ecosystems approach to a global nitrous oxide budget. *Bioscience*, **40**, pp. 667-672.
Matthews, J.A., 1980. Some problems and implications of ^{14}C dates from a podzol buried beneath an end moraine at Haugabreen, southern Norway. *Geogr. Annlr.*, **62A**, pp. 185-298.
McBean, G. & McCarthy, J., 1990. Narrowing the uncertainties. In: *Climate Change: The IPCC Scientific Assessment*, Intergovernmental Panel on Climate Change (IPCC), Houghton, J.T., Jenkins, G.J. & Ephraums, J.J. (eds.). Cambridge University Press, Cambridge, pp. 311-328.
Melillo, J.M., Callaghan, T.V., Woodward, F.I., Salati, E. & Sinha, S.K., 1990. Effect on Ecosystems. In: *Climate Change: The IPCC Scientific Assessment*, Intergovernmental Panel on Climate Change (IPCC), Houghton, J.T., Jenkins, G.J. & Ephraums, J.J. (eds.). Cambridge University Press, Cambridge, pp. 281-310.
Melillo, J.M., McGuire, A.D., Kicklighter, D.W., Moore III, B., Vorosmarty, C.J. & Schloss, A.L., 1993. Global climate change and terrestrial net primary production. *Nature*, **363**, pp. 234-240.
Milankovitch, M.M., 1941. *Canon of Insolation and the Ice Age Problem*. Königlich Serbische Academie, Belgrade. English translation by the Israel Program for Scientific Translations, United States Department of Commerce and the National Science Foundation, Washington D.C.
Mitchell, J.F.B., Senior, C.A. & Ingram, W.J., 1989. CO_2 and climate: A missing feedback? *Nature*, **341**, pp. 132-143.
Mitchell, J.F.B., Manabe, S., Meleshko, V.P. & Tokioka, T., 1990. Equilibrium climate change - and its implications for the future. In: *Climate Change: The IPCC Scientific Assessment*, Intergovernmental Panel on Climate Change (IPCC), Houghton, J.T., Jenkins, G.J. & Ephraums, J.J. (eds.). Cambridge University Press, Cambridge, pp. 131-172.
Mitchell, J.M., 1976. An overview of climatic variability and its causal mechanisms. *Quaternary Research*, **6(4)**, pp. 481-494.
Mitchell, J.M., Stockton, C.W. & Meko, D.M., 1979. Evidence of a 22 year rhythm of drought in the western United States related to the Hale solar cycle since the 17th century. In: *Solar-Terrestrial Influxes on Weather and Climate*, McCormack, B.M. & Seliga, T.A. (eds.). D. Reidel, Dordrecht, pp. 125-144.
Mitchell, J.M. *et al.*, 1966, *Climate Change. Technical Note No. 79*. World Meteorological Organisation.
Moore, R.M. & Tokarczyk, R., 1993. Volatile biogenic halocarbons in the north-west Atlantic. *Global Biogeochem. Cycles*, **7(1)**, pp. 195-210.
Moore, T.C., Loutit, T.S. & Grenlee, S.M., 1987. Estimating short term changes in eustatic sea level. *Palaeoceanog.*, **2**, pp. 625-637.
Morgan, V.I., 1982. Antarctic Ice Sheet surface oxygen isotope values. *J. Glaciol.*, **28**, pp. 315-323.
Mueller, C.C. & Kidder, E.H., 1972. Rainguage catch variations due to airflow disturbances around a standard rainguage. *Water Resources Res.*, **8**, pp. 1077-1082.
Mysak, L.A. & Manak, D.A., 1989. Arctic sea-ice extent and anomalies, 1953-1984. *Atmospher-Ocean*, **27**, pp. 376-405.
Nakazawa, T., Machida, T., Tanaks, M., Fujii, Y., Aoki, S. & Watanabe, O., 1993. Differences of the atmsopheric CH_4 concentration between Arctic and Antarctic regions in pre-industrial/agricultural era. *J. Geophys. Res. Lett.*, **20**, pp. 943-946.
Neftel, A., Oeschger, H., Staffelbach, T. & Stauffer, B., 1988. CO_2 record in the Byrd ice core 50,000 - 5,000 years BP. *Nature*, **331**, pp. 609-611.
Newell, N.E., Newell, R.E., Hsiung, J. & Zhongxiang, W., 1989. Global marine temperature variation and the solar magnetic cycle. *Geophys. Res. Lett.*, **16(4)**, pp. 311-314.
Nicholson, S.E., 1989. Long term changes in African rainfall. *Weather*, **44**, pp. 46-56.
Nicolis, C., 1984. Self-oscillations, external forcings, and climate predictability. In: *Milankovitch & Climate*, Berger, A.L., Imbrie, J., Hays, J., Kukla, G. & Saltzman, B. (eds.). D. Reidal, Dordrecht, Netherlands, pp. 637-652.
Nicolis, C. & Nicolis, G., 1984. Is there a climate attractor? *Nature*, **311**, pp. 529-532.
Nye, J.F., 1965. A numerical modelling method of inferring the budget history of a glacier from its advance and retreat. *J. Glaciol.*, **5**, pp. 589-607.
O'Riordan, T. & Jordan, A., 1995. *Precautionary Principle, Science, Politics & Ethics*, Working Paper PA 95-02. Centre for Social & Economic Research on Global Environment (CSERGE).

Oeschger, H., Stauffer, B., Bucher, P. & Loosli, H.H., 1977. Extraction of gases and dissolved and particulate matter from ice in deep boreholes. In: *Isotopes and impurities in Snow and Ice*, International Association of Scientifc Hydrology Publication No. 118. IASH, Washington D.C., pp. 307-311.

Officer, C.B. & Drake, C.L., 1983. The Cretaceous-Tertiary transition. *Science*, **219**, pp. 1383-1390.

Olderman, L.R., 1987. Characterisation of main experimental sites and sub-sites and questions of instrumentation. In: *Agricultural Environments*, Bunting, A.H. (ed.). Commonwealth Agricultural Bureau International, pp. 101-112.

Palmer, T., 1989. A weather eye on unpredictability. *New Scientist*, **124**, pp. 56-59.

Parrish, J.T., Ziegler, A.M. & Scotese, C.R., 1982. Rainfall pattern and the distribution of coals and evaporites in the Mesozoic and Cenozoic. *Palaeogeog., Palaeoclimatol., Palaeoecol..*, **40**, pp. 67-101.

Parry, M.L., 1990. *Climate Change & World Agriculture*. London: Earthscan Publications Ltd.

Parry, M.L. & Carter, T.R., 1988. The assessments of the effects of climate variations on agriculture: aims, methods and summary of results. In: *The Impact of Climate Variations on Agriculture, Volume 1, Assessments in Cool Temperate and Cold Regions*, Parry, M.L., Carter, T.R. & Konijn, N.T. (eds.). Kluwer, Dordrecht, The Netherlands.

Parry, M.L., Carter, T.R. & Porter, J.H., 1989. The Greenhouse Effect and the Future of UK Agriculture. *J. Royal Agric. Soc. Eng.*, pp. 120-131.

Paterson, W.S.B., Koerner, R.M., Fisher, D., Johnsen, S.J., Clausen, H.B., Dansgaard, W., Bucher, P. & Oeschger, H., 1977. An oxygen isotope climatic record from the Devon Ice Cap, Arctic Canada. *Nature*, **266**, pp. 508-511.

Petit, J.R., Briat, M. & Royer, R.A., 1981. Ice age aerosol content from East Antarctica ice core samples and past wind strength. *Nature*, **293**, pp. 391-394.

Petit, J.R., Mounier, L., Jouzel, J., Korotkevich, Y.S. & Kotlyakov, V.M. & Lorius, C., 1990. Palaeoclimatological and chronological implications of the Vostok core dust record. *Nature*, **343**, pp. 56-58.

Pickering, K.T. & Owen, L.A., 1994. *An introduction to global environmental issues*. Routledge, London, 390pp.

Porter, S.C., 1979. Hawaiian glacial ages. *Quat. Res.*, **12**, pp. 161-187.

Porter, S.C., 1986. Pattern and forcing of northern hemisphere glacier variations during the last millenium. *Quat. Res.*, **26**, pp. 27-48.

Porter, S.C. & Orombelli, G., 1985. Glacier contraction during the middle Holocene in the western Italian Alps: Evidence and implications. *Geology*, **13**, pp. 296-298.

Posey, J.W. & Clapp, P.F., 1964. Global distribution of normal surface albedo. *Geofis. Int.*, **4**, pp. 33-48.

Powell, C. McA. & Veevers, J.J., 1987. Namurian uplift in Australia and South America triggered the main Gondwanan glaciation. *Nature*, **326**, pp. 117-119.

Prather, M., Derwent, R., Ehhalt, D., Fraser, P., Sanhueza, E. & Zhou, X., 1995. Other trace gases and atmospheric chemsitry. In: *Climate Change 1994: Radiative Forcing of Climate Change, and An Evaluation of the IPCC IS92 Emission Scenarios*, Intergovernmental Panel on Climate Change (IPCC), Houghton, J.T., Meira Filho, L.G., Bruce, J., Lee, H., Callander, B.A., Haites, E., Harris, N. & Maskell, K. (eds.). Cambridge University Press, Cambridge, pp. 73-126.

Quinn, W.H. & Neal, V.T., 1992. The historical record of El Niño events. In: *Climate Since A.D. 1500*. Bradley, R.S. & Jones, P.D. (eds.). Routledge, London, pp. 623-648.

Rampino, M.R. & Volk, T., 1988. Mass extinctions, atmospheric sulphur and climatic warming at the K/T boundary. *Nature*, **332**, pp. 63-65.

Raval, A. & Ramanathan, V., 1989. Observational determination of the greenhouse effect. *Nature*, **342**, pp. 758-761.

Raymo, M.E. & Ruddiman, W.F., 1992. Tectonic forcing of late Cenozoic climate. *Nature*, **359**, pp. 117-122.

Raymo, M.E., Ruddiman, W.F. & Clement, B.M., 1986. Pliocene-Pleistocene palaeoceanography of the North Atlantic at Deep Sea Drilling Project 609. *Initail Report of the Deep Sea Drilling Project*, **94**, pp. 895-901.

Raymo, M.E., Ruddiman, W.F. & Froelich, P.N. 1988. Influence of late Cenozoic mountain building on ocean geochemical cycles. *Geology*, **16**, pp. 649-653.

Raymo, M.E., Ruddiman, W., Shackleton, N.J. & Oppo, D.W., 1990. Evolution of Atlantic-Pacific $\delta^{13}C$ gradints over the last 2.5 m.y. *Earth Plan. Sci. Lett.*, **97**, pp. 353-368.

Raynaud, D. & Lorius, C., 1973. Climatic implications of total gas content in ice at Camp Century. *Nature*, **243**, pp. 283-284.

Ritchie, J.C. & Haynes, C.V., 1987. Holocene vegetation zonation in the eastern Sahara. *Nature*, **330**, pp. 126-128.

Ritchie, J.C., Cwynar, L.C. & Spear, R.W., 1983. Evidence from north-west Canada for an early Holocene Milankovitch thermal maximum. *Nature*, **305**, pp. 126-128.

Robock, A., 1978. Internally and externally caused climate change. *J. Atmos. Sci.*, **35**, pp. 1111-1122.

Rosenzweig, C., Parry, M.L., Fischer, G. & Frohberg, K., 1993. *Climate Change and World Food Supply*,

Research Report No. 3. Univeristy of Oxford, Environmental Change Unit.
Rotmans, J. & den Elzen, M.G.J., 1993. Modelling feedback mechanisms in the carbon cycle: balancing the carbon budget. *Tellus*, **45B**, pp. 301-320.
Ruddiman, W.F. & Kutzbach, J.E., 1989. Forcing of late Cenozoic northern hemisphere climates by plateau uplift in southeast Asia and the American southwest. *J. Geophys. Res.*, **94**, pp. 18409-18427.
Ruddiman, W.F. & Kutzbach, J.E., 1991. Plateau uplift and climatic change. *Scientific American*, **264(3)**, pp. 42-50.
Sachs, M.H., Webb, T. & Clark, D.R., 1977. Palaeoecological transfer functions. *Ann. Rev. Earth Planet Sci*., **5**, pp. 159-178.
Saltzman, B., 1985. Palaeoclimate modeling. In: *Palaeoclimate Analysis and Modeling*, Hecht, A.D. (ed.). Wiley-Intersience, New York, pp. 341-396.
Sanberg, P.A., 1983. An oscillating trend in Phanerozoic nonskeletal carbonate mineralogy. *Nature*, **305**, pp. 19-22.
Sarmiento, J. L. & Toggweiler, J.R, 1984. A new model for the role of the oceans in determining atmospheric pCO_2. *Nature*, **308**, pp. 621-624.
Savin, S.M., 1977. The history of the Earth's surface temperature during the past 100 million years. *Ann. Rev. Earth. Plan. Sci.*, **5**, pp. 319-355.
Schimel, D., Enting, I.G., Heimann, M, Wigley, T.M.L., Raynaud, D., Alves, D. & Siegenthaler, U., 1995. CO_2 and the carbon cycle. In: *Climate Change 1994: Radiative Forcing of Climate Change, and An Evaluation of the IPCC IS92 Emission Scenarios*, Intergovernmental Panel on Climate Change (IPCC), Houghton, J.T., Meira Filho, L.G., Bruce, J., Lee, H., Callander, B.A., Haites, E., Harris, N. & Maskell, K. (eds.). Cambridge University Press, Cambridge, pp. 35-71.
Schneider, S.H., 1992. Introduction to climate modelling. In: *Climate System Modeling*, Trenberth, K.E. (ed.). Cambridge University Press, Cambridge, pp. 3-26.
Schneider, S.H., Thompson, S.L. & Barron, E.J., 1985. Mid-Cretaceous continental surface temperatures: Are high CO_2 concentrations needed to simulate above freezing winter conditions? In: *The Carbon Cycle and Atmospheric CO_2: Natural Variations Archean to Present.* Sundquist, E.T. & Broecker, W.S (eds.). Geophys. Mon., **32**, Amer. Geophys. Union, Washington D.C., pp. 554-560.
Schwarzkopf, M.D. & Ramaswamy, V., 1993. Radiative forcing due to ozone in the 1980s: Dependence on altitude of ozone change. *Geophys. Res. Lett.*, **20**, pp. 205-208.
Schweingruber, F.H., Fritts, H.C., Bräker, O.U., Drew, L.G. & Schär, E., 1978. The X-ray technique as applied to dendroclimatology. *Tree Ring Bull.*, **38**, pp. 61-91.
Sear, C.B., Kelly, P.M., Jones, P.D. & Goodess, C.M., 1987. Global surface-temperature responses to major volcanic eruptions. *Nature*, **330**, pp. 365-367.
Seidov, D.G., 1986. Numerical modelling of the ocean circulation and palaeocirculation. In: *Mesozoic and Cenozoic Oceans.* Hsü, K. (ed.). Geodynamic Series, **vol. 15**, Amer. Geophys. Union, Washington D.C., pp. 11-26.
Shackleton, N.J., 1974. Attainment of isotopic equilibrium between ocean water and the benthonic foraminifera Genus *Uvigerina*: isotopic changes in the ocean during the last glacial. In: *Les méthodes quantitatives d'étude des variations du climat au cours du Pleistocène*, Colloques Internationaux de Centre National de la Recherce Scientifique No. 219. CNRS, Paris, pp. 4-5,
Shackleton, N.J., 1977. The oxygen isotope stratigraphic record of the late Pleistocene. *Phil. Trans. Royal Soc. B*, **280**, pp. 169-179.
Shackleton, N.J., 1988. Oxygen isotopes, ice volume and sea level. *Quat. Sci. Rev.*, **6**, pp. 183-190.
Shackleton, N.J. & Kennett, J.P., 1975. Palaeotemperature history of the Cenozoic and the initiation of Antarctic glaciation: Oxygen and carbon isotope analysis in DSDP sites 277, 279 and 281. In: *Initail Report of the Deep Sea Drilling Project*, **vol. 5**, Kennet, J.P. *et al.* (eds.). US Govt. Print. Office, Washington D.C., pp. 743-755.
Shine, K., Derwent, R.G., Wuebbles, D.J. & Morcette, J.J., 1990. Radiative forcing of climate. In: *Climate Change: The IPCC Scientific Assessment*, Intergovernmental Panel on Climate Change (IPCC), Houghton, J.T., Jenkins, G.J. & Ephraums, J.J. (eds.). Cambridge University Press, Cambridge, pp. 41-68.
Shine, K.P., Fouquart, Y, Ramaswamy, V, Solomon, S. & Srinivasan, J., 1995. Radiative Forcing. In: *Climate Change 1994: Radiative Forcing of Climate Change, and An Evaluation of the IPCC IS92 Emission Scenarios*, Intergovernmental Panel on Climate Change (IPCC),. Houghton, J.T., Meira Filho, L.G., Bruce, J., Lee, H., Callander, B.A., Haites, E., Harris, N. & Maskell, K. (eds.). Cambridge University Press, Cambridge, pp. 163-203.
Shukla, J. & Mintz, Y., 1982. Influence of land-surface evapotranspiration on the Earth's climate. *Science*, **215**, pp. 1498-1501.
Siegenthaler, U. 1988. Causes and effects of natural CO_2 variations during the glacial-interglacial cycles. In: *Lecture Notes in Earth Sciences, **16**. Long & Short Term Variability of Climate*. Wanner, H. & Siegenthaler, U. (eds.). Spring-Verlag, Berlin, Heidelberg.

Siegenthaler, U. & Wenk, T., 1984. Rapid atmospheric CO_2 variations and ocean circulation. *Nature*, **308**, pp. 624-625.

Solomon, A.M., 1986. Transient response of forests to CO_2-induced climate change: simulation modelling experiments in eastern North America. *Oecologia*, **68**, pp. 567-579.

Solomon, A.M. & West, D.C., 1986. Atmospheric carbon dioxide change: agent of future forest growth or decline? In: *Effects of Changes in Stratospheric Ozone and Global Climate; Volume 3: Climate Change*, Titus, J.G. (ed.). United Nations Environment Programme, and US Environmental Protection Agency, Washington D.C., pp. 23-28.

Spencer, R.W. & Christy, J.R., 1990. Precise monitoring of global surface temperature trends from satellites. *Science*, **247**, pp. 1558-1561.

Stauffer, B., Neftel, A., Oeschger, H. & Schwander, J., 1985. CO_2 concentration in air extracted from Greenland ice samples. In: *Greenland Ice Core: Geophysics, Geochemistry & the Environment.* Langway, C.C., Oeschger, H. & Dansgaard, W. (eds.). Geophys. Mono., **33**, Amer. Geophys. Union. Washington D.C., pp. 85-90.

Stauffer, B., Lochbronner, E., Oeschger, H. & Schwander, J., 1988. Methane concentration in the glacial atmosphere was only half that of the pre-industrial Holocene. *Nature*, **332**, pp. 812-814.

Stouffer, R.J., Manabe, S. & Bryan, K., 1989. On the climate change induced by a gradual increase of atmospheric carbon dioxide. *Nature*, **342**, pp. 660-662.

Strain, B.R. & Cure, J.D., eds., 1985. *Direct Effect on Increasing Carbon Dioxide on Vegetation*, DOE/ER-0238, U.S. Department of Energy, Washington D.C.

Street-Perrott, F.A. & Perrott, R.A., 1990. Abrupt climate fluctuations in the tropics: The influence of Atlantic circulation. *Nature*, **343**, pp. 607-612.

Stuiver, M., Braziunas, T.F., Becker, B. & Kromer, B., 1991. Climatic, solar, oceanic and geomagnetic influences on late-glacial and Holocene atmosphere $^{14}C/^{12}C$ change. *Quat. Res.*, **35(1)**, pp. 1-24.

Sullivan, W., 1974. *Continents in Motion: The New Earth Debate*. McGraw Hill, New Yok.

Taylor, K. & Penner, J.E., 1994. Response of the climate system to atmospheric aerosols and greenhouse gases. *Nature*, **369**, pp. 734-737.

Tickell, Sir Crispin, 1989. Environmental Refugees. National Environment Research Council Annual Lectures, at the Royal Society , London, UK, 5 June, 1989.

Titus, J.G., Kuo, C.Y., Gibbs, M.J., LaRoche, T.B., Webb, M.K. & Waddell, J.O., 1987. Greenhouse effect, sea level rise, and coastal drainage systems. *Journal of Water Resources Planning & Management*, **113 (2)**.

Toggweiler, J.R. & Sarmiento, J. L., 1985. Glacial to interglacial changes in atmopsheric carbon dioxide: The critical role of ocean surface water in high latitudes. In: *The Carbon Cycle and Atmospheric CO_2: Natural Variations Archean to Present.* Sundquist, E.T. & Broecker, W.S. (eds.). Geophys. Mon., **32**, Amer. Geophys. Union, Washington D.C., pp. 163-184.

Trewartha, G.T. & Horn, L.H., 1980. *An Introduction to Climate*. New York: McGraw-Hill.

Turner, R.K., 1995 *Sustainable Development and Climate Change*, Working Paper PA 95-01, Centre for Social & Economic Research on the Global Environment (CSERGE).

Umbgrove, J.H.F., 1947. *The pulse of the Earth*, 2nd edn. Nijhoff, The Hague.

United Nations Environment Programme (UNEP), 1989. *Criteria for Assessing Vulnerabilty to Sea Level Rise: A Global Inventory to High Risk Areas*, UNEP and the Government of the Netherlands. Draft report, 57pp.

United Nations Environmental Programme / World Meteorological Organisation (UNEP/WMO), 1992. *Framework Convention on Climate Change*. United Nations, Geneva.

Urey, H.C., 1947. The thermodynamic properties of isotopic substances. *J. Chem. Soc.*, **152**, pp. 190-219.

Urey, H.C., 1948. Oxygen isotopes in nature and in the laboratory. *Science*, **108**, pp. 489-496.

Vail, P.R., Mitchum, R.M. & Thompson, S., 1977. Seismic stratigraphy and global changes in sea level. In: *Seismic Stratigraphy: Application to Hydrocarbon Exploration*. Amer. Soc. Pet. Geol. Mem., **26**, Amer. Soc. Pet. Geol. Mem., Tulsa, Oklahoma, pp. 83-97.

Vinnikov, K.Ya., Groisman, P.Ya. & Lugina, K.M., 1990. Empirical data on global climate changes (temperature and precipitation). *J. Climate*, **3**, pp. 662-677.

Wang, W.C., Zhuang, Y.C. & Bojkov, R.D., 1993. Climate implications of observed changes in ozone vertical distribution at middle and high latitudes of the Northern Hemisphere. *Geophys. Res. Lett.*, **20**, pp. 1567-1570.

Warren, S.G., Hahn, C.J., London, J., Chervin, R.M. & Jenne, R.L., 1988. Global distribution of total cloud cover and cloud type amounts over the oceans. NCAR Tech. Note TN 317+STR, 44pp, Boulder, Colorado, USA.

Washburn, A.L., 1979. Permafrost features as evidence of climatic change. *Earth Sci. Rev.*, **15**, pp. 327-402.

Washington, W.M. & Meehl, G.A., 1989. Climate sensitivity due to increased CO_2: experiments with a coupled atmosphere and ocean general circulation model. *Climate Dynamics*, **4**, pp. 1-38.

Watson, R.T., Rodhe, H., Oeschger, H. & Siegenthaler, U., 1990. Greenhouse gases and aerosols. In: *Climate Change: The IPCC Scientific Assessment*, Intergovernmental Panel on Climate Change (IPCC), Houghton, J.T., Jenkins, G.J. & Ephraums, J.J. (eds.). Cambridge University Press, Cambridge, pp. 1-40.

Watson. R.T., Meira Filho, L.G., Sanhueza, E. & Janetos, A., 1992. Greenhouse gases: Sources and sinks. In: *Climate Change 1992: The supplementary report to the IPCC Scientific Assessment*, Intergovernmental Panel on Climate Change (IPCC), Houghton, J.T., Callander, B.A. & Varney, S.K. (eds.). Cambridge University Press, Cambridge, pp. 25-46.

Wayne, 1991. *Chemistry of Atmosphere*. Oxford University Press, Oxford.

Weiss, N.O., 1990. Periodicity and aperiodicty in solar magnetic activity. *Phil. Trans. Royal Soc. London, A*, **330**, pp. 617-626.

Wenk, T. & Siegenthaler, U. 1985. The high latitude ocean as a control of atmospheric CO_2. In: *The Carbon Cycle and Atmospheric CO_2: Natural Variations Archean to Present.* Sundquist, E.T. & Broecker, W.S (eds.). Geophys. Mon., **32**, Amer. Geophys. Union, Washington D.C., pp. 185-194.

Wigley, T.M.L., 1976. Spectral analysis and the astronomical theory of climate change. *Nature*, **264**, pp. 629-631.

Wigley, T.M.L., 1978. Climate change since 1000 AD. In: *Evolution des atmosphères planétaires et climatologie de la terre*. Centre National D'Etudes Spatiales, Toulouse, pp. 313-324.

Wigley, T.M.L., 1987. Relative contributions of different trace gases to the greenhouse effect. *Climate Monitir*, **16**, pp. 16-29.

Wigley, T.M.L., 1988. The climate of the past 10,000 years and the role of the Sun. In: *Secular Solar and Geomagnetic Variations in the Last 10,000 years*, Stephenson, R.F. & Wolfendale, A.W. (eds.). Kluwer Academic Publishers. Dordrecht, pp. 209-224.

Wigley, T.M.L., 1989. Possible climatic change due to SO_2-derived cloud condensation nuclei. *Nature*, **339**, pp. 365-367.

Wigley, T.M.L. & Kelly, P.M., 1990. Holocene climatic change: ^{14}C wiggles and variations in solar irradiance. *Phil. Trans. Royal Soc. London A*, **330**, pp. 547-560.

Wigley, T.M.L. & Barnett, T.P., 1990. Detection of the Greenhouse Effect in the Observations. In: *Climate Change: The IPCC Scientific Assessment*, Intergovernmental Panel on Climate Change (IPCC), 1990a. Houghton, J.T., Jenkins, G.J. & Ephraums, J.J. (eds.). Cambridge University Press, Cambridge, pp. 239-255.

Wigley, T.M.L. & Raper, S.C.B., 1990. Natural variability of the climate system and detection of the greenhouse effect. *Nature*, **344**, pp. 324-327.

Wigley, T.M.L., Holt, T. & Raper, S.C.B., 1991. STUGE (An Interactive Grenhouse Model): User's Manual, Climate Research Unit, Norwich, UK, 44pp.

Wilkinson, B.H. & Given, R.K., 1986. Secular variation in abiotic marine carbonates: Constraints on Phanerozoic atmospheric carbon dioxide contents and oceanic Mg/Ca ratios. *J. Geol.*, **94**, pp. 321-333.

Williams, D.F. & Johnson, W.C., 1975. Diversity of recent planktonic foraminifera in the southern Indian Ocean and late Pleistocene temperatures. *Quat. Res.*, **5**, pp. 237-250.

Williams, G.E., 1975a. Possible relation between periodic glaciation and the flexure of the Galaxy. *Earth Planet. Sci. Lett.*, **26**, pp. 361-369.

Williams, G.E., 1975b. Late Precambrian glacial climate and the Earth's obliquity. *Geol. Mag.*, **112**, pp. 441-465.

Williams, R.B.G., 1975., The British climate during the last glaciation; an interpretation based on periglacial phenomena. In: *Ice ages: ancient and modern*, Wright, A.E. & Moseley, F. (eds.). Geological Journal Special Issue No. 6, Liverpool University Press, Liverpool, pp. 95-120.

Winget, C.H., 1987. Forest management strategies to address climate change. In: *Preparing for Climate Change - Proceedings of the First North American Conference on Preparing for Climate Change: A Cooperative Approach*. The Climate Institute, Washington D.C., pp. 328-333.

World Meteorological Organisation (WMO), 1991. Scientific Assessment of Ozone Depletion; 1991. *Global Ozone Research and Monitoring Project Report No. 25*. WMO, Geneva.

World Meteorological Organisation (WMO), 1994. Scientific Assessment of Ozone Depletion; 1994. *Global Ozone Research and Monitoring Project Report No. 37*. WMO, Geneva.

Yao, Shan-yu, 1943. The geographical distribution of floods and droughts in Chinese history 206 BC - AD 1911. *Far East. Qtly*, **2**, pp. 357-378.