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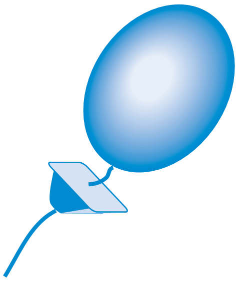
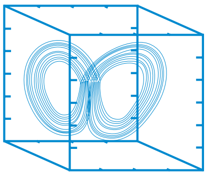
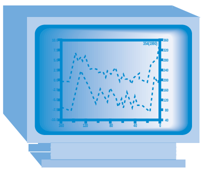
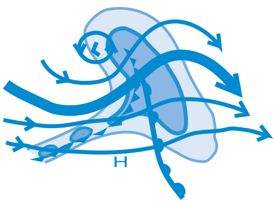


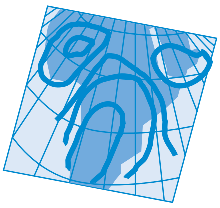
INTRODUCTION TO CLIMATE CHANGE: LECTURE NOTES FOR METEOROLOGISTS



Prepared by
David D. Houghton







WMO-No. 926

Secretariat of the World Meteorological Organization
Geneva – Switzerland

WMO TECHNICAL PUBLICATIONS

relating to education and training

WMO No.

114 — Guide to qualifications and training of meteorological personnel employed in the provision of meteorological services for international air navigation. 2nd edition, 1974. (French–Spanish)

258 — Guidelines for the education and training of personnel in meteorology and operational hydrology. 4th edition, 2001, in preparation. (English)

266 — Compendium of lecture notes for training Class IV meteorological personnel. Volume I—Earth science; 1970. (English); Volume II—Meteorology; 1984. (English–French)

364 — Compendium of meteorology for use by Class I and Class II meteorological personnel.
Volume I, Part 1—Dynamic meteorology. (French–Spanish), Part 2—Physical meteorology. (French–Spanish), Part 3—Synoptic meteorology. (English–French), Volume II, Part 1—General hydrology. (English), Part 2—Aeronautical meteorology. (English–French–Spanish), Part 3—Marine meteorology. (English–French–Spanish), Part 4—Tropical meteorology. (English), Part 5—Hydrometeorology. (English), Part 6—Air chemistry and air pollution meteorology. (English–French–Spanish)

182 — International meteorological vocabulary. Second edition, 1992. (E/F/R/S)

385 — International glossary of hydrology. Published jointly by WMO and UNESCO; 2nd edition, 1992

407 — International cloud atlas. Volume I—Manual on the observation of clouds and other meteors. Reprinted in 1995. Volume II (plates), 1987.

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738 — Meteorological and hydrological risk assessment and disaster reduction—Lectures presented at the forty-first session of the WMO Executive Council (1989). 1991. (E/R)

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795 — Scientific lectures presented at the Eleventh World Meteorological Congress (1991). 1993

798 — Climate change issues—Lectures presented at the forty-fourth session of the WMO Executive Council (1992). 1994. (English)

805 — Lectures presented at the forty-fifth session of the WMO Executive Council (1993). 1994. (E/F)

822 — Lectures presented at the forty-sixth session of the WMO Executive Council (1994). 1995. (E/F)

845 — Lectures presented at the Twelfth World Meteorological Congress (1995). 1997. (English)

866 — Scientific lectures presented at the forty-eighth session of the WMO Executive Council (1996). 1997. (English)

910 — Lectures presented at the forty-ninth session of the WMO Executive Council (1997). 2000 (English)

911 — Lectures presented at the fiftieth session of the WMO Executive Council (1998), 2000. (English)

916 — Forecasting in the 21st Century. 2000. (English)

WORLD METEOROLOGICAL ORGANIZATION

INTRODUCTION TO CLIMATE CHANGE: LECTURE NOTES FOR METEOROLOGISTS

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NOTE

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FOREWORD

The atmosphere is the essential physical and chemical environment for life. Changes, anthropogenic or otherwise, to the physical and chemical properties of the atmosphere have the potential of affecting directly the quality of life and even the very existence of some forms of life.

Human-induced climate change, in particular, as well as other global environmental issues such as land degradation, loss of biological diversity and stratospheric ozone depletion, threatens our ability to meet very basic human needs, such as adequate food, water and energy, safe shelter and a healthy environment.

A great majority of the scientific community, whilst recognizing that scientific uncertainties exist, believe that human-induced climate change is inevitable. Further proof of the reality of climate change was made available through the work of the IPCC, established by WMO and UNEP in 1988. The recently released IPCC's Third Assessment Report forms the most comprehensive picture of the state of the climate and the global environment yet published, confirming that earlier judgements and projections of global mean temperature increases were underestimated. The IPCC also concludes that human activity is having a discernible effect on the environment, and that global temperatures are projected to increase at a rate unprecedented in the last thousand years.

A majority of experts believe that important reductions in net greenhouse gas emissions are technically feasible due to a wide range of technologies and policy measures in the energy supply, energy demand and agricultural and forestry sectors. Besides, the anticipated adverse effects of climate change on socio-economic and ecological systems can, to some degree, be reduced through proactive adaptation measures. Consequently discussions are taking place under the UNFCCC and the Kyoto Protocol seeking how to best cope with this issue, particularly in developing mitigation and adaptation strategies to prevent future generations from excessively negative impacts and to reduce the world vulnerability to these changes.

The meteorological community and, in particular, the Weather Services are now being increasingly solicited by the media, the general public and national and private institutions for information and guidance on climate issues. It is then essential that the technical and professional staff of weather services have the necessary background and knowledge of basic concepts and approaches of the issue of climate change in order to provide authoritative responses.

Professor David D. Houghton, from the Department of Atmospheric and Oceanic Sciences, University of Wisconsin – Madison, USA, kindly agreed, as part of his sabbatical leave, to prepare these lecture notes. They aim at enabling meteorologists, hydrologists and oceanographers to gain a more comprehensive understanding of the topics related to the issue of climate change.

I wish to express my sincere appreciation and gratitude to Professor Houghton for the high quality and comprehensiveness of his text in such a difficult and complex subject. I feel fully confident that these notes will be highly appreciated by meteorologists and related professionals interested in expanding their knowledge and understanding of the science of climate change.

(G.O.P. Obasi)
Secretary General

ACKNOWLEDGEMENTS

The cooperation and support of many persons and institutions were essential for producing this book. I am deeply indebted to the University of Wisconsin–Madison and its Department of Atmospheric and Oceanic Sciences for a sabbatical leave which gave me the time needed for the project and to the Education and Training Programme (ETRP) at the World Meteorological Organization (WMO) in Geneva for providing work facilities. I thank Dr Gustavo Necco, Director of WMO's Education and Training Department, for his support and guidance and Eliette Tarry, also from the Education and Training Department, for her handling of logistical arrangements.

I am very grateful for the comprehensive scientific reviews by Dr Hartmut Grassl, Director of WMO's World Climate Programme (WCP), and by Professor Igor Karol, Voeikov Main Geophysical Laboratory, St. Petersburg, Russian Federation. They provided essential calibration, constructive comments and additional information for the book.

I appreciated the warm welcomes and constructive conversations with many persons in the climate research groups at the Hadley Centre, Bracknell, United Kingdom, and the Max Planck Institute for Meteorology, Hamburg, Germany. In particular, I wish to thank Geoff Jenkins and Lennart Bengtsson for hosting me on these visits. Thanks are also due to Lev Karlin, who hosted my visit to the WMO Training Center at the Russian State Hydrometeorological Institute in St. Petersburg, and his assistant, Edward Podgaisky, and to Vilma Castro for hosting my visit to the WMO Training Center at the University of Costa Rica, San Jose, Costa Rica. These visits provided me with useful insights on the environments in which this book may be used as well as constructive comments on the science of climate change.

Many other persons contributed to this project. They included: Maurice Blackmon, Francis Bretherton, Jay Fein, Richard Hallgren, Ben Santer, Kevin Trenberth and Warren Washington, for their comments in the formative stages of the project; David Parker, Thomas Spence, Paul Try and William Rossow for information they provided on meteorological observations; Ulrich Cubasch for his insights on climate change education; Walter Fernandez for calling to my attention a recent book about the role of the sun in climate change; Steve Hammond for information on computers; and Linda Hedges and Jean Phillips for helping with clarifying references; and Pete Pokrandt for helping me to scan figures into electronic files. I also want to thank Heather McCullough for her work on the index.

Finally, I am very appreciative for the comprehensive editorial proofing of the entire book by Linda Keller, my son Eric Houghton, and my wife Barbara Houghton and all others who assisted in editing. Barbara provided support and constructive review throughout the project that were critical for successful completion of the project.

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10 July 2001

INTRODUCTION

Climate change and the need for environmental protection are global problems and call for a knowledgeable response from all countries in order to be effectively addressed. In recognition of this, the ten-year plan (1996–2005) of the Education and Training Programme (ETRP) of the World Meteorological Organization (WMO) places a high priority on enhancing a global system approach in its education and training work in member countries, particularly in developing countries. The plan states that personnel will need to have available an integrated training programme for understanding the ocean-land-atmosphere system, whether for monitoring purposes under the Global Atmosphere Watch (GAW), for learning about physical concepts and understanding patterns, or for prediction of tomorrow's weather and long-term climate changes.

Considerable attention has been given to climate change by the scientific community, government bodies and the public media. However, many issues are not fully understood. It is important that the professional operational community of meteorologists, hydrologists, and oceanographers become more knowledgeable on this subject in order to be able to respond to the needs to monitor climate change, to incorporate climate change perspectives into their own work, to help governing bodies to understand the scientific issues, and to provide information to the general public.

These lecture notes are intended to enhance familiarity with the broad scope of topics related to climate change. They provide material on the science of climate change assuming that the users already have a basic understanding of geophysical fluid dynamics, and relevant physical processes such as radiation transfer, diffusion, the hydrological cycle, and cloud physics along with some understanding of air chemistry, hydrology, and oceanography. Individuals needing advanced level material for climate change studies or research should refer to the basic references listed at the beginning of the reference section on page 115.

The term 'climate change' is used with different meanings and perspectives. In some cases it may refer to all environmental change or include natural variability. It is most useful to think of climate change as one of several symptoms of human-produced environmental change with both global and local perspectives. A global perspective is appropriate in recognition of the global interactions involving the component physical systems fundamental to climate change. The local perspective is essential because it is the local impacts which are of significance to individuals and communities, and because it is at the local level where measurements must be obtained from all parts of the world in order to properly describe climate and predict its changes.

With the current vigorous research programs and advances in observation technology and analysis it is clear that the specific assessments of climate change will be continually evolving. It is hoped that these lecture notes will give a background which remains relevant to understanding the advancing science of climate change.

There is a review of the characteristics and physical processes of the climate system in Chapter 1 followed by a discussion of its variability first from natural causes in Chapter 2 and then from human activity in Chapter 3. The description of and advances made with numerical climate models are presented in Chapter 4 followed by a focus on climate predictability in Chapter 5. Chapter 6 presents important requirements for observations needed to identify and understand climate change. Progress in the isolation and analysis of recent climate change is discussed in Chapter 7 followed by examples of climate change impacts in Chapter 8.

The primary seven references for these lecture notes are listed at the top of the list of references on page 115. They are labeled with numbers running from one to seven, and are cited in the text, particularly in tables and figures. Note that the book on long-term climate monitoring, the three reports of the Intergovernmental Panel on Climate Change (IPCC), and the climate system modelling book, numbered 2, 3, 4, 6 and 7, respectively, contain material from a very large number of contributors. A few other references are cited in the text. Figures and tables adapted from the source material contain numerous reference citations. A complete list of references is provided at the end of the text.

The third report of Working Group I of the IPCC which updates the scientific analysis and conclusions for climate change appeared just at this book was being finalized (July 2001). That report (cited as Reference no. 7 here) basically confirmed and strengthened the scientific conclusions from the previous reports used in this book. Some material was incorporated into this book; however, it was not considered necessary to include all of the details.

CHAPTER 1

UNDERSTANDING THE CLIMATE SYSTEM

1.1 DEFINITIONS

1.1.1 CLIMATE

Climate is generally defined as the average state of the atmosphere for a given time scale (hour, day, month, season, year, decade and so forth) and generally for a specified geographical region. The average-state statistics for a given time scale including all deviations from the mean are obtained from the ensemble of conditions recorded for many occurrences for the specified period of time. Thus the mean temperature for the month of May in Ankara, Turkey, is obtained from measurements considered representative for Ankara averaged over the month of May from a record of many years. Climate descriptors also include conditions at the Earth's surface such as ocean temperatures and snow cover.

The average-state description involves a wide range of variables depending on what is of interest. Temperature and precipitation are the most commonly used; however the list may include wind, cloudiness and sunshine, pressure, visibility, humidity and elements with noteworthy human impacts such as severe storms, excessively high and low temperatures, fog, snow and hail. The method of description focuses on statistical parameters, the mean and measures of variability in time such as the range, standard deviation, and autocorrelations.

It is important to identify the difference between weather and climate. Weather involves the description of the atmospheric condition at a single instant of time for a single occurrence. In general, climate may be thought of as an average of weather conditions over a period of time including the probability for distributions from this average.

1.1.2 CLIMATE SYSTEM

The climate system is defined as the five components in the geophysical system, the atmosphere and four others which directly interact with the atmosphere and which jointly determine the climate of the atmosphere. The five components are listed below:

(*a*) Atmosphere;
(*b*) Ocean;
(*c*) Land surface;
(*d*) Ice and snow surfaces (both land and ocean areas); and,
(*e*) Biosphere (both terrestrial and marine).

Figure 1.1 shows the scope of the climate system. Note that the two-way arrows in the diagram identify explicit interactions between the atmosphere and other components.

At this point it is appropriate to recognize that there are other factors, also variable in nature, which contribute to determining the climate. These are considered 'external' forcing factors and include the sun, Earth orbital parameters, land-ocean distribution, Earth topography (land and ocean), and basic composition of the atmosphere and ocean. These are important determiners of the climate which, except for the basic composition of the atmosphere and oceans, are not affected in return by the climate conditions.

1.1.3 CLIMATE CHANGE

Climate change in these lecture notes is defined as the change in climate attributed directly or indirectly to human activity which, in addition to natural climate variability, is observed over comparable time periods. The definition adopted by the United Nations Framework Convention on Climate Change (UNFCCC) focuses only on the human activity that alters the composition of the global atmosphere and excludes other human activity effects such as changes in the land surface. Sometimes the term 'climate change' is used to include all climate variability, which can lead to considerable confusion. Climate has variability on all time and space scales and will always be changing.

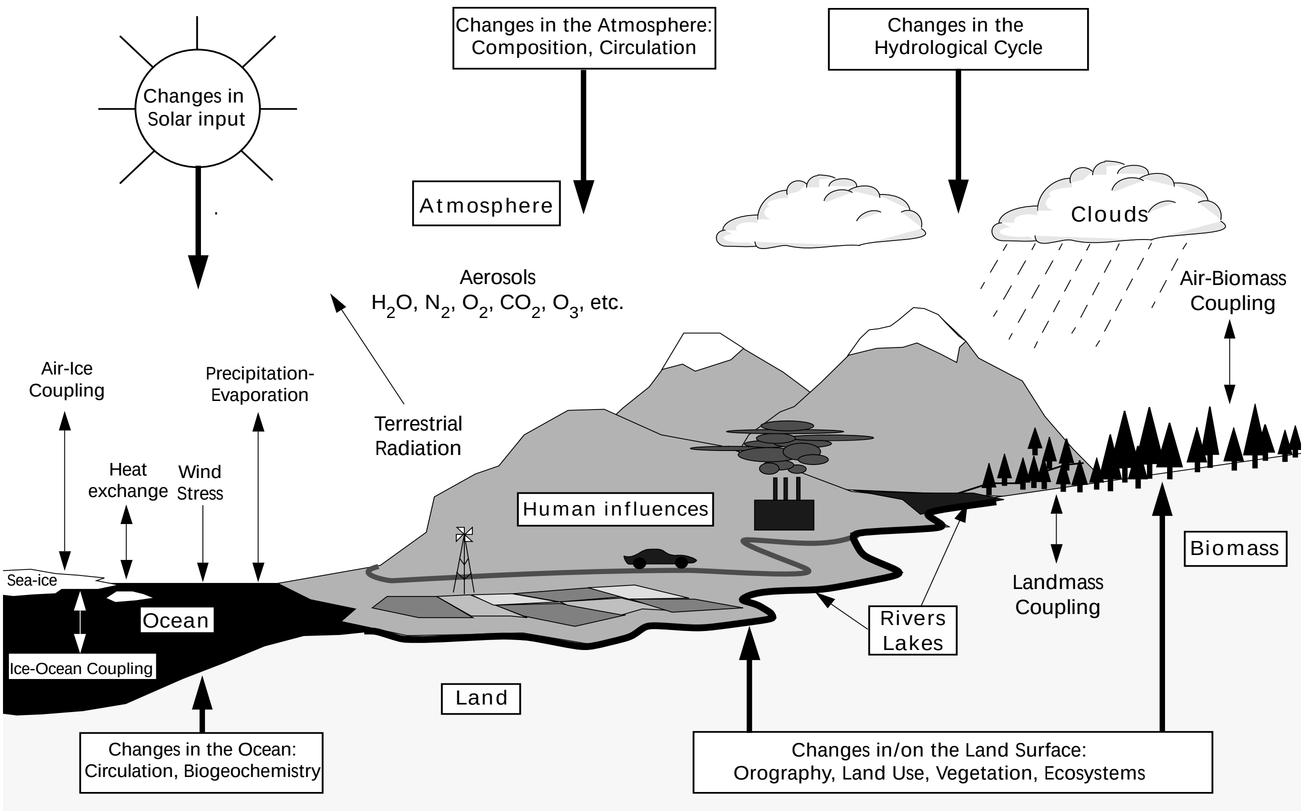


Figure 1.1 — Schematic view of the components of the global climate system (bold), their processes and interactions (thin arrows) and some aspects that may change (bold arrows). [from page 55, Reference no. 3].

1.2 GENERAL OVERVIEW

The definition for the climate system makes it clear that one has to have an understanding of all of that system's components (atmosphere, ocean, land surface processes, cryosphere, and biosphere) in order to understand it. In reality one needs to know a limited amount, dependent on the time scales considered, about the non-atmospheric components to understand the interactions of those components with the atmosphere. In general, these interactions occur primarily at physical interfaces so that, for example, for ocean interactions, it is necessary to know only the conditions at the oceanic upper boundary and for cryosphere interactions only at the surface of the ice. To know such conditions, of course, it is necessary to understand how they vary in relationship to conditions within the ocean and ice. Unlike the other interactive components, the ocean is an easily-movable fluid, as is the atmosphere, so that understanding the ocean for climate system applications requires dealing with geophysical fluid dynamic and thermodynamic relationships as complex as those for the atmosphere. Hence, it becomes necessary to use numerical model representation for the ocean comparable to that used for the atmosphere. Current climate system research depends strongly on coupled atmosphere-ocean numerical models.

This chapter focuses on the forcing and interaction processes of particular relevance to climate change. Basic material on topics such as large-scale geophysical fluid dynamics, synoptic-scale weather systems, turbulence, or the hydrological cycle is not covered here. Radiation processes play a key role in the climate change scenario and are discussed first in some detail followed by discussion of relevant aspects of the five climate system components. Examples demonstrating the global connections of the climate change processes are then presented, and finally, there is discussion of regional-scale aspects of climate variability and climate change.

1.3 RADIATION PROCESSES

1.3.1 INTRODUCTORY COMMENTS

Electromagnetic wave energy transfer (radiation) accounts for nearly all energy transfer from the sun, and is the primary source of energy for the atmosphere and the entire climate system. Such transfer is also the only way in which significant amounts of energy can leave the climate system. The energy of the global climate system is nearly in balance with incoming and outgoing radiation transfers. A change in one component will produce a different balanced state. The primary human impact on the energy balance is to alter the radiative properties of the atmosphere with respect to these two energy streams. This effect far out shadows other anthropogenic energy sources and sink effects such as the heating due to combustion and nuclear processes. Understanding the impacts of human environmental change on radiation transfer processes in the atmosphere and on the Earth's surface is crucial to understanding climate change. Because of this central role of radiation in climate change, a brief review of relevant radiation principles is given here even though it is assumed that the student already has a basic understanding of radiation.

Radiation principles cover the production (emission) of radiation energy from the internal energy (heat) of material substance and the transformation of radiation into the internal energy of material substance (absorption). These radiation principles also govern a number of processes that change the nature of the radiation, but do not convert its energy to internal energy of matter: reflection, refraction, diffraction and scattering. The production (emission) of radiation energy depends upon internal energy (temperature) as well as other properties of the emitting material substance. The destruction (absorption) of radiation depends on the amount of incident radiation energy and properties of the absorbing material substance except for its temperature. The properties of radiation depend on its wavelength. Radiation can exist for a wide and continuous range of wavelengths referred to as the radiation spectrum.

There are two primary forms of radiation relevant to the energy balance properties of the climate system. The first is the 'solar' or 'short-wave' form predominant in the radiation from the sun. This is primarily in the wavelength range from 0.2 to 4.0 microns (a micron is one-millionth of a meter) which encompasses the visible part of the spectrum. This short-wave radiation provides a source of energy for the climate system as it is absorbed in the atmosphere, clouds, ocean, land surface, and by living matter. The second form is the 'terrestrial' or 'long-wave' type predominant in the radiation emitted by matter in the climate system. The primary wavelength range for this form is from 4 to 60 microns which is entirely in the invisible infrared part of the spectrum. Sometimes the solar and terrestrial radiation forms are called 'visible' and 'invisible,' respectively.

The overall differences in predominant wavelengths are due to the differences in temperature of the emission sources for the radiation, about 6000 K for the sun and in the range of 190-330 K for the climate system components emitting terrestrial radiation. The terms 'short-wave' and 'long-wave' refer to the wavelengths of these radiation forms relative to each other and should not be confused with the same terminology used to describe the wavelengths of radiation used for communications (radio and television).

The relative roles of the two primary forms of radiation in the energy balance are complicated by the fact that the components of the climate system absorb as, well as emit, long-wave infrared radiation. This leads to a very complex description of the long-wave radiation energy transfer processes (Schwarzchild's equation) which is complicated to solve for the atmosphere situation.

Large-scale radiation effect variations in the climate system are most pronounced with respect to height and latitude. Horizontal variations in radiation transfer do exist at the small scale as demonstrated by the difference in solar heating at the Earth's surface on the two sides of a hill, one facing the sun and the other facing away from the sun. For general applications to the climate system, only the vertical component of radiation and is considered. Energy transfer magnitudes are discussed with reference to energy crossing horizontal surfaces.

1.3.2 RADIATIVE ENERGY BUDGET

1.3.2.1 Solar radiation

Solar radiation incident upon the Earth system coming into the atmosphere from above leads to heating as it is absorbed by gases, aerosols and clouds in the atmosphere, and by the ocean, land, ice and biosphere elements at the Earth's surface. The absorption is proportional to the intensity of the incident solar radiation and depends on the properties of these substances. As discussed above, the relevant intensity is the component in the vertical direction. As the solar radiation is absorbed, there is less radiation available to be absorbed at lower levels. Although the radiation is initially in a narrow beam traveling in one direction from the sun, reflection at surfaces and scattering within the atmosphere sends the solar radiation all directions. For this reason, when one looks outdoors in the daytime one sees light coming from all directions. The complete description of the absorption effects (heating) must include the cumulative effects from radiation propagating in from all directions.

The overall heating effects due to solar radiation absorption on a horizontal surface in the climate system relate to the intensity of the solar beam coming into the atmosphere from space and the angle of the solar beam to the local vertical. The intensity depends on the temperature of the sun and the distance from the sun to the Earth. The angle of the solar beam to the local vertical varies according to a number of astronomical factors: latitude on Earth (distance from the equator), longitude on Earth (time of day), and the orientation of the Earth's axis with respect to the sun (the solar declination angle) which varies according to the time of year. Variations in these factors lead to large variations in heating from day to night, from equatorial to polar regions, and from summer to winter. Figure 1.2 shows the variations in total daily solar radiation energy received at the top of the atmosphere as a function of latitude and time of year. Variations in the Earth's orbit and solar conditions result in additional longer-term variations to be discussed later.

The details of solar radiation absorption (heating) within the atmosphere and at the Earth's surface depend strongly on the properties of the absorbing substance. The albedo (reflectivity) of sunlight from the Earth's surface is indicative of (inversely related to) the absorption of radiation by that surface. A surface with a high albedo (high visible brightness) is heated much less than one with a low albedo (low visible brightness). At the Earth's surface, the albedo ranges from about five per cent for ocean surfaces (with the sun high in the sky) and the top surface of dark thick coniferous forests to 90 per cent for fresh snow. Thick clouds in the atmosphere can also have an albedo nearly as high as fresh snow. Since much of the reflected and back-scattered solar radiation travels back out to space, it is never converted to heat in the climate system. The atmosphere (gases, aerosols, and clouds) absorbs less of the incident radiation than the Earth surface, so that solar heating effects are greater at the Earth surface than in the atmosphere.

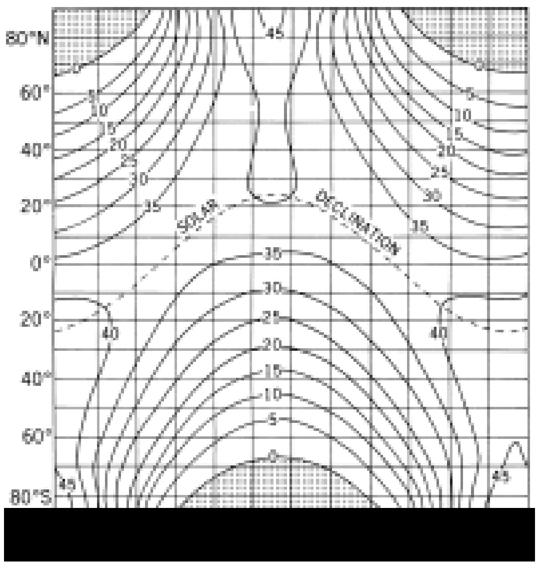


Figure 1.2 — Daily total of the solar radiation incident on a unit horizontal surface at the top of the atmosphere as a function of latitude and date in 10^6 J m^{-2} for one day (11.6 W m^{-2}). Shaded areas represent the areas that are not illuminated by the sun (from Wallace and Hobbs, 1977; after List, 1951). [from page 100, Reference no. 5, with permission of Springer-Verlag].

Terrestrial (long-wave) radiation is both emitted and absorbed by material substances in the climate system. The absorption depends on the incident radiation intensity and physical properties of the substances (except for temperature) whereas the emission depends on the temperature and other physical properties of the substances. The Earth's surface and clouds have radiative properties that tend to produce the maximum amount of terrestrial radiation given by 'black body' values and to absorb incident terrestrial radiation completely. On the other hand, the radiation emission and absorption characteristics of atmospheric gases have a large variability depending on wavelength, as shown in Figure 1.3. The strongest effects are exhibited by minor constituents in the atmosphere: water vapour, carbon dioxide, ozone, nitrous oxide, and methane. These gases occur naturally and are known as 'greenhouse gases.'

1.3.2.3 The 'greenhouse effect'

The 'greenhouse gas' radiative properties just noted are much more pronounced for terrestrial radiation than for solar radiation. Panel b in Figure 1.3 shows the magnitude of absorptivity for the entire depth of the atmosphere. Note that the absorptivity effects are much larger for terrestrial radiation (wavelength range 4 to 60 microns) than for the solar radiation (wavelength range 0.15 to 4 microns). The large absorptivity of the atmospheric gases for the terrestrial radiation together with atmospheric temperatures in the 210-310K range [optimal for terrestrial radiation emission as shown in Figure 1.3, Panel (a)] leads to emission of significant

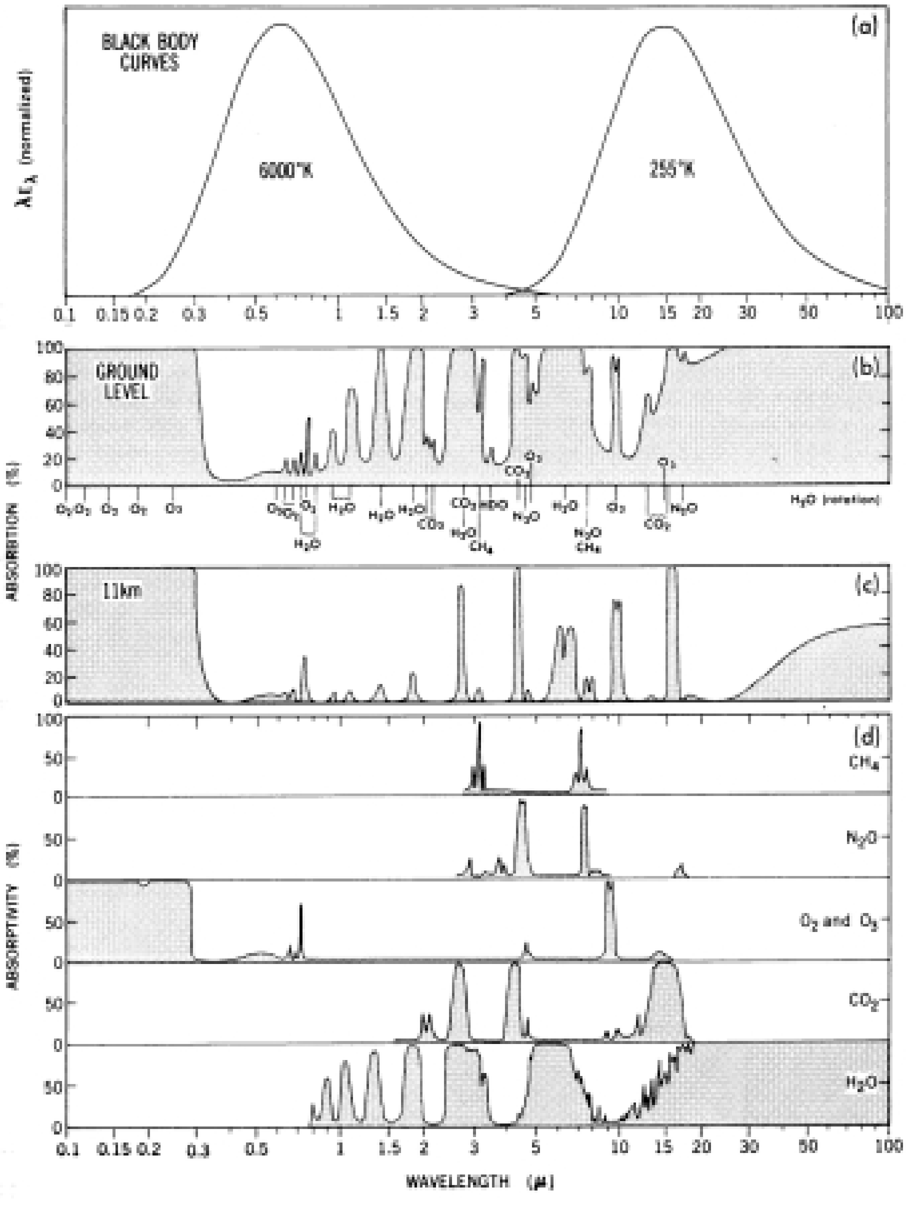


Figure 1.3 — (a) Black body curves for the solar radiation (assumed to have a temperature of 6000 K) and the terrestrial radiation (assumed to have a temperature of 255 K); (b) low resolution absorption spectra for the entire vertical extent of the atmosphere; and (c) for the portion of the atmosphere above 11 km after Goody (1964); and, (d) the absorption spectrum for the various atmospheric gases between the top of the atmosphere and the earth's surface after Howard et al. *[updated with data from Fels and Schwarzkopf (1988, personal communication) between 10 and 100 μm].* [adapted from page 93, Reference no. 3].

amounts of terrestrial radiation in all directions even where clouds are not present. [Students should recall the Kirchhoff and the Wien displacement radiation laws.] It is the downward component which retains energy in the climate system and keeps equilibrium temperatures at the Earth's surface and in the lower atmosphere higher than would otherwise be the case. This enhanced temperature is said to result from the 'greenhouse effect.'

On a globally-averaged basis the observed surface temperature is about 33K above the 255K expected with no atmosphere at all. This enhancement value would be even greater (over 80K) if radiative effects for the clear atmosphere were the only modifying factors [Manabe and Strickler, 1964]. Sensible and latent heat fluxes from the Earth's surface to the atmosphere along with atmospheric convection partially offset the surface temperature enhancement due to radiation transfer back from the atmosphere.

1.3.2.4 Role of radiation in the overall energy balance

When considering only the vertical transfers of energy, radiation energy transfers have a dominant role in the overall energy balance of the globally-averaged atmosphere and the Earth's surface. Figure 1.4 summarizes these energy transfers. The top of the diagram represents the top of the atmosphere, the bottom the Earth's surface, and the atmosphere is in the middle. The solar radiation components are on the left side. The terrestrial radiation components are on the right side, and the sensible and latent heat transfers are shown in the center.

Note that about 30 per cent of the incoming solar radiation is returned to space without being converted to heat (an albedo of about 30 per cent for the Earth–atmosphere system); about half is absorbed at the Earth surface, and only about 20 per cent is absorbed in the atmosphere. For the terrestrial radiation

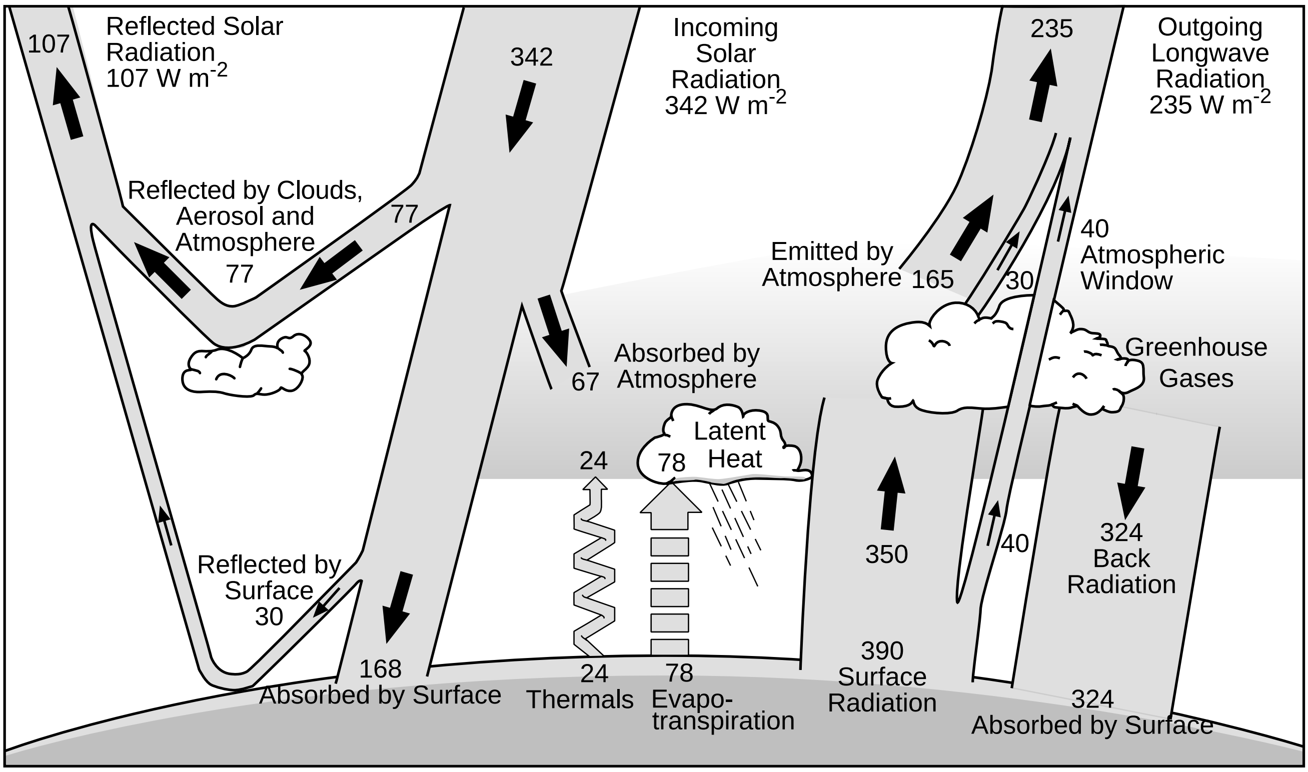


Figure 1.4 — The Earth's radiation and energy balance. The net average incoming solar radiation of 342 Wm^{-2} is partially reflected by clouds and the atmosphere, or at the surface, but 49 per cent is absorbed by the surface. Some of that heat is returned to the atmosphere as sensible heating and most as evapotranspiration that is realized as latent heat in precipitation. The rest is radiated as thermal infrared radiation and most of that is absorbed by the atmosphere which in turn emits radiation both up and down, producing a greenhouse effect, as most of the thermal radiation lost to space comes from cloud tops and parts of the atmosphere much colder than the surface. The partitioning of the annual global mean energy budget and the accuracy of the values are given in Kiehl and Trenberth (1997). [from page 58, Reference no. 3].

emitted from the Earth only about 10 per cent is transmitted directly to space; the remaining part is absorbed in the atmosphere.

The energy component labeled ‘back radiation’ is a key indicator of the greenhouse effect. Note also that the magnitude of terrestrial radiation emitted downward from the atmosphere to Earth and absorbed at the Earth’s surface is nearly equal to the total solar radiation incident at the top of the atmosphere and is about double the amount of solar radiation absorbed at the Earth’s surface. In general, the magnitudes of radiation energy transfer are considerably larger than those associated with the sensible and latent heat transfers (see Figure 1.4).

A basic aspect of radiation forcing is the systematic variation with latitude. As had been shown in Figure 1.2, there is generally an overall reduction with distance from the equator in the daily total solar radiation coming into the Earth–atmosphere system, being more extreme in the winter season and nearly absent at the time of the summer solstice. Seasonal and annual means for solar radiation absorbed in the Earth-atmosphere system show poleward decreases in both the summer and winter hemispheres (see Figure 1.5).

The net radiative forcing, of course, depends on both the input from solar radiation and losses due to terrestrial radiation emission to space. Latitude variations of terrestrial radiation emission are much less than for solar radiation (Figure 1.5). This terrestrial emission depends on the temperature (on the absolute Kelvin scale) both at the Earth’s surface, and in the atmosphere which has a smaller percentage variation than that for the zenith angle factor change with latitude which affects solar radiation absorption amounts.

The resulting net radiation forcing for the Earth–atmosphere system (see the last panel in Figure 1.5) has a net excess in the tropical latitudes and a deficit in the polar latitudes. If radiation transfer were the only process occuring, the equatorial regions would be hotter than observed and the polar regions colder than observed. However, the transport of heat from the equatorial to polar regions by atmospheric and oceanic circulations offsets this radiation imbalance and provides an overall energy balance at each latitude.

In conclusion and as stated before, the primary connection between human activity and climate change is the alteration of the radiation transfer characteristics of the atmosphere. The change in greenhouse gas concentration and the addition of other gases with similar characteristics will change the terrestrial radiation transfer. In addition, a change in aerosol concentration and perhaps related change in cloud cover will change the solar radiation transfer. Except on a very local scale, the energy transferred by radiation is far greater than any production rate of energy due to human activity.

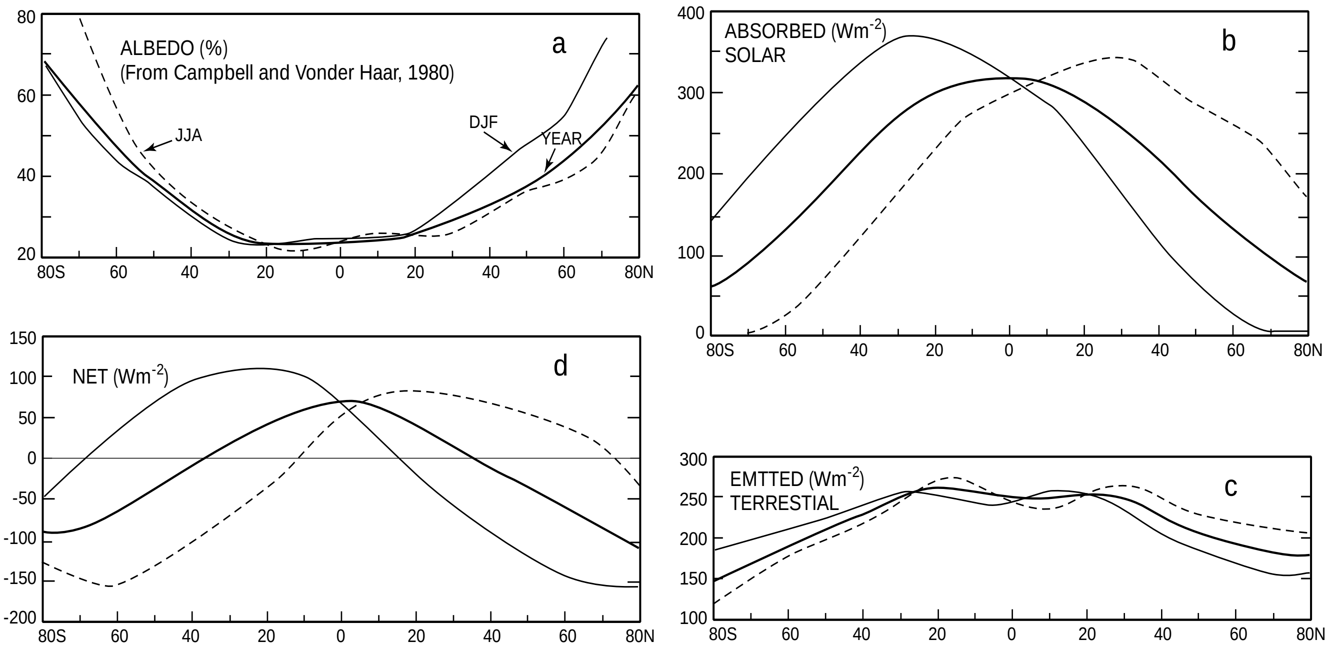


Figure 1.5 — Meridional profiles at the top of the atmosphere, for annual, DJF and JJA mean conditions, of: (a) the zonal-mean albedo; (b) absorbed solar radiation; (c) emitted terrestrial radiation; and (d) net radiation; (based on data from Campbell and Vonder Haar, 1980). No corrections were made for global radiation balance. [from page 128, Reference no. 5. With permission of Springer-Verlag].

1.4 CHARACTERISTICS OF CLIMATE SYSTEM COMPONENTS

1.4.1 INTRODUCTORY COMMENTS

A brief description of each of the five components of the climate system is presented. The information on the atmosphere is more extensive as climate is largely defined by conditions in the atmosphere. It is expected that students will already have background knowledge on the atmosphere, its circulations and physical processes. It is understood that students may have little or no background for the other four components. Therefore the material presented here deals only with aspects relevant to interactions with the atmosphere.

1.4.2 ATMOSPHERE

A number of factors make the atmosphere a very complex fluid system. As a gas, it has compressibility and great mobility. It extends well above topographic barriers and can sustain global-scale circulations. A number of forcing factors including radiative heating and cooling, latent heat sources and sinks due to phase change of water, and variations in Earth's surface temperature give rise to significant temperature variations in all three space dimensions and in time. These temperature variations give rise to horizontal pressure gradient forces, approximately consistent with the hydrostatic relationship, which are the basis for horizontal atmospheric motions. Temperature variations also affect vertical pressure gradient forces which, in instances of 'static instability,' can lead to large local-scale vertical motions. The large-scale horizontal motions are greatly modified by Coriolis effects arising from the rotation of the Earth and attain sufficient magnitudes to force smaller-scale transient motions. Fluid transport and nonlinear processes together with interactions among the physical factors listed above lead to a complex global general circulation system and embedded smaller-scale systems with space and time scales all the way down to atmospheric turbulence which we see in the gustiness of winds and the dispersion of smoke plumes.

The hydrological cycle is an important part of the atmospheric system. Evaporation and condensation of water can transfer considerable amounts of energy by both vertically and horizontally. The cloud component of the hydrological cycle strongly affects transfers of both solar and terrestrial radiation. The precipitation is the source of fresh water needed for life on land surfaces.

In the climate perspective, where statistics of atmospheric conditions are considered, the general circulation and associated temperature, cloudiness and precipitation patterns provide the primary bases for the mean climate conditions. Some of the transient features have systematic variations in time which are associated with the diurnal and annual cycles which are described directly in climate descriptions. Examples include average high and low temperatures for the day, monthly mean temperatures for each month of the year, and the annual range in monthly mean temperature. The random transient features such as extratropical cyclones, moist convection in the tropics and middle latitudes, and turbulence contribute to the climate descriptions for extremes and also for the mean states if appropriate correlations exist among the variables of the transient systems.

This last point deserves to be developed and illustrated. Let us, for example, take the relatively random transient atmospheric feature, the cumulus cloud. If the vertical circulations associated with cumulus convection have upward motions with systematically higher temperature than the downward motions, it would be expected that an ensemble of these weather systems would give a net upward transport of sensible heat. In the same way, if in association with extratropical cyclones, the winds from the south were typically warmer than winds from the north, an ensemble of these storm systems would give a net northward transport of sensible heat. In many situations randomly-occurring transient systems contribute significantly to the larger-scale conditions which determine the longer-term climate mean states.

It is assumed that the student is aware of the general climate conditions over the Earth. Many texts and atlases present climate maps, including the books referenced in the introduction. Nevertheless, a few climate charts are shown here to illustrate differences over the world as well as to present seasonal differences. It is important to remember that climate conditions vary considerably over the world, climate change would therefore affect people in different places of the world quite differently. Warming of summers in Canada may be welcomed as an enhancement of the growing season whereas warming of summers in the Sahara, where it is already hot, may be an entirely negative development.

Figures are shown for the climatology of surface temperature in January and July (Figure 1.6), precipitation in December-February and June-August (Figure 1.7), surface airflow (and pressure field) in December-February and June-August (Figure 1.8), and upper tropospheric airflow in December-February and June-August (Figure 1.9). The surface temperature and precipitation are key parameters for surface living conditions. The wind fields highlight atmospheric transport conditions at the surface where we live and in the upper troposphere where the primary maxima of kinetic energy exist in the circulation (the subtropical and polar jet streams). Both topography and ocean–land temperature differences result in wavy patterns in the horizontal mean flow particularly in the Northern Hemisphere.

The circulation in the atmosphere is sufficiently vigorous that material injected into one part of the atmosphere can be spread quickly over broad regions. It may take just days for a volcanic smoke plume or radioactive products to circle the Earth. Constituents with sufficiently long lifetimes, such as carbon dioxide, would be expected to have relatively uniform concentration throughout the atmosphere. The large-scale north-south circulations are less vigorous than those in the the east-west direction (see Figure 1.9). Vertical motions generally are much smaller than horizontal motions so the vertical transport of atmospheric constituents may be quite limited. This accentuates the build-up of atmospheric pollutants in the lower layers of the atmosphere especially in local areas with large sources of the pollutants.

The internal instabilities, feedbacks and nonlinear nature of the atmospheric system can result in circulation features that appear quite unrelated to the basic forcing due to energy fluxes at the Earth's surface and to radiation energy transfer. Examples of these include tropical cyclones and extratropical weather fronts. Furthermore, circulations may have more than one equilibrium for a given external forcing. Chaos theory deals with the variability characteristics in systems with such multiple equilibria. This characteristic of the atmosphere adds additional challenges and uncertainty to understanding and determining the outcomes for climate change scenarios.

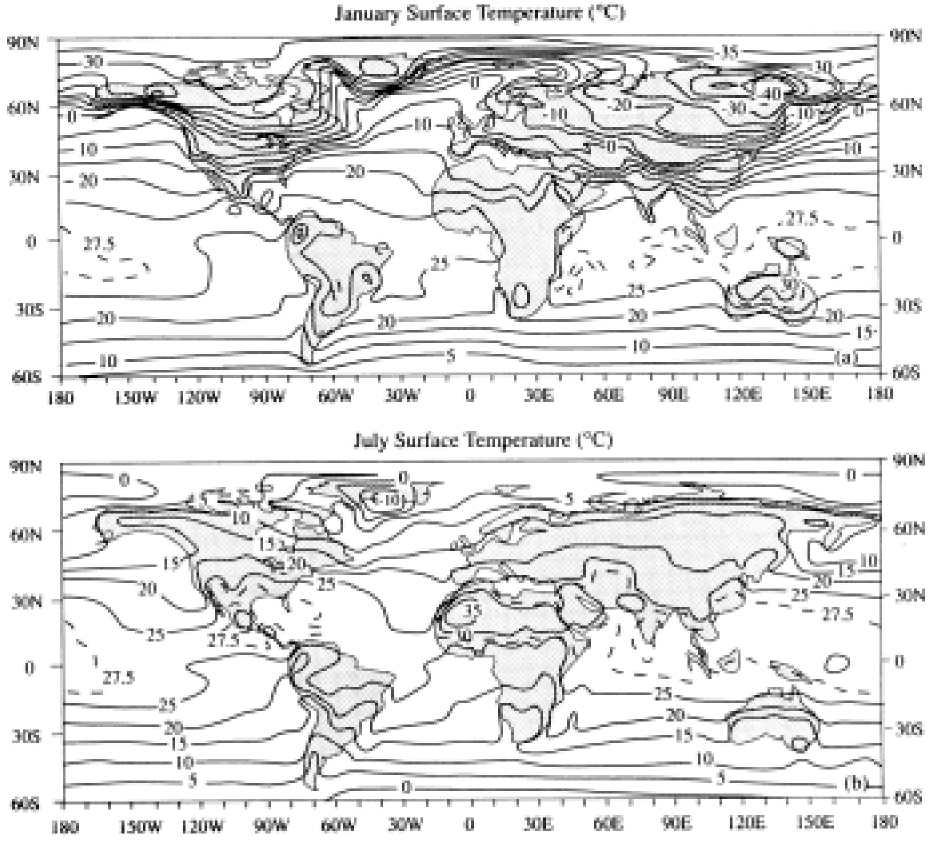


Figure 1.6 — Global map of (a) the January surface temperature January and (b) July surface temperature. (From Shea (1986), reproduced with permission from the National Center for Atmospheric Research). [from page 7, Reference no. 1, with permission of Academic Press].

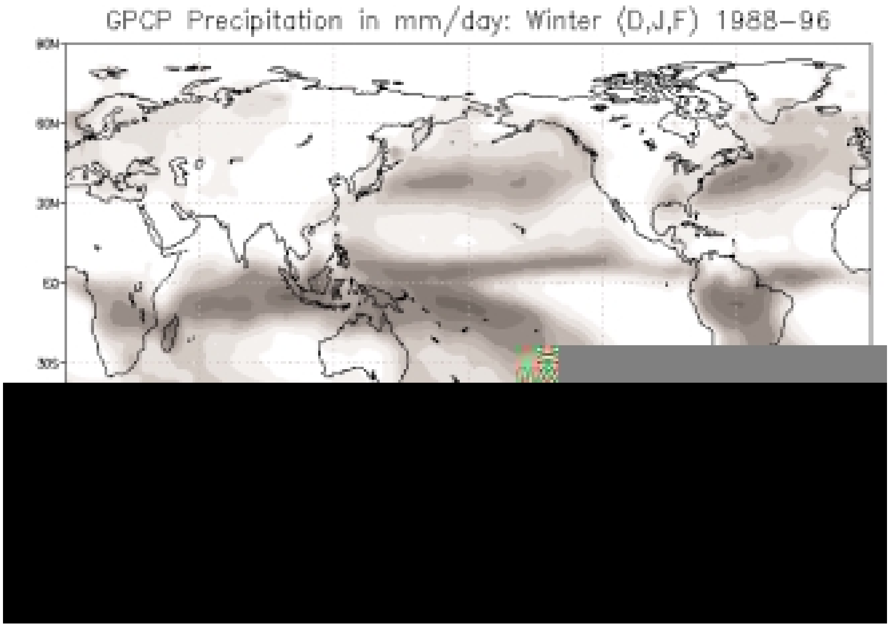


Figure 1.7 — Global distribution of average precipitation rate for December-January-February (D,J,F; left panel) and June-July-August (J,J,A; right panel) for 1988-1996
(from the Global Precipitation Climatology Project [GPCP] of the Global Energy and Water Cycle Experiment [GEWEX].

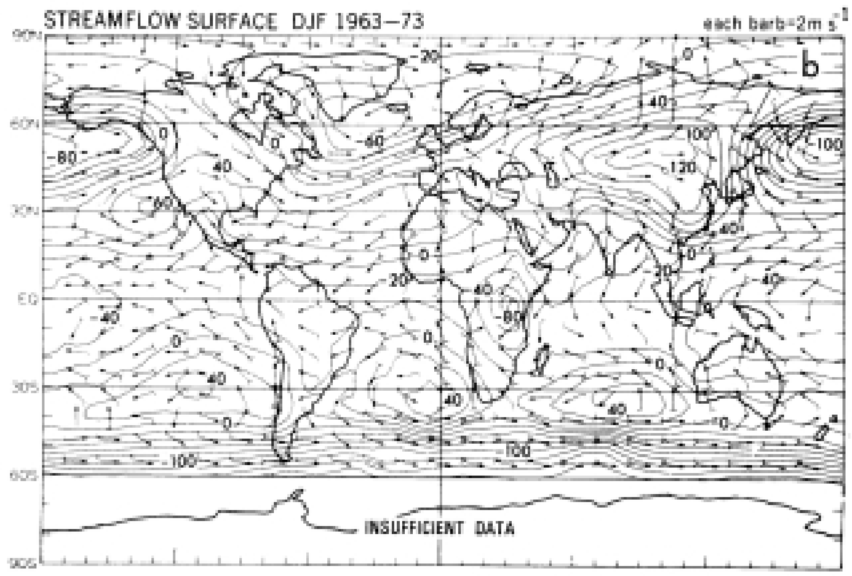


Figure 1.8 — Global distributions of the height anomalies of the 1000-mb pressure field in gpm from the standard atmosphere height, 113 gpm, and vector plots of the surface winds for northern winter (left panel) and northern summer (right panel) mean conditions. Each barb on the tail of an arrow represents a wind speed of 2 m s^{-1}. The isoheight lines can be interpreted as isobars for surface pressure reduced to sea level.
[adapted from page 134, Reference no. 5, with permission of Springer-Verlag].

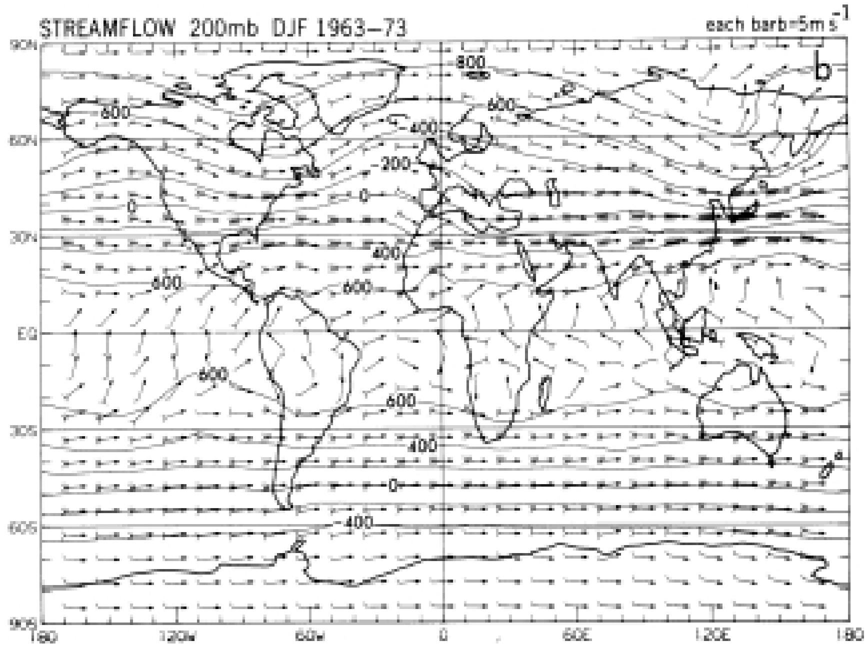
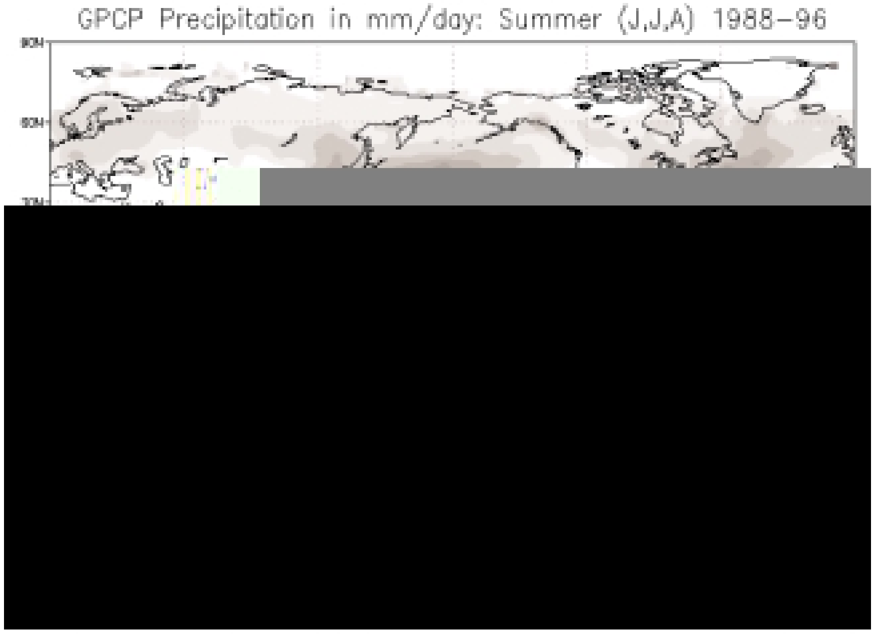
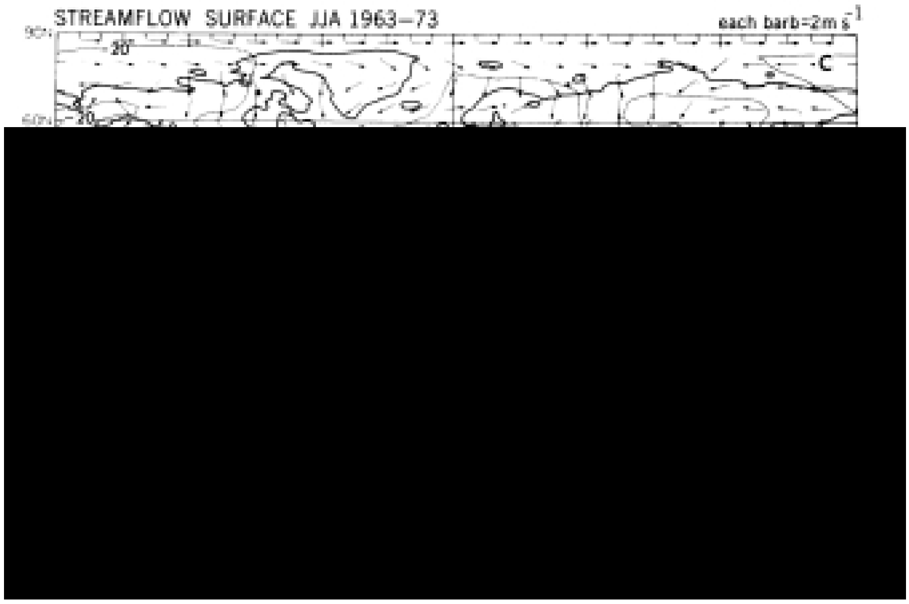
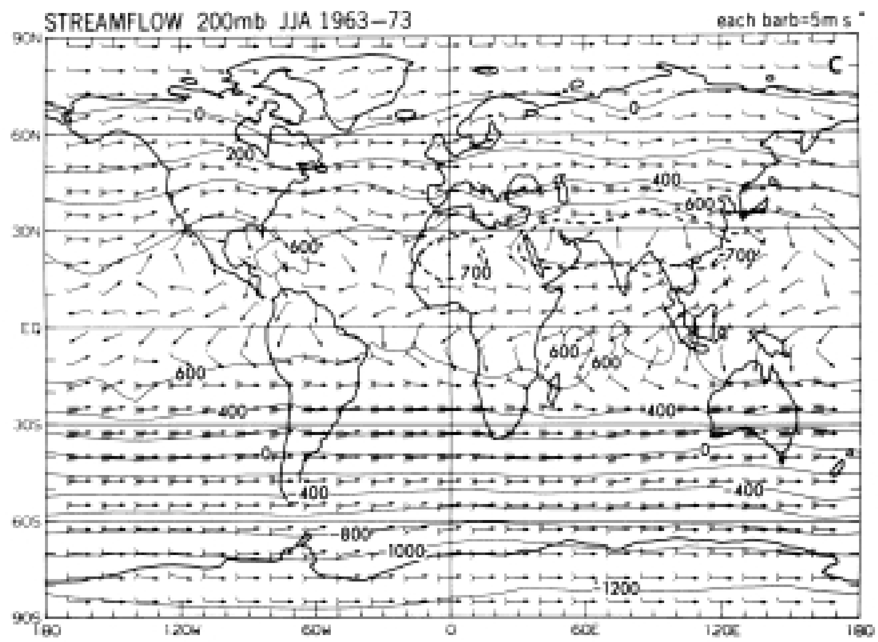


Figure 1.9 — Global distributions of the height difference of the 200-mb pressure field in gpm from 11,784 gpm and vector plots of the 200-mb winds for northern winter (left panel) and northern summer (right panel) mean conditions. Each barb on the tail of an arrow represents a wind speed of 5 m s^{-1}.
[adapted from pages 151 and 152, Reference no. 5, with permission of Springer-Verlag].


GPCP Precipitation in mm/day: Summer (J,J,A) 1988–96
90N
60N


STREAMFLOW SURFACE JJA 1963–73
each barb=2m s-1
90N
60N
20
C


STREAMFLOW 200mb JJA 1963–73
each barb=5m s
90N
60N
30N
EQ
30S
60S
90S
180
120W
60W
0
60E
120E
180
C
0
200
400
600
700
800
1000
1200

1.4.3 OCEAN

The ocean has a major impact on the climate of the atmosphere. It covers approximately 71 per cent of the Earth's surface and thus has a dominant role for transfers of energy and other properties between the atmosphere and the Earth's surface. Its large heat capacity, made accessible for surface energy transfers by circulations within the ocean, provides a moderating effect on temperature variability in the atmosphere. Oceanic currents transfer large amounts of heat energy away from equatorial regions. Finally, the ocean is an important source for atmospheric water vapour, as well as a source and sink for other greenhouse gases.

The ocean's heat capacity exceeds that of the atmosphere by a factor of the order of 1000. This is due to differences both in heat capacity per unit mass (the specific heat of liquid water is about four times that of air), and in total mass between the ocean and atmosphere. The ocean's heat capacity bears upon atmospheric temperature through oceanic transports (both horizontal and vertical) that produce and maintain surface water temperatures warmer or colder than the atmosphere resulting in large heat transfers. The depth to which the oceans interact with the atmosphere depends on the time scale under consideration. For diurnal variations the depth is small, of the order of five to 10 meters. For seasonal variations the depth is 20-200 meters (the depth of the well-mixed oceanic surface layer). The ocean is a major component in determining the climate and its variations for annual, annually-averaged and longer period conditions.

The strong influence of the ocean on surface air temperature is clearly evident. Figure 1.10 shows climatological sea-surface temperature conditions for January and July. Note that these temperatures are similar to those in the surface air over the oceans shown in Figure 1.6. Ocean temperatures are maximum in equatorial regions and decrease poleward. However, the poleward decrease in the

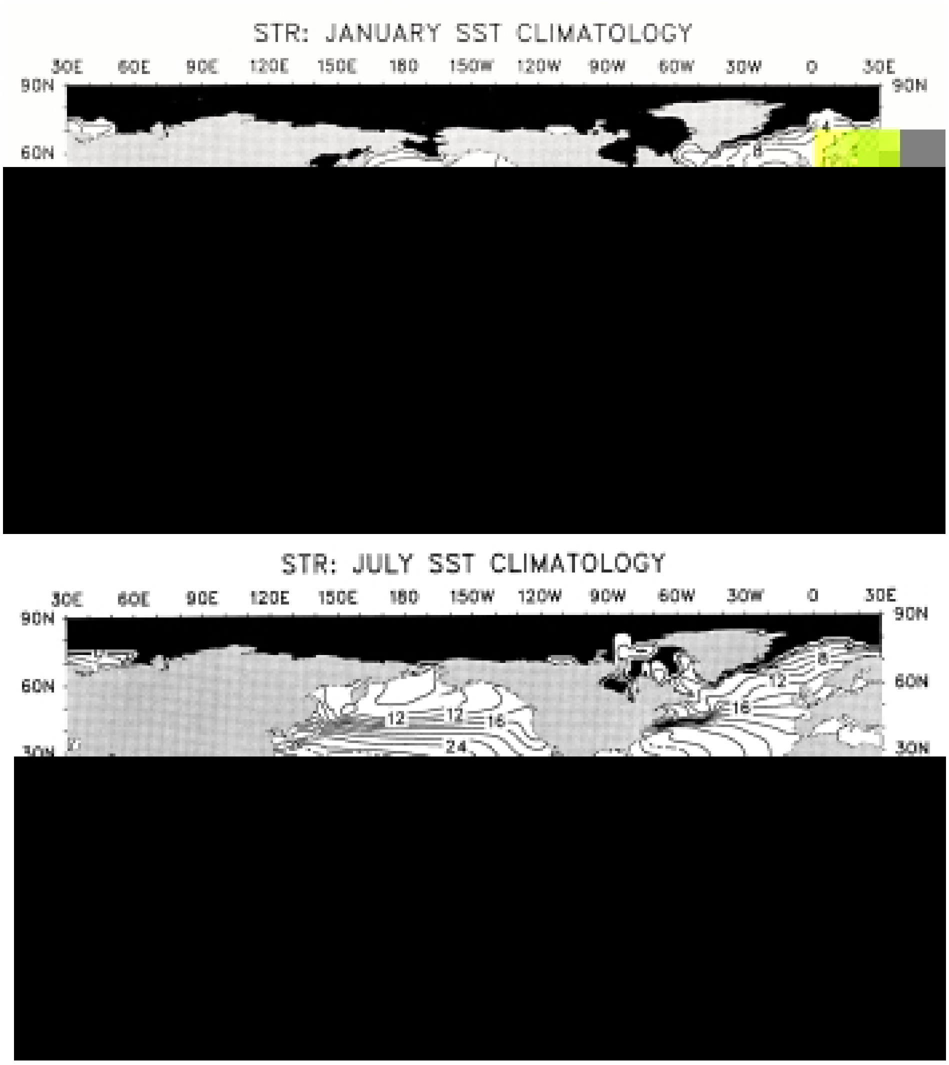


Figure 1.10 — Monthly mean Sea-Surface Temperatures (SSTs) for January and July. Dark areas indicate sea ice; stippling indicates land areas. The contour interval is 2°C except for dashed contours of 27 and 29°C. (from Shea *et al.*, 1990).

winter hemisphere is not as rapid as that experienced for land surface temperatures. The large ocean heat capacity results in summer-to-winter temperature differences over the ocean are generally much less than the summer-to-winter differences over land areas at comparable latitudes. Adjacent land areas downwind of the ocean, in fact, experience a moderation of winter cold temperatures and summer hot temperatures compared to other land areas at the same latitude. Note, for example, that in January at 50°N the mean surface air temperatures are well below freezing in eastern north America, well above freezing in the central Atlantic (as is the sea-surface temperature), and still above freezing in western Europe (see Figure 1.6).

Ocean currents have a major influence on sea-surface temperature. It is important to understand these currents to fully appreciate the ocean's interactions in the climate system. The laws of basic geophysical fluid dynamics apply as they do for atmospheric motions. As for the atmosphere, spatial variations in heating, primarily in the ocean surface layer, lead to horizontal pressure gradients which cause motion.

However, the oceanic condition is different from the atmosphere in two fundamental ways. First, the primary forcing of the ocean is at the upper boundary, whereas the primary forcing for the atmosphere is at its lower boundary. Atmospheric winds above the ocean are a major factor in causing ocean surface currents through surface friction processes. In contrast, for the atmosphere frictional conditions at its lower boundary tend to reduce atmospheric motion. Second, the density of the ocean water is determined primarily by its salinity and temperature instead of pressure, temperature and water vapour content as in the atmosphere. The water vapour factor is relatively unimportant for atmospheric density except in hot and humid conditions. On the other hand, salinity can play a major role for ocean density especially when temperatures are near freezing in which case density changes very little with temperature. Salinity conditions in polar ocean regions are important for determining whether or not significant vertical motions occur in local areas. The variations in ocean water density according to temperature and salinity are shown in Figure 1.11. The density value is shown as the difference from 1000 kg m^{-3}. Thus, for example, for a temperature of 10°C and a salinity of 16 parts per thousand, the seawater density is approximately 1016 kg m^{-3}.

The major surface-layer ocean currents are shown in Figure 1.12. Ocean currents result in significant equator-to-pole transport of thermal energy. The strong northward-moving currents off the east coasts of Asia and north America (the Kuroshio and Gulf Stream currents) take warm water away from the tropics. Currents towards the equator on the west side of continents (the California, Peru, and Benguela currents) transport cold water towards the equator.

Currents at deep ocean layers form primarily due to pressure gradients from density variations (thermohaline currents). The density variations are strongly influenced by fresh water sources from land surface runoff and sea-ice melt and

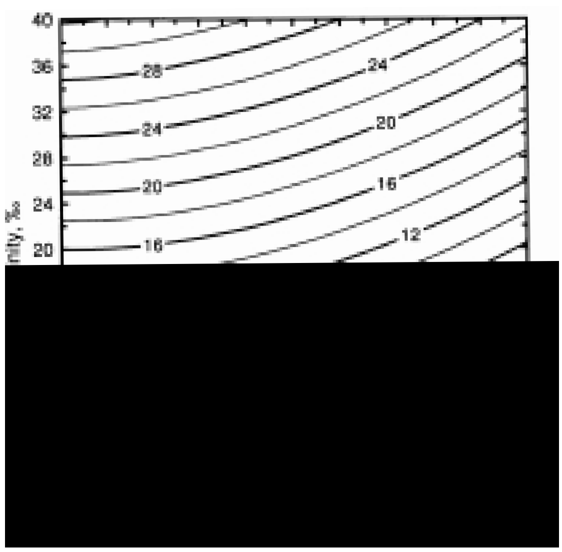


Figure 1.11 — Contours of seawater density anomalies (difference from a reference density of 1000 kg m^{-3}) in kg m^{-3} plotted against salinity and temperature. [from page 175, Reference no. 1 with permission of Academic Press].

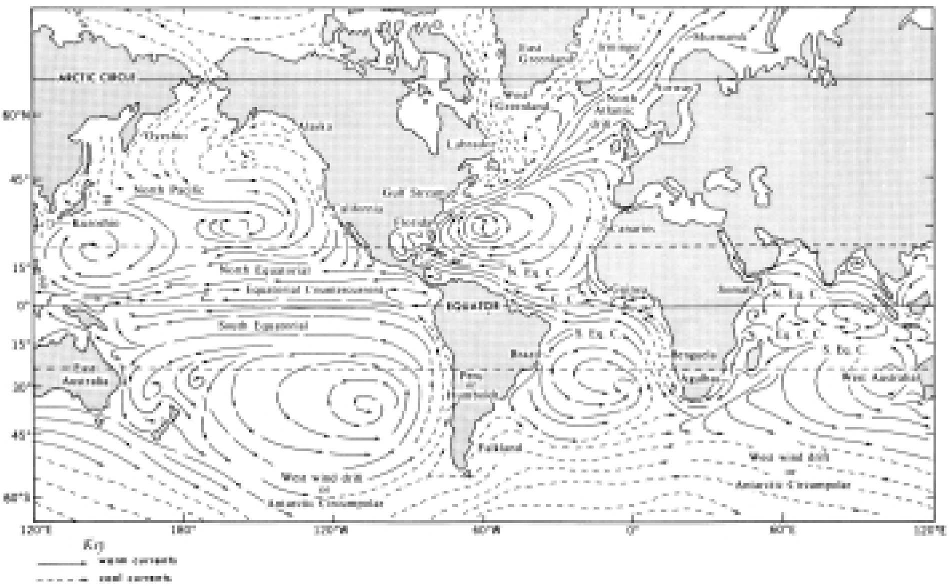


*Figure 1.12 — A map of the major surface currents in the world ocean during the northern winter (*from Tolmazin, 1985). [from page 177, Reference no. 5, with permission of Springer-Verlag].

fresh water evaporation from the sea surface. These currents provide a coupling between deep and surface ocean waters that involves a large portion of the ocean and provides ocean impacts on climate system variability on time scales of centuries, millennia and even longer.

The oceans play a significant role as a source and sink for atmospheric gases, including greenhouse gases. Changes in ocean temperature can change the holding capacity for gases and can result in a net outflow or intake from the atmosphere. Particularly noteworthy is the case for carbon dioxide. It is estimated that the carbon dioxide dissolved in the upper layers of the ocean is nearly 50 per cent more than the total amount in the atmosphere (1020 versus 750 gigatons of carbon content). See Figure 1.13. Thus, there is much potential for effects on the atmospheric carbon dioxide concentrations and the resulting radiation impacts due to changes in the ocean.

In conclusion, the ocean is a very important and interactive component in the climate system. The atmosphere forces oceanic motions through surface friction and affects oceanic temperature through surface sensible, latent, and

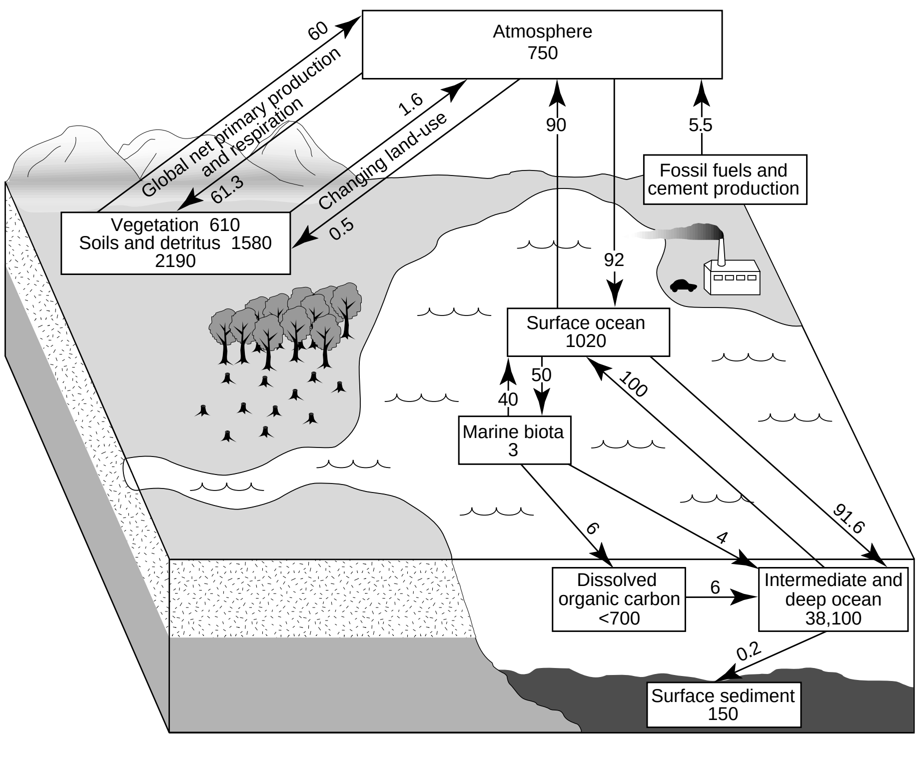


Figure 1.13 — The global carbon cycle, showing the reservoirs (in GtC) and fluxes (GtC/yr) relevant to the anthropogenic perturbation as annual averages over the period 1980 to 1989 (Eswaran et al., 1993; Potter et al.*, 1993, Siegenthaler and Sarmiento, 1993). The component cycles are simplified and subject to considerable uncertainty. In addition, this figure represents average values. The riverine flux, particularly the anthropogenic portion, is currently very poorly quantified and so is not shown here. Evidence is accumulating that many of the key fluxes can fluctuate significantly from year to year (terrestrial sinks and sources: INPE, 1992; Ciais* et al.*, 1995; export from the marine biota; Wong* et al.*, 1993). In contrast to the static view conveyed by figures such as this one, the carbon system is clearly dynamic and coupled to the climate system on seasonal, interannual and decadal time-scales (e.g. Schimel and Sulzman, 1995).* [from page 77, Reference no. 3].

radiative energy transfers at the ocean–atmosphere interface. The ocean affects atmospheric temperature by virtue of its large heat capacity which is enhanced by circulations that distribute its heat energy internally. As will be seen elsewhere in this chapter, it is also is a source and sink for atmospheric water vapour and other greenhouse gases. The ocean also has an important biosphere component.

1.4.4 LAND SURFACE

Land surface is an important interactive component of the climate system. It covers 29 per cent of the Earth's surface. Significant exchanges of heat, moisture, and momentum occur between the atmosphere and the land surface, including its biosphere. It is also the surface on which people live. The heat storage factor of land surface with respect to atmospheric temperature variations is much less than that for the oceans. Land has a lower specific heat than the ocean, and its rigidity restricts heat transport to deeper levels. As a result, the depth of the soil layer which is important for energy exchange interactions with the atmosphere is only several meters for the annual-cycle time scale. A cave 20 meters underground will remain at the same temperature all year round. Because of the small heat capacity of the land surface, variations in atmospheric temperature just above the surface are much larger over the land than over the ocean.

The energy and momentum exchanges between land surfaces and the atmosphere are similar to those for an ocean surface. Heat and latent heat (water vapour) exchanges depend on temperature and water vapour pressure differences between the land surface and the lower atmosphere, roughness of the land surface, and surface atmospheric wind speed. The latter may be characterized by wind conditions in the lowest ten meters of the atmosphere (the atmospheric 'mixed layer').

Radiation transfer is the other important energy exchange. The amount of solar radiation absorbed by a land surface depends on both the amount of solar radiation coming through the atmosphere (a highly variable quantity as discussed before) and the albedo (reflectivity) of the land surface which is also highly variable. The albedo ranges from five to 90 per cent and depends on the type of cover for the land surface as shown in Table 1.1. The infrared radiation transfer is the net of the infrared radiation emitted by the land surface (which is close to the maximum 'black body' value and thus dependent only on temperature) and the total downward infrared radiation produced by the atmosphere. Because of the small heat capacity of the land surface, the radiative, sensible, and latent energy transfers come close to balancing most of the time.

Topography of the land surface has a pronounced effect on large-scale atmospheric circulations. particularly in the Northern Hemisphere. The Rocky Mountains, which are oriented north-south transect the Northern Hemisphere westerlies, and the Tibetan Plateau with its extreme height and aerial extent affects flow over a large area. Topography is a factor in the wave patterns in the upper tropospheric horizontal wind flow (shown in Figure 1.9), and also has major effects on surface temperature and rainfall.

Alteration of land surface by human activity is an important factor in climate change that adds to the effects of human-produced changes in the radiative characteristics of the atmosphere.

Urbanization, cultivation for agriculture, irrigation, and deforestation change the albedo of land surfaces and the surface sensible and latent heat transfers. These factors can also greatly influence the local aspects of climate change.

1.4.5 CRYOSPHERE

The cryosphere — the ice component — has significant impacts on the climate system in several ways. It affects radiative and sensible heat transfers at the Earth's surface. It influences temperatures in the ocean and at the Earth's surface due to transfers between latent and sensible energy during melting and freezing. Finally, its melting and freezing influences water runoff from land and ocean salinity. Ice and snow exist primarily in the latitudes poleward of 30 degrees latitude and are thus are unfamiliar to the majority of the world's human population. Although only about two per cent of all the water on Earth is frozen,

Table 1.1 — Albedos for various surfaces. [from page 88 in Reference no. 1, with permission of Academic Press].

Surface type	*Range*	*Typical value (in per cent)*
Water		
Deep water; low wind, low altitude	5–10	7
Deep water; high wind, high altitude	10–20	12
Bare surfaces		
Moist dark soil, high humus	5–15	10
Moist gray soil	10–20	15
Dry soil, desert	20–35	30
Wet sand	20–30	25
Dry light sand	30–40	35
Asphalt pavement	5–10	7
Concrete pavement	15–35	20
Vegetation		
Short green vegetation	10–20	17
Dry vegetation	20–30	25
Coniferous forest	10–15	12
Deciduous forest	15–25	17
Snow and ice		
Forest with surface snowcover	20–35	25
Sea ice, no snowcover	25–40	30
Old, melting snow	35–65	50
Dry, cold snow	60–75	70
Fresh, dry snow	70–90	80

it covers an average of 11 per cent of the world's land surface and seven per cent of its oceans.

There are many constituents to the cryosphere: land ice in polar ice sheets, glaciers, permafrost, frozen ground, seasonal snow cover and sea ice. Table 1.2 summarizes the amounts of ice in the various categories. Note that although Antarctica and Greenland between them account for 98 per cent of the world's land ice, the total area covered by ice and snow can be much larger in the Northern Hemisphere winter. Figure 1.14 shows maps of ice coverage. The Northern Hemisphere has a much larger seasonal range than the Southern Hemisphere because of the larger amount of land area.

The albedos of ice and snow are higher than the albedo of the land or ocean surface that they cover (see Table 1.1). Thus, their seasonal variations in coverage will cause important seasonal variations in the Earth's surface albedo. The impact is less than might be first thought because in the winter season when coverage is maximum, the solar radiation is at a minimum and solar energy is less important in the atmospheric energy balance. Nevertheless, ice and snow introduce a process of positive feedback as the expansion of ice and snow coverage increases the albedo which, in turn, decreases solar heating. The resultant cooling acts to further enhance the ice and snow cover.

Snow and ice cover have strong insulation effects for sensible heat transfer which reduces heat transfer from the Earth, oceans, and lakes to the atmosphere. Over the ocean (and lakes), snow and ice cover effectively cut off the moderating effects of the water so that air temperature over the ocean ice cover can fall well below freezing point.

Cryosphere conditions may not be directly evident to most people of the world. However, because of the global nature of the climate system, the cryosphere component also influences lower latitudes.

*Table 1.2 — Estimated global inventory of land and sea ice**. [After Untersteiner, 1984 from page 15 in Reference no. 1, with permission of Academic Press].

* Not included in this table is the volume of water in the ground that annually freezes and thaws at the surface of permafrost ('active layer'), and in regions without permafrost but with subfreezing winter temperatures.

			Area (km^2)	*Volume (km^3)*	*Per cent of total ice mass*
Land ice	Antarctic ice sheet		13.9×10^6	30.1×10^6	89.3
	Greenland ice sheet		1.7×10^6	2.6×10^6	8.6
	Mountain glaciers		0.5×10^6	0.3×10^6	0.76
		Continous	8×10^6	(ice content)	
	Permafrost			$0.2–0.5 \times 10^6$	0.95
		Discontinous	17×10^6		
		Eurasia	30×10^6		
	Sea snow		$2-3 \times 10^3$		
	(average maximum)	America	17×10^6		
Sea ice		Max.	18×10^6	2×10^4	
	Southern Ocean				
		Min.	3×10^6	6×10^4	
		Max.	15×10^6	4×10^4	
	Arctic Ocean				
		Min.	8×10^6	2×10^4	

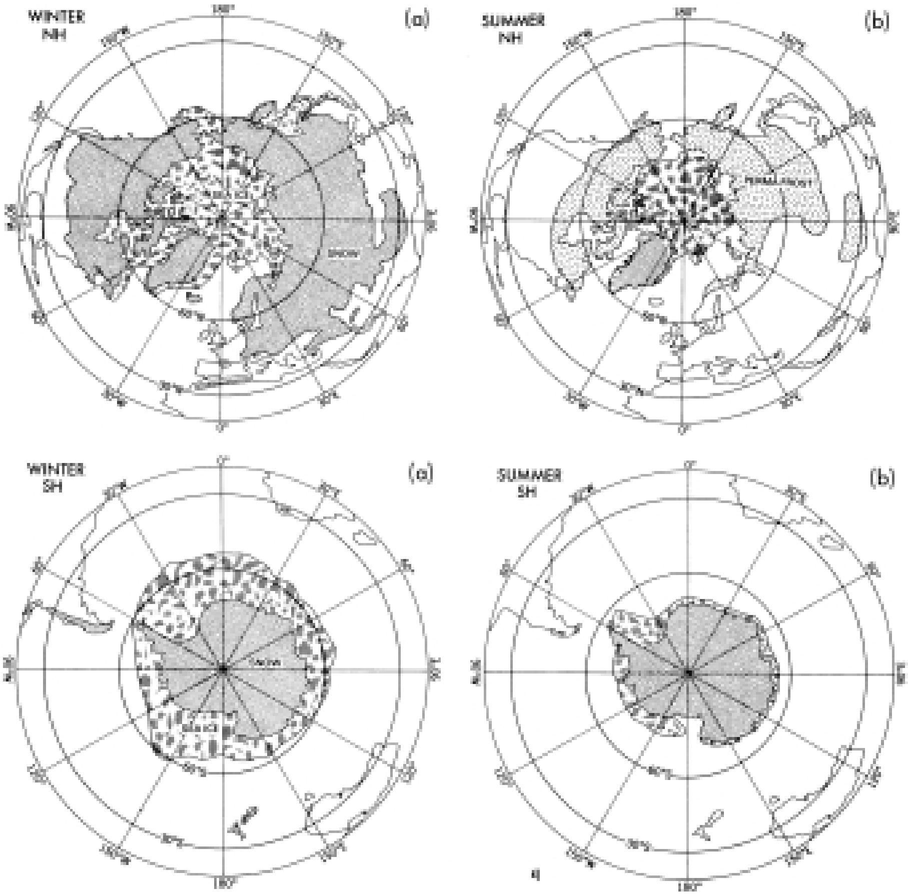


Figure 1.14 — Maximum extent of snow and sea ice during winter (a-panels) and maximum extent of snow and sea ice during summer (b-panels) in the Northern Hemisphere (upper panels) and Southern Hemisphere (lower panels) after Untersteiner (1984). Permafrost regions are also shown in the Northern Hemisphere summer. The ice limit was taken to be at concentrations 15 per cent in the Southern Hemisphere. [from pages 208 and 209 in Reference no. 5, with permission of Springer-Verlag].

1.4.6 BIOSPHERE

The biosphere is a component of the climate system that has a distinct role in the interactions of both the oceans and land surface with the atmosphere. Vegetation on the land surface and both plant and animal life in the oceans are all relevant elements of the biosphere component that interact with the atmosphere. Important exchanges between the terrestrial vegetation and the atmosphere are summarized in Figure 1.15.

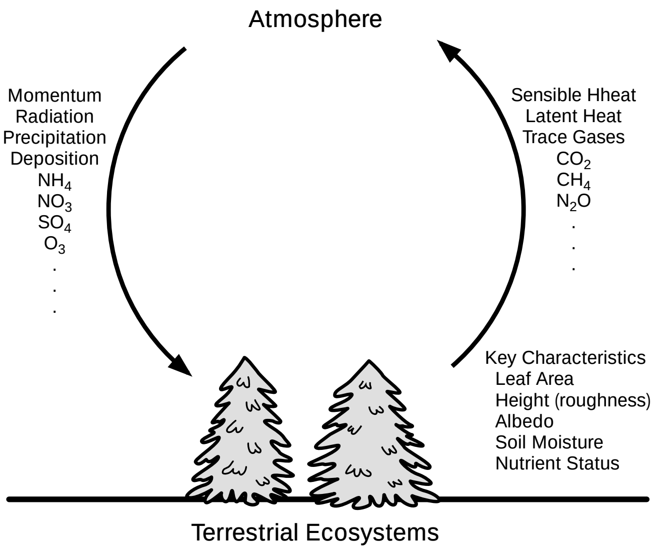


Figure 1.15 — Important exchanges between the atmosphere and terrestrial ecosystem. [from page 174, Reference no. 6, with permission of Cambridge University Press].

Climate conditions of the atmosphere have a direct effect on the type of terrestrial plant growth at the Earth's surface, as summarized in Figure 1.16. The nature of the plant cover in turn feeds back on the atmospheric condition by influencing the sensible and latent energy transfers from a land surface, as well as surface layer turbulence in the atmosphere (through its roughness properties). Furthermore, land vegetation is a significant reservoir for carbon with a total carbon content nearly equal to that in the atmosphere (Figure 1.13). Changes in the amount of land vegetation due, for instance, to forest cutting and burning or simply seasonal changes have a direct impact on the carbon dioxide concentration in the atmosphere. Along with dissolved inorganic carbon and calcium carbonate solids, plant and animal life have key roles in the ocean, in

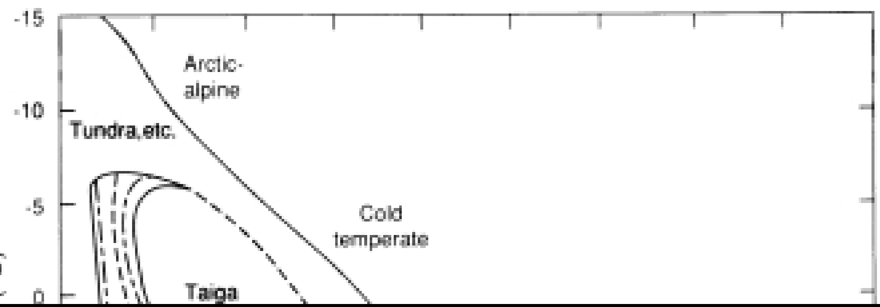


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Figure 1.16 — An example of a simple classification of vegetation types of the world based on annual precipitation and mean annual temperature (Whittaker, 1975). [from page 176, Reference no. 6, with permission of Cambridge University Press].

the carbon cycle which influences the concentration of the greenhouse gas, and carbon dioxide in the atmosphere and results in a loss of carbon due to sedimentation of carbonates at the ocean bottom.

The interactions between the biosphere and atmospheric climate have produced a record of past climate conditions. Tree rings, fossil patterns, pollen counts in ocean and lake bottom sediments, and coal and oil deposits are records which give us information on past climates.

Humans, themselves, are members of the biosphere. Humans alter the biosphere directly by agricultural and forestry activities and indirectly by altering the climate system in which the biosphere exists. It is important to understand these various impacts of human activity in order to understand climate change.

The biosphere must be included in the climate system analysis in order to understand climate change. It is a component that interacts with other climate system components, and is the component where the effects of climate change will be clearly evident to people.

1.5 FEEDBACKS IN THE CLIMATE SYSTEM

There are numerous significant physical interactions among the components of the climate system that are relevant to climate change. A brief overview of these components is presented below, they include: radiation energy transfer, heat energy transfer, and biosphere interactions.

1.5.1 RADIATION ENERGY TRANSFER

Radiation is a primary mechanism for energy transfer in the climate system. At the same time characteristics of the climate system itself have a great impact on the magnitudes of radiative energy transfer. There are two key feedbacks as described below.

1.5.1.1 Temperature feedback

The interrelationship between temperature and radiation provides a negative feedback whereby radiation transfer tends to reduce variations in temperature and to stabilize temperature conditions. This situation arises for two reasons. First, the magnitude of radiation emission from substance depends on the (absolute) temperature of the substance raised to the fourth power. Second, radiation emission represents a loss of energy from the substance causing its temperature to decrease. In the climate system the radiation involved is of the infrared (terrestrial) type. Thus increasing temperature will lead to increased radiative cooling.

1.5.1.2 Albedo feedback

A primary energy source for the climate system is the absorption of solar (visible) radiation. The amount absorbed is dependent on the reflectivity (albedo) properties of the substance. Since there is a large variation of albedo for substances in the climate system, significant feedbacks exist based on variations in amounts of specific substances. Important albedo feedbacks exist for ice cover, cloud cover, and land-surface characteristics. The feedback is positive for the ice cover because an increase in ice cover raises the overall albedo at the Earth's surface which tends to reduce surface temperature thereby making it possible for the ice cover to increase even more. An increase in cloud cover would also tend to cool the Earth's surface temperatures based on albedo effects since cloud albedo tends to be higher than that of the Earth's surface. However, clouds also affect infrared radiation transfer so that the net effect on the Earth's surface temperatures may be a warming or a cooling. Overall land surface albedo varies according to land use, the type of plants and ice cover. This provides an important albedo feedback related to human activity, climatic conditions and biosphere cycles.

1.5.2 HEAT ENERGY TRANSFER

Sensible and latent heat energy transfers provide for important energy-related transfers between the components of the climate system and involve important feedbacks. Vertical energy transfers between the ocean and atmosphere have already been mentioned. To this must be added the energy transfers between land and atmosphere and between ice and ocean water. The latent heat component arises from the phase change of water between its vapour, liquid, and solid forms. In many cases the transfer of latent heat energy can be as significant as that of sensible heat. Recall that in the global mean, the latent energy transfer from the Earth to the atmosphere was much larger than the sensible heat transfer (Figure 1.4). The dependence of the equilibrium saturation vapour pressure of water on temperature introduces a significant role of temperature into the latent heat energy feedback.

Because of their fluid nature, both the atmosphere and ocean transfer significant amounts of heat energy by horizontal motions. An important feedback between the atmosphere and ocean exists with regard to the latitudinal (poleward) energy transfer. If, for instance, the ocean poleward heat transport were to change due to internal conditions, there would be a change in the oceanic latitudinal temperature variations which would affect the latitudinal temperature variations in the atmosphere. This, in turn, would alter atmospheric circulations and poleward heat transports.

1.5.3 BIOSPHERE INTERACTIONS

Important two-way feedbacks between the atmosphere and biosphere were discussed earlier in Section 1.4.6. On one hand, the biosphere, as a central component in the carbon cycle, is a key determiner of greenhouse gas concentration in the atmosphere. On the other hand, the atmosphere, in particular its temperature and precipitation, has a major influence on the biosphere. The biosphere in both the atmosphere and oceans plays a significant role in the carbon cycle which includes the atmospheric carbon dioxide.

1.6 GLOBAL NATURE OF THE CLIMATE SYSTEM

1.6.1 INTRODUCTION

The circulations in the atmosphere and ocean transmit changes in one region of the climate system to broad sectors of the world. This means that many aspects of both natural climate variability and climate change are global in nature. This makes the climate change issue a global one which will require the understanding of people everywhere and the participation of all countries in dealing with its impacts. A few examples of this global scope are presented here.

1.6.2 OZONE HOLE IN THE STRATOSPHERE

The depletion of the stratospheric ozone layer in recent years has been shown to be due to chemical effects arising from the introduction of chlorofluorocarbons (CFCs) into the atmosphere. The CFCs were primarily used in the manufacture of refrigeration systems, in plastics blowing agents and in aerosol spray-can propellants. Sources of these gases may have been originally in the industrialized countries (primarily in the Northern Hemisphere); however, the primary effect has been seen in the reduction of the stratospheric ozone concentrations at polar latitudes, particularly in the Southern Hemisphere, with associated impacts on human life in Australia and southern parts of South America. There has also been significant stratospheric ozone reduction in the high latitudes of the Northern Hemisphere. The long lifetime of CFCs means that the impacts are felt for many years after the gases were released into the atmosphere. Thus, overall, the impacts of CFC emissions by human activity extend far, both in time and space, from their source points.

1.6.3 EL NIÑO — SOUTHERN OSCILLATION (ENSO)

The El Niño phenomenon in its original definition referred to warmer-than-normal temperature conditions on the ocean surface off the coast of Peru. In recent times the definition has been expanded to refer to warmer-than-normal conditions on and near the equator in the eastern half of the Pacific Ocean. Figure 1.17 shows the typical pattern for the sea-surface temperature in the eastern tropical Pacific during El Niño. The specific example is for the very strong El Niño in 1997-98. This condition ties into an atmospheric oscillation called the Southern Oscillation to cause anomalous conditions in both the ocean and atmosphere in the tropical Pacific and Indian Ocean area. Figure 1.18 shows the patterns of anomalies in surface atmospheric pressure of the Southern Oscillation which accompanies El Niño. Notice that pressure in the eastern tropical Pacific area is lower than the mean where the sea-surface temperatures are higher than the mean. Figure 1.18 actually shows correlations of surface pressure to that observed in Darwin, Australia, where the pressure is higher than the local mean value during El Niño.


[redacted]

Figure 1.17 — Anomalous sea-surface temperature for December 1997. Contour interval is 1°C. Dashed contours indicate negative anomalies. Anomalies are departures from the adjusted optimum interpolation climatology (Reynolds and Smith, 1995). [from page 24, Climate Prediction Center, 1997].

Observational data reveals 'teleconnections' (significant correlations) between ENSO conditions in the Pacific Ocean areas and variability in many other parts of the world. Correlations have been found not only in atmospheric conditions but also in ocean conditions such as for Indian Ocean sea-surface temperature anomalies as shown for the 1997-98 El Niño in Figure 1.17. Correlations are quite evident for weather conditions in the western coastal areas and southern region of the United States, the northeastern part of Brazil, and the eastern part of Asia as will be discussed in Chapter 2. Clearly the fluid-dynamic processes in the atmosphere and ocean do much to give global-scale perspective to what may appear as a regional phenomenon.

1.6.4 MONSOON

The monsoon refers to quasi-stationary circulation patterns and associated weather that exist on a seasonal basis due to surface temperature contrasts between continents and surrounding oceans. Monsoons cause large rainfall amounts over tropical and subtropical regions of Asia and Africa in summertime. Note, for example, the difference in precipitation amounts over south-east Asia between July (summer) and January (winter) as shown earlier in Figure 1.7. This regional rainfall pattern is a well-recognized aspect of a monsoon. However, further examination shows that the Asian monsoon, in particular, has impacts which extend over most of Asia, the Pacific Ocean and down into the Australian and Indian Ocean areas; truly a global scale. A schematic example of the low-level aspects of Asian monsoon circulation into these areas is shown in Figure 1.19. Variation in conditions in one part of the monsoon region may relate to conditions at quite distant locations. For instance, correlations have been found between monsoon rain intensities over south-east Asia and sea-surface temperatures in the eastern Pacific Ocean south of the equator.

1.6.5 VOLCANOES

The relatively recent major eruptions of El Chichon in Mexico in 1982 and Mount Pinatubo in the Philippines in 1991 serve as good examples of the global impacts of volcanic eruptions. In both cases the aerosols and gases from the eruptions spread around the world in the latitude bands of the volcanoes within a few

[redacted]

Figure 1.18 — Horizontal distribution of the correlation coefficient between annual-mean sea-level pressure anomalies over the globe and the corresponding pressure anomalies in Darwin, Australia (12°S, 131°E) as a measure of the Southern Oscillation. The map shows that global shifts of atmospheric mass take place during ENSO episodes (adapted from Trenberth and Shea, 1987). Areas with anomalies greater than 0.4 are stippled. [from page 422, Reference no. 5, with permission of Springer-Verlag].

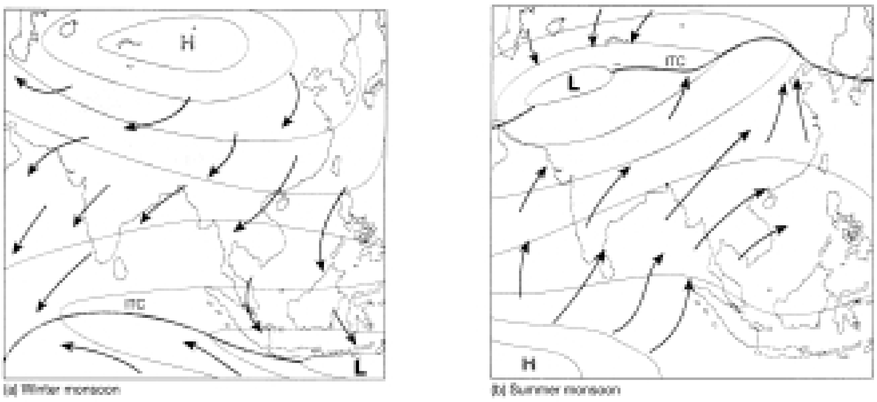


Figure 1.19 — Asia's monsoon circulation occurs in conjunction with the seasonal shift in the Inter-Tropical Convergence Zone (ITCZ) [Shown as ITC *in the diagram]. (a) In January, a strong high pressure develops over Asia and cool, dry continental air generates the dry winter monsoon. (b) With the onset of summer, the ITCZ migrates northward and draws warm moist air onto the continent.* [from page 179, Lutgens and Tarbuck, 1995, with permission of Pearson Education Publications].

weeks and, eventually over several months, to most other latitudes. In many parts of the world these aerosols resulted in unusual colorations in the morning and evening skies. The dust from Mount Pinatubo persisting at stratospheric levels resulted in a decrease of solar radiation at the Earth surface and a measurable decrease in global-mean surface temperature of several tenths of a degree Celsius for the following year in the Northern Hemisphere.

1.7 REGIONAL NATURE OF THE CLIMATE SYSTEM

1.7.1 INTRODUCTION

Climate has a global nature, but it also has local-scale variability which is very important for its impacts on life. Local variations are caused by topography, differences in ground cover at the Earth's surface and organization of weather systems resulting from the atmospheric general circulation in tropical regions, such as the Inter-Tropical Convergence Zone (ITCZ). Mean temperature and rainfall conditions change markedly with elevation of the land. Variations also exist upstream and downstream of topography with effects such as 'rain forests' or 'rain shadows' where the mean rainfall is greater or less, respectively, than in surrounding areas. At the very small scale, farmers know that the slope of the surface and small valleys in fields can noticeably alter growing conditions. Even the areas of shade and sun around one's home provide for very small-scale climate variations (microclimates) which affect plant growth. These local variations are significant for defining one's personal environment.

1.7.2 GEOGRAPHY OF CLIMATE

There are a wide range of climate conditions on Earth. It is useful to characterize them in terms of surface temperature and precipitation because these aspects relate directly to the biosphere and human activities. A classification system that has evolved from the original work of the Russian-born German climatologist, Wladimir Köppen, in the early 20th century is commonly used (Köppen, 1931). The original classification has five basic categories denoted by the first (upper case) letter in the classification code. Subcategories denoted by letters after the basic category letter provide details on seasonal variations in temperature and rainfall. Note that temperature criteria refer to monthly mean values. An update by Trewartha and Horn (1980) to this climatic classification is summarized below. Note that it has two additional basic categories.

A **(Tropical humid):** Warm enough and sufficient moisture (on an annual basis) for plant growth year round; essentially no winter with no frost in continental areas and mean temperature of coldest month 18°C or greater in maritime areas.

B **(Dry):** Plant growth limited by moisture supply alone (Steppe and Desert climates)

C **(Humid middle-latitude with long growing season):** Sufficient moisture (on an annual basis) for plant growth; monthly-mean temperature equal to or greater than 10°C for at least eight months.

D **(Humid middle-latitude with short growing season):** Sufficient moisture (on an annual basis) for plant growth; monthly mean temperature equal to or greater than 10°C for at least four months, but less than eight months.

E **(Boreal subarctic):** Very short summer with monthly mean temperature equal to or greater than 10°C for only one to three months.

F **(Polar):** Essentially no growing season; for the warmest month the mean-monthly temperature is less than 10°C in the Tundra climate and less than 0°C in the Ice Cap climate.

H **(Highland):** Not a type of climate; high elevations are an important factor in climatic conditions

Figure 1.20 shows the overall climate characteristics for the land areas of the world in terms of the classification summarized above together with subcategories. The key in the diagram describes the subcategories.

1.7.3 LOCAL VARIATIONS

1.7.3.1 Rainfall

Mean rainfall distributions in tropical areas can have large variability over small distances. The rainfall in Africa is one example. Mean annual and seasonal rainfall in the Sahel area of Africa can change rapidly over small distances in the north-south direction due to the quasi-stationary and small-scale structure of the ITCZ. Changes in annual rainfall amounts are as large as from roughly 1000 mm at 10°N to 50 mm at 17°N (a distance of roughly 700 km). Similar variations occur in the tropical region of South America. The largest variations are between the western-most portion of the Amazon River basin and the west coast of South America due to the Andes Mountains.

As an example for subtropical areas the mean annual rainfall in the India-Pakistan area has large variations due to the topographic influences of the Tibetan mountains and lower mountain ranges and from the structure and persistence of the monsoon circulation. Rainfall values are as low as 200 mm in Pakistan and reach 1000 mm in the western part of India only 400 km to the south-east. Three hundred kilometers further down the west coast of India, rainfall values reach 3000 mm.

Ocean islands, especially those in the trade-wind areas, can have large variations in mean annual rainfall from one side to another depending on topography. The Hawaiian Islands are a good example.

1.7.3.2 Temperature

Temperature variations due to altitude or proximity to oceans can be large over short distances. The nature of agricultural crops in tropical countries can change rapidly with elevation in mountainous areas. This is illustrated in countries such as Honduras and regions such as eastern Africa. Mean wintertime temperatures in Juneau, Alaska are 20°C warmer than places 100 km inland.

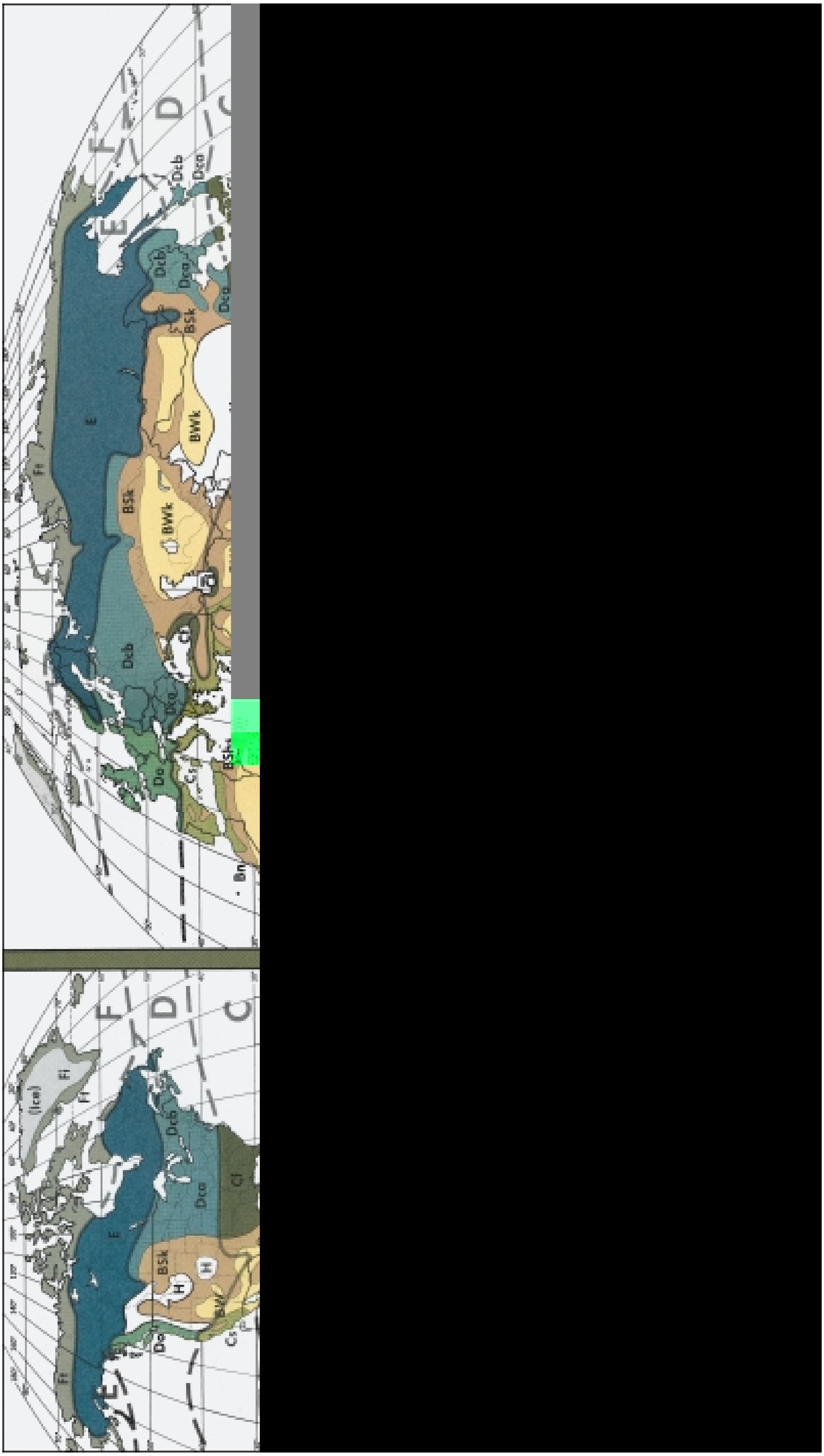


Figure 1.20 — Climate of the earth [from G.T. Trewartha and L.H. Horn, 1980, inside cover, with permission from McGraw-Hill).

CHAPTER 2

NATURAL TEMPORAL VARIABILITY IN THE CLIMATE SYSTEM

2.1 INTRODUCTION

The Earth's climate exhibits natural variability on all time scales. Some of this, for instance, that has occurred in surface temperature is much larger than anything envisioned due to human impacts. It is a continuing challenge to demonstrate variations that are due to anthropogenic causes rather than natural causes.

Variability magnitudes differ widely over the range of time scales. There are a number of time scales for which the variability is markedly larger than it is on slightly larger or smaller time scales. Many of these local peaks in magnitude can be ascribed to identifiable forcing processes. An idealized variance spectrum for the Earth's surface temperature is presented in Figure 2.1 with identification of the time scales of the maxima, where time scale is measured by the period of oscillation. The daily and annual cycles can be clearly seen. Many of the narrow peaks relate to other astronomical and geological effects such as variations in the Earth's orbit, continental drift and mountain formation (forcing mechanisms 'external' to the climate system). Some of these are discussed in the next section. The broader peaks relate to variability enhancements that include the effects of interactions within the climate system (internal forcing mechanisms). The temperature spectrum, as for other variables in the climate system, is 'red' meaning that amplitudes of the variations are larger for the longer time-scales.

The natural variabilities in the range of seasonal to millennial (three months to thousands of years) may be considered as the most relevant in our discussion about climate change. However, even the diurnal temperature variability is an important factor in climate change as it is affected by changes in greenhouse gas concentration.

In this chapter basic forcing mechanisms for climate variability are discussed, covering both external and internal types. Specific examples of variability are then presented.

2.2 BASIC FORCING MECHANISMS

2.2.1 EXTERNAL FORCING

2.2.1.1 Astronomical effects

2.2.1.1.1 Variations in solar radiation emission

The sunspot cycle is a well-defined variation in the solar condition with a period of about 11 years. The three-hundred year record shown in Figure 2.2 shows the regularity in the periodicity of the variation. The amplitude of the peaks varies by a factor of two over the record and the periodicity itself ranges between 10

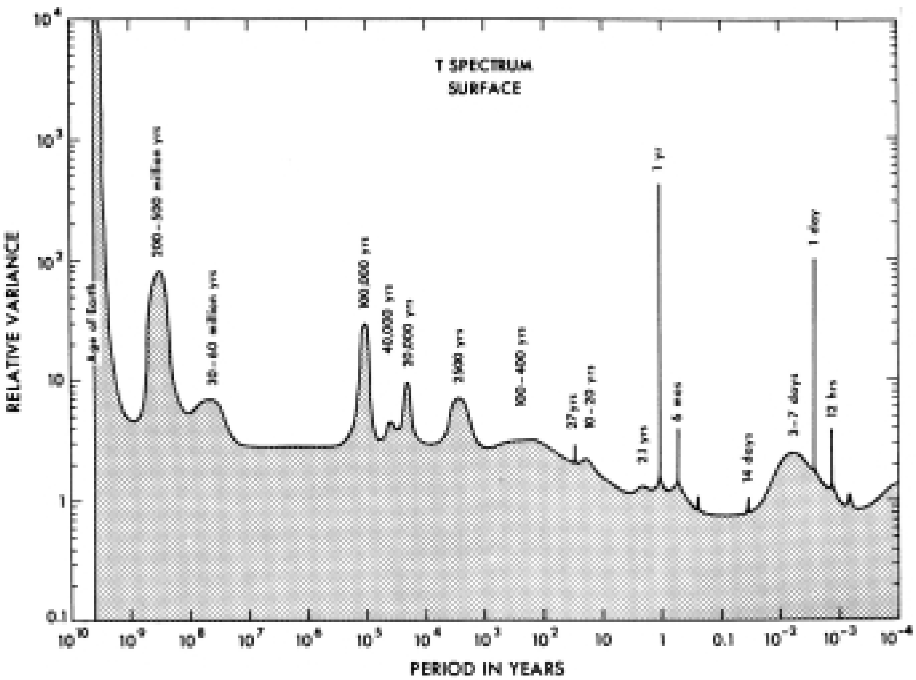
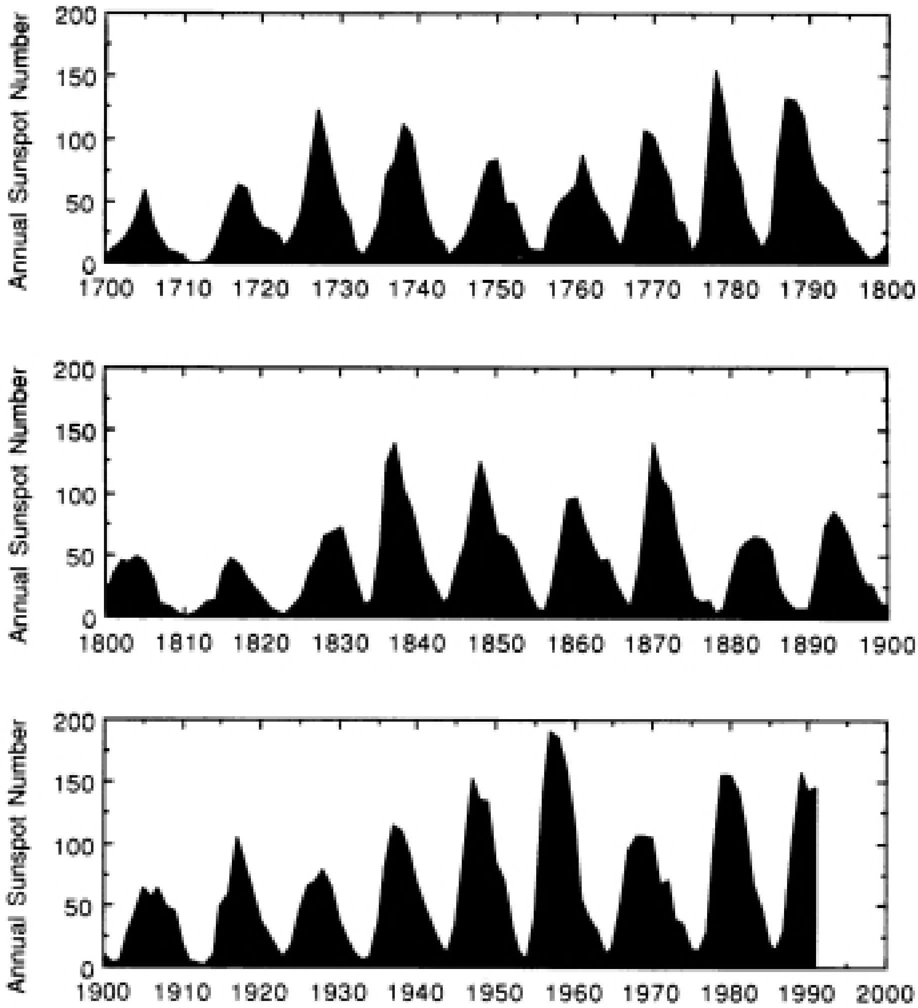


Figure 2.1 — Idealized, schematic spectrum of atmospheric temperature between 10^{-4} and 10^{10} yr adapted from Mitchell (1976). [from page 25, Reference no. 5, with permission of Springer-Verlag].

Figure 2.2 — Annual mean sunspot numbers from 1700 to 1991. [from page 288, Reference no. 1, with permission of Academic Press].



and 12 years. Although the presence of a sunspot itself (a relatively darkened area on the sun's surface) causes a reduction in solar radiation output, there are extra bright regions called faculae, that are found in conjunction with sunspots, that lead to an overall increase of solar radiation.

There is a direct relationship between the number of sunspots and the increase in solar radiation. At the top of the atmosphere the difference in the solar constant between the sunspot minimum and maximum is of the order of 1.5 Wm^{-2}. This difference results in an average change of solar radiation absorbed at the Earth surface of about 0.2 Wm^{-2}. This number is small, but not negligible, in comparison with the 2.45 Wm^{-2} estimated as the overall change in greenhouse gas forcing due to human-produced increases since the pre-industrial era.

The variation in the sunspot periodicity itself has also been shown to relate to Earth surface mean temperatures, with shorter periods corresponding to warmer temperatures. An overall review of solar effects on atmospheric climate is presented by Hoyt and Schatten (1997).

2.2.1.1.2 Diurnal and annual cycles of solar radiation input

The diurnal and annual cycles are both large amplitude variations. The rotation of the Earth around its axis leads to the well-pronounced diurnal cycle which is strongest in equatorial regions and does not exist at all at the poles where the sun remains low in the sky (or just below the horizon) for the entire day.

The annual cycle of solar energy input is primarily caused by the tilt of the Earth's axis with respect to the plane of the Earth's revolution around the sun. This tilt is at present about 23.5° to the normal of the plane and leads to seasonal variations in solar sunbeam zenith angle for all parts of the globe as well as variations in the length of the daylight period. Since the Earth's orbit around the sun is elliptical rather than circular, the variation in distance from the sun causes

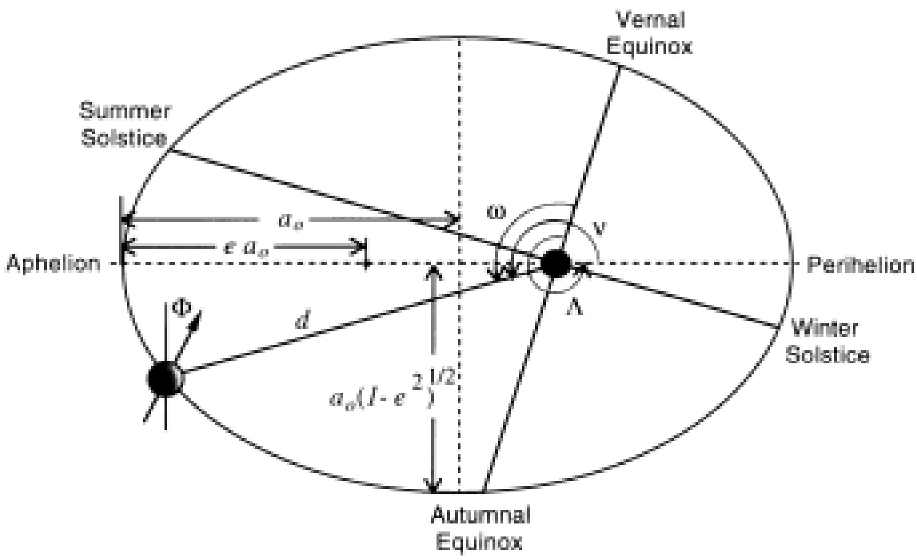


Figure 2.3 — Schematic diagram of the Earth's elliptical orbit about the sun showing the critical parameters of eccentricity (e), obliquity (Φ), and longitude of perihelion (Λ) defined relative to the vernal equinox. The size of the orbit is defined by the greatest distance between the ellipse and its center point, which is called the semi-major axis length, a_o. The Earth–sun distance at any time (d), the angle between the position of Earth and perihelion that we call the true anomaly (v), and the angle between the position of Earth and the vernal equinox (ω) are also shown. [from page 303, Reference no. 1, with permission of Academic Press].

additional fluctuations in the amount of solar radiation received on Earth. This distance factor produces a variation of 6 per cent in solar radiation intensity between the July 4 minimum and the January 3 maximum, an annual variation that is largely masked by the axis-tilt effects for daily total solar radiation input.

The Earth is currently closest to the sun in the Northern Hemisphere in winter and furthest in summer. Thus, in the Northern Hemisphere this distance variation tends to offset the seasonal variability in solar radiation input. In about 11 000 years the sun will be the closest in the Northern Hemisphere summer and furthest away in the winter tending to make the Northern Hemisphere summers hotter and winters colder than they are at present.

The annual cycle of daily total solar energy input varies considerably with latitude as shown in Figure 1.2 earlier. The range is most extreme at the poles where there are six months with no energy input. At the equator the range of the annual cycle is very small and a small semi-annual variability is observed. The seasonal shift of the latitude where the sun shines straight down at noon (shown by the dashed line in Figure 1.2) results in significant shifts in the Intertropical Convergence Zone (ITCZ) and its associated weather. As a result, land areas very near to the equator (in parts of tropical Africa, for example) where the ITCZ crosses over twice in the year may have two rainy seasons each year.

2.2.1.1.3 Variations in orbital parameters of the Earth

Three of the Earth's orbital parameters have long-term variations that can cause large variations in the range of solar energy input over the annual cycle. These are the eccentricity of the orbit, tilt of the Earth's axis (obliquity) and the positioning of the Earth's axis (see Figure 2.3). The periods for oscillations in these three parameters are approximately 100 000, 41 000, and 22 000 years respectively. Milutin Milankovitch was the first person to theorize that these orbital variations could be responsible for climate variability related to ice ages. The positioning involves a precession of the Earth's axis that leads to changes in the season of the year when the Earth is closest to the sun. This precession effect may be described by changes in the longitude of perihelion relative to the vernal equinox.

Variations in these three orbital parameters cause significant effects on the amount of solar radiation received on the Earth as a function of season and latitude. It is possible for all three to be in phase and cause variations in seasonal insolation as large as 30 per cent in high latitudes (p. 307, Reference no. 1).

2.2.1.1.4 Meteors

A far less predictable astronomical forcing effect is due to meteors. The impact of large meteors with the Earth can cause a major short-term variation in the Earth's climate due to the production of large amounts of dust and smoke. Adding large amounts of dust and smoke can greatly reduce the amount of solar radiation that gets to the Earth surface causing surface temperatures to decrease. A current theory for the extinction of dinosaurs is that a giant asteroid hit the Earth about

65 million years ago. This introduced so much dust that surface conditions became much darker and much colder; the significant effect lasting for about three years.

2.2.1.2
Geological effects
2.2.1.2.1
Tectonics

Tectonic effects such as continental drift and changes in mountains produce very significant changes in climate. Their evaluation over time is important for determining past climates. The time scales associated with these changes are the order of millions of years or greater. Since they are far greater than those relevant for anthropogenic climate change effects, they are not discussed here.

2.2.1.2.2
Volcanoes

Volcanoes can inject vast amounts of dust and gases into the atmosphere. An individual eruption may affect climate conditions for up to three years. On average newsworthy eruptions occur every 20 years; major eruptions with a significant impact on the global climate will perhaps take place every 100 years or so. (The statistics can be better restated as: in a given year the probabilities of newsworthy and major eruptions are five and one per cent, respectively.) Volcanoes provide intermittent forcing impacts and are not generally associated with longer-term climate variations.

Volcanic products which get into the stratosphere can have a significant impact. Products that remain in the troposphere, such as the dust, are subject to rather rapid removal processes by gravitational settling and washout by precipitation. On the other hand sulphates formed from the sulphur dioxide injected into the stratosphere by more severe eruptions can have lasting effects. They are small particles with very small settling speeds. The stratosphere has little up-and-down motion and little precipitation to remove the sulphates. The primary effect of sulphate aerosols on radiation transfer is to reduce the short-wave solar radiation reaching the Earth's surface. This causes the cooling associated with volcanic events.

2.2.2
INTERACTION OF CLIMATE SYSTEM COMPONENTS
2.2.2.1
Ocean effects

For variability time-scales of less than a month, the ocean generally has a damping effect on amplitudes of climate system oscillations. Its heat capacity is large compared with that of other interactive components in the climate system, and horizontal transport effects due to currents are small on these time-scales. Thus, ocean temperature temporal variations are small, and variability of energy exchange processes that depend on temperature differences will be primarily a function of atmospheric variations.

On a seasonal time-scale, the ocean can increase the synoptic-scale transient fluctuations in the extratropical atmosphere by virtue of its slow temperature variations. By remaining nearly constant in temperature as the continents cool off in winter, surface temperature contrasts at the coasts of continents in the Northern Hemisphere increase. This fosters baroclinic processes in atmospheric disturbances and enhances synoptic-scale variability. Synoptic-scale activity can have a net effect on the general circulation due to correlations among synoptic-scale flow and temperature components. For longer time scales of variability, processes within the ocean begin to provide mechanisms that allow for two-way interactions with atmospheric processes. These interactions foster significant amplitudes of variability at the interannual and longer time scales. A striking example of this is the ENSO phenomenon which introduces into the climate system a pronounced global-scale variability on an interannual time-scale with periods ranging mostly from two to seven years.

ENSO has both ocean and atmosphere components. The primary two-way interactions occur in the tropical area of the Pacific Ocean. The ocean component has variations in tropical sea-surface temperature that are related to vertical motion and changes in the depth of the ocean thermocline. The oceanic vertical motions (upwelling) are associated with horizontal ocean currents forced by atmospheric surface winds. The atmospheric component has variations in convective storm activity in the equatorial region caused by changes in the sea-surface temperature. The convective activity influences the surface atmospheric pressure and the associated surface winds. Processes involving horizontal advection, upwelling, and wave propagation in the ocean are the primary determinants for the time scale of the ENSO cycle.

In the climatological mean, the ocean temperature in the western equatorial region of the Pacific is warmer than in the eastern equatorial region. This is accompanied by a rainfall maximum, surface pressure minimum in the western Pacific and surface trade wind flow in the atmosphere from east to west. During El Niño events the temperature in the eastern Pacific is increased which enhances rainfall in the east, reduces surface pressure difference between the east and west, and reduces (or even reverses) the low-level atmospheric flow component from east to west. In the opposite phase of the oscillation, termed 'La Niña,' the eastern Pacific sea-surface temperature becomes colder than normal, reducing precipitation in the region, enhancing the surface pressure difference between the east and west, and accelerating the east to west flow component at low levels.

It is noteworthy that the atmospheric general circulation in the tropical Pacific area has important east-west variations along with the north-south variations normally considered. The east-west variations are associated with the 'Walker Circulation' and the north-south variations are associated with the 'Hadley Circulation.'

Primary indicators for ENSO are the anomalies in the sea-surface temperature of the eastern tropical Pacific Ocean (east of the international date line) and averaged atmospheric surface pressure difference between the western and eastern tropical Pacific Ocean regions (see Figure 1.17). The latter has traditionally been represented by the monthly mean pressure differences between Darwin, Australia and Easter Island in the eastern Pacific. Sometimes Tahiti, to the west of Easter Island, is used instead of Easter Island. Both of these are south of the equator. The surface pressure difference also provides a measure of the east to west surface airflow and the stress force that it applies to the ocean surface currents. During El Niño events the pressure difference between Darwin and Easter Island decreases, the surface wind stress decreases and the eastern Pacific sea-surface temperature increases.

Observation for the ENSO indicators described above are shown in Figure 2.4 for the period from 1949 to 1988. In the Figure it is easy to see the frequency of occurrence of the El Niño and La Niña phases of the ENSO cycle. The major El Niño event years are marked with arrows in the pressure difference graphs. Note that the ENSO frequency increased in the 1990s with El Niño events in 1991, 1994, and 1997.

Earlier reference was made to the global influences of the ENSO cycle in terms of correlated variations that have been observed at great distances (teleconnections) from the tropical Pacific area. Figure 2.5 shows some of these ENSO-related anomalies (for the El Niño or warm phase), primarily for precipitation. The sign of the anomaly is shown (wetter, dryer, or warmer) indicating the length of time of duration and the time of year. The plus sign refers to the year after the El Niño year.

There are negative precipitation anomalies in the west Pacific area and positive values in the central Pacific as the main precipitation area is displaced eastward. Also, conditions are wetter than normal on the west coast of South America. These wetter conditions and the warmer than normal sea-surface temperatures along the west coast of South America have major impacts on people living there, especially fishermen. The more distant impacts include altered monsoon precipitation in India, altered rainfall in central and southern Africa, reduced rainfall in Australia, reduced rainfall in Central America and the northern part of South America, increased rainfall in the Gulf of Mexico area, and warmer temperatures in Alaska and western Canada. Figure 2.5 also shows the locations of Darwin and Easter Island, locations that have been used to define a pressure gradient indicator for ENSO.

The ocean has a major role in decadal and longer time scales of variability of climate. For these time scales, ocean circulations in the thermocline and deep waters over vast regions of the ocean are important factors. There is observational evidence of decadal variability in sea-surface temperatures in the Pacific and

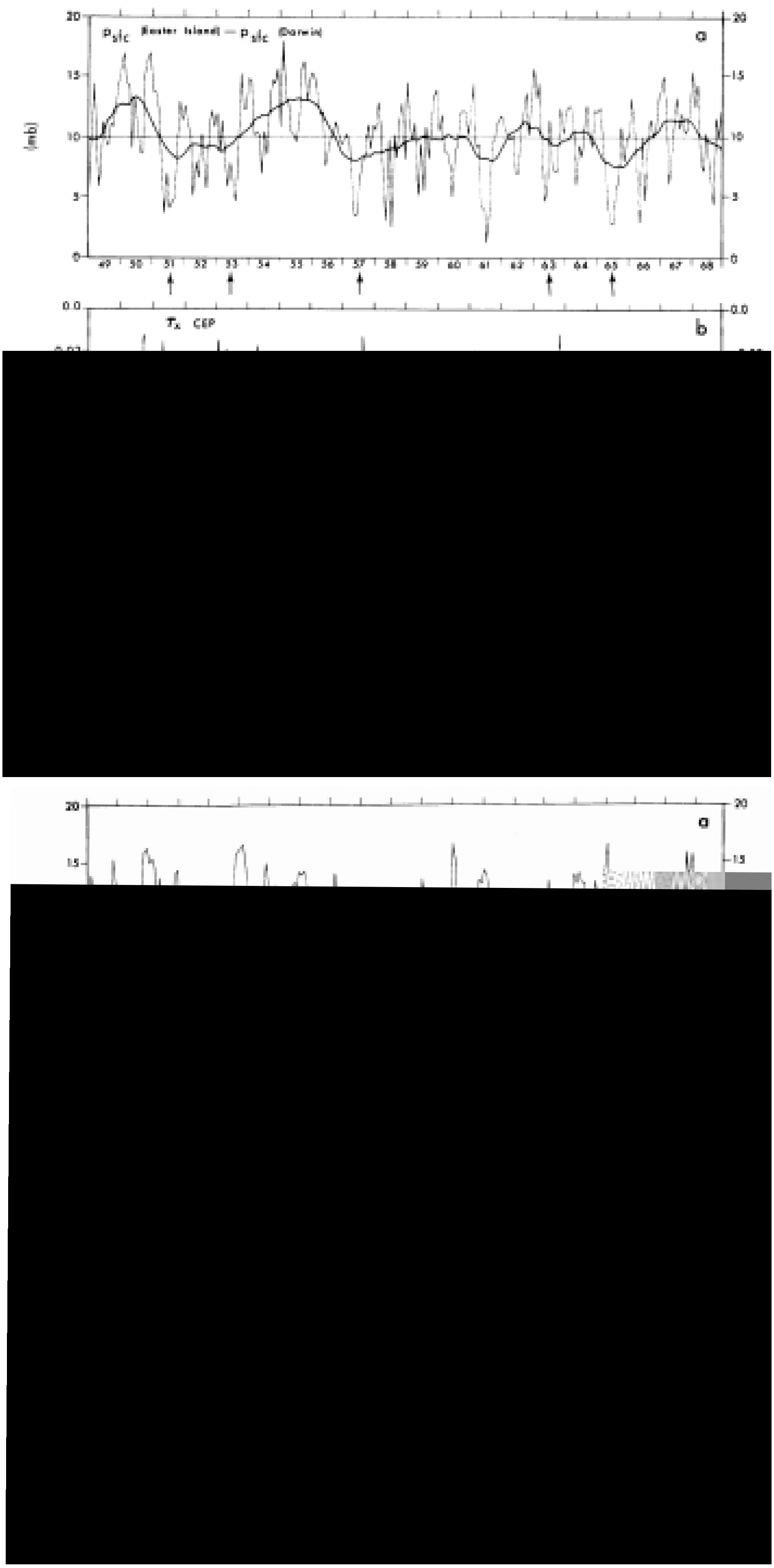


Figure 2.4 — Two sets of monthly-mean time series of (a) a classical index of the Southern Oscillation, i.e., the pressure difference between Easter Island and Darwin in mb (after Wyrtki, 1982, updated with data from Ropelewski, 1987; (b) the zonal wind stress over the central τ_x^{CEP} equatorial Pacific region (8°S-4°N, 160°E-130°W) in units of Pa (Pa = 10 dyn cm $^{-2}$); and (c) the sea-surface temperature anomalies in the eastern equatorial Pacific region T_s^{EEP}(20°S-20°N, 180-80°W) in °C.

The upper set of three panels cover the years 1949 to 1968, the lower set covers 1969 to 1988. The most important El Niño events (i.e., when warm water covers the entire eastern equatorial Pacific Ocean) according to Rasmusson and Carpenter (1982, updated) are indicated by arrows at the bottom of (a). The thick solid lines indicate a 12-month running mean in (a) and 15-month weighted means in (b) and (c) (obtained by using a Gaussian-type filter with eights 0.012, 0.025, 0.040, 0.061, 0.083, 0.101, 0.117, and 0.122 at the central point). [from pages 424 and 425, Reference no. 5, with permission of Springer-Verlag].

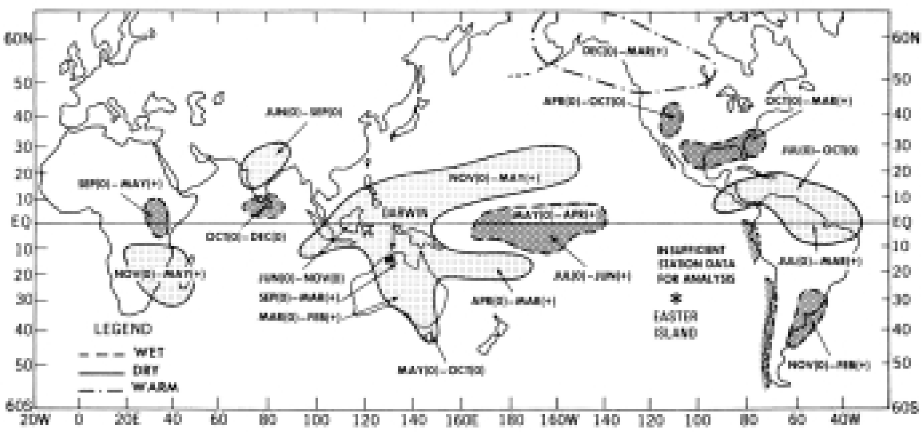


Figure 2.5 — Schematic representation of typical ENSO-related precipitation anomalies over the globe. Solid contours enclose relatively dry regions (light shading) and dashed contours enclose relatively wet regions (heavier shading). The approximate period of extreme conditions relative to the typical El Niño (0) year is also shown for the various regions (adapted from Ropelewski and Halpert, 1987). [from page 427, Reference no. 5, with permission of Springer-Verlag].

Atlantic Oceans. Oceanic processes that have a decadal time scale include advective effects in the primary central gyre in the north Pacific and subduction processes within the thermocline layer between the middle latitude and tropical regions. It will require analysis with numerical models to sort out the role of these processes as there is only a short record of observations for subsurface currents in the ocean.

Currents in the deep ocean layers below the thermocline owe their existance to thermohaline density effects. Observational evidence has been obtained to show that the deep ocean currents have a large scale structure and connect the major ocean basins of the world. Figure 2.6 shows a schematic of this circulation which is referenced as the 'great ocean conveyor belt.' The interconnection of the deep ocean and surface waters brings the heat storage of the entire ocean into play and produces century- and millennium-scale variability in sea-surface temperature and the atmospheric climate. A key part of the deep ocean circulation system is the downwelling regions where dense water sinks to deep or even bottom layers of the ocean in relatively small areas of rapid downward motion in the vicinity of Greenland and Antarctica.

2.2.2.2 Cryosphere effects

There are a number of important interactions of the cryosphere with the other components of the climate system that can affect variability characteristics. None of these introduce easily-defined time scales of oscillation, as is the case for the ocean interactions. Ice cover is an indicator for changes in climate. The net melting of mountain glaciers observed since 1850 is a clear example.

As earlier discussed, ice cover on the Earth's surface has a dramatic feedback effect on climate due to its high albedo. Briefly, an increase in ice cover by virtue of increasing surface albedo causes less solar energy to be absorbed at the Earth's

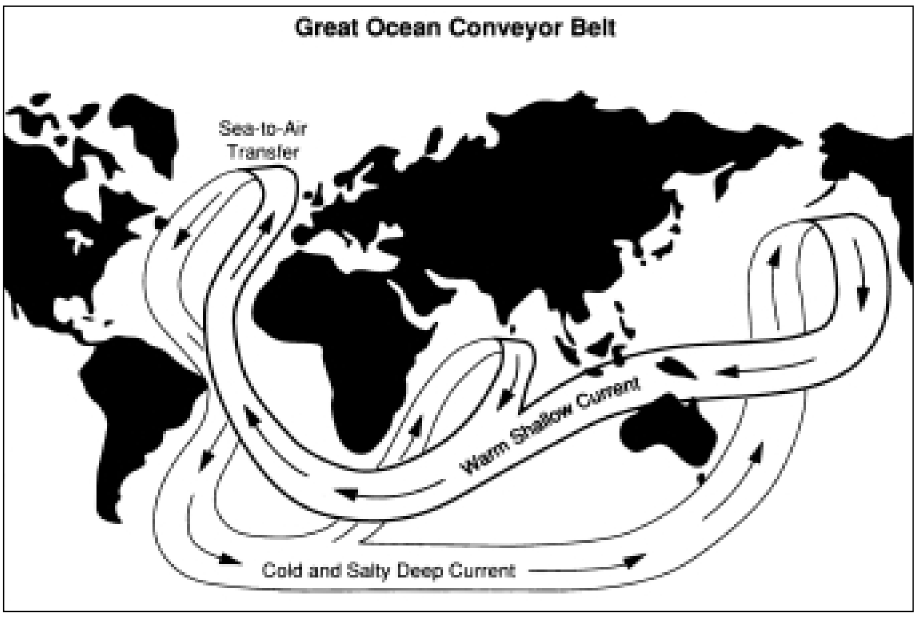


Figure 2.6 — Schematic diagram of the global 'conveyor belt' depicting global thermohaline circulation (after Broecker, 1987). [from page 580, Reference no. 6, with permission of Cambridge University Press].

surface which helps to maintain the cool temperatures needed to sustain or increase the ice cover. This positive feedback process is considered to be a factor in the maintenance of previous ice age conditions.

There are several other interactive processes. The melting of ice over and near the oceans in the polar regions can increase the supply of fresh water to the oceans. This water will tend to stay at the surface because its low salt content tends to make it less dense than sea water. Sufficient amounts of fresh water at the ocean surface in the polar regions have the potential to reduce or even prevent the occurrence of strong downward convective currents that supply cold water to the deep waters and maintain the great ocean conveyor belt. Suppression or alteration of the great ocean conveyor belt would have major global, long-term climate impacts.

Another interactive process involves the polar ice caps and the oceans. Most of the ice in the cryosphere is found in the polar ice caps over Antarctica and Greenland. As these ice areas lie on solid ground and are not floating in the ocean, their melting will contribute directly to sea-level changes. It is estimated that if these two ice fields melted completely, the sea level would rise more than 75 metres. It is also estimated that in the last major ice age (during the Pleistocene about 18 000 years ago) sea levels were lower by about 100 metres. Such changes in sea level would significantly affect the shape of ocean coastlines and the impacts of ocean bottom topography leading to changes in the basic ocean circulation. Interactions involving the cryosphere have time scales that can range upwards to many thousands of years.

2.2.2.3 Biosphere interactions

The biosphere, both on land and in the ocean, has a number of interactions that can lead to climate variability. We are just beginning to understand some of the aspects that could lead to variability cycles. Biosphere relationships with albedo and greenhouse-gas concentrations provide two important factors for such climate system variability. Many investigations have focused on the impacts of human alterations of the terrestrial biosphere.

Albedo impacts on vegetation ground cover are shown in Table 1.1. Vegetation can have albedos that are higher or lower than ground cover without vegetation. It has been possible to show examples of the biosphere albedo interactive effects on the climate system through very simple models. The classical example is the ‘Gaia’ model developed by Watson and Lovelock (1983) summarized very nicely on pages 251-253 in Reference no. 1.

The ‘Gaia’ model has a world with three regions covered by black daisies (low albedo), white daisies (high albedo), and Earth bare soil (intermediate albedo), respectively. The growth and death rates of the daisies are prescribed functions of regional temperature. The temperature in turn is determined from equilibrium conditions for the overall global radiation budget (considering both the solar and terrestrial radiation components) which also satisfies local energy balance for each region. The local energy balance for each daisy type includes assumed rates of energy transfer between the local areas depending on temperature differences. The model defines an equilibrium solution in terms of the percentage of the area covered by each type of daisy for a prescribed value of solar luminosity. No oscillating solutions are produced in this simple model.

Daisy regions exist only for an intermediate range of solar luminosity values. Values too low or too high give conditions which are too cold or too warm for growth. For solar luminosity values at the low end of the range the black daisies predominate. For luminosity values at the high end of the range the white daisies predominate. The albedo of the daisy is a key factor in determining the temperature of the daisy’s growth environment.

The land-surface biosphere interacts directly with the greenhouse gases (water vapour, carbon dioxide and methane) in the atmosphere. Water intake and water vapour supply are part of the terrestrial biosphere life processes, and methane is a product of decay. The marine biosphere interacts with the carbon dioxide absorbed in the water (dissolved inorganic carbon) and converts it to solid carbonates. The carbon dioxide gas can be transferred directly to and from the atmosphere. The global carbon cycle presented earlier in Figure 1.13 shows components for both the terrestrial and marine biosphere.

Time scales associated with biosphere interactions are short for responses of the atmosphere to biosphere changes in comparison to those for the response of the biosphere to changes in atmospheric conditions. The former involves changes in albedo and surface energy and moisture transfer rates, whereas the latter involves life cycles of plant life. The terrestrial biosphere response to the annual cycle is clearly evident in the differences between summer and winter and between wet and dry seasons. Changes in climate can cause changes in the types of plants growing in a particular place and the gradual migration of the area where a given plant species exists. A good example is the northward advance of the boreal forests in Canada and Siberia in the last 10 000 years following the recent ice age. Associated time scales vary from decadal to centennial to even longer. The terrestrial biosphere is also affected by human activity (such as agriculture, forest cutting, and urbanization) which introduces time scales dependent on the rate of change due to human activity.

2.2.2.4 Internal atmospheric processes

There are a number of low-frequency oscillations in the atmospheric flow which are largely independent of interactions with other components of the climate system or effects of external forcing. These have a variety of oscillation periods and preferred latitudes for their existence, and may or may not have an impact on climate. Many of them include nonlinear processes.

The stratospheric Quasi-Biennial Oscillation (QBO) is a variability in the stratospheric flow field most apparent in the upper and middle stratosphere in tropical latitudes. It has oscillations periods ranging from 22 to 34 months and has minimal impacts on tropospheric climate. It is believed to result from the interaction of vertically propagating gravity waves and the mean horizontal flow.

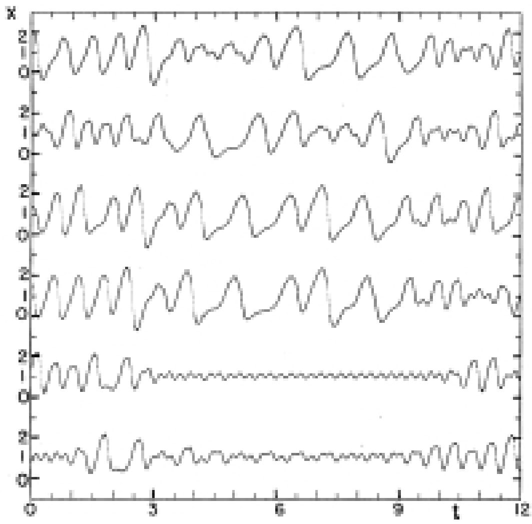
Another oscillation most pronounced in the tropical region is a tropospheric intraseasonal oscillation, sometimes referred to as the 'Maddan-Julian Oscillation.' This oscillation originates in the African area, develops as it moves eastward over the Indian Ocean, and then diminishes as it reaches the eastern Pacific Ocean. It has a period of oscillation ranging from 30 to 60 days. The oscillation modulates tropical convective activity and is related to break periods in the Indian monsoon. It results from interactions between eastward-moving Kelvin waves in the atmosphere and convective activity.

A variability centered in the extratropical troposphere regions results from nonlinear processes in the tropospheric jet streams. Nonlinear effects give rise to 'index cycles' and 'weather regimes' which involve persistent trough and ridge patterns in the tropospheric jet streams and related modulations in extratropical cyclone activity. The periods of oscillation are generally in the range of 10-20 days. A notable aspect of this variability is the development of a 'blocking' pattern where a persistent and anomalous anti-cyclonic flow pattern prevents cyclones and fronts from reaching given regions for a sustained period. This may cause a drought situation to become very severe.

Nonlinear aspects of the atmospheric flow such as described for the extratropics above seemingly lead to erratic variability characteristics. A focus on the nonlinear aspects alone is described by mathematical chaos theory. Solutions found by Lorenz (1990) for an idealized system of three components (a zonal jet flow and two superimposed transient eddy components) showed some irregular characteristics that resembled extratropical latitude flow variability.

His solutions for the strength of the zonal flow forced by a smoothly varying annual cycle are shown in Figure 2.7 for a six-year period. The winter season is on the left and right hand sides of the diagram and the summer season is in the middle. The irregular oscillations appear to have some systematic aspects which change as a function of season. The overall nature of the oscillation pattern changes from year to year. In general chaos theory the variability is of 'fractal' nature meaning that it has similar complex structure for any time scale. Such theory may be relevant to describing some aspects of climate variability.

Figure 2.7 — The variations of X (dimensionless variable representing the zonal flow speed) with time, t (months) in a 6-year numerical solution of an idealized three-component nonlinear circulation model. Each row begins on 1 January, and, except for the first, each row is a continuation of the previous one. [from page 383, Lorenz, 1990, with permission of Munksgaard].



2.3 OBSERVED CLIMATE VARIABILITY

For any time scale there are variations in the climate variables such as temperature, precipitation, and severe storms and in related climate system parameters such as sea-surface temperature and sea level. A few examples are shown here to highlight magnitude, range, periodicity, and extremes in the variations. In the examples you will see both systematic and irregular types of variability. It needs to be emphasized that observational information itself has quantitative uncertainty. The proper use of quantitative data requires being aware of their uncertainties and what region and time scale they actually represent.

2.3.1 SURFACE TEMPERATURE

For the geological time scales a schematic for the Earth's overall surface temperature levels for the past 100 million years is shown in Figure 2.8. The curve shown for the future (on the left side) is speculative and shows an early numerical model estimate for effects where a doubling of CO_2 was the only human impact considered. It is noted that climate change projections for temperature increases may be less important than natural variations over time; however, the rate of change might be larger in anthropogenic climate change compared to any natural rates of change.

Primary features of temperature variations in the past 100, 1 000, 10 000, and 100 000 years (up to about 1970) shown in Figure 2.9 reveal irregular variations at all these time scales. Note that the region covered and the data source vary from panel to panel. Some of the recent variations relevant to the climate change discussion are identified: the warm period 'thermal maximum of 1940s covering the 1930s-1950s; the particularly cold period 'little ice age' particularly in the 18^{th} and 19^{th} centuries; and the 'Younger Dryas cold interval' about 10 000 years ago. The range for temperature variations increases with the time scale which is consistent with the 'red' nature of the variance spectrum discussed at the beginning of the chapter. For the last 15 000 years the range is 10°C in the middle latitudes in the Northern Hemisphere (4-5°C for the global mean) whereas the temperature range is only 1.5°C for the last 1 000 years over eastern Europe.

The details of surface global-mean temperature variations for the period since 1860 (up to 2000), used as a reference for current climate change impact studies, is shown in Figure 2.10. The values are actually deviations from the 1961-1990 time period. Note that both yearly and smoothed depictions are used. The level of uncertainty in the values is suggested by the plotted standard error bars and the differences between two estimates, the dashed and solid curves, respectively. Note that the standard error is larger for the earlier years.

Values for each hemisphere and the globe all show the general upward trend in temperature. Based on this data it is now estimated that global mean surface temperature has increased about 0.6 degrees C since the beginning of the 20^{th} century. It is also considered, with very high statistical confidence, that the last decade (1900-1999) has been the warmest decade of the period. Furthermore,

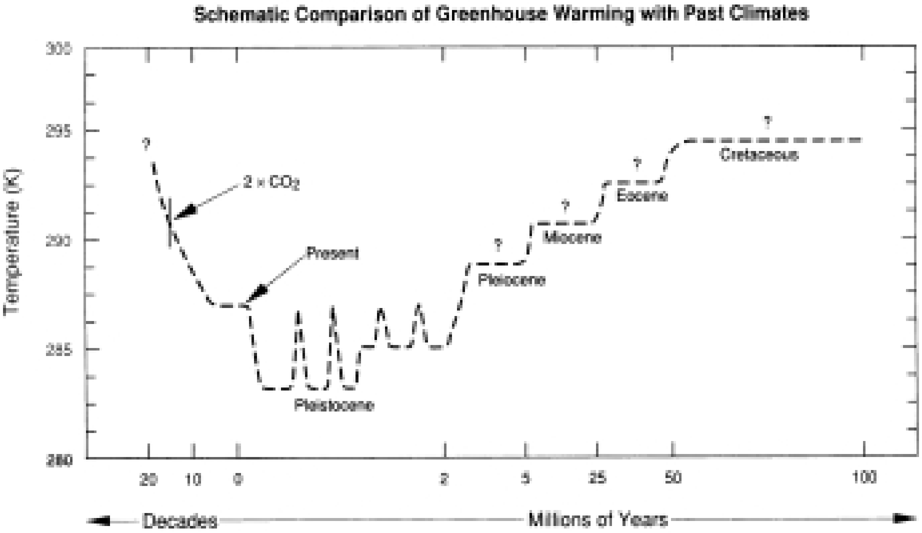


Figure 2.8 — Rough schematic comparison of possible future greenhouse warming with estimates of past changes in temperature. Pleistocene glacial-inter-glacial cycles are more numerous than shown. The characteristic amplitude of global temperature change during glacial-interglacial cycles is 3-4 K. Note that pre-Pleistocene changes are not well fixed in magnitude, but their relative warmth is approximately correct. Maximum warming in the Cretaceous is based on estimates by Barron and colleagues. Time intervals in between have been scaled accordingly (Crowley, 1989). [from page 671, Reference no. 6, with permission of Cambridge University Press].

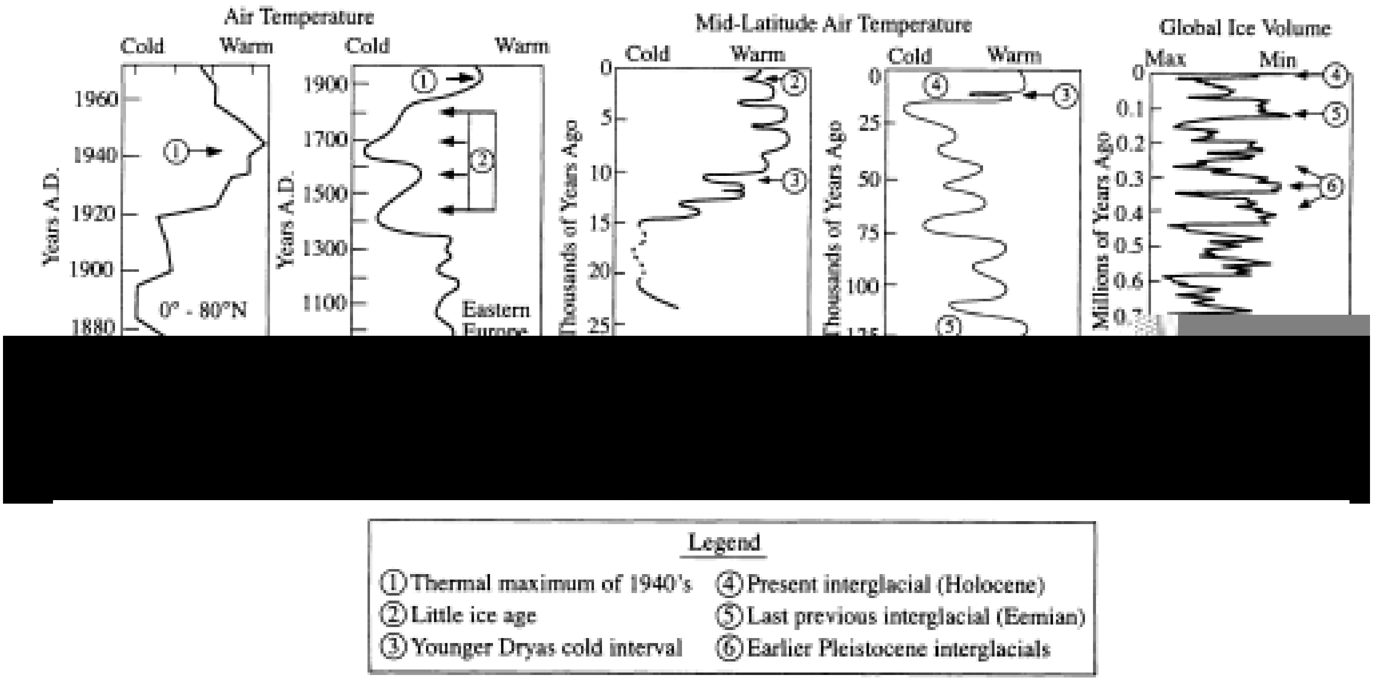
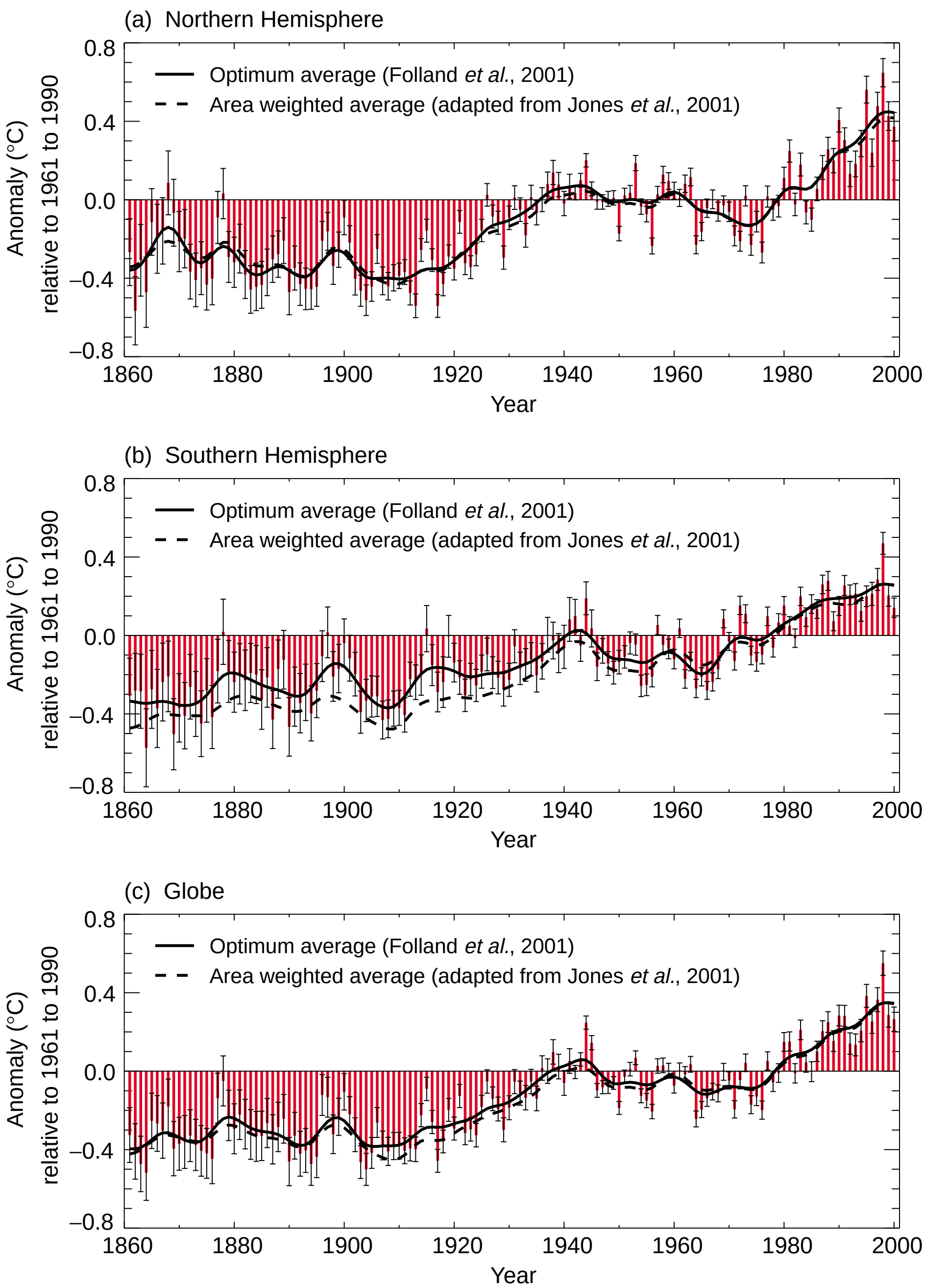


Figure 2.9 — General trends in global climate for a variety of time scales. (a) Changes in the 5-year average surface temperatures from about 1875 to 1970 averaged from instrumental records over the region 0-80°N (from Mitchell, 1963). (b) Winter severity index for eastern Europe during the last 1000 years up to about 1970 (from Lamb, 1969). (c) Mid-latitude Northern Hemisphere air temperature trends during the last 15 000 years based on changes in tree lines [from La Marche (1974), marginal fluctuations in alpine and continental glaciers (from Denton and Karlén, 1973), and shifts in vegetation patterns recorded in pollen spectra (from van der Hammen et al.*, 1971), (d) Northern Hemisphere air temperature trends during the last 100 000 years based on mid-latitude sea-surface temperature and pollen records and on worldwide sea-level records. (e) Fluctuations in global ice volume during the last million years as recorded in changes in isotopic composition of fossil plankton in deep-sea core V28-238 (from Shackleton and Opdyke, 1973).* [from page 256, Reference no. 1, with permission of Academic Press].

Figure 2.10 — (a) to (c): Combined annual land-surface air and sea-surface temperature anomalies (°C) from 1986 to 1999, relative to the 1961 to 1999 mean value temperatures (vertical solid bars) for Northern Hemisphere, Southern Hemisphere, and Globe respectively; twice their standard error (denoted by the thinner vertical lines with small horizontal bars at the top and bottom, where one end may be partially obscured by the solid bars); and time-smoothed curves calculated using optimum averages and area weighted averages (solid and dashed lines, respectively). Information is from data bases at the U.K. Met. Office and the Climatic Research Unit at the Hadley Centre, both in the United Kingdom. [from page 114, Reference no. 7]



based upon indicators for temperature, it is considered likely that the increase in temperature in the 20th century has been larger than that for any century in the past 1 000 years.

2.3.2 PRECIPITATION

Precipitation variability has quite different features from temperature variability. Figure 2.11 presents precipitation variability for all of the main land areas of the globe except for Antarctica for the past 90 years. Decadal variability is quite evident as it was for temperature. The data suggests systematic variations of land-area precipitation with latitude with increases in precipitation in the middle- and high-latitudes of the Northern Hemisphere and decreases in the tropics. It is difficult to see correlations between the temperature and precipitation variability shown in Figures. 2.10 and 2.11, respectively. Also the precipitation does not have a clear trend. Differences in precipitation variability from one area to another in the tropics and subtropics is shown in Figure 2.12. Note that in northern Africa multidecadal variations are quite evident whereas in the regions of Mexico and India higher frequency oscillations are more evident.

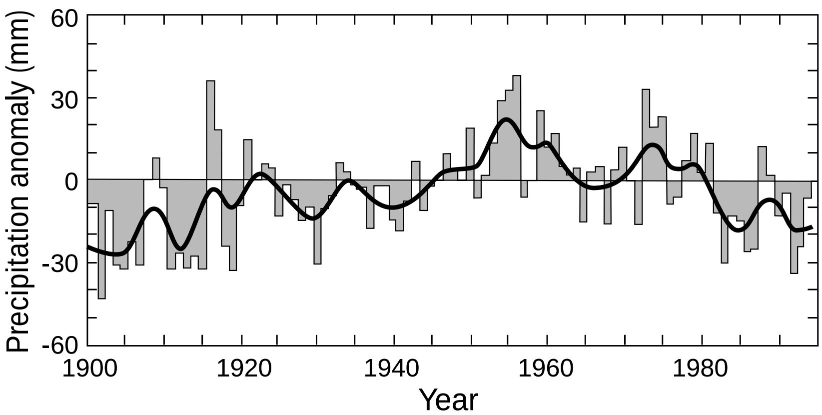


Figure 2.11 — Changes in land-surface precipitation averaged over regions between 55°S and 85°N. Annual precipitation departures from the 1961-1990 period are depicted by the hollow bars. The continuous curve is a smoothing of the same data. [from page 28, Reference no. 3].

2.3.3 SEVERE WEATHER

It is important to be aware of the variability in extreme events such as droughts, floods and tropical storms. Drought conditions similar to those experienced in Africa over several years in the 1980s can be devastating in semi-arid areas. Long-term changes in water supply result in significant changes in areas covered by semi-arid and arid conditions with important impacts on civilization. An integrated picture of rainfall and evaporation in certain watershed regions may be provided by lake level information such as for Lake Victoria, Tanzania and the Great Salt Lake in the western United States. A 97-year record for lake level in the Great Salt Lake shows a dramatic drop from 1925 to 1935 at the time of the very dry period in the central and western part of the country and a 4 metre increase during the wet 1982-85 period.

As a second example, climatic variations in severe storm activity for the past 100 years up to 1988 are shown for the case of tropical storms and hurricanes in

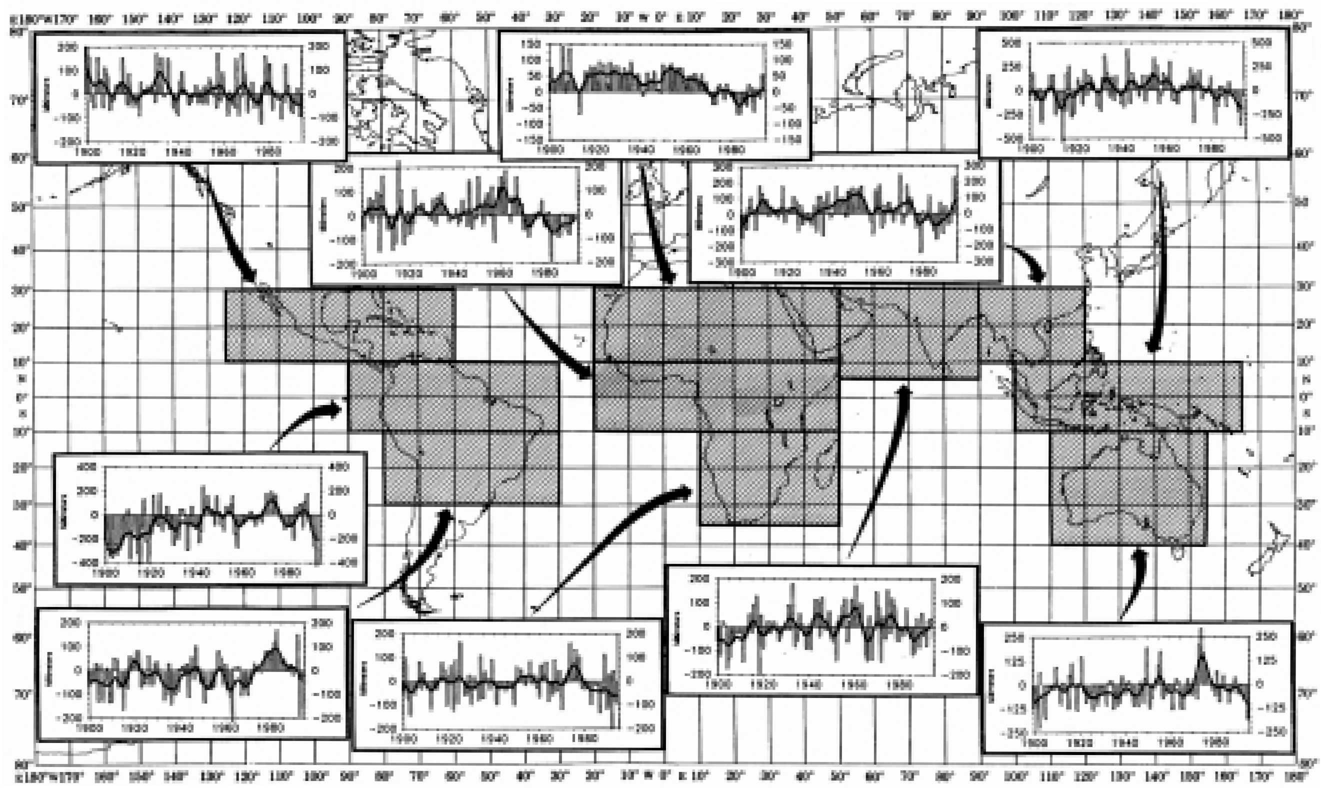


Figure 2.12 — Variations of tropical and subtropical land-surface precipitation anomalies based on the average of the anomalies relative to 1961-1990 means from Hulme, 1991, Hulme et al.*, 1994, and the Global Historical Climate Network (GHCN) [Vose* et al.*, 1992; Eischeid* et al.*, 1995]. Smooth curves are generated from nine-point binomial filters of the annual anomalies.* (from page 154, Reference no. 3).

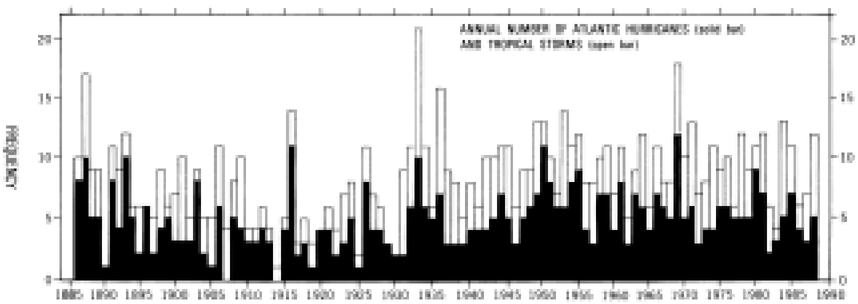


Figure 2.13 — Time series of the annual number of Atlantic tropical cyclones reaching at least tropical storm strength (open bar) and those reaching hurricane strength (solid bar) for 1886-1988. The average numbers of tropical storms and hurricanes per year are 8.4 and 4.9, respectively (adapted from Neumann et al., 1981, updated). [from page 448, Reference no. 5, with permission of Springer-Verlag].

the Atlantic ocean area in Figure 2.13. The overall pattern in variations over time is quite different from that for global surface temperature as shown above in Figure 2.10.

2.3.4 OCEAN CONDITIONS

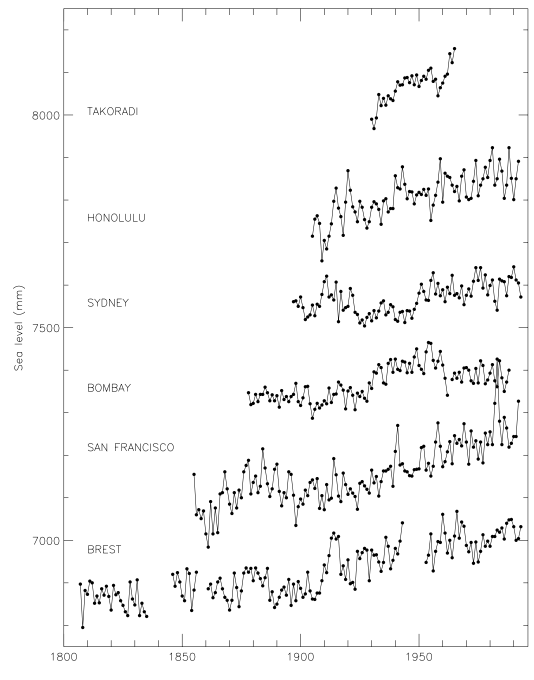
Significant variations in sea-surface temperature exist at annual, interannual, decadal and longer time scales which are important for atmospheric climate. Pronounced interannual variability exists in the tropical Pacific sea-surface temperature as part of ENSO described earlier in Chapter 1 (Figures 1.17 and 1.18) and in section 2.2.2.1 (Figure 2.4) above. Oceanic variability at decadal scales has been hard to detect and describe because of the short observational record.

Recently decadal variability in sea-surface temperature has been detected in the central and northern Pacific Ocean area. In the north Atlantic the largest oceanic decadal signal detected so far has been found in association with the atmospheric North Atlantic Oscillation (defined by the atmospheric surface pressure difference between Iceland and the Azores or Lisbon). The decadal component of the North Atlantic Oscillation correlates with decadal variations in sea-surface temperature, currents, salinity and sea ice in the north Atlantic. The involvement of deep thermohaline currents can link this variability with other parts of the ocean system.

Sea level itself is of interest to all shore land areas and ocean island inhabitants. Long-term records for sea level show a general rising trend in the major ocean regions in the range of 1-3 mm/yr along with decadal variations as large as 200 mm (see Figure 2.14).

The latest IPCC assessment (2001) concluded that the global mean sea-level rise was between 10 and 20 cm in the 20th century and that this was mainly the result of a rise in ocean temperature.

Figure 2.14 — Six long sea level records from major world regions: Takoradi (Africa), Honolulu (Pacific), Sydney (Australia), Bombay (Asia), San Francisco (North America) and Brest (Europe). Each record has been offset vertically for presentational purposes. The observed trends (in mm/yr) for each record over the 20th century are, respectively, 3.1, 1.5, 0.8, 0.9, 2.0, and 1.3. The effect of post-glacial rebound (lowering relative sea level) as simulated by the Peltier ICE-3G model is less than, or of the order of, 0.5 mm/yr at each site. [from page 367, Reference no. 3].



CHAPTER 3

HUMAN IMPACTS ON THE CLIMATE SYSTEM

3.1 INTRODUCTION

As discussed in Chapter 1, the primary human activities which cause climate change are those which influence radiative transfer in the atmosphere and radiation absorption on the Earth's surface. Other human activities, such as heating the atmosphere through combustion and changing surface wind flow by deforestation and building construction, are of secondary importance. There are four specific impacts to be considered: increasing greenhouse gas concentrations in the atmosphere, adding aerosols to the atmosphere, changing cloudiness and changing surface conditions. The first two clearly have global effects. There is incomplete understanding for climate impacts from changes in cloudiness due to human activity. More observational and modelling studies are needed. The effect could be large and global, particularly if the cloud changes are related to the aerosols added by human activity. Land surface changes may have large impacts on local climate; but have less impact globally as land covers only 29 per cent of the earth's surface.

There are several measures of the magnitude of human impact. An overall measure is accomplished by comparison of current conditions of the atmosphere and the Earth's surface with those that existed before the industrial revolution (considered to be before 1750). A measure used for gases and aerosols is the current rate of increase in atmospheric concentration. This is expected to relate directly to level of possible impact in the future. A convenient way to describe and compare the overall impact of human-caused changes in greenhouse gases or aerosols on radiation exchanges is by their 'radiative forcing.'

Radiative forcing here is defined as the change in average net total radiation in the planetary radiation budget at the top of the troposphere due to changes in solar and infrared radiation with fixed vertical structure for temperature. Radiative forcing can be used to describe the impacts of both anthropogenic and natural changes in the physical system. A positive radiative forcing (increase in net downward radiation) tends to cause average warming of the Earth's surface and a negative radiative forcing, average cooling. This definition was made by the IPCC in 1995. As stated in that report: "For a range of mechanisms there appears to be a similar relationship between global mean radiative forcing and global mean temperature change. However, the applicability of global mean radiative forcing to mechanisms such as changes in ozone or tropospheric aerosol (concentrations), which are spatially very inhomogeneous, is unclear." [p. 16, IPCC, 1995].

3.2 ATMOSPHERIC GREENHOUSE GAS ENHANCEMENT

The natural atmosphere contains greenhouse gases (primarily water vapour, carbon dioxide, ozone, nitrous oxide, and methane) which have a major impact on determining temperatures in the atmosphere and at the Earth's surface. Human activity has provided additional sources for these and other gases that have greenhouse-gas characteristics. The result is an enhanced greenhouse effect which is expected to force increased temperature at the Earth's surface and in the lower atmosphere.

All greenhouse gases have sources and sinks which may include chemical conversions in the atmosphere. The effect of human activity on concentration levels in the atmosphere depends on the cumulative amounts added by human activity and the strength of the sinks.

For each greenhouse gas constituent one can define an 'adjustment time' which describes the rate of reduction of a concentration enhancement. For instance, if an unusual event were to suddenly increase carbon dioxide concentration in the atmosphere by 100 parts per million by volume (ppmv), the adjustment time may be defined as the time it takes to reduce the concentration enhancement to 1/e of its initial value (where e is the base of natural logarithms or about 2.71) or to roughly 37 per cent of its initial value, namely 37 ppmv. This definition assumes that the rate of depletion of the gas concentration is proportional to the concentration. Sometimes the adjustment time is equated to the ‘lifetime’ of the constituent. The lifetime is defined as the total content of the constituent divided by the current rate of removal. The lifetime definition is used for stationary input of a tracer into the atmosphere.

3.2.1 NATURAL GREENHOUSE GAS CONSTITUENTS

Three of the five natural greenhouse gas constituents are undergoing well-documented increases in concentration due to human activity: carbon dioxide, methane, and nitrous oxide. The other two naturally occurring greenhouse gases, water vapour and ozone, are less directly linked to human activity. The role of ozone is small. The relevant concentration for ozone includes both tropospheric and stratospheric components and not just the stratospheric concentration values for which there has been an observed decrease in polar regions. Water vapour is the dominant greenhouse gas and accounts for about 75 per cent of the overall greenhouse effect. However, its modification results primarily from changes in evaporation rates from the oceans due to temperature and surface wind variability and not from human activity.

Carbon dioxide enhancement is the most important human impact on the greenhouse gases. This enhancement accounts for more than half of the total enhanced greenhouse effects due to human activity. A continuous observational record for atmospheric carbon dioxide has been obtained since 1958 at Mauna Loa, Hawaii and since 1957 at the South Pole. Concentrations prior to that time have been measured from air trapped in ice in Antarctica. Both of these measurements are considered to be representative of global mean concentration. Carbon dioxide concentration is rather uniform in the atmosphere.

The change in concentration values from the pre-industrial period to about 1989 reveals that the rate of increase itself is increasing with time as shown in Figure 3.1. The figure also shows estimates for the production of carbon dioxide from fossil fuel combustion and cement manufacturing since 1860. On a linear scale the shape of the production curve would be similar to that shown for carbon dioxide concentration. The rate of production has increased by nearly a factor of 10 since 1900.

The overall human impact on carbon dioxide concentration has been to increase it from roughly 278 ppmv to 365 ppmv (the value in 1998), an increase of 87 ppmv or almost 31 per cent. The current level is estimated to be higher than at any time since the last interglacial warming about 120 000 years ago. At the current rate of increase, the present (1998) carbon dioxide concentration will double in less than 100 years.

Estimates for the budget for the carbon introduced into the atmosphere by human activity from 1980 to 1989 are presented in Table 3.1. Shown are

Table 3.1 (below) — Average annual budget of CO_2 perturbations for 1980 to 1989. Fluxes and reservoir changes of carbon are expressed in GtC/yr, error limits correspond to an estimated 90 per cent confidence interval. [from page 79, Reference no. 3].

	IPCC 1992†	*IPCC 1994**	*IPCC 1995*
	Estimates for 1980s budget		
CO_2 sources			
(1) Emissions from fossil fuel combustion and cement production	5.5 ± 0.5$^{\Delta}$	5.5 ± 0.5	5.5 ± 0.5§
(2) Net emissions from changes in tropical land-use	1.6 ± 1.0$^{\Delta}$	1.6 ± 1.0	1.6 ± 1.0§
(3) Total anthropogenic emissions = (1) + (2)	7.1 ± 1.1	7.1 ± 1.1	7.1 ± 1.1
Partitioning amongst reservoirs			
(4) Storage in the atmosphere	3.4 ± 0.2$^{\Delta}$	3.4 ± 0.2	3.3 ± 0.2§
(5) Ocean uptake	2.0 ± 0.8$^{\Delta}$	2.0 ± 0.8	2.0 ± 0.8§
(6) Uptake by Northern Hemisphere forest regrowth	not accounted for	0.5 ± 0.5	0.5 ± 0.5§
(7) Other terrestrial sinks = (3)–((4)+(5)+(6)) (CO_2 fertilization, nitrogen fertilization, climatic effects)	1.7 ± 1.4	1.4 ± 1.5	1.3 ± 1.5

† Values given in IPCC (1990, 1992).

* Values given in IPCC (1994).

Δ Values used in the carbon cycle models for the calculations presented in IPCC (1994).

§ Values used in the carbon cycle models for the calculations presented here.

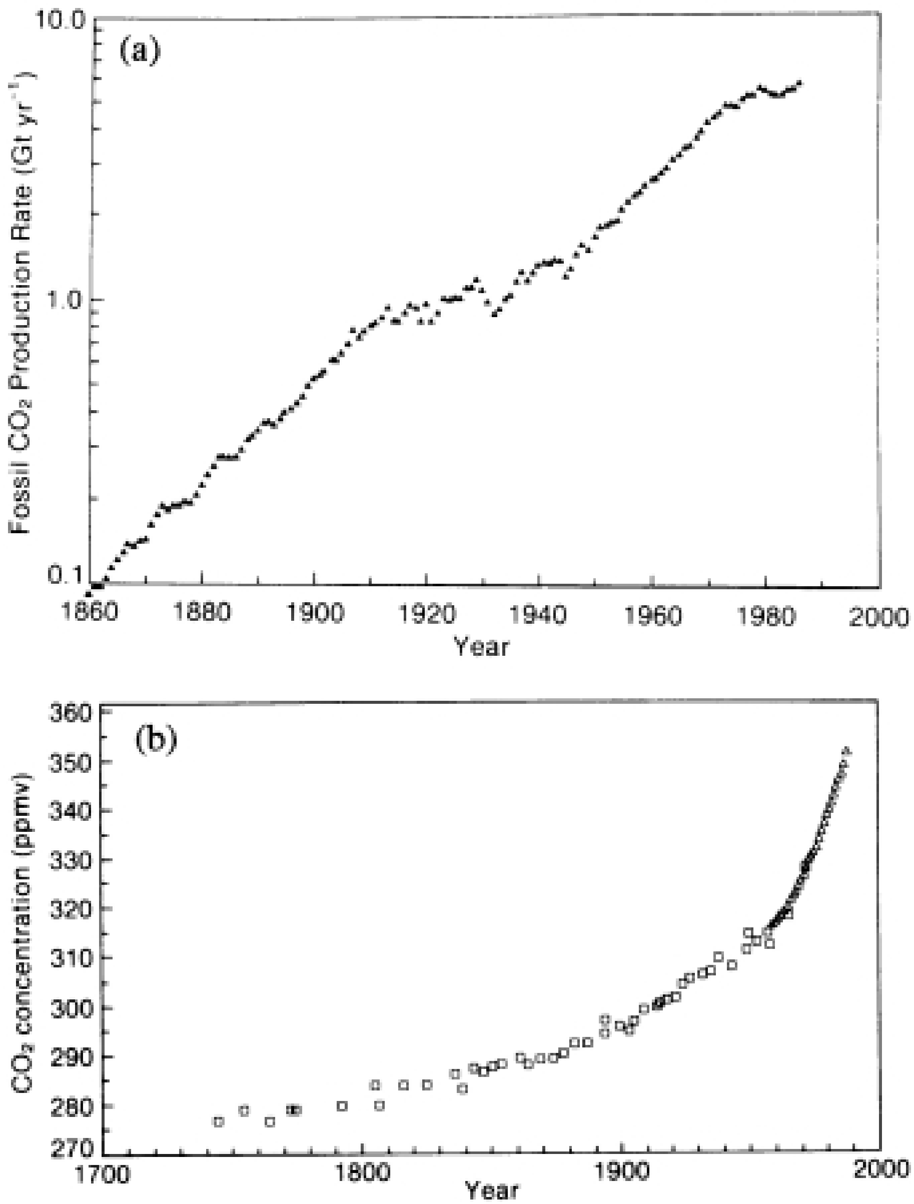


Figure 3.1 — (a) Global annual emissions of CO_2 from fossil fuel combustion and cement manufacturing expressed in GtC year^{-1} (Rotty and Marland, 1986; Marland, 1989). The average rate of increase in emissions between 1960 and 1910 and between 1950 and 1970 is about 4 per cent per year. Note: the ordinate scale is logarithmic. (b) Atmospheric CO_2 concentration for the past 250 years as indicated by measurements in air trapped in ice from Siple Station, Antarctica (squares, Neftel et al.*, 1985; Friedli* et al.*, 1986), and by direct atmospheric measurements at Mauna Loa, Hawaii (triangles, Keeling* et al.*, 1990). Note: ppmv means part per million by volume.* [from Watson *et al.*, 1990, from page 323 in Reference no. 1, with permission of Academic Press].

apportionments for the sources and reservoir dispositions. Fossil fuel combustion and cement manufacturing account for more than 75 per cent of the total carbon dioxide input. The remainder (less than 25 per cent) is due to net effects of deforestation and other land-clearing operations mainly in tropical areas. Of the total carbon dioxide input into the atmosphere, almost half remains in the atmosphere, an estimated 30 per cent goes into the ocean, seven per cent into Northern Hemisphere forest regrowth, and the remainder is assumed to go into other parts of the terrestrial biosphere. Note that the remainder component in the budget estimates has been reduced between the 1992 and 1995 IPCC reports. The 2001 IPCC report (Reference no. 7) shows that for the decade 1990-1999 the land ecosystem uptake of CO_2 is larger and is now nearly as large as the ocean uptake. The estimates for the land and ocean were 1.4 and 1.7 GtC/yr respectively.

Two other 'natural' greenhouse gases that are increasing due to human activity are methane and nitrous oxide. As shown in Table 3.2, their relative increases between pre-industrial times and 1990 are significant especially for methane which has more than doubled in concentration since preindustrial times. Even though the absolute values of the concentration of both these gases are less than one per cent of that of carbon dioxide, the radiative effects of the human-caused increases in these two gases together amount to nearly 40 per cent of that of carbon dioxide.

Table 3.2 — Characteristics of some key greenhouse gases that are influenced by human activities [a].

Parameter	*CO_2*	*CH_4*	*CFC-11*	*CFC-12*	*N_2O*
Pre-industrial atmospheric concentration (1750–1800)	280 ppmv[b]	0.8 ppmv	0	0	288 ppbv[b]
Current atmospheric concentration (1990)[c]	353 ppmv	1.72 ppmv	280 pptv[b]	484 pptv	310 ppbv
Current rate of annual atmospheric accumulation	1.8 ppmv (0.5%)	0.015 ppmv (0.9%)	9.5 pptv (4%)	17 pptv (4%)	0.8 ppbv (0.25%)
Atmospheric lifetime[d] (years)	(50–200)	10	65	130	150

(from page 320, Reference no. 1).

a. Ozone has not been included in the table because of a lack of precise data.

b. ppmv = parts per million by volume; ppbv = parts per billion by volume; pptv = parts per trillion by volume.

c. The current (1990) concentration have been estimated from an extrapolation of measurements reported for earlier years, assuming that the recent trends remained approximately constant.

d. For each gas in the table (except CO_2), the 'lifetime' is defined here as the ratio of the atmospheric content to the total rate of removal. This time scale also characterizes the rate of adjustment of the atmosphereric concentrations if the emission rates are changed abruptly. Carbon dioxide is a special case since it has no real sinks, but is merely circulated between various reservoirs (atmosphere, ocean, biota). The 'lifetime' of CO_2 given in the table is a rough indication of the time it would take for the CO_2 concentration to adjust to changes in the emissions.

3.2.2 NEW GREENHOUSE GASES

The introduction of new greenhouse gases by human activity has had a noticeable impact on the greenhouse effect, accounting for over 10 per cent of the total human impact on the greenhouse effect. Many different gases have been introduced. They are mainly halocarbons (compounds containing carbon together with halogens such as chlorine, fluorine, bromine, and iodine) such as chlorofluorocarbons (CFCs) and hydrofluorocarbons (HFCs). As stated before, these compounds were manufactured for use in refrigeration units, foaming agents and solvents.

The halocarbons are strong greenhouse gases and their lifetimes are possibly longer than those of the long-lived natural greenhouse gases. In addition, ultraviolet radiation from the sun can disassociate their molecules and release chlorine and bromine which will interact with and cause the destruction of stratospheric ozone. The potential harmful effects were so obvious that international agreements have already been put in place to reduce the production of these gases, e.g. the 1987 Montreal protocol as part of the Vienna Convention to Protect the Ozone Layer and its subsequent amendments.

A nearly-complete inventory of human impacts on the greenhouse gases is presented in Table 3.3 (pages 46–47). This list includes a large variety of halocarbons and hydrocarbons. The entire list is presented to emphasize that there are a large number of gases human impacts and to give some terminology for the halo- and hydro-carbons. Note the numbers used to identify the halocarbon types in the first column. The list does not contain data for ozone itself because much of the change in ozone is not due directly to emissions from human activity but from subsequent chemical reactions in the atmosphere. The list identifies: the lifetime of each gas; concentration change since the pre-industrial era; current growth rate; and radiative forcing due to changes since the pre-industrial era. Note that the concentrations of some of the gases are so small that no concentration is listed. Several already show reductions in concentration through compliance with international agreements.

Key information for the relative importance of the gases listed in Table 3.3 for greenhouse effects is given by the radiative forcing shown in the very last column of the table. With this information alone one can easily sort out the small subset of gases important for the enhancement of the greenhouse effect. Numbers are omitted for radiative forcing if they are less than .001 Wm^{-2}. As stated before, the radiative forcing defines the change in the amount of radiative energy transfer at the tropopause due to the changes in concentrations of the gas in the atmosphere, in this case since the pre-industrial era. A positive sign means increased energy in the downward direction which would lead to higher temperatures at the ground on average.

The gases CFC-11 and CFC-12 are responsible for most of the radiative forcing due to the new greenhouse gases. They account for a total of 0.20 Wm^{-2} or 8 per cent of the total radiative forcing (+2.45 Wm^{-2}) caused by human enhancement of the greenhouse gases. Note from Table 3.2 that the percentage rate of increase of these two halocarbons was much larger than for the naturally occurring greenhouse gases before the Montreal Protocol was enacted. All of the halocarbons together result in a radiative forcing of about 0.27 Wm^{-2} or 11 per cent of the total radiative forcing in the early 1990s.

3.3 ATMOSPHERIC AEROSOL ENHANCEMENT

3.3.1 TYPES OF AEROSOLS

Fossil fuel combustion and biomass burning are the primary sources of aerosols due to human activity. These sources produce both soot (particulate black carbon aerosols), gaseous sulphur dioxide and nitrogen oxides. The latter two are partially transformed by chemical processes into sulphate and nitrate aerosols. Additional sources for human-produced aerosols include dust from changes in land use. All of these aerosol products remain primarily in the troposphere (unlike those produced by major volcanic eruptions) and thus are subject to rapid removal by precipitation and settling processes. Table 3.4 summarizes source strength, atmospheric loading, the radiative mass extinction coefficient and radiative optical depth for the main aerosol constituents for both natural- and human-produced components.

The mass extinction coefficient is a measure of the effectiveness of radiative absorption and scattering per unit mass of aerosol, and the optical depth is a measure of the overall reduction in radiation due to absorption and scattering while passing through the atmosphere. The optical depth depends on both the mass extinction coefficient and the total mass of the aerosol in a vertical column of atmosphere. The optical depth gives a measure of the overall direct radiative forcing on the atmosphere due to the aerosol.

The human production of sulphate aerosols can be measured directly by sulphur dioxide emissions from fossil fuel combustion. The dominant source regions are in the Northern Hemisphere. This emission has increased dramatically over the past 130 years as shown in Figure 3.2. Concerns about other impacts such as acid rain have led to a levelling off of production rates in recent years in some areas such as Europe and north America.

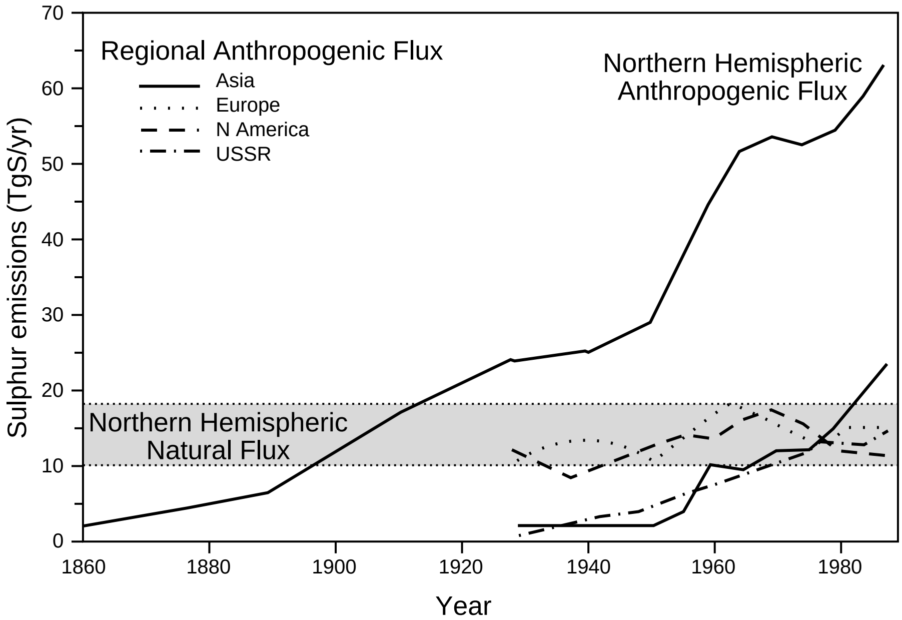


Figure 3.2 — Natural and fossil fuel combustion sources of SO_2 in the Northern Hemisphere (after Dignon and Hameed, 1989; 1992). [from page 106, Reference no. 3].

Species		Lifetime		Concentration (ppbv)		Current growth	Radiative forcing	
		Year	Uncert.	1992	pre-ind.	ppbv/yr	Wm^{-2}/ppbv	Wm^{-2}
Natural and anthropogenically influenced gases								
carbon dioxide	CO_2	variable		356 000	278 000	1 600	1.8×10^{-5}	1.56
methane	CH_4@	12.2	25%	1714	700	8	3.7×10^{-4}	0.47
nitrous oxide	N_2O	120		311	275	0.8	3.7×10^{-3}	0.14
methyl chloride	CH_3Cl	1.5	25%	~0.6	~0.6	~0		0
methyl bromide	CH_3Br	1.2	32%	0.010	<0.010	~0		0
chloroform	$CHCl_3$	0.51	300%	~0.012		~0	0.017	
methylene chloride	CH_2Cl_2	0.46	200%	~0.030		~0	0.03	
carbon monoxide	CO	0.25		50-150		~0		$
Gases phased out before 2000 under the Montreal Protocol and its amendments								
CFC-11	CCl_3F	50	10%	0.268	0	+0.000**	0.22	0.06
CFC-2	CCl_2F_2	102		0.503	0	+0.007**	0.28	0.14
CFC-113	CCl_2FCClF_2	85		0.082	0	0.000**	0.28	0.02
CFC-114	$CClF_2CClF_2$	300		0.020	0		0.32	0.007
CFC-115	CF_3FCClF_2	1700		<0.01	0		0.26	<0.003
carbon tetrachloride	CCl_4	42		0.132	0	-0.0005**	0.10	0.01
methyl chloroform	CH_3CCl_3	4.9	8%	0.135[#]	0	-0.010**	0.05	0.007
halon-1211	$CBrClF_2$	20		0.007	0	.00015		
halon-1301	$CBrF_3$	65		0.003	0	.0002	0.28	
halon-2402	$CBrF_2CBrF_2$	20		0.0007	0			
Chlorinated hydrocarbons controlled by the Montreal Protocol and its amendments								
HCFC-22	$CHClF_2$	12.1	20%	0.100	0	+0.005**	0.19	0.02
HCFC-123	CF_3CHCl_2	1.4	25%		0		0.18	
HCFC-124	CF_3CHClF	6.1	25%		0		0.19	
HCFC-141b	CH_3CFCl_2	9.4	25%	0.002	0	0.001**	0.14	
HCFC-142b	CH_3CF_2Cl	18.4	25%	0.006	0	0.001**	0.18	
HCFC-225ca	$C_3HF_5Cl_2$	2.1	35%		0		0.24	
HCFC-225cb	$C_3HF_5Cl_2$	6.2	35%		0		0.28	
Perfluorinated compounds								
sulphur hexafluoride	SF_6	3200		0.032	0	+0.0002	0.64	0.002
perfluoromethane	CF_4	50000		0.070	0	+0.0012	0.10	0.007
perfluoroethane	C_2F_6	10000		0.004	0		0.23	
perfluoropropane	C_3F_8	2600			0		0.24	
perfluorobutane	C_4F_{10}	2600			0		0.31	
perfluoropentane	C_5F_{12}	4100			0		0.39	
perfluorohexane	C_6F_{14}	3200			0		0.46	
perfluorocyclobutane	c-C_4F_8	3200			0		0.32	
Anthropogenic greenhouse gases not regulated (proposed or in use)								
HFC-23	CHF_3	264	45%				0.18	
HFC-32	CH_2F_2	5.6	25%				0.11	
HFC-41	CH_3F	3.7					0.02	
HFC-43-10mee	$C_5H_2F_{10}$	17.1	35%				0.35	
HFC-125	C_2HF_5	32.6	35%				0.20	
HFC-134	CF_2HCF_2H	10.6	200%				0.18	
HFC-134a	CH_2FCF_3	14.6	20%				0.17	
HFC-143	CF_2HCH_2F	3.8	50%				0.11	
HFC-143a	CH_3CF_3	48.3	35%				0.14	
HFC-152a	CH_3CHF_2	1.5	25%				0.11	
HFC-227ea	C_3HF_7	36.5	20%				0.26	
HFC-236fa	$C_3H_2F_6$	209	50%				0.24	
HFC-245ca	$C_3H_3F_5$	6.6	35%				0.20	
HFOC-125e	CF_3OCHF_2	82	300%					
HFOC-134e	CHF_2OCHF_2	8	300%					
trifluoroiodomethane	CHF_3I	<0.005					0.38	

Table 3.3 — Lifetimes for radiatively active gases and halocarbons. [from pages 92 and 93, Reference no. 3].

Notes:

This table lists only the direct radiative forcing from emitted gases. The indirect effects due to subsequent changes in atmospheric chemistry, notably ozone (see below), are not included. The Wm^{-2} column refers to the radiative forcing since the pre-industrial, and the Wm^{-2}/ppbv column is accurate only for small changes about the current atmospheric composition (see Section 2.4 of Reference no. 3 and IPCC, 1994). In particular, CO_2, CH_4 and N_2O concentration changes since pre-industrial times are too large to assume linearity; the formulae reported in IPCC (1990) are used to evaluate their total contribution.

A blank entry indicates that a value is not available. Uncertainties for many lifetimes have not been evaluated. The concentration of some anthropogenic gases are small and difficult to measure. The pre-industrial concentrations of some gases with natural sources are difficult to determine. Radiative forcings are only given for those gases with values greater than 0.001 Wm^{-2}.

@ Methane increases are calculated to cause increases in tropospheric ozone and stratospheric H_2O; these indirect effects, about 25 per cent of the direct effect, are not included in the radiative forcings given here.

$ The direct radiative forcing due to changes in the CO concentration is unlikely to reach a few hundredths of a Wm^{-2}. The direct radiative forcing is hard to quantify.

** Gases with rapidly changing growth rates over the past decade, recent trends since 1992 are reported.

The change in CH_3CCl_3 concentration is due to the recalibration of the absolute standards used to measure this gas.

Stratospheric ozone depletion due to halocarbons is about -2 per cent (globally) over the period 1979 to 1990 with half as much again occurring both immediately before and since; the total radiative forcing is thus now about –0.1 Wm^{-2}. Tropospheric ozone appears to have increased since the 19th Century over the northern mid-latitudes where few observational records are available; if over the entire Northern Hemisphere, tropospheric ozone increased from 25 ppb to 50 ppb at present, then the radiative forcing is about +0.4 Wm^{-2}.

Table 3.4 — Source strength, atmospheric burden, extinction efficiency and optical depth due to the various types of aerosol particles (after IPCC, 1994; Andreae, 1995; and Cook and Wilson, 1996). [from page 104 in Reference no. 3].

Source	*Flux (Tg/yr)*	*Global mean column burden (mg m^{-2})*	*Mass extinction coefficient (hydrated) (m^2g^{-1})*	*Global mean optical depth*
Natural				
Primary				
Soil dust (mineral aeorosol)	1500	32.2	0.7	0.023
Sea salt	1300	7.0	0.4	0.003
Volcanic dust	33	0.7	2.0	0.001
Biological debris	50	1.1	2.0	0.002
Secondary				
Sulphates from natural precursors, as $(NH_4)_2SO_4$)	102	2.8	5.1	0.014
Organic matter from biogenic VOC	55	2.1	5.1	0.011
Nitrates from NO_x	22	0.5	2.0	0.001
Anthropogenic				
Primary				
Industrial dust, etc.	100	2.1	2.0	0.004
Soot (elemental carbon) from fossil fuels	8	0.2	10.0	0.002
Soot from biomass combustion	5	0.1	10.0	0.001
Secondary				
Sulphates from SO_4 as $(NH_4)_2SO_4$	140	3.8	5.1	0.019
Biomass burning	80	3.4	5.1	0.017
Nitrates from NO_x	36	0.8	2.0	0.002

3.3.2 RADIATIVE IMPACTS

3.3.2.1 Direct impacts

Aerosols have a pronounced effect on solar radiation transfers in the atmosphere and a smaller one on terrestrial radiation transfers in the atmosphere. They cause scattering and absorption of solar radiation with the result that less solar energy gets to the Earth's surface, more is absorbed in the atmosphere, and more is backscattered to space. Only for the case of soot (black carbons) is there such strong absorption of solar radiation in the lower atmosphere that solar radiation backscattered to space is reduced, producing a positive radiative forcing. Aerosols also enhance absorption and emission of terrestrial radiation and thus slightly enhance the greenhouse effect. However, the effect on solar radiation is generally the dominating factor so that the net aerosol effect is estimated to be opposite to the greenhouse effect.

Since the lifetime of tropospheric aerosols is short, their concentration is not uniform over the globe; they are concentrated over the land areas. A model simulation for lower tropospheric sulphate aerosol concentrations in the 1980s is shown in Figure 3.3. (In Europe, the current values are much lower than in the 1980s). Figure 3.3 highlights the areas where another environmental effect of sulphates has occurred, the acidification of rainwater leading to 'acid rain.' Acid rain became a serious problem in places such as eastern north America, Europe and eastern Asia. The short lifetime of aerosols implies that if the human sources were turned off, the human-produced aerosol concentrations would decrease very fast, with a time scale of about a week.

In summary the overall direct radiative forcing from human-produced aerosols is negative, estimated as -0.5 Wm^{-2}. Estimates (from numerical models and not observations) for the radiative forcing for the three main components, sulphate aerosols, fossil fuel soot, and biomass burning are, -0.4 Wm^{-2}, +0.1 Wm^{-2} and -0.2 Wm^{-2}, respectively. For comparison, the estimated peak values for radiative forcing of stratospheric aerosols put there by volcanic activity at times during the last 130 years have nearly ten times the current negative radiative forcing due to tropospheric aerosols from human activity, as shown in Figure 3.4. The last major peak of roughly -3.5 Wm^{-2} was in 1991 due to the eruption of Mt. Pinatubo. However, volcano-produced negative forcing lasts only for a few years and it occurs only sporadically. Over multi-year periods, the mean volcanic forcing is comparable to anthropogenic aerosol.

3.3.2.2 Indirect impacts

Assessment of the overall impact of human aerosol enhancement is complicated by transformations that occur due to chemical processes with other gases and with the water droplets in clouds. These transformations result in 'indirect' impacts. The impact on clouds and their optical properties is discussed below.

3.4 CHANGE OF RADIATIVE EFFECTS OF CLOUDS

Since aerosol particles act as cloud condensation nuclei and freezing nuclei, they have important impacts on the radiative properties of clouds. Aerosols may change the drop-size distribution in clouds which can influence the optical properties. The magnitude of this effect depends on the properties of the aerosols present which can vary from region to region. Nucleation leading to an increase

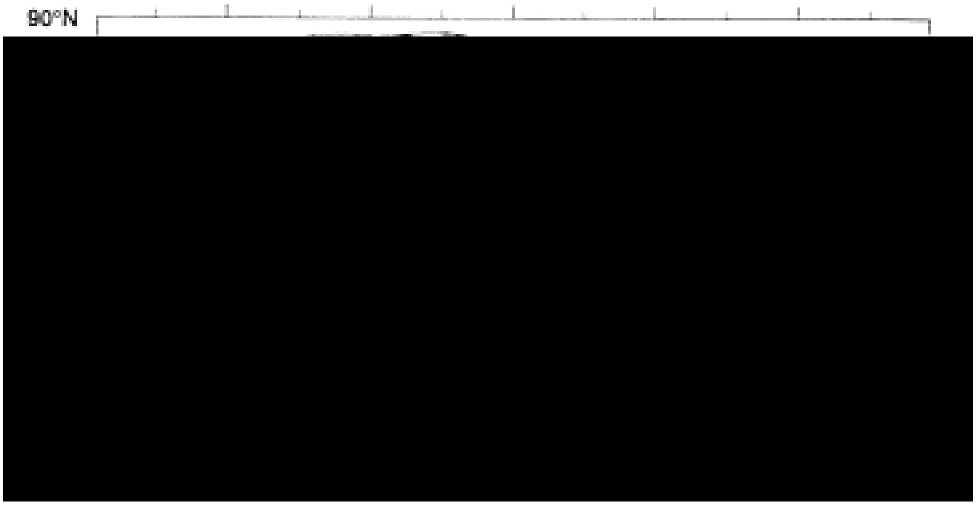


Figure 3.3 — Model simulation of the annual mean sulphate (SO_4^{2-}) concentration at 900 mb. Contours are shown at 25, 50, 100, 250, 500, 1000, and 2500 pptv. Concentrations over eastern North America and eastern Europe exceeded natural (non-anthropogenic) levels by a factor of 10. [from Langner and Rodhe, 1991, from page 328, Reference no. 1, with permission of Academic Press].

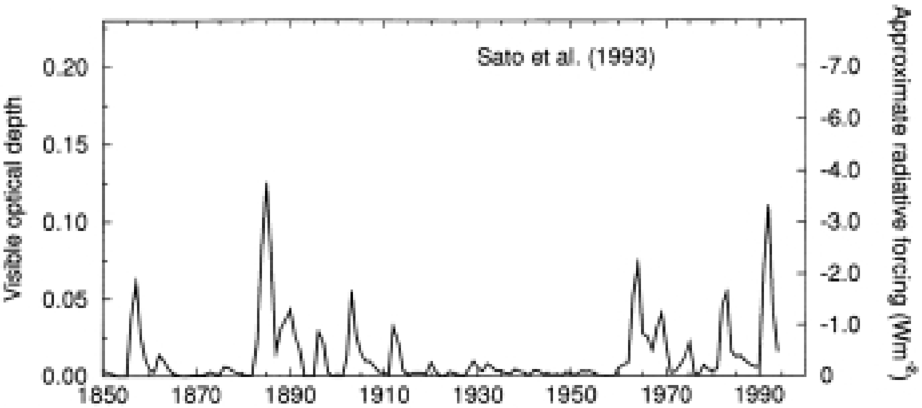


Figure 3.4 — Variation of global mean visible optical depth, and the consequent radiative forcing (Wm^{-2}) resulting from stratospheric aerosols of volcanic origin from 1850 to 1993, as estimated by Sata et al.*, 1993. The radiative forcing has been estimated using the simple relationship given in Lacis* et al.*, 1992, where the radiative forcing is -30 times the visible optical depth.* [adapted from page 116, Reference no. 3].

in cloudiness and/or the prolonging of existing clouds are other possible influences of aerosols on clouds. Effects are believed to be quite different between ice clouds and water clouds. The increase in high-level cloudiness due to jet aircraft contrails is a specific example of increase in cloudiness due to the introduction of aerosols and water vapour.

Additional observations and theoretical studies are needed before the changes in the optical properties of clouds can be quantified. At the present time, the IPCC assessment judges that, overall, the altered properties of clouds due to aerosols produced by human activity will give a negative radiative forcing on the global scale. Estimates of magnitudes range as high as -1.50 Wm^{-2} which would be much larger than the direct aerosol impact and could offset the positive forcing attributable to greenhouse gases by nearly a half.

3.5 CHANGE OF RADIATIVE PROPERTIES OF THE LAND SURFACE

It is recognized that humans have greatly altered the Earth's land surface by settlements, deforestation and agriculture. These changes have resulted in both increases and decreases in local albedo values. Replacing forests with crops that are present only part of the year would likely increase albedo effects. Urbanization could either increase or decrease the albedo depending on the reduction in trees and on the materials used for roofs and streets.

These albedo changes clearly have a role in the local energy balance and contribute to microscale climate changes such as the heat 'islands' over urban regions. However, other factors such as evaporation, precipitation and wind flow changes could have large impacts on local or even regional climate change. On a global basis the change in the radiative properties of land surfaces is not considered to be a major factor in the energy balance.

3.6 SUMMARY OF HUMAN IMPACTS

Figure 3.5 provides an overall summary of the relative importance of the many ways in which human activity can alter radiative energy transfers in the atmosphere. Confidence in the quantitative values is indicated at the bottom. The depiction includes ozone effects and, for comparison, solar variability effects due to the sun spot cycle. Note that increases in tropospheric ozone are estimated to give a radiative forcing value of +0.4 Wm^{-2} which is equal and opposite to the forcing due to tropospheric sulphate increases. The cloud change impacts due to aerosols are shown in terms of the range of estimated values of radiative forcing. Further understanding is needed to establish a mid-range estimate as is done for all the other radiative forcing components. No entry is made for volcanic effects, which are quite variable in time, or for the impacts of changes in surface albedo as these effects are considered important primarily on a regional scale. Much research is needed to reduce these uncertainties especially with respect to the indirect aerosol effect on clouds.

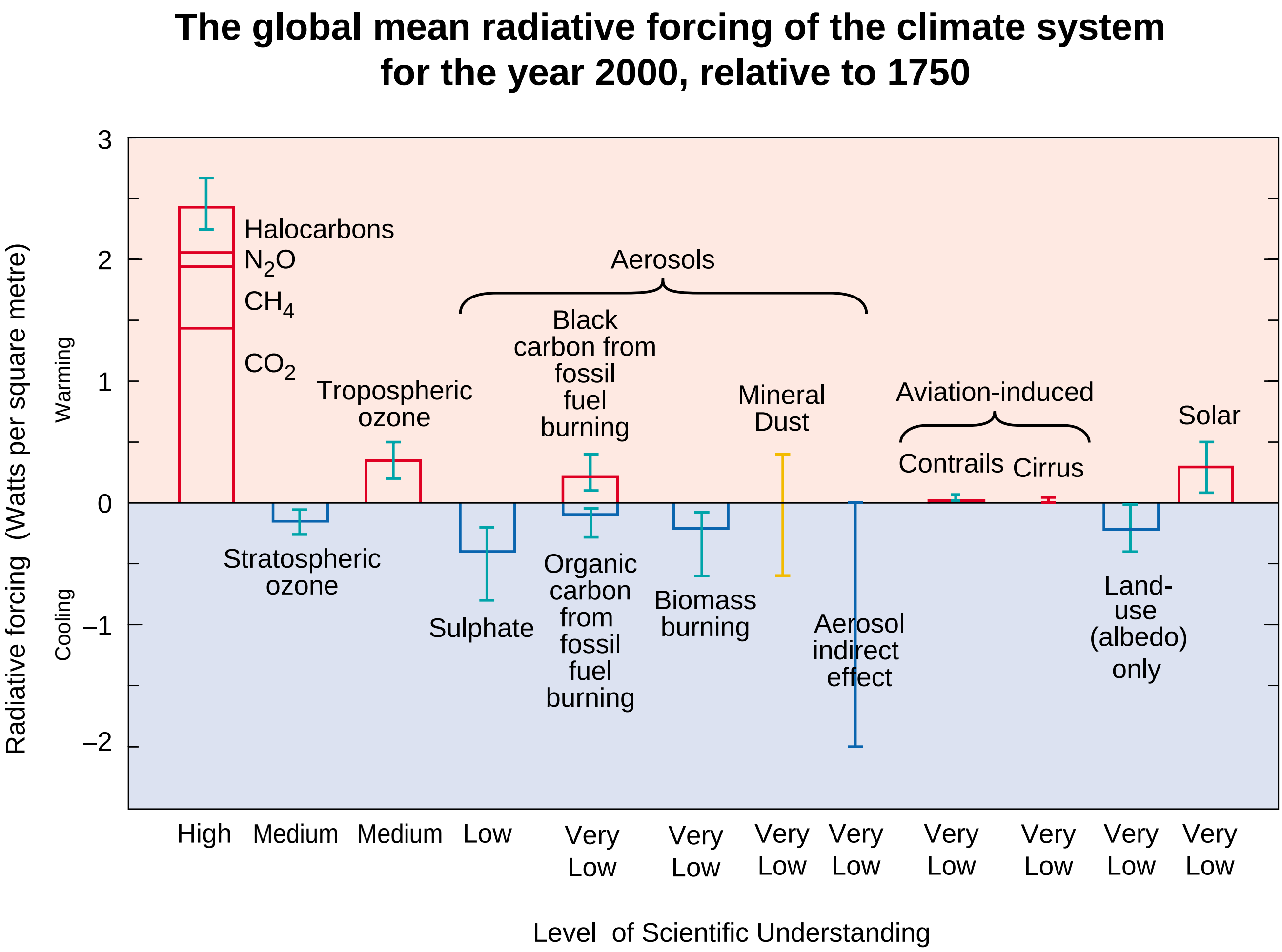


Figure 3.5 — Global, annual mean radiative forcings (Wm^{-2}) due to a number of agents for the period from pre-industrial (1750) to present (late 1990s; ~2000). The height of the rectangular bar denotes a central or best-estimate value while its absence denotes no best estimate is possible. The vertical lines capped with horizontal lines indicate an estimate of the uncertainty range, for the most part guided by the spread in the published values of the forcing. The uncertainty range specified here has no statistical basis. At the bottom a 'level of scientific understanding' index is accorded to each forcing. This represents a subjective judgement about the reliability of the forcing estimate, involving factors such as the assumptions necessary to evaluate the forcing, the degree of knowledge of the physical/chemical mechanisms determining the forcing, and the uncertainties surrounding the quantitative estimate of the forcing. The well-mixed greenhouse gases are grouped together into a single rectangular bar with the individual mean contributions due to CO_2, CH_4, N_2O and halocarbons. The sign of the effects due to mineral dust is itself an uncertainty. The forcing due to stratospheric aerosols from volcanic eruptions is highly variable over the period and is not considered for this plot. All the forcings shown have distinct spatial and seasonal features such that the global, annual-means do not yield a complete picture of the radiative perturbations. They are intended to give a first-order perspective on a global, annual-mean scale and cannot be readily employed to obtain the total response to forcings. It is emphasized that the positive and negative forcings cannot be added up and viewed a priori as providing offsets in terms of the complete global climate impact. [from page 8, Reference no. 7].

CHAPTER 4

MODELLING CLIMATE CHANGE

4.1 INTRODUCTION

This chapter has two basic purposes. The first is to summarize briefly some of the fundamentals for numerical modelling of climate. This provides literacy for terminology used in the discussion of climate models and some understanding of the numerical modelling approach including its strengths and weaknesses. The second is to present assessments of current climate model performance. This is very important as most projections for future climate change are based on numerical model solutions.

Mathematical simulation of the climate system is essential for understanding climate change outcomes from human impacts such as those described in the last chapter and for making estimates of future climate change. The climate system is very complex with many interacting components and a large number of variables and processes that need to be represented. A quantified description of the system and its changes requires obtaining solutions to a large number of governing mathematical equations.

The physical component systems are continuous in space and the scales of phenomena that exist range from global to molecular. The governing equations include partial differential and integral relationships. Many processes are not fully described in the equations such as turbulence in the atmosphere and ocean, precipitation growth in clouds, cumulus convection, radiation transfer in and around clouds, and CO_2 transfer processes in heterogeneous biosphere canopies. The complexity of the system precludes obtaining general solutions by analytical mathematical methods.

In order to obtain solutions, it is necessary to make approximations in both the governing equations and the numerical methods. Approximations in the equations are guided by understanding the relative importance of various processes represented in the equations. Computer capacity limits the spatial resolution available for the physical system in the numerical model. A climate model must represent the entire globe, and the resolution limitation means that there is a lower limit on the size of scales that can be explicitly represented in a model. Current computer capacities limit overall atmospheric resolution to the order of 100 km in the horizontal for full climate system models. The effects of all smaller scales must be parameterized, i.e. represented in terms of conditions which are resolved by the model.

The most effective climate modelling work utilizes a hierarchy of models including those which are simplified more than required by the computer system. The simpler models make it easier to isolate physical processes in the climate system and to give a general indication for overall impacts of climate change forcing. The complete models provide the best quantitative estimates of climate change on a detailed regional basis. Simple, one-dimensional or volume-integrated models of the atmosphere with prescribed ocean, cryosphere, biosphere, and land surface specifications, or with an interactive ocean using simple diffusion, radiation and convection, and energy balance relationships are important. They provide useful first indications of atmospheric changes due to human enhancements of radiatively-active constituent concentrations in the atmosphere. The model hierarchy also includes models that represent only a part of the climate system to give more details on that part. Examples are regional models which use information from the global models as external forcing conditions.

It is essential that numerical climate models be calibrated by comparing solutions with observational information. This should be done for as wide a range of variables in as many locations as possible. A numerical model may give a reasonable solution for surface atmospheric temperature, sea-surface temperature, or sea-ice coverage while at the same time its solution for precipitation over a given area can be very poor.

Obtaining and analysing the relevant observational information is a considerable undertaking. Paleoclimatologists have made great progress in meeting the challenge to make analyses of conditions from geological evidence for comparison with model predictions of past climates. For current-day analyses, observational data from many different sources are merged together with internal consistency provided by numerical model solutions. This process, termed data assimilation, is used by many operational weather prediction groups in the world today. As a result, the observational information used to compare with the model prediction solutions may not be fully independent of the model.

The basic modelling strategy for climate change prediction has the following key components. The first step is to understand the quality of model performance based on comparison of model solutions with observations of existing conditions for a variety of situations. The second step is to obtain model solutions where anticipated future forcing of the climate system due to human activity is introduced into the model. The final step is to interpret, evaluate, and establish the level of uncertainty for the model predictions based on the understanding of model performance.

4.2 BASICS FOR MODELING

4.2.1 GOVERNING PHYSICAL EQUATIONS

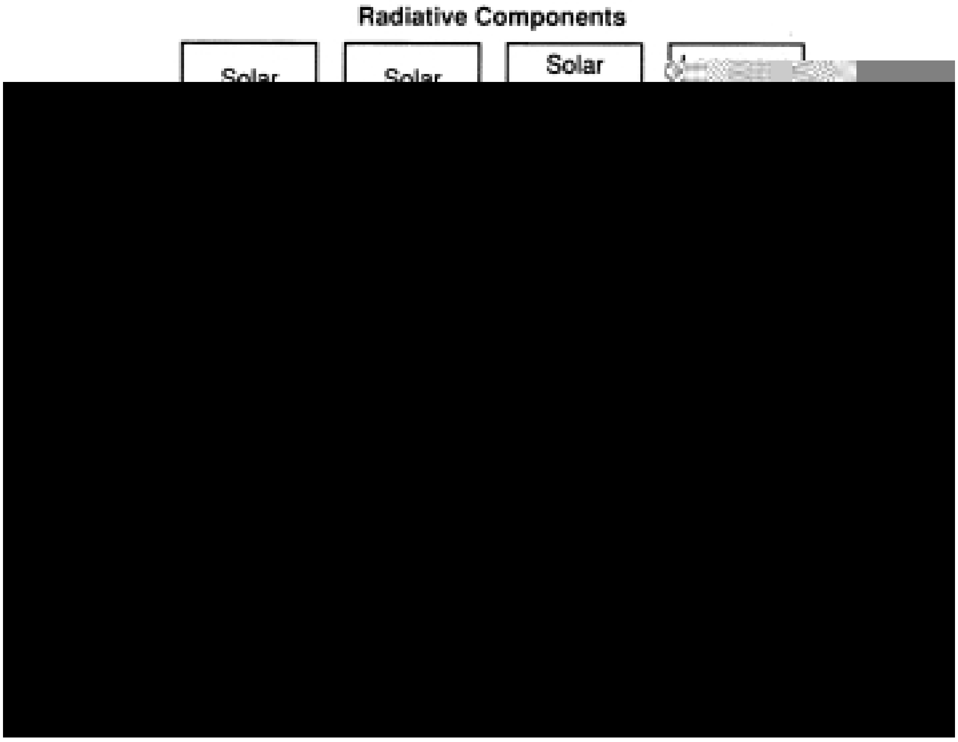
The equations used to define the climate model are those associated with each of the five components of the climate system plus additional relationships for the interactions between the components. Needless to say, the list of governing equations is extensive. Formulation of basic equations is still under way particularly for the cryosphere, biosphere and land-surface components.

Many of the equations for the atmospheric and oceanic components have been published in textbooks including the basic references listed in the introduction to these lecture notes. The governing equations include those which describe mass continuity, Newton’s laws of motion for a fluid (with the vertical component equation replaced by the hydrostatic relationship), the equation of state, first law of thermodynamics, radiation transfer and, for the atmosphere, the hydrological cycle. The equation of state for the atmosphere is that for a compressible gas including water vapour effects. The equation of state for the ocean is that for an incompressible fluid including salinity effects.

Equations for chemistry must be included with the atmospheric equations. Chemistry is very important as it describes the concentrations, life cycles, and interactions involving the greenhouse gases and aerosols, key forcing agents of the climate system. An example of the important areas and processes involving chemistry relationships in the climate system for water vapour is shown in Figure 4.1. This figure outlines the many chemical and physical interactions in the climate system. No attempt is made to summarize here the governing equations for all the chemical processes.

Equations for the land surface component in the climate system include much more detail for variations in surface composition than are used in atmospheric weather prediction models. Terrestrial biosphere conditions are closely tied to the overall land surface condition. The low heat capacity of the surface means that small changes in albedo can have large impacts on surface temperature. Thus, governing equations are needed for ground cover canopy types and coverages, and for the resultant albedo. Equations to describe details for hydrological processes are essential both for the condition of the surface biosphere and for evaporation rates into the atmosphere. Key parameters for discussion of the Earth’s surface biosphere properties are the ‘resistances’ (resistance to transfer) for water vapour and carbon dioxide. Comprehensive data on land cover types and the details of vertical structure in the terrestrial biosphere are still needed to calibrate a full set of equations for the land surface.

A land surface model must represent and quantify an intricate array of processes to define the key surface parameters (albedo, roughness length and soil moisture) which can lead to large-scale climate system feedbacks. There are numerous physical interactions related to each of these quantities from which it



[redacted]

Figure 4.1 — Schematic diagram of the physical and chemical interactions involving water vapour that might be included in a comprehensive global climate model. The key physical climate parameters are shown in boxes. Atmospheric chemical species are enclosed in elipses. Chemical mechanisms that may involve water vapour are divided into classes, identified at the bottom of the figure, and the pathways and modes of interactions are indicated by arrows. 'Photo' encompasses those chemical processes driven by solar radiation or involving the reaction of species produced when ambient air is exposed to solar ultraviolet radiation; 'hetero' refers to chemical processes occurring in aqueous solutions, principally in cloud droplets, or on particle surfaces, particularly solid aerosols and ice crystals; 'anthropo' to emissions and processes associated mainly with human activities; 'bio' refers to processes related to the assimilation and respiration of atmospheric constituents by living organisms; and 'geo' refers to chemical processes occurring on surfaces or in media at the interface between the atmosphere and the oceans and land. [from page 213, Reference no. 6, with permission of Cambridge University Press].

is possible to argue both positive and negative feedbacks. It remains a challenge to develop a land surface model that can resolve quantitatively the interaction impacts.

Representation of marine biochemical processes is important to define the gas exchanges with the atmosphere for the radiatively-active gases such as carbon dioxide, nitrous oxide, and dimethyl sulphide. Geological data document variations in the marine biosphere associated with carbon dioxide variations that have accompanied large climate changes in the past. There is a wide range of processes, including biosphere interactions, associated with the carbon cycle in the oceans. Four forms of carbon are involved: gaseous carbon dioxide, its dissolved counterpart (Dissolved Inorganic Carbon, DIC), solid organic compounds and solid inorganic compounds (carbonates). Ocean temperature and motions, and solar radiation penetrating below the ocean surface are important factors in the oceanic carbon cycle.

The cryosphere has several components that must be represented by geophysical governing equations. For glaciers and ice domes on land, there are the equations for motion that can be simplified because acceleration terms can be neglected. The motion of glaciers and ice fields is an important factor in determining coverage and total mass content for ice. Hydrodynamic equations are needed to describe the motion of sea ice as it responds to ocean currents, surrounding ice and atmospheric winds. There is an important instability hypothesis (positive feedback mechanism) for the west Antarctic ice sheet that could occur with an increase in sea-surface temperature. In this hypothesis, warming temperatures increase the slippage of ice into the ocean which raises sea level thus increasing further the slippage of ice into the ocean. It will be a challenge to represent this with deterministic equations.

The linkage between the ocean and atmosphere is central for the climate model. It is necessary to accurately represent sensible and latent energy transports. The sensible energy transport depends on turbulent boundary layer processes in both the ocean and atmosphere. Latent energy (water vapour) transport depends on the boundary layer processes only in the atmosphere. In addition, atmospheric forcing of ocean currents must be represented. This requires proper specification of wind stress at the surface — another aspect of the atmospheric boundary layer. Radiation transfer must take into account that solar radiation penetrates into the ocean, thereby distributing the heating effect into the upper layer of the ocean. The sensible and latent energy transport specifications involve approximations because of parameterization for turbulence in the boundary layer. Because of such approximations in both the atmosphere and ocean, special attention must be given to make sure that the vertical energy transports are the same on both sides of the oceanic-atmospheric interface.

4.2.2 PARAMETERIZATION OF PHYSICAL PROCESSES

The last section gave a brief overview of the large number of mathematical relationships needed for a complete climate model. It was noted that some of the relationships have yet to be developed. In the equations there are numerous processes, such as turbulence, that cannot be described because of limitations in the numerical model resolution. These processes are described as subgrid-scale processes. Parameterization is the representation of the effects of the subgrid-scale processes on the model-resolved conditions. It is based on the model variables and specified proportionality constants.

The best spatial resolution attained in current climate models is on the order of 100 km in the horizontal direction and up to 40 layers in the vertical direction in both the ocean and atmospheric components. This means that many phenomena which have important influences on the climate system cannot be fully resolved explicitly and must be parameterized. Examples of these include:

- Radiation (both solar and terrestrial);
- Cumulus convection, including thunderstorms;
- Oceanic convection;
- Surface layer (including plant canopies) momentum, water vapour and heat transfers;
- Planetary boundary layer momentum, water vapour and heat transfers;
- Turbulence in the free atmosphere and ocean;
- Precipitation processes;
- Atmospheric gravity waves over mountains and in the stratosphere;
- Weather fronts;
- Oceanic jet streams;
- Sea-ice thickness changes; and,
- Marine biosphere processes.

Improvement of parameterization is an essential part of improving climate models. Parameterization schemes may be interrelated so that changing one scheme may require simultaneously changing another to get overall improvement in the model. An example is the interconnection between cumulus convection parameterization and planetary boundary layer parameterization. Changes in parameterization can have significant effects on the simulated climate.

4.2.3 MATHEMATICS

The mathematics required for solving the equations for the current comprehensive climate system models are quite complicated in practice. Thus, many scientists as well as other persons do not have a working knowledge of such details and must rely upon specialists. Much of the modelling complexity arises from dealing with the atmospheric and oceanic components of the climate system. As noted before, the atmospheric and oceanic general circulation equations include a large number of nonlinear, partial differential equations including integrals and do not have exact analytical solutions.

Different numerical approximation approaches have been used to represent the partial differential equations as a finite set of algebraic equations that can be handled by the computer. These approaches include *finite-difference*, *spectral* and *finite-element* methods. Basic variables of the system such as pressure, temperature, wind, water vapour and radiation are prescribed by a discrete and finite set of numbers in the spatial and temporal domain of the climate system. These numbers represent values of the variables in different ways. In the finite-difference method these number are grid point values for basic variables of the system. In the spectral and finite-element methods the numbers are the amplitudes (transform values) of specified continuous functions which when added together describe the spatial variations of the basic variables. For each method of representation appropriate algebraic expressions are derived for the spatial and temporal partial derivatives and integral functions in the governing equations.

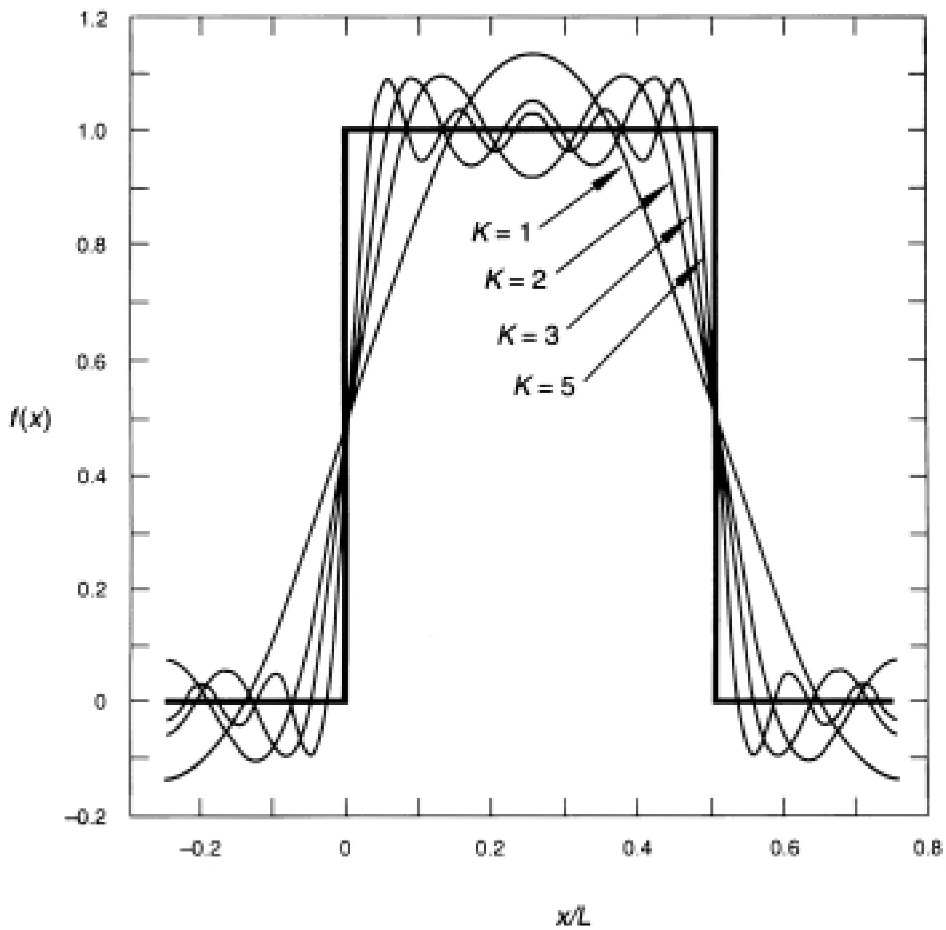
The ‘resolution’ of the model is an indicator of the number of grid point positions or specified continuous functions that are used in the model. Resolution increases as the number of the positions or functions used is increased. It is expected that the accuracy of the solution will be improved as the resolution is increased.

The *finite-difference* method is the easiest to understand and to use. Take, for example, the algebraic expression used for the first derivative. The first derivative (gradient) for a variable f in the x-direction at position x may be expressed as the difference in value between f at the grid point position on the plus side of x (position x+1) and f at the grid point position on the minus side of x (position x-1) divided by the distance between these two points. For an integral over a given spatial range the finite difference expression could be a summation of products of all f grid point values within this range, each multiplied by the increment of distance for which that value is representative. There are many choices for the finite-difference formulation depending on the accuracy desired in the approximation.

Of the other two methods, the *spectral* method is more commonly used in climate models. The concept of this approximation and definition of its resolution are illustrated in Figure 4.2. The thick line with right-angle bends, an example of a physical variable functional relationship, is represented by sine and cosine functions of appropriate amplitudes (transform values) so that the function obtained by adding them together (represented by thin lines) is as close as possible to the original function (thick line). Results are improved if one increases the resolution by increasing the number (K) of the sine and cosine functions added together. Figure 4.2 shows results using 1, 2, 3, and 5 functions, respectively.

If the square function in Figure 4.2 were considered representative of a sharp mountain on earth, note that the spectral approximation will give regions where the ground level actually is lower than anywhere in the original representation.

Figure 4.2 — Spectral approximations (thin lines) to the top-hat function (thick lines) for several different wave number truncations. [from page 296, Reference no. 6, with permission of Cambridge University Press].



In all fairness, it should be noted that the finite difference method will also have an obvious deficiency in this case. Namely, at each of the positions 0 and 0.5 on the x-axis in Figure 4.2 only one value can be used for a grid point value so that the vertical lines must be replaced by (steeply) sloping lines.

The mathematics for the solution of the spectral method equations are very complex. Originally it was not even possible to use them for accurate representation of nonlinear or local phenomena because of extensive computer requirements. However, new mathematical procedures such as the 'fast Fourier transform' and the combination of the finite-difference method with the spectral method have made it possible to incorporate the spectral method into the comprehensive climate models. The 'fast Fourier transform' requires a grid point array with grid point spacing specified according to the number of spectral components in the model. This grid is called the transform, Gaussian, or equivalent grid. Many climate models incorporate a mixture of the finite difference and spectral formulations. In these cases the resolution may be described in terms of either the number of spectral components or the spacing of grid points in the equivalent grid.

There are a number of mathematical properties of the numerical model equations that must be considered to produce a satisfactory climate model. Numerical instability is the one feature that needs to be avoided at all cost. This instability can result in very rapid and unrealistic increases in magnitudes of the model variables which render the solution meaningless. This instability arises from the mathematical formulation itself. Methods to prevent it include decreasing the size of the time steps, adopting finite difference equation formulations that have special energy-conserving properties, and adding smoothing properties to the equation system to reduce amplitudes of the solution components most likely to be unstable.

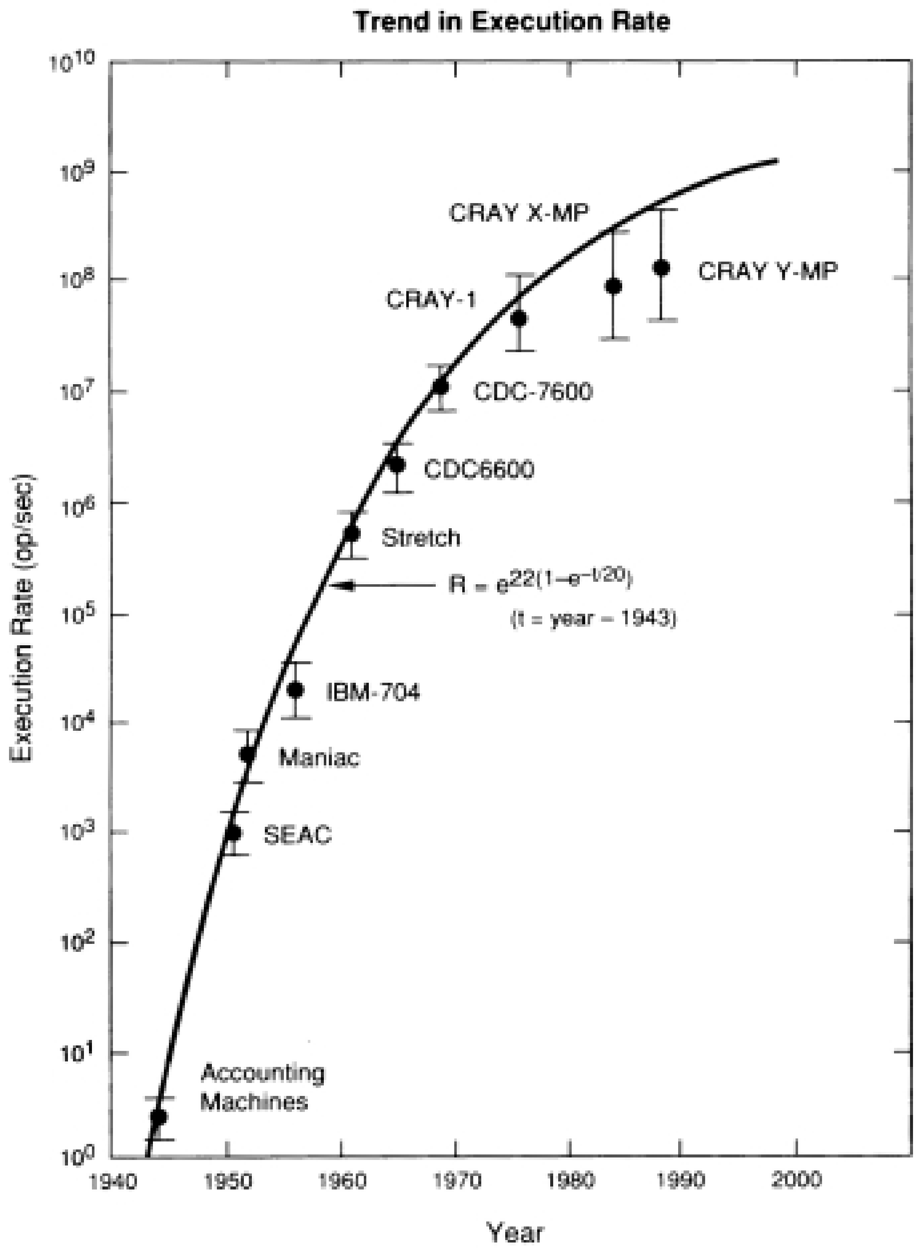
It is desirable to have mathematical formulations so that the model solutions have similar properties to those of the partial differential equations. A basic feature is the constancy of total mass and total energy expected if the only physical process operating is transport (advection due to fluid motion) within the model domain. Also, it is known that transport due to a uniform flow field will move another dependent variable feature with the velocity of the uniform flow and without distortion of the structure (shape and peak magnitude value) of the other dependent variable. Numerical model errors with respect to these two characteristics are referred to as 'phase' (position) and 'amplitude' errors.

4.2.4 COMPUTERS

Electronic computing machines which included stored-programme technology came into being in the late 1940s. This was followed by a rapid growth in the capability of single-processor machines starting in the early 1950s and still continuing today. Application to weather prediction was one motivating factor in this development. The first operational numerical weather prediction model began operation in 1955. Speed of single-processor machines, one measure of computer power, has increased by many orders of magnitude from the first electronic machines to the present time. Figure 4.3 shows the increase in electronic computer speeds from the mid-1950s to 1990. The pace of increase in computer speed has increased even more since 1990 and current (1998) single-processor machines have reached sustained speeds as high as 600 million operations per second (600 megaflops) [Hammond, 1998, private communication].

Advances in computer power continued in the 1980s with the development of parallel-processor vector machines wherein single-processor machines were combined to work on parts of the same problem simultaneously. This interconnection capability has made it possible to achieve computer system speeds as high as 20 billion operations per second by 1998. Projects are now under way in the United States and Japan to increase computer system peak speeds to 32 trillion operations per second (32 teraflops) by the year 2004 [private communication H. Grassl,1998 with strategy document for ECMWF 1999-2008]. At the same time communications and networking capabilities have led to systems that may be physically separate but can combine to give large computer capability.

Figure 4.3 — Trend in single processor computational performance up to 1990 (after Worlton, 1987, personal communication). [from page 288 , Reference no. 6, with permission of Cambridge University Press].



General circulation modelling became an important use of computer power by 1960. Expansion of such models to climate models and the subsequent growth of climate modelling work has continued the pressure to expand computer power even further.

4.3 CURRENT CLIMATE MODELS AND THEIR PERFORMANCE

4.3.1 INTRODUCTION

Both simple and comprehensive numerical climate models have been used extensively in climate change research. Simple climate models based on (oceanic) Upwelling Diffusion and Energy Balance (UD/EB) have played a key role in sensitivity studies for climate change. These UD/EB models represent the atmosphere and ocean by several large domains respectively. Mean temperatures are found which balance the energy transfers between the domains for a prescribed radiative forcing in the atmosphere. Energy transfers within the ocean domains are represented by simple upwelling and diffusion processes. Descriptions of the UD/EB model used in climate change studies appear in publications by Wigley and Raper (1987, 1992).

It has been possible to get solutions for global mean temperatures in the UD/EB model that correspond to comprehensive climate model results by adjusting the structure and parameter values in the UD/EB model. Since the UD/EB models have much smaller computer requirements than the comprehensive models, it has been possible to perform a large number of sensitivity experiments to quantify the range in global temperature changes due to uncertainties in radiative forcing effects and future concentrations of aerosols and greenhouse gases. These results have been comparable to the results for global mean temperatures from the comprehensive climate models.

Comprehensive climate models containing representation of both the ocean and atmosphere have been developed by over two dozen research groups worldwide for climate change research and assessments. These models differ in details of model formulation including resolution and parameterization. The representation of the ocean is comprehensive including currents, temperature and salinity distributions, sea ice, turbulent mixing processes, radiation transfer and bottom topography. Systematic efforts have been made to compare model simulations with current and past observed conditions and with each other. Such model intercomparisons help us to isolate and understand the impact of different model approaches and to document the range of uncertainty in climate model simulations.

Three kinds of model evaluation have been made. The first is an evaluation of the overall full climate model where the components are interactive. The second approach is to examine individual components of the climate model. The third is to look at the sensitivity of the model results to its formulation, boundary conditions and parameterizations. These three approaches are presented in the next three sections. The final part of the section on model sensitivity summarizes the model factors which are currently believed to be most responsible for the uncertainty in climate model simulations.

4.3.2 OVERALL CLIMATE MODEL EVALUATION

4.3.2.1 Current climate conditions

Sixteen comprehensive climate models presented in the 1995 IPCC report included full two-way interactions between the atmospheric and oceanic components have provided a basis for overall climate model evaluation and intercomparison. These evaluations have been based on the climate produced by the models, all with the same forcing parameters corresponding to current conditions. Extended time simulations ranging in length from 100 to 1000 years have been made to establish what was considered to be equilibrium climate conditions. Model results have been compared to observations and to each other.

The sixteen groups are listed in Table 4.1 along with a few key descriptors for each model such as:

(*a*) Resolution of the atmospheric general circulation model component (AGCM).

Terminology for horizontal resolution is given by the latitude-longitude grid point spacing for the grid point models and with numbers representing the number of spectral components for the spectral models (R and T refer to rhomboidal and triangular, respectively, descriptors for the details of the spectral components). The vertical resolution is given by the number of layers written after 'L.' Note that the model summary given in the 2001 IPCC report (Reference no. 7) shows improved resolution for many of these models.

(*b*) Resolution of the Ocean General Circulation Model Component (OGCM).

Terminology is the same as for the atmospheric component. (Note: only grid point formulations are used since the spectral approach is not practical for a domain broken up by continents.).

(*c*) Descriptions for the sea-ice and land-surface components.

(*d*) Indication if 'flux corrections' are used at the ocean-atmosphere interface.

The use of flux corrections has been necessary for some of the models to offset discrepancies in the vertical momentum, heat, and water vapour fluxes at the ocean-atmosphere interface which could cause an unnatural drift away from observed climate conditions. It is anticipated that the need for flux corrections will decrease as models are further improved.

A number of variables from these climate simulations are compared with observations and with each other in the group of the 11 climate models that had completed the simulation at the time of the 1995 IPCC report. The variables are surface air temperature, precipitation, Northern Hemisphere snow cover, sea-ice cover, mean sea-level pressure, surface heat flux over the ocean, and strength of the north Atlantic ocean thermohaline circulation. Comparisons are made in

Table 4.1 — Coupled model control simulation for current climate. [from page 236, Reference no. 3].

Group	*Model No.*	*Country*	*AGCM Resolution*	*OGCM Resolution*	*Sea Ice*[@]	*Flux correction*[+]	*Land-surface*[$] *scheme*	*Initial state*[^]	*Notes*[§]
BMRC	1	Australia	R21 L9	3.2° × 5.6° L12	T	none	B	E	d
CCC	2	Canada	T32 L10	1.8° × 1.8° L29	T	H, W, T	BB	E	d
CERFACS	3	France	T42 L31	1° × 2° L20		none	B	E	
COLA	4	USA	R15 L9	3° × 3° L16	T	none	Cr_S (SSiB)		d
CSIRO	5	Australia	R21 L9	3.2° × 5.6° L12	T/R	H, W, τ, T	Cr_S (CSIRO)	E	d
GFDL	6	USA	R30 L14	2° × 2° L18	T/Dr	H, W	B	E	d
GISS	7	USA	4° × 5° L9	4° × 5° L13	T	none	C	I	d (1)
GISS	8	USA	4° × 5° L9	4° × 5° L16	T	none	C	E	(2)
IAP	9	China	4° × 5° L2	4° × 5° L20	T	H, W	B	E	NPOGA
LMD/OPA	10	France	3.6° × 2.4° L15	1° × 2° L20	T	none	B (SECHIBA)		
MPI	11	Germany	T21 L19	5.6° × 5.6° L11	T	H, W, τ, T	B	E	d E1/LSG
MPI E2/OPYC	12	Germany	T21 L19	2.8° × 2.8° L9	T/R	H, W, τ, T	Cr_S (ECHAM)	E	d
MRI	13	Japan	4° × 5° L15	(0.5–2°) × 2.5° L21	T/Dr	H, W	B	E	d
NCAR	14	USA	R15 L9	1° × 1° L20	T/R	none	B	I	d
UCLA	15	USA	4° × 5° L9	1° × 1° L15		none	fixed wetness		NPOGA
UKMO	16	UK	2.5° × 3.8° L19	2.5° × 3.8° L20	T/Dr	H, W	Cr_S (UKMO)	U	d

@ T refers to 'thermodynamic' and R to 'dynamic' sea ice with rheology. Dr stands for 'free drift' sea ice.

\+ H, W, τ, T stand for flux adjustment of heat, fresh water, surface stress and ocean surface temperture, respectively.

$ B refers to a 'simple bucket"; BB refers to 'modified bucket'; C includes canopy processes; r_s denotes inclusion of stomatal resistence. The name of the land-surface scheme is given in parentheses.

^ The method of initialising the coupled model is indicated by E for an equilibrium of the coupled system, U for equilibrium of the upper ocean and I for initial conditions specified from available observations.

§ NPOGA: stands for 'no polar ocean with global atmosphere' and reflects models which either exclude the polar oceans and ice by specifying climatological values or which control polar deep ocean quantities with a relaxation to observed values. E1 and E2 refer to the ECHAM1 and ECHAM2 AGDMs, (1) and (2) refer to two versions of the GISS model, and LSG and OPYC refer to ocean models. d indicates that data from this stimulation are included in Reference no. 3.

terms of global means or global coverage, spatial distributions, and zonal averages. Mean values were shown for all variables. Also examined is surface temperature variability on monthly, seasonal, interannual and decadal time scales.

Some comparisons of these climate descriptors are shown in the following table and two figures with information on observed values: Table 4.2 for global average surface air temperature and precipitation for December-February and June-August; Figure 4.4 for zonally-averaged surface temperature fields for December-February and June-August; and Figure 4.5 for zonally-averaged mean precipitation fields for December-February and June-August.

These figures illustrate the differences between models. For global mean temperature, differences are on the order of several degrees Celsius (Table 4.2) which is in the same range as the overall temperature increases discussed for global climate change. For zonally-averaged surface temperature, model errors tend to be larger at higher latitudes (Figure 4.4).

Global mean precipitation simulation values vary from observed values by up to 30 per cent (Table 4.2). Spatial variations in model-simulated precipitation differences from observed values tend to be largest in the tropical latitudes.

For the models using surface heat flux corrections over the oceans, the typical values required to give equilibrium consistent with current climate conditions were on the order of tens of watts per square metre. These values are larger than any associated with radiative forcing associated with climate change. Since 1995 great progress has been made to reduce or eliminate the need for surface heat flux corrections (Reference no. 7).

These examples show that model errors and variability between models in simulations for current climate exceed changes expected with climate change. The proper use of models for climate-change determination is to examine the changes in model climate when radiative and other forcing associated with climate change are added, instead of comparing the model prediction of future climate directly to current observed climate. It is assumed that as long as the overall climate conditions simulated by the model are close to observed conditions, then the modelled changes will be valid indicators for climate change in the real world.

The variability in the model-simulated climates has also been examined. This is an important but more difficult aspect of the climate system to simulate. In one study it was shown that the modelled intermonthly standard deviation of lower tropospheric temperature was larger than observed in the tropics and smaller than observed in the southern high latitudes. In another study it was found that the model simulation of the ENSO oscillations tended to underestimate the magnitudes of the oscillations. Decadal variability also has been found in model simulations for current climate conditions having amplitude increases with latitude consistent with observations.

Table 4.2 — Coupled model simulations of global average temperature and precipitation. [from page 238, Reference no. 3].

	Surface air Temperatures (°C)		*Precipitation (mm/day)*	
	DJF	*JJA*	*DJF*	*JJA*
BMRC*	12.7	16.7	2.79	2.92
CCC	12.0	15.7	2.72	2.86
COLA	12.6	15.5	2.64	2.67
CSIRO	12.1	15.3	2.73	2.82
GFDL	9.6	14.0	2.39	2.50
GISS (I) *	13.0	15.6	3.14	3.13
MPI (LSG)	11.0	15.2		
MPI (OPYC)	11.2	14.8	2.64	2.73
MRI	13.4	17.4	2.89	3.03
NCAR*	15.5	19.6	3.78	3.74
UKMO	12.0	15.0	3.02	3.09
Observed	12.4	15.9	2.74	2.90

* models without flux adjustment

Here the observed surface air temperature is from Jenne (1975) and the observed precipitation from Jaeger (1976).

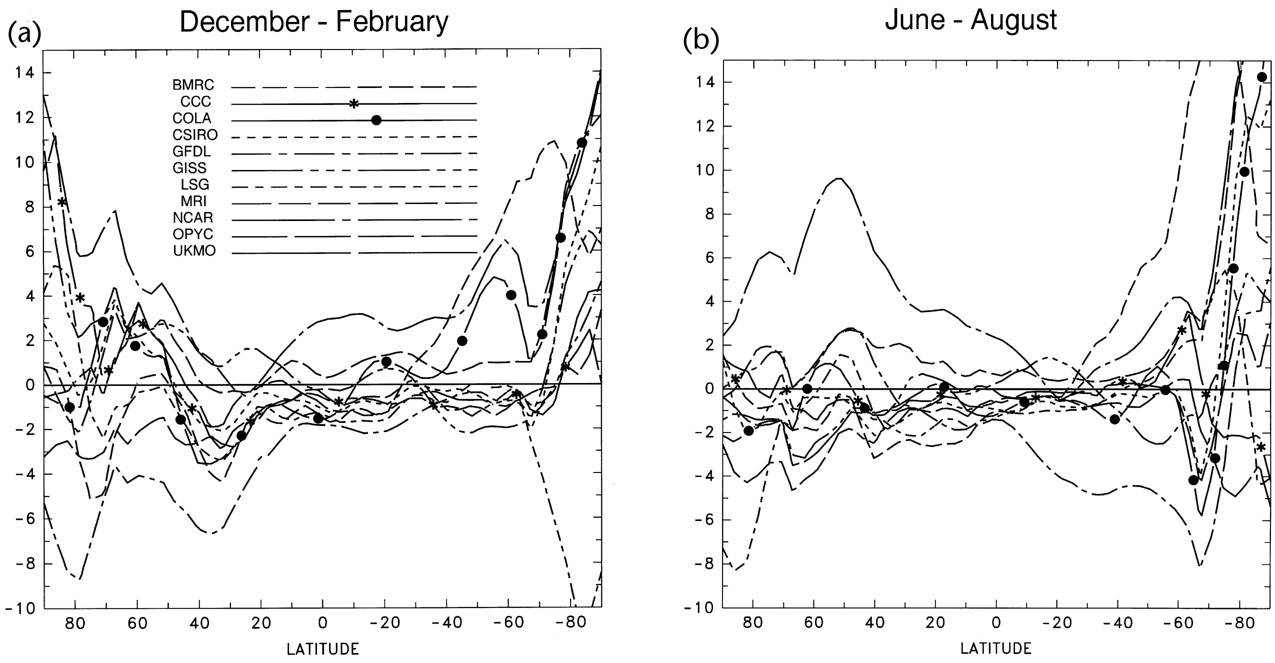


Figure 4.4 – (a) The zonally-averaged difference of 11 coupled models' surface air temperatures from observations for December-February; (b) as in (a) but for June-August. Units °C. [from page 239, Reference no. 3].

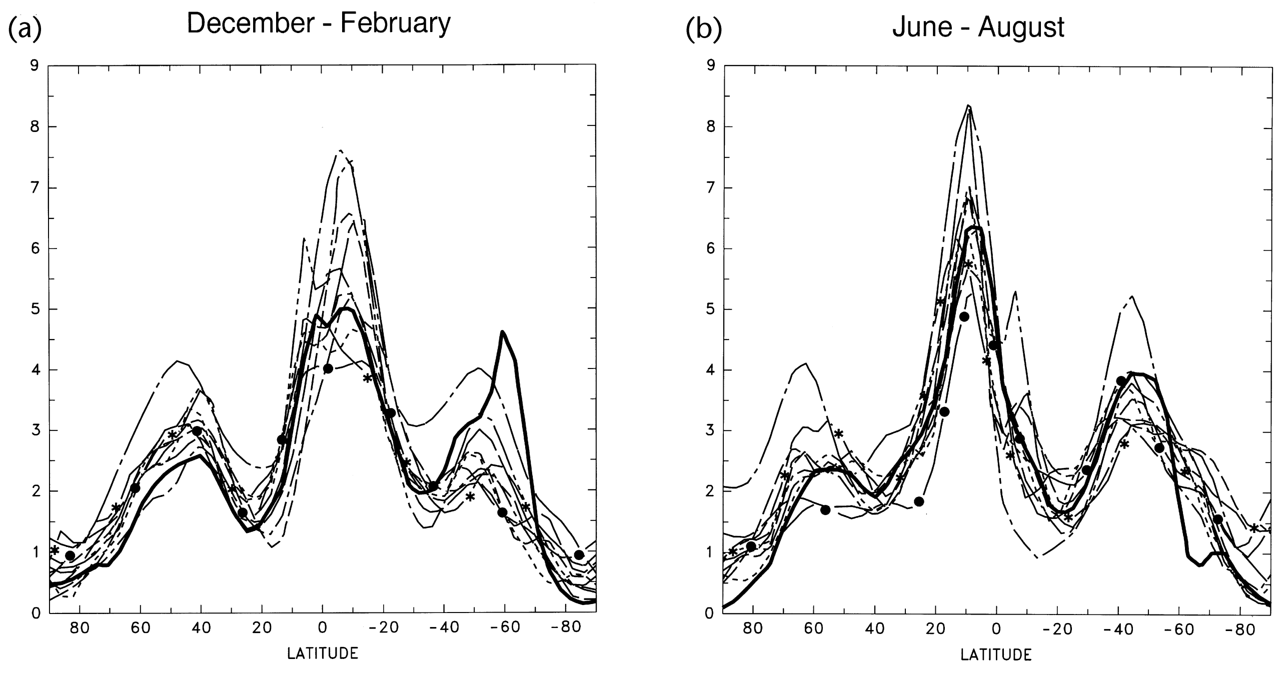


Figure 4.5 – (a) The zonally-averaged precipitation rate from 11 coupled models and that from observations according to Jaeger (1976) (solid line) for December-February; (b) as in (a) but for June-August. Units mm/day. (See Fig 4.4(a) for model identification.) [from page 241, Reference no. 3]

In conclusion, as already judged in the 1995 IPCC report (Reference no. 3), climate models have been shown to simulate satisfactorily as a group the large-scale features of current climate. The IPCC report also concludes that the: "different coupled models simulate the current climate with varying degrees of success, and this affects the confidence that can be placed in their simulations of climate change."

4.3.2.2 Past climate conditions

Testing climate models on past climate situations (paleoclimate) makes it possible to evaluate model performance for a wider range of conditions of forcing of the climate system. Chapter 2 reviewed variability of past climates in terms of surface atmospheric temperature. Changes since the last glacial maximum (about 20 000 years ago) have been documented from ice cores, tree rings, fossil pollen, mountain glaciers, ancient soils, closed-basin lakes, and sediments in lakes and oceans. These records have provided enough information for climate in terms of time changes of spatial patterns for temperature and precipitation so that validation of the general comprehensive climate models is possible.

The general approach has been to examine model-simulated equilibrium climates applying fixed forcing conditions for given times in the past. By this means the evolution of climate changes corresponding to changes in the forcing conditions has been examined. It has only been possible to evaluate the general climate regimes produced by the model and not the details of processes and local variations because of the limitations in the paleoclimatic data.

The validation of climate models for paleoclimate situations has proven to be useful for testing and understanding climate models. This has led to the use of more sophisticated climate models which include interactions with simple 'mixed-layer' oceans and the terrestrial biosphere. The establishment of a comprehensive Paleoclimate Modelling Intercomparison Project (PMIP) will provide even more understanding of model performance in the future.

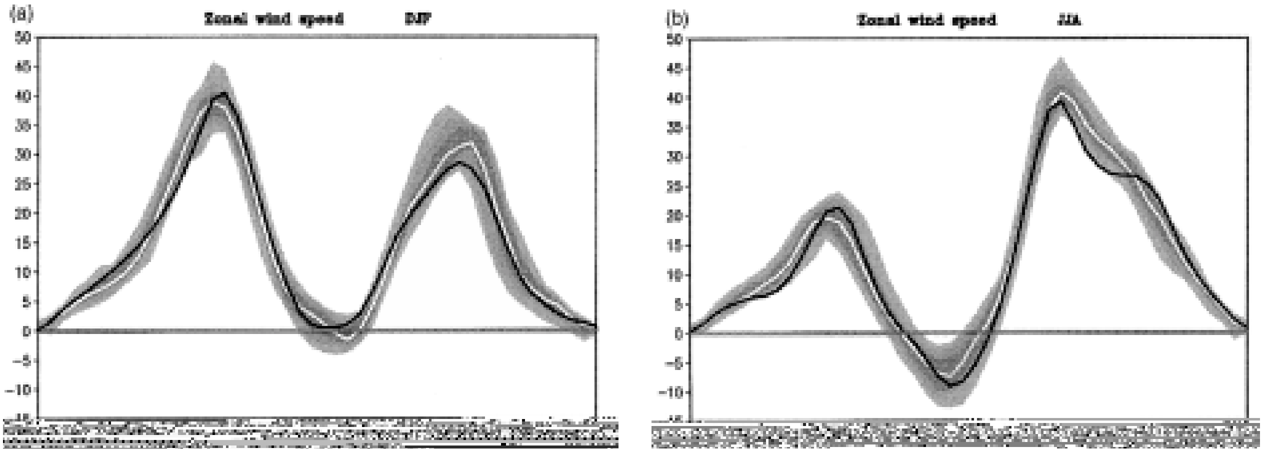
4.3.3 EVALUATION OF CLIMATE MODEL COMPONENTS

4.3.3.1 Atmospheric component

Modelling of the atmospheric component is the most comprehensive and developed of all the climate system components. Atmospheric general circulation models have been developed over a 40-year period in connection with weather prediction and climate modelling. More than two dozen model versions exist, and these have been subject to extensive intercomparison such as in the international Atmospheric Model Intercomparison Project (AMIP) [Gates, 1992]. In the simulations, sea-surface temperature is normally prescribed with climatological mean annual cycle values. This means that surface atmospheric temperature over the oceans is constrained to be close to climatological values because of the strong influence of ocean temperature on the temperature of the atmospheric surface layer.

Model simulations for surface air temperature over land, precipitation, and sea-level pressure have the same general quality as obtained in the coupled climate models discussed in the last section. Simulated tropospheric temperature is close to that observed, but there is a tendency for the models to be too cold at lower levels in the troposphere in the tropics and too warm in the tropical lower stratosphere. The associated zonally- and seasonally-averaged zonal winds at the upper troposphere represent the observations well. The mean zonal wind values found by averaging all the model results together have errors of less than 5 ms^{-1} at all of the latitudes including the winter hemisphere latitudes where the jet stream magnitudes reach 40 ms^{-1}. There are differences among the models for jet stream speeds at given latitudes also in the order of 5 ms^{-1}. Figure 4.6 shows the observed and AMIP-model distribution percentiles for the December-January and June-August zonal wind speed mean values.

Figure 4.6 — The zonally-averaged zonal wind (ms^{-1}) at 200 hPa as observed (black line) and as simulated by the AMIP models for (a) DJF and (b) JJA. The mean of the models' results is given by the full white lines, and the 10, 20, 30, 70, 80 and 90 percentiles are given by the shading surrounding the model mean. The observed data are from Schubert et al. (1992). [from page 250, Reference no. 3].



Model simulations for cloudiness are not so good. Cloudiness is an important component in a climate model because of its effects on radiation transfer. On a zonally- and seasonally-averaged basis, errors in cloud cover and differences between models have magnitudes in the order of 20 per cent. Generally the models tend to underestimate the cloudiness in the winter and summer seasons in low and middle latitudes, and overestimate cloudiness in polar regions.

The quality of the simulated radiation transfers is a key to the overall climate of the atmosphere and the capability of the atmosphere model to simulate properly the effects of climate change forcing.

An important measure is the amount of terrestrial radiation emitted to space from the atmosphere-Earth system. In terms of zonally- and seasonally-averaged values, the models simulate outgoing terrestrial radiation values that generally differ by no more than an order of 10 Wm^{-2} from observations at any latitude. It is noted that the models are able to represent the local minimum of outgoing terrestrial radiation values near the equator. This minimum results from the effects of deep convective clouds in tropical areas. This convective activity is an important component for the general circulation of the atmosphere.

The overall effect of the clouds on the radiation budget of the atmosphere is a key factor in the climate system. Clouds affect both the solar and terrestrial radiation transfer, and reflect solar radiation which leads to a reduction of input energy; whereas they absorb terrestrial radiation which leads to a reduction in output energy. These effects offset each other so that the net radiative forcing due to clouds can be positive or negative.

Model simulations for the net radiative forcing due to clouds are compared to observations in Figure 4.7. The overall latitudinal structure is reproduced by the models fairly well. However, differences at individual latitudes such as those: 1) between the average of the model results and observed conditions; and 2) among the model results typically range between 10 and 20 Wm^{-2}. Global and annual mean values of the differences are less because of offset between plus and minus values. Nevertheless, values are comparable to those for global mean radiative forcing due to the greenhouse gases. This shows that cloud cover is an important factor for model simulations for climate. The most obvious discrepancy in model performance is the systematic underestimation of negative radiative forcing due to clouds in the tropical and subtropical latitudes (30°S to 30°N) in both the December-February and June-August periods and the overestimation of the negative forcing near 60° latitude in the summer hemisphere.

Overall the atmospheric general circulation model simulations show realistic variability for atmospheric surface temperature. The observed and simulated diurnal ranges in surface temperature over land are similar, although simulated values are too large over the high northern latitude areas in January and over the northern continents and deserts in July. The simulated annual cycle for surface temperature over the continents is generally reasonable although there is an overestimation of the seasonal amplitude in the drier

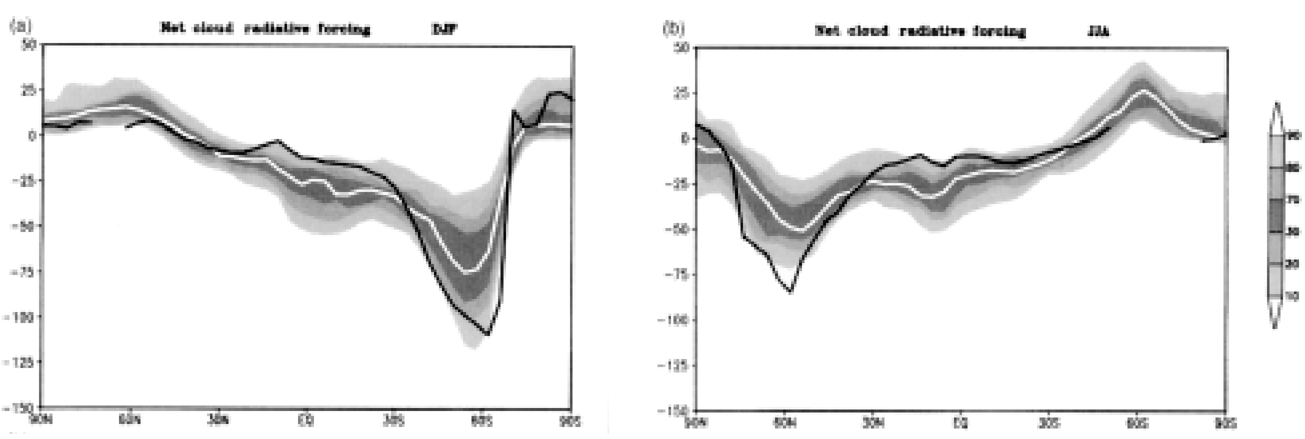


Figure 4.7 — As in Figure 4.6 except for net cloud radiative forcing (Wm^{-2}). The observational estimates are from ERBE data for 1985-88 (Harrison et al., 1990). [from page 251, Reference no. 3].

climate regions. There are large differences among the models for the seasonal amplitudes in the higher latitudes.

Variability associated with processes internal to the atmosphere is also present in the models but with amplitudes smaller than observed. This is true both for extratropical variability due to synoptic storm activity and for the Madden-Julian 30-60 day oscillation in the tropics.

Finally, the recent trends in surface temperature have been approximately represented in atmospheric model simulations for the past 45 years using observed values of sea-surface temperature and, in some cases, observed carbon dioxide and tropospheric aerosol changes. The results obtained by five modelling groups for mean surface temperature over land areas compared to each other and to observations are shown in Figure 4.8.

4.3.3.2 Ocean component

Ocean general circulation models have been developed by a number of research groups. The highest resolution versions of these models have horizontal resolutions as small as 20 km and vertical resolution up to 60 layers (as of 1995). At present, computer size limits the resolution that can be used in coupled ocean-atmosphere models.

Ocean general circulation models have been able to simulate interannual variability of the type related to ENSO. In the coupled ocean-atmosphere models the ENSO-like variability is much weaker and much more regular than in observed conditions. Some ocean general circulation models have been able to simulate the Rossby waves and equatorial Kelvin waves that are part of the ENSO variability cycle. The newest coupled models show ENSO rather near to reality both in frequency and intensity. Decadal variability has also been simulated in ocean basins where deep water formation occurs (due to downward convective motions of cold water).

As concluded in the 1995 IPCC report [Reference no. 3], ocean general circulation models are able to portray realistically the large-scale structure of the oceanic gyres and the main features of the thermohaline circulation. Primary deficiencies are in the representation of mixing processes and the structure and strength of the western boundary currents, the simulation of meridional heat transport, and the simulation of convection and subduction. Accordingly, a number of problem areas and deficiencies remain to be addressed, these include:

(*a*) Representation of geometry and bathymetry;

(*b*) Parameterization of subgrid-scale processes as convection, mixing, and mesoscale eddies;

(*c*) Errors in defining surface forcing by the atmosphere;

(*d*) A thermocline that is too deep and too diffuse;

(*e*) Weak poleward heat transport;

(*f*) Distortion of upper ocean and deep boundary currents; and

(*g*) Temperature and salinity errors in deep water.

An important challenge, adding to the list above, for further improving and understanding ocean general circulation modelling is obtaining observational data for the oceans. Observations for the ocean are insufficient in coverage especially in the deep waters for comprehensive description of currents and thermohaline structure and for their variability. In many parts of the ocean the observational record is insufficient to define the key decadal and longer-term variability characteristics.

4.3.3.3 Land-surface component

The land-surface component is dependent on the parameterization for a number of processes that involve water vapour, evaporation, liquid water distribution and storage, soil moisture, heat and momentum transfer, ground temperature, and radiative exchanges. The parameterizations depend in part on resolved variables in the atmospheric model, so that deficiencies in the atmospheric model will influence the performance of the land-surface parameterizations.

Differences between parameterization methods have been shown by the Project for Intercomparison of Land Surface Parameterization Schemes (PILPS). In this project 20 land-surface models were compared for computed values of surface heat and moisture fluxes and liquid water runoff and drainage. All of the models

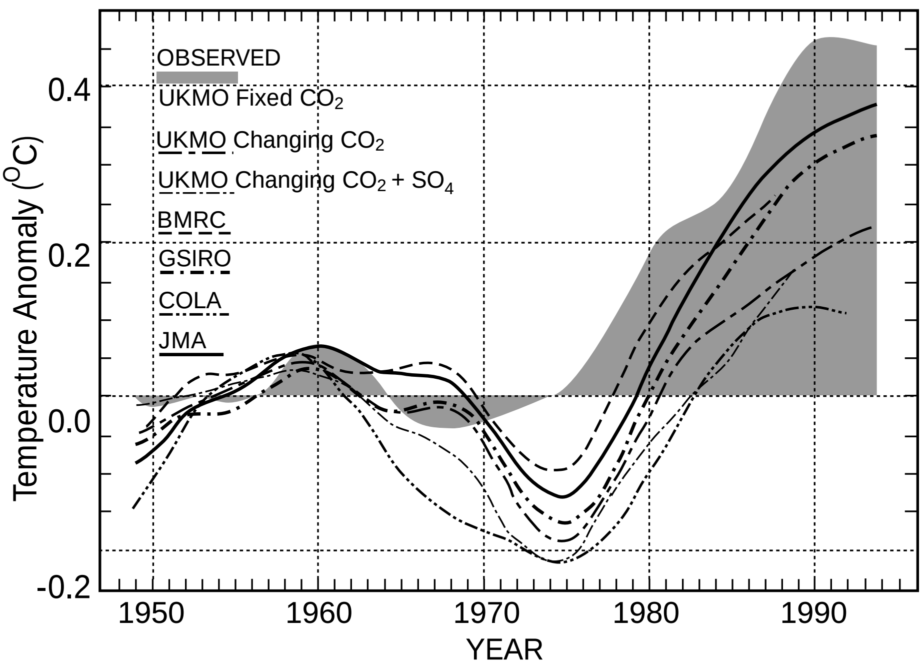


Figure 4.8 — The observed global annual 1.5 m land air temperature for 1949 to 1993 from the Jones (1994) dataset (shaded) and the corresponding modelled 1.5 m air temperature deviations from the 1950 to 1959 average for: (i) the average of four simulations with the UKMO model without progressive changes in radiative forcing; (ii) the average of four simulations with changing CO_2; (iii) the average of four models with a representation of tropospheric aerosols; and (iv) the averages from four other models with no changes in radiative forcing. [from page 258, Reference no. 3].

were subjected to the same observed atmospheric forcing inputs for rainfall and surface radiation, and for low-level temperature, moisture and winds. One year of atmospheric observations at Cabauw, the Netherlands, was used for the forcing inputs and for verification of the land-surface quantities computed. Each model was run for as many annual cycles as necessary to obtain a near-constancy for the annual averages of the computed quantities.

General results are shown in Figure 4.9. The variations have certain correlations. For instance, an increased sensible heat flux relates to decreased latent heat flux implying that the total of sensible plus latent heat flux has less variation. Similarly, the increase in runoff relates directly to decreases in evapotranspiration. Nevertheless, the range in sensible and latent heat fluxes individually was on the order of 20 Wm^{-2} with values roughly centered on the observed amounts. The range in runoff and evaporation amounts was in the order of 300 mm (for a year). Note that for the runoff and evapotranspiration, results are included for several cases where the land surface model was forced by results from atmospheric models (referenced AMIP/PILPS). The generally wider spread of the later points illustrates the error contribution from the atmospheric model.

Another intercomparison experiment was made for the annual cycle of soil moisture using 13 land-surface models and observational data from an experimental site in southern France. The range for computed soil moisture amounts among the simulations was about 100 mm. The range was the greatest in the summer when the schemes tended to underestimate soil moisture.

4.3.3.4 Cryosphere component (sea-ice models)

Sea-ice models have been developed that can simulate sea-ice thickness and coverage. An example of capability is shown in Figure 4.10. In current climate models, only very approximate sea-ice models have been used. Major simulation discrepancies exist in climate models such as the ice edge being too far south in the Atlantic Ocean and not far enough north in the southern ocean.

It is necessary to develop more comprehensive sea-ice models for climate modelling. Sea-ice affects ocean salinity and oceanic convection that sends surface water down to the ocean depths. Furthermore, sea-ice transport results in important heat transport. There is some evidence from preliminary studies that incorporating sea-ice dynamics reduces the climate (change) sensitivity of the climate model. Recently the Sea-Ice Model Intercomparison Project (SIMIP), which was undertaken within the Arctic Climate System Study (ACSYS) has completed an intercomparison of sea-ice models and promoted a 'viscous-plastic' sea-ice model for climate models [Lemke *et al.*, 1997].

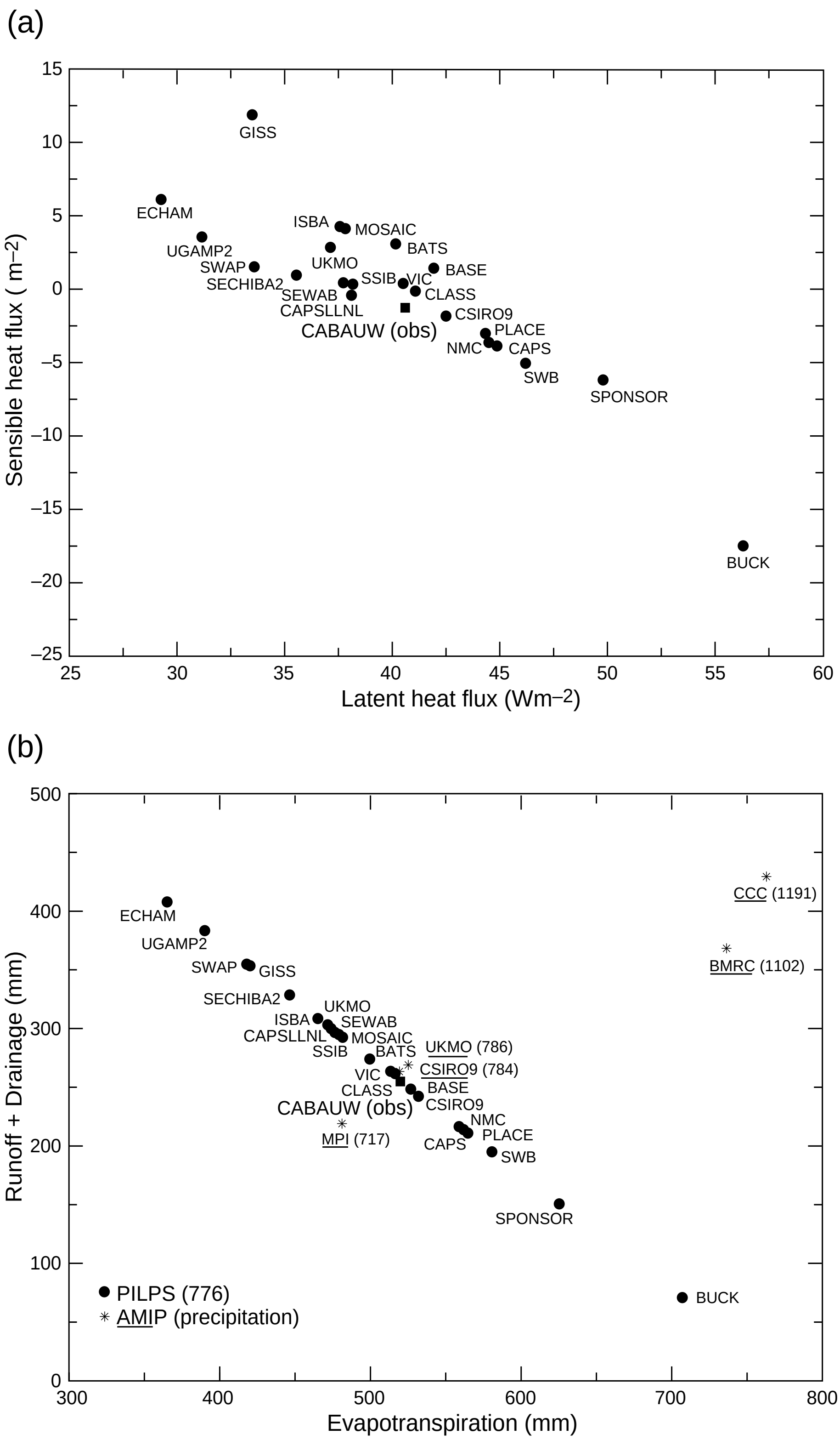


Figure 4.9 — (a) The annually averaged latent and sensible heat fluxes predicted by the PILPS land-surface schemes. A single year's observations from Cabauw, The Netherlands, were used for as many annual cycles as was required for each land-surface scheme to conserve energy (≤3 Wm^{-2}) and water (≤3 mm/yr).

(b) As in (a) but for the annual totals of evaporation and runoff plus drainage. The AMIP/PILPS models' 10-year mean values are the weighted average for the closest GCM grid point and as many of the surrounding eight grid points that are designated as land. (The AMIP precipitation totals are given in parentheses and differ from those prescribed for the off-line forcing.) [from page 259, Reference no. 3].

4.3.4 SENSITIVITY OF CLIMATE MODELS TO MODEL FORMULATION, BOUNDARY CONDITIONS, AND PARAMETERIZATION

The climate model is a very complex entity. Improvement in its solutions requires improving many of the subparts. Improvements cannot be made in isolation as the parts affect each other, and the final solution includes approximations that balance each other. A parameterization scheme that is improved in isolation may, in fact, reduce the quality of the solution of the model system because it changes inputs to other parameterization schemes and that could offset existing compensations within the model.

The key strategy is to focus on improving the parts of a model system which contribute the most uncertainty to the model simulations. This avoids giving

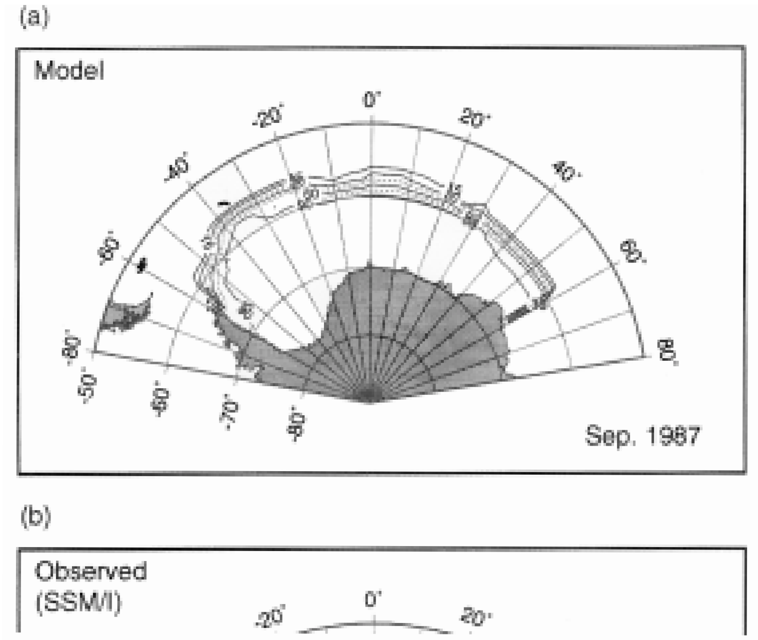


Figure 4.10 — (a) Simulated, and (b) observed sea-ice concentrations (in per cent) in the Weddell Sea for September 1987 (Fischer and Lemke, 1994). [from page 268, Reference no. 3].

excess attention to improving parts of a model which would not make much difference to the quality of the solution. An important part of model evaluation is to identify the aspects of a model most responsible for its simulation deficiencies, i.e., 'model sensitivity.' The 1995 IPCC report (pp. 271-274, Reference no. 3) lists five key areas of sensitivity given below.

(*a*) Representation of water vapour

Water vapour content has a large range of values in the troposphere. Because of this, numerical approximation errors for advection of water vapour can lead to large errors in water vapour distribution. Such errors can lead to errors not only in cloud coverage and precipitation but also in latent heating and radiation transfer. Special numerical techniques are needed to deal with this problem.

(*b*) Model resolution

Model resolution in both the atmosphere and the ocean model is considered to have important impacts on simulation quality. Resolution affects both the numerical errors in the resolved phenomena of the model and the behavior of parameterization schemes. It may be hard to separate these two effects. It is appropriate to consider a resolution that varies in space such as is usually done with vertical resolution near the atmosphere-ocean interface and near the tropopause in the atmosphere.

(*c*) Convection and clouds

The sensitivity of climate model solutions to the cumulus convection parameterization scheme has already been demonstrated. Cumulus convection parameterization quality is very important in tropical regions as it affects precipitation, cloud cover, cloud radiative forcing, and the overall hydrological cycle. Climate model solutions are also sensitive to microphysical processes for all clouds.

(*d*) Land surface processes

The earlier discussion showed the wide range of solutions given by current land surface model parameterization methods. The overall climate solution may be less sensitive to the details of the land surface because land covers only 29 per cent of the Earth's surface; however, the land surface is where people live so it will be necessary to have more detailed description of climate change effects for the land areas.

(*e*)
Initial conditions and surface boundary conditions

In order to offset the effects of chaos sensitivity due to the nonlinear character of the fluid dynamic equations for the atmosphere and ocean, it is considered necessary to employ ensemble simulation techniques to reduce the sensitivity to initial conditions. This is considered important for aspects of the climate such as interannual variability. It is believed that the boundary conditions at the ocean surface play an important role in the stability of a simulated thermohaline circulation in the ocean.

Of the points listed above, it is felt that clouds, convection, the hydrological cycle, and land surface processes — points (c) and (d) above — are the areas of largest uncertainty in climate models.

4.3.5 UPDATE FROM 2001 IPCC REPORT (Reference no. 7)

The 2001 IPCC report summarized recent advances in modelling studies. These have improved even further confidence in the ability of models to predict future climates. These advances include increased computer power to enable higher resolution models and multiple runs (ensembles) of models, better coupled model simulation of lower frequency events such as the El Niño-Southern Oscillation (ENSO), and model calibration for climate changes using paleoclimate simulations.

CHAPTER 5

CLIMATE PREDICTION

5.1 INTRODUCTION

Climate prediction covers many ranges of time. Strategies for handling these can be organized into three time-scale categories: short-range (covering seasonal and interannual time scales), medium-range (covering time scales larger than interannual up to a century), and long-range (covering time scales on the order of 10 000 years) [see Bengtsson in Chapter 23 of Reference no. 6]. Prediction strategy can build on experience of day-to-day weather prediction, and it must also consider the nonlinear and chaotic characteristics of the atmospheric and oceanic components in the climate system. 'Predictability,' defined as the length of time for which the model can provide useful, deterministic information, is greatly influenced by the nonlinear characteristics of the system.

5.2 PREDICTABILITY

Predictability limits in numerical weather prediction models have been measured by the growth in differences, or conversely the reduction in correlations, between the model forecast and the corresponding observational information. Commonly the atmospheric variable examined has been the 500 mb height field. Typically, the growth of model errors in time is exponential when the errors are very small compared to overall spatial variability. The error growth gradually decreases to zero as error magnitudes reach values similar to overall spatial variability. Conversely, the correlation between the time-dependent spatial patterns (anomaly structure) in the model field and those in the observations will decrease with time and eventually approach zero.

The predictability of the model can be measured by the length of time the above correlation remains greater than a specified base value. In setting the base value, an effort is made to distinguish model prediction skill in the forecast from other factors which might happen to give a correct forecast. For instance, a persistence prediction (predicting continuation of existing conditions) can be correct some of the time. The skill in the model forecast is represented by the increase in anomaly structure correlations between the model forecast and observations over those obtained with a persistence or climatological forecast.

The predictability of a numerical weather prediction model depends on the accuracy of its numerical formulation, the accuracy of the initial conditions, and the rate at which smaller scales unresolved in the model actually affect the resolvable scales. No matter how well the first two conditions are met, predictability will always be limited by the third factor in a geophysical fluid system because the model can never resolve all scales. Even in the geophysical fluid system itself there is no complete determinism from an initial state because of the inherent uncertainty in the initial state and of forcing effects. Lorenz (1969) gave a graphic example of the concept of predictability limits by stating that it could be argued that the flapping wings of a single butterfly could completely alter the details of the entire atmospheric system if given a sufficiently long time.

For weather prediction, the inherent limit to predictability has been estimated to be on the order of two to three weeks based on theoretical studies by Lorenz (1969, *ob cit.*). The predictability achieved by weather prediction models for the entire region north of 20°N has improved greatly over the past 30 years as shown in Figure 5.1. From data sources such as those shown in Figure 5.1, it has been estimated that the 'useful' predictability skill achieved by models increased from slightly more than half a week in the mid-1960s to nearly ten days in the 1980s.

Weather prediction models have become so accurate for one-day forecasts that it is now possible to approximate closely the predictability limits to which weather prediction models may be compared. The procedure proposed by Lorenz (1982) compares forecasts of one-day difference in length for each day in the

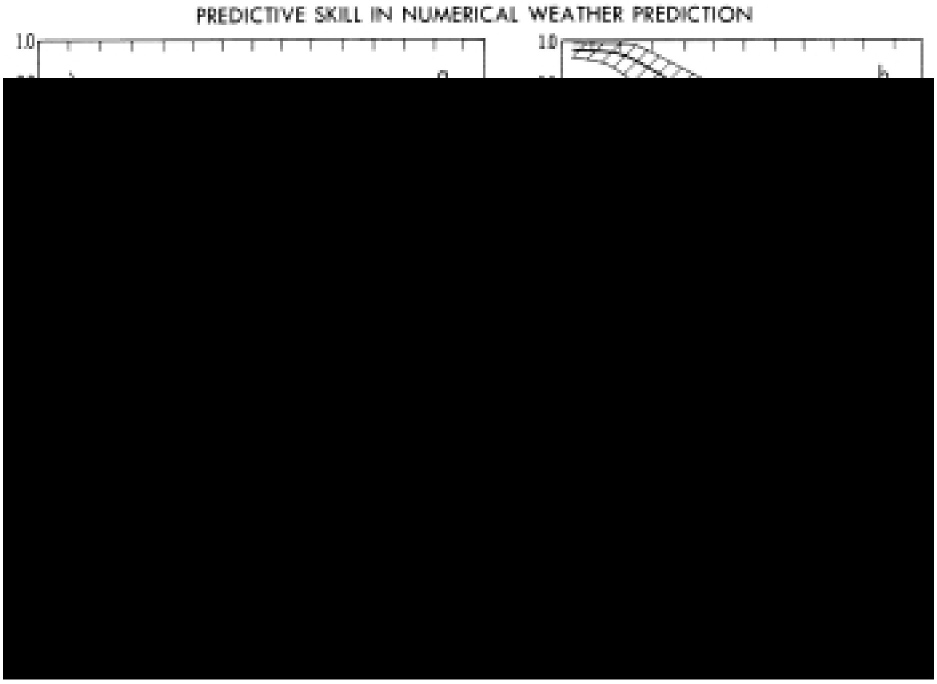


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Figure 5.1 — The predictive skill in numerical weather prediction and its improvement from (a) the mid-1960s to (b) the late 1980s. Shown is the decrease with time of the correlation coefficient between predicted and observed 500-mb geopotential height anomalies for 12 two-week predictions in (a) and 90 ten-day predictions in (b). The correlation coefficients were computed using all grid-point values north of 20°N. The thick solid lines show the mean correlation values averaged over all individual cases, whereas the shaded areas give a measure of the range of the individual cases (i.e., about 5 per cent of the predictions lie below the shaded area and 5 per cent above it). For comparison, the persistence curve (dashed) indicates the no-skill forecast. The growth in (useful) predictive skill from less than one week in the mid-1960s to about ten days in the 1980s is evident. The data in (a) are from Miyakoda et al., *1972, and in (b) from unpublished ECMWF statistics (courtesy L. Bengtsson).* [from page 461, Reference no. 5, with permission of Springer-Verlag].

forecast period. Likewise this procedure specifically compares the two-day and one-day forecasts for tomorrow, the three-day and two-day forecasts for the day after tomorrow, the four-day and three-day forecasts for the day after that, and so forth. This provides a predictability upper-limit estimate containing only one-day forecast errors for the whole period of time. Figure 5.2 shows an example of forecasts for a winter season. This procedure provides estimates on how close the ten-day forecast is to the best that could be obtained (the solid and dashed lines, respectively, in Figure 5.2).

Predictability considerations for climate are quite different from those for weather. Climate prediction is made for time periods longer than the two to three week weather predictability limits. This is possible because climate prediction is for statistical descriptions of the weather conditions, such as monthly means, instead of instantaneous conditions. Climate predictability extends to periods longer than two or three weeks; this is especially true in situations influenced by slowly varying oceanic conditions such as ENSO.

However, nonlinear processes add complexity to climate predictability. Specifically, there may be more than one climate for a given set of forcing parameters. In such a case, the climate is not unique and is termed ‘intransitive.’ If the climate is unique, it is termed ‘transitive.’ The first case is an example of chaos theory where ‘multiple attractors’ exist. The ‘attractor’ is mathematical terminology that may be considered a frame of reference for a ‘given climate.’ The theory goes on to describe how the shift from one ‘attractor’ (or climate) to another may be random and unpredictable.

It is not known what, if any, aspects of climate are intransitive. However, this characteristic would imply that a small additional forcing, such as that due to human activity, could potentially cause a general shift to a new climate pattern which is not necessarily reversible even if the small additional forcing is removed. Clearly, this is an important area for research.

An ‘ensemble forecasting’ modelling strategy has been developed to deal with numerical predictability in situations which are highly sensitive to nonlinear processes. In this method a large number of numerical simulations are made for a forecast period, using the same model. The simulations differ slightly only in details of the initial conditions or forcing effects. Then statistics for model forecast variability can be determined as a function of space and time. These statistics show where the model solution is more variable and therefore less reliable for estimating values. The statistics also provide information on the probability of occurrence for specific forecast states.

This approach has been used for operational weather prediction in which up to 50 separate forecasts may be made for a given forecast period. It is just as valid for climate models. Ensembles of up to ten forecasts are currently used in short range climate prediction (seasonal forecasts). The intercomparison tests among climate models, such as those discussed in Chapter 4, serve a similar function. However, in such intercomparison experiments, it is not possible to separate internal nonlinear processes from effects due to differences in model formulation and parameterization.

Short-term climate forecasting (seasonal to interannual time scales) is a direct extension of deterministic long-range weather forecasting. Currently, long-range weather forecasting has been extended to seasonal forecasting by focusing on the pronounced ENSO variability in the tropical Pacific ocean area and its relation-

- Winter precipitation in central Chile;
- Central England summer precipitation (based on Atlantic Ocean sea-surface temperature);
- Tropical Pacific and Indian Ocean sea-surface temperatures;
- Northern tropical Atlantic sea-surface temperatures;
- Atlantic tropical storm activity;
- Southern Oscillation index time series;
- Summer monsoon rainfall in central-east China;
- Tropical Pacific island precipitation;
- Canadian wintertime temperature and precipitation; and
- U.S. temperature and precipitation.

A range of methods is often used in seasonal forecasting. For example, a number of groups currently supply seasonal forecasts for rainfall in north-east Brazil. These forecasts use January conditions for sea-surface temperature, particularly in the Atlantic Ocean, to predict rainfall anomalies in the following spring. Methods used include statistical linear regression, discriminant analysis and dynamic approaches. Correlation skills are generally in the 0.9 range for this seasonal forecast.

The long-range forecasts for tropical Pacific sea-surface temperatures show skill. As an example, forecasts from a coupled ocean-atmosphere model at the Climate Prediction Center, U.S. National Centers for Environmental Prediction (NCEP), are presented in Figure 5.3. These are mean values of ensemble forecasts for sea surface temperature anomalies in the eastern tropical Pacific for the area between 5°N and 5°S latitude and longitudes ranging from 150° to 90°W for the 'NIÑO' index and 170° to 120°W for the 'NIÑO3.4' index. Forecasts made three, six, and nine months in advance are compared with observed values. Note that the forecasts for late 1997 included the large El Niño event that did actually occur.

Since January 1998 global seasonal forecasts have been made available from the European Centre for Medium Range Weather Forecasts (ECMWF) on the Internet at the web site: <http://www.ecmwf.int/html/seasonal/info/info.html>. An example of their six-month forecast for the NIÑO-3 index anomaly during the onset of the major 1997/98 El Niño event is shown in Figure 5.4. The monthly forecast ensemble is based on forecasts made three times a week, which gives an ensemble of 12 to 15 members for each month.

Forecasts for climate conditions in other parts of the world which have some observed correlation with the ENSO generally have limited, but useful, accuracy over subregions of the forecast area. The results of an experimental programme to predict seasonal rainfall over northern Africa one month prior to the season are are a good example. (A joint programme among operational weather prediction units in England, France, and ECMWF.) An ensemble prediction method was

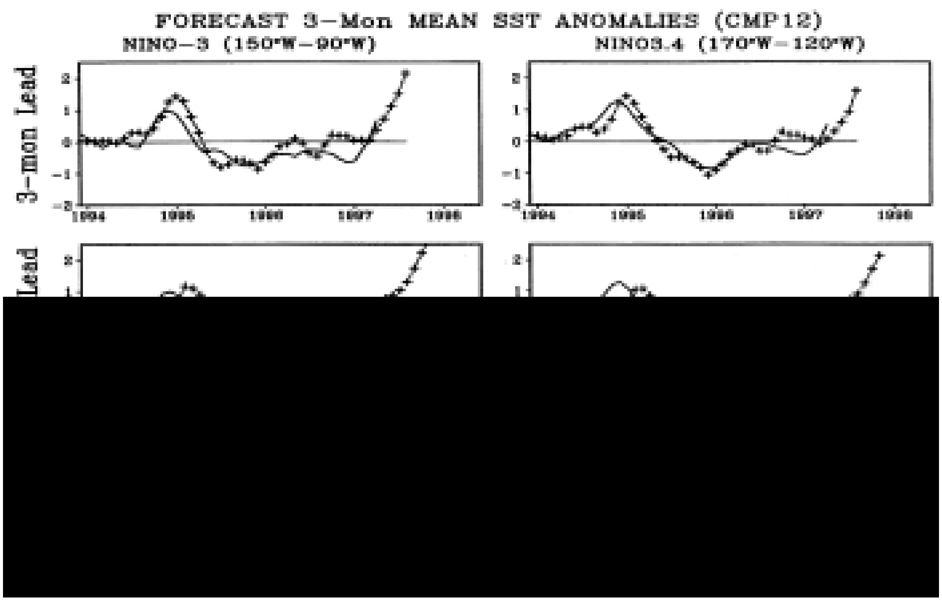


Figure 5.3 — NCEP coupled-model, monthly-mean SST anomaly forecast time series for Niño-3 and Niño-3.4 forecasts from 1994 to 1998 for three, six, and nine month lead times in the upper, middle, and lower panels respectively. These monthly values were used in three-month mean SST anomaly forecast products. The predictions represent the mean of three ensemble-mean forecasts, each for one of the three most recent months, respectively, and each produced by forecasts from two to three individual one to two-week-apart initial conditions per month. [from page 13, Ji. et al. (1997)].

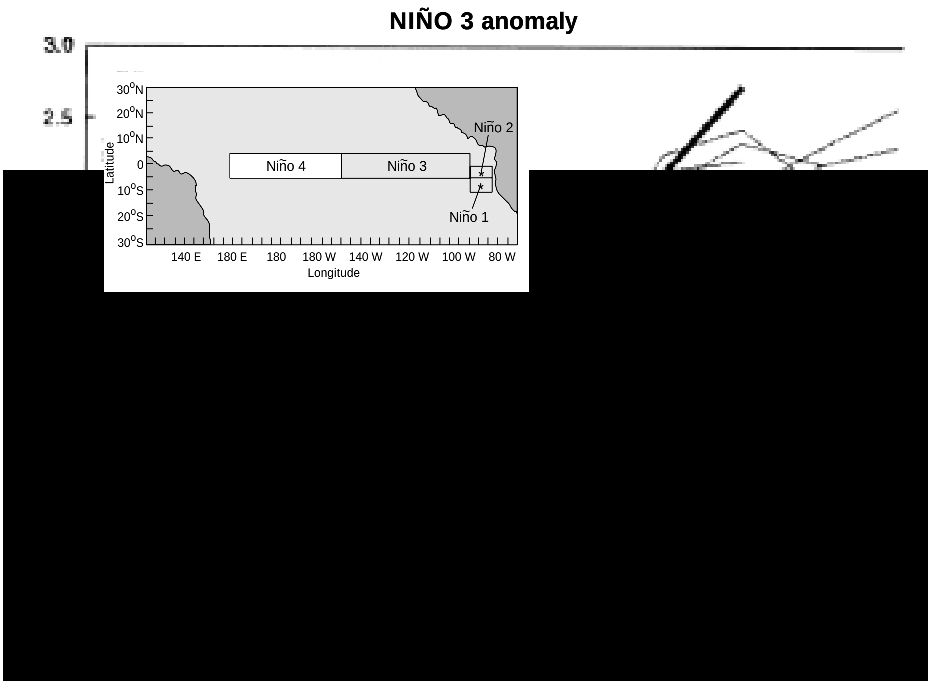


Figure 5.4 — Monthly sea-surface temperature (SST) ensemble predictions for the eastern tropical Pacific (the NIÑO-3 region) for the onset of the 1997-98 El Niño. Six-month predictions from a coupled ocean-atmosphere model (fine lines) compared with observations (heavy line) are shown for the November 1996 to September 1997 time period. Predictions are initiated over an extended period from November 1996 to March 1997. Three forecasts are made each week (only one is plotted in order to avoid clutter) which gives an ensemble of 12 to 15 members each month. Note that the onset of El Niño in April 1997 is well forecast. The subsequent evolution is also well forecast, although one can see a tendency to underpredict the amplitude of the event. [from ECMWF, 1998].

applied to the 15-year period from 1979 to 1993. Correlations between the ensemble mean rainfall and the observed rainfall were made by Harrison, *et al.* [1997]. After calibrating the ensemble forecast variance, they found that in some areas the observed rainfall was within the ensemble range for most of the 15 years while in others this was the case for only ten of the 15 years.

In summary, useful seasonal prediction is now possible for selected regions of the globe. The prediction is generally for anomalies in monthly mean temperature or precipitation. The skill level in some cases may be sufficient to be of value as an operational product for the public. It is important that operational meteorologists become acquainted with possible benefits for their area of service.

The World Meteorological Organization has instituted a Climate Information and Prediction Services (CLIPS) project which will make short-range climate forecast information available to the world meteorological community. It is important to understand the strengths and limitations in forecasting to be able to integrate CLIPS most effectively with national service and policy.

5.4 MEDIUM-RANGE CLIMATE FORECASTS

Forecasting climate variability for decadal to century time scales does not deal with specific variability phenomena but rather with simulation of overall climate processes and changes due to external forcing that causes climate change. It is expected that the modelling of all components of the climate system will have an important role in forecast quality. For such forecasts, major variations in some of the external forcing conditions, for example changes in the major ice sheets that occurred in the last ice age, may be excluded.

It is important that the climate models used for simulating this time scale have reasonable equilibrium characteristics for the control state. Climate drift due to energy transfer discrepancies must be counteracted with ‘flux adjustments’ as discussed in Chapter 4.

For this scale of forecasting and certainly for long-range climate prediction, the possible intransitive nature of the ocean thermohaline circulation can present a challenge. There is evidence that a significantly different circulation existed in the Atlantic Ocean around 11 000 years ago. The increase in fresh water runoff from north America into the north Atlantic caused oceanic subduction and downward convection to stop. This in turn prevented the formation of cold deep ocean waters and reduced currents that provide the northward transport of heat by the oceans. This resulted in a lowering of atmospheric temperatures in the northern polar latitudes. Modelling studies have shown that this large change in circulation could have occurred on a decadal time scale.

Calibration of medium-range climate forecast products is provided by testing the model on the climate of the recent past, for a period for which sufficient observational data are available. An example was shown in Figure 4.9. A model simulation of global mean surface temperature for the entire period of climate change since the pre-industrial era is shown in Figure 5.5. In this simulation the greenhouse gas and aerosol forcing is based on the observational estimates of the concentration of these forcing constituents. Note that the model forecast for overall trends is reasonable; however, decadal variability, especially in the period from 1920 to 1960, is not forecast well.

5.5 LONG-RANGE CLIMATE PREDICTION

Forecasting for time scales of 10 000 years and greater requires climate models with the full representation of the components of the climate system. Effects due to the deep circulations in the oceans (with time scales of 1 000 years or more) and major changes in the cryosphere (glaciers and ice fields) must be included. A number of processes, such as the oceanic thermohaline circulation, may cause the climate to be intransitive. This would mean that there is not a unique climate for a specified climate forcing, so that the model prediction would depend on the initial conditions.

Model studies to simulate past climates have demonstrated that climate models can represent different climate conditions. However, these studies have not reproduced all of the process cycles in the evolution of the climate system. As discussed by Bengtsson [p. 721, Reference no. 6], the time scales that must be handled for long-range prediction range from hours for the atmosphere to centuries and millennia for the cryosphere, ocean, and astronomical forcing.

Bengtsson goes on to describe a modelling technique originally suggested by Hasselmann (1988) whereby the ‘slow system’ representing the land ice, deep ocean, and astronomical forcing are solved separately from the ‘fast system’ consisting of the atmosphere, land surface and the upper ocean. An equilibrium condition would be found in the fast system for a given state of the slow system. The forcing effects of the fast system due to this equilibrium condition would be maintained as constant on the slow system for a time duration comparable to the time scale for changes in the slow system. Periodically a new equilibrium would be computed for the fast system to be consistent with changes in the slow system.

5.6 PREDICTABILITY FOR REGIONAL CLIMATE

The primary climate simulation models are global. Currently available computer power limits the resolution of the global climate models to a few hundred kilometers in the horizontal. This means that local climate conditions and variations cannot be represented. Yet the local conditions are those which relate most directly to climate-impact assessments. Local variations and changes in climate

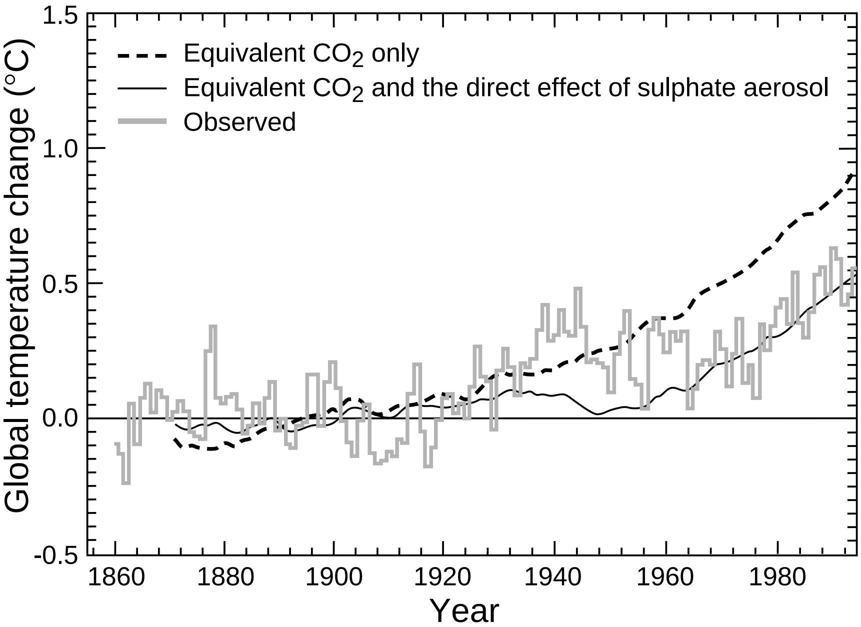


Figure 5.5 — Simulated global annual mean warming from 1860 to 1900, allowing for increases in equivalent CO_2 only (dashed curve) and allowing for increases in equivalent CO_2 and the direct effects of sulphates (flecked curve) (Mitchell et al.*, 1995). The observed changes are from Parker et al. (1994). The anomalies are calculated relative to 1880-1920.* [from page 297, Reference no. 3].

are much larger than averages measured over large spatial scales. Below, an example of the limitations of the global climate models for regional climate prediction is presented followed by a brief discussion of two methods that can provide useful predictive information on local climate from climate models: statistical 'downscaling' and higher resolution regional models.

5.6.1 GLOBAL CLIMATE MODELS

Analysis of regional results in global simulation models demonstrates the large variability in comparing model results with each other and with observations. Nine modelling groups examined seasonal climatologies for surface temperature and precipitation in control runs intended to represent existing current conditions. Comparisons were made for seven regions over land areas with dimensions of very roughly 2 000 km by 2 000 km. They were from central north America (CNA), south-east Asia (SEA), the Sahel (SAH), southern Europe (SEU), Australia (AUS), northern Europe (NEU), and east Asia (EAS). Results shown in panels b, d, f, and h of Figure 5.6 illustrate the large difference between the global models.

5.6.2 STATISTICAL DOWNSCALING TECHNIQUE

The statistical method has the following two-step approach: 1) Development of statistical relationships between local climate variables and large-scale predictors from the global climate model; and 2) Application of such relationships to the output from the climate models to estimate local climate characteristics. This approach has been quite successful in relating weather-prediction model simulations to local surface weather conditions. The method requires a large number of realizations where model output is related to corresponding observed surface conditions in order to determine the appropriate model predictors and statistical coefficients.

This method can be applied to any predicted parameter that has a physical relationship to the model variables, as long as there is an observational record for

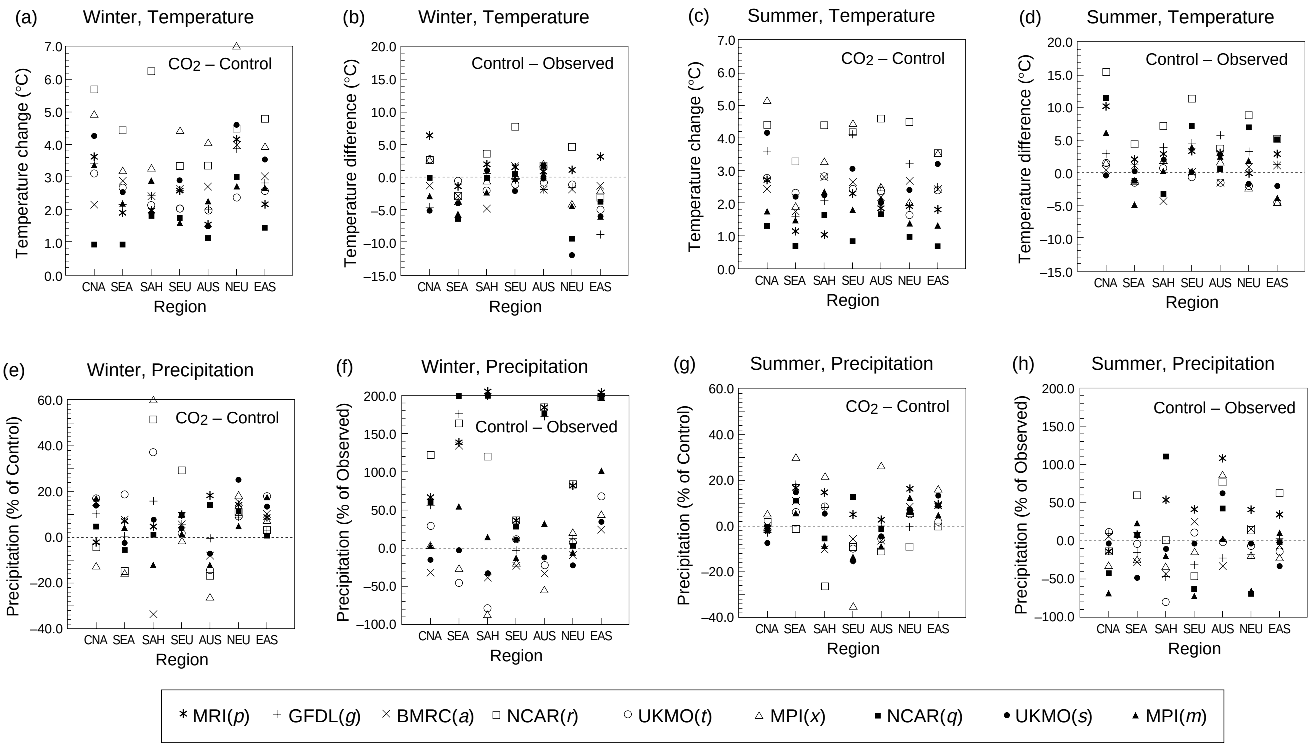


Figure 5.6 — Differences between averages at time of CO_2 doubling and control run averages (CO_2–Control) and difference between control-run averages and observed averages (Control–Observed) as simulated by nine AOGCM runs over seven regions. (a), (b) Temperature, winter; (c), (d) temperature, summer; (e), (f) precipitation, winter; (g), (h) precipitation, summer. Units are °C for temperature and percentage of control run or observed averages for precipitation. In (f) and (h) values in excess of 200 per cent have been reported at the top end of the vertical scale. In (e) values in excess of 60 per cent have been reported at the top end of the vertical scale. CNA=central north American, SEA=south-east Asia; SAH=Sahel; SEU=southern Europe; AUS=Australia; NEU=northern Europe; EAS=east Asia. [from page 338, Reference no. 3].

calibration. The method can provide results for situations where small spatial structure is expected, such as for temperature and precipitation in regions with large topography. One example of application is the prediction of sea level at tidal gauges in Japan from model-simulated sea-level pressure anomalies [Maochange *et al.*, 1995].

5.6.3 REGIONAL CLIMATE MODELS

The regional modelling approach uses output from the global climate model to provide initial and boundary conditions for a regional climate model. These regional models are forced 'one way' by the global climate model, i.e. the global climate model determines forcing conditions for the regional model but the regional model does not, in turn, influence the global model.

These models can have a much higher resolution and can incorporate physical processes not in the global model. Regional climate models have included coupling to lake models, dynamic sea ice models, coastal ocean models and ecosystem models.

Experiments using regional climate models for present-day climate experiments have shown the following results for regional climate models [see pp. 340-341, Reference no. 3]. These models had horizontal resolutions ranging from 15 to 125 km.

(*a*) Realistic synoptic events have been simulated with small biases in temperature and precipitation when initial and boundary conditions were provided by observations. Biases were in the range of a few °C for temperature and 10-40 per cent for precipitation.

(*b*) Performance was degraded when the quality of the model forcing (i.e. specification of initial and boundary conditions) was reduced by using general circulation model simulations instead of observations to force the regional climate model.

(*c*) Simulations produced more realistic detail than in the global climate models used to force the regional models; however, regionally-averaged values could be more or less realistic than those in the driving climate model.

(*d*) Models performed better at mid-latitudes than in tropical regions.

(*e*) Model performance improved as the resolution of the driving global climate model increased.

(*f*) Seasonal as well as diurnal temperature ranges were simulated reasonably well.

(*g*) Validation data from adequately dense observational networks was lacking, especially in mountainous areas.

5.7 UPDATE HIGHLIGHTS ON CLIMATE MODELING FROM THE 2001 IPCC REPORT

The 2001 IPCC report lists ten highlights in modelling advances since the 1995 IPCC report. These are listed below.

- Coupled models can provide credible simulations of both the present annual mean climate and the climatological seasonal cycle over broad continental scales for most variables of interest for climate change. Clouds and humidity remain sources of significant uncertainty but there have been incremental improvements in simulations of these quantities.
- Confidence in model projections is increased by the improved performance of several models that do not use flux adjustment. These models now maintain stable, multi-century simulations of surface climate that are considered to be of sufficient quality to allow their use for climate change projections.
- There is no systematic difference between flux-adjusted and non flux-adjusted models in the simulation of internal climate variability. This supports the use of both types of model in detection and attribution of climate change.
- Confidence in the ability of models to project future climates is increased by the ability of several models to reproduce the warming trend in 20th century surface air temperature when driven by radiative forcing due to increasing greenhouse gases and sulphate aerosols. However, only idealised scenarios of sulphate aerosols have been used.
- Some modelling studies suggest that inclusion of additional forcings such as solar variability and volcanic aerosols may improve some aspects of the simulated climate variability of the 20th century.
- Confidence in simulating future climates has been enhanced following a systematic evaluation of models under a limited number of past climates.

- The performance of coupled models in simulating the El Niño-Southern Oscillation (ENSO) has improved; however, the region of maximum sea-surface temperature variability associated with El Niño events is displaced westward and its strength is generally underestimated. When suitably initialised with an ocean data assimilation system, some coupled models have had a degree of success in predicting El Niño events.
- Other phenomena previously not well simulated in coupled models are now handled reasonably well, including monsoons and the North Atlantic Oscillation.
- Some palaeoclimate modelling studies and some land-surface experiments (including deforestation, desertification and land cover change) have revealed the importance of vegetation feedbacks at sub-continental scales. Whether or not vegetation changes are important for future climate projections should be investigated.
- Analysis of, and confidence in, extreme events simulated within climate models is emerging, particularly for storm tracks and storm frequency. 'Tropical cyclone-like' vortices are being simulated in climate models, although enough uncertainty remains over their interpretation to warrant caution in projections of tropical cyclone changes.

CHAPTER 6

OBSERVATIONS FOR LONG-TERM CLIMATE MONITORING

6.1 INTRODUCTION

Observations are a critical component in the discussion, understanding, and identification of climate change. There are uncertainties in the observational data which contribute to the overall uncertainty about climate variability both natural and anthropogenic. Many persons might assume that the current observational network is sufficient. A global observing system has been in place for years to support operational weather prediction; it forms part of WMO's World Weather Watch (WWW) programme. In recent decades, the observations have been substantially reinforced by observations from new systems such as satellites.

Ocean measurements are an indispensable part of the global climate observing system. Measurements are needed to provide spatial and temporal descriptions of temperature, salinity and currents. Measurements for sea-surface temperature have been comprehensive; however, measurements for conditions below the ocean surface are currently insufficient. A fully four-dimensional observational system is needed for the ocean as exists for the atmosphere.

There are serious deficiencies in, and critical issues for, the observational system for climate monitoring for the atmosphere, as well as the ocean. It is important that these deficiencies and critical issues should be understood and addressed. In recognition of the needs for climate monitoring, a new international programme was established by the Second World Climate Conference in 1990, the Global Climate Observing System (GCOS). Local weather observations provide a key foundation for climate-information systems. Meteorologists throughout the world need to understand the importance of long-term and well-documented local observations for climate monitoring.

There are several underlying principles for a climate monitoring system. First, the system must be underpinned by the scientific community in terms of development, calibration, and monitoring. This is true even for automated observing systems. Second, it must be understood that climate observations have requirements that go beyond those for weather observing. Third, the observational records need to be long-term and have a consistent, homogeneous, and documented frame of reference to be able to detect trends in climate conditions.

This chapter leads off with elaboration on the key specific principles for long-term climate monitoring. Examples of the status of selected observations relevant to climate change are then presented followed by discussion of strategies for improving long-term climate monitoring. The primary reference for this chapter is Reference no. 2 listed in the introduction.

6.2 KEY PRINCIPLES FOR LONG-TERM CLIMATE MODELING

The list of ten principles below covers the details for assuring that observations will be of value for long-term climate monitoring. They apply to observations taken anywhere in the world. Staff meteorologists, weather station managers, meteorological service directors, and the appropriate government agencies need to become aware of and understand the importance of these principles. The principles apply to both small and simple and large and elaborate observing operations. The list is presented in complete form on pages 86-87 in Reference no. 2.

(*a*) System changes

The effects on the climate record due to changes in instrumentation, observing practices, observation location, etc., must be known before implementing such changes. These effects can be determined by a period of overlapping measurements between the old and new system or by comparison of the old and new systems with a standard reference. Sites chosen for *in-situ* measurements should have expectations of long and uninterrupted use and expectations that there will be little change of the nearby physical environment over time.

(*b*) Processing algorithm description — Processing algorithms for determining data values from the instrument system must be well documented and archived with the original data.

(*c*) Observing system description — Information on instrument, station, platform history, changes in sampling time and local environmental conditions, and all other factors relevant to interpretation of the data should be recorded as part of the observing routine and archived with the original data.

(*d*) Length of record — Observations that have a long, uninterrupted record should be maintained and kept uniform (homogeneous) in terms of measurement procedures. Long-term for space-based measurements is measured in decades, but long-term records for more conventional measurements may be a century or more.

(*e*) Climate record homogeneity — Calibration, validation and system maintenance should be used to provide a constant frame of reference for the climate record.

(*f*) System backup — Some form of 'low-technology' back-up to 'high-technology' observing systems should be put in place to safeguard against unexpected operational failures.

(*g*) Observing system priorities — Highest priority should be given to the design and implementation of new climate observing systems for; 1) data-poor regions; 2) variables and regions sensitive to climate change; and 3) key measurements with inadequate spatial and temporal resolution.

(*h*) Network design — Long-term climate requirements should be made known to the designers and engineers at the outset of designing a new network.

(*i*) New observation system development — A long-term association and cooperation commitment is needed between the research group whose needs require developing a new instrument system and the group that will eventually handle the system in operational mode. A clear plan for the transition from research to operational applications should be made.

(*j*) Data management — It is essential to have data management systems that facilitate access, use and interpretation of the data. Data management should have freedom of access, low cost, and user-friendly interfaces (directions, catalogues, browsers, metadata on station history, algorithm accessibility, documentation, etc.). International cooperation is very important.

6.3 STATUS OF SELECTED OBSERVATIONS CRITICAL FOR CLIMATE CHANGE

6.3.1 OBSERVATIONS FOR BASIC FORCING FACTORS

6.3.1.1 Solar radiation

The solar radiation entering the Earth's atmosphere is the primary input forcing factor for the climate system and its measurement is essential for analysis of climate. Variations in the magnitude of this irradiance have a direct impact on equilibrium energy levels in the climate system. Satellite systems are optimal for its measurement because they are above the atmosphere. However, measurement of the solar radiation made by multiple satellite systems may show differences from system to system that are nearly as large as 10 Wm^{-2} . A difference of 10 Wm^{-2} is quite significant. It exceeds by quite a bit the 1-2 Wm^{-2} variation in irradiance due to the sun spot cycle as discussed in Chapter 2. A solar irradiance difference of 10 Wm^{-2} is equivalent to a mean radiative forcing factor of 1.75 Wm^{-2} if the Earth's sphericity (a factor of 0.25) and albedo (a factor of 0.7) are taken into account. The estimated overall change in radiative forcing due to human-produced greenhouse gas enhancements is of the same order of magnitude. Clearly, great care is required to obtain a valid (homogeneous) long-term climate record of solar radiation intensity with calibration to a fixed reference.

A new 20-year record has been produced with appropriate calibration corrections which has less variability and clearly depicts the 11-year sunspot cycle (see Figure 6.1).

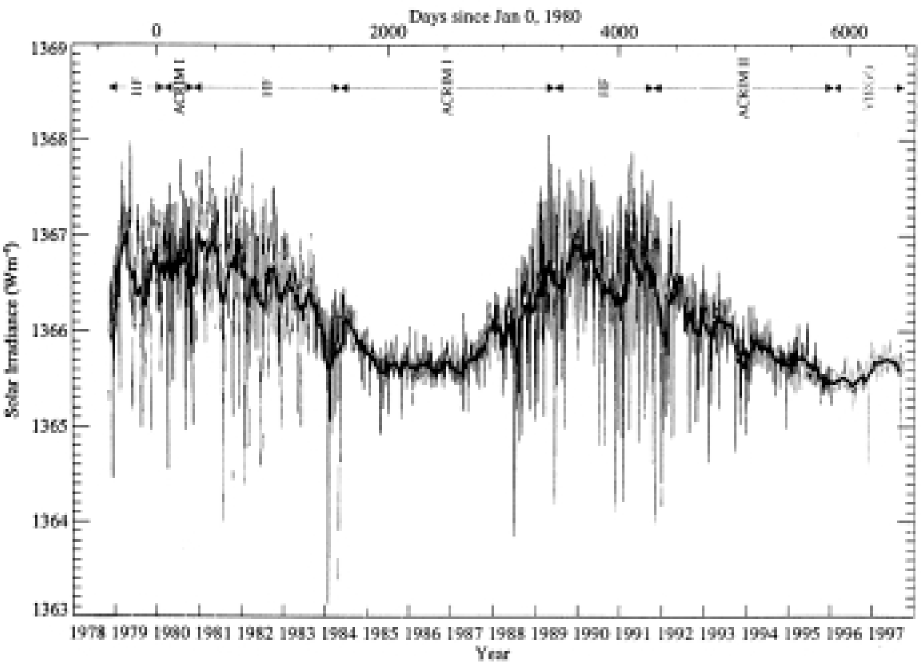


Figure 6.1 – Composite total solar irradiance for 1978 to 1997. The whole time series is adjusted to the Space Absolute Radiometer Reference (SARR) which does not improve absolute accuracy, but allows comparison of repeated space experiments with the same radiometer. [from Fröhlich and Lean, 1988, with permission of Robert B. Lee III and Kluwer Academic Publishers].

6.3.1.2 Greenhouse gases

6.3.1.2.1 Carbon dioxide

Carbon dioxide concentration in the atmosphere has been measured at Mauna Loa Observatory, Hawaii and the South Pole from around 1957 onward. This has provided consistent long-term, single-point records of an important climate-forcing parameter. In climate change study it is necessary to expand the measurement programme to improve our understanding of the global carbon-cycle budget. This requires measuring spatial variations in carbon dioxide concentration in the atmosphere to identify the details of sources and sinks with respect to the biosphere and ocean. Investigation of the sources and sinks of carbon dioxide will also be facilitated by having measurements for spatial variations of O_2 concentrations. The extremely high accuracy needed for the O_2 measurements has restricted these records to the last few years at only a few stations.

6.3.1.2.2 Ozone

Both vertical and horizontal distributions of ozone need to be measured in order to understand implications for climate change forcing. Total ozone (in a vertical column) has been measured by surface-based Dobson spectrophotometers and by satellites. It has been necessary to calibrate Dobson measurements and to adjust for changes in the calibration factor for satellite measurements. Now it is important to develop measurement systems to determine ozone concentrations in the troposphere. There are also satellite instruments measuring the vertical profiles of ozone.

6.3.1.2.3 Water vapour

Water vapour is the most important greenhouse gas, accounting for roughly 80 per cent of the total greenhouse effect. Water vapour concentration is expected to increase if temperature increases because of enhanced evaporation from natural sources. It will be important to observe variations in water vapour concentrations on a global basis to understand observed climate change trends and to validate climate model simulations.

Current water vapour measurements are not as accurate as those for other basic variables of the atmosphere. The slow response to relative humidity of the sensors in radiosondes means that vertical variation of water vapour are smoothed out. In addition, processing procedures have changed over time. For instance, starting in 1993 the calculation for relative humidity from VIZ sondes, which are part of the global network, was modified to include calculation of humidity values below 20 per cent and upward adjustments in humidity values over 80 per cent. The net effect was to increase the measured value of total water vapour and to add an artificial discontinuity to the time-series record for water vapour. The number of different kinds of radiosonde systems currently used in the global observing system also makes it necessary to apply calibration adjustments in the data processing.

Satellite observations are good for showing horizontal variations of humidity; however, vertical variations are smoothed out even more than with radiosonde measurements. Intercomparison of satellite measurements with *in-situ* (radiosonde) measurements will be required to assure the validity of trend studies [page 70, Reference no. 7].

6.3.1.3 Aerosols

Improved measurements are needed for aerosol concentrations in the atmosphere. Such measurement systems must include monitoring the aerosol type and its size spectrum. Turbidity measurements alone are not sufficient. Monitoring of source magnitudes including those from biomass burning will be essential for understanding the budgets for aerosols. It is expected that remote sensing from satellites will be essential for making the required observations.

6.3.2 OBSERVATIONS FOR FEEDBACKS FROM CLIMATE SYSTEM COMPONENTS

6.3.2.1 Clouds

Cloud observations are critical for understanding recent and future climate change. The measurements need to include not only total cloud cover, but also quantative data on the level and composition of the cloud. Ice clouds have quite different radiative properties than water clouds. Continuity in the climatology record for clouds has been degraded by changes in observational methods. Viewing clouds from the Earth's surface is quite different from viewing them from space. Conversion to automated cloud-observing systems introduces a major discontinuity into the climate record.

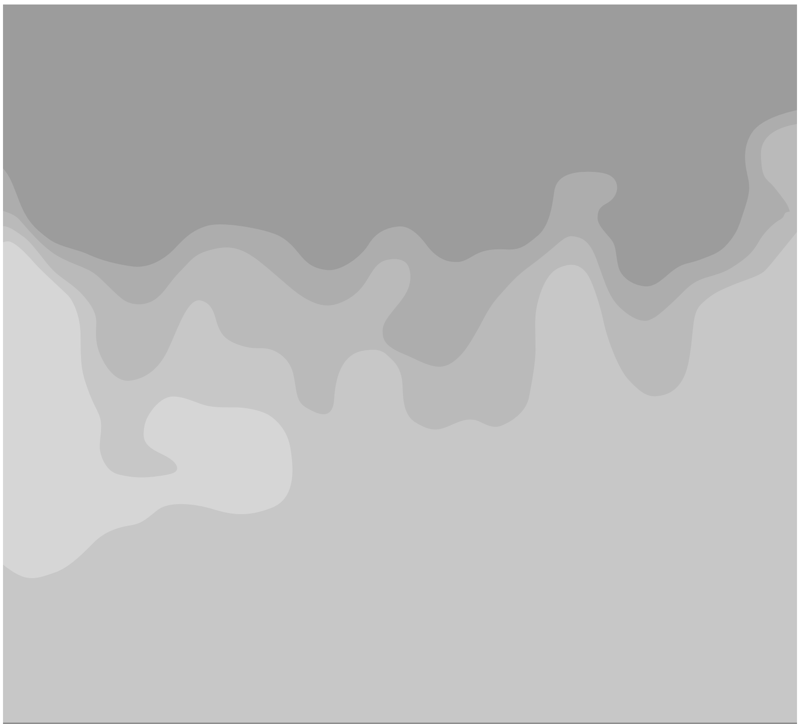
The International Satellite Cloud Climatology Project (ISCCP) has been established to construct a valid climatology of cloud coverage. Observational records from the 1980s had shown considerable uncertainty in the observations even when made by satellite systems. Reductions in percentage cloud cover of approximately three per cent over a seven-year period were found for both nimbostratus and deep convective clouds. The changes were primarily in the periods when there was conversion from one satellite system to another. This conversion effect has now been eliminated by reprocessing the data. Cloud data are now considered reliable for studies of shorter term and regional variations of clouds, even if we cannot monitor long-term trends.

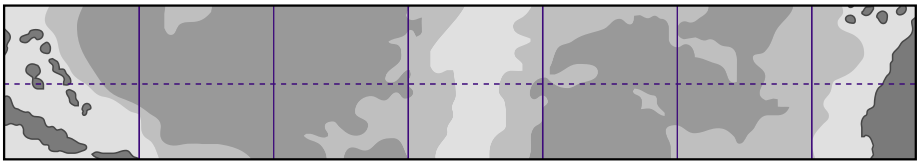
Attempts have been made to use surface solar radiation measurements to give an indication of cloud cover. However, the two are poorly correlated because of measurement system changes for both methods and because of the interference effect of air pollution.

6.3.2.2 Oceans

Measurements of temperature, salinity and currents are required to understand oceanic processes and overall how much heat the oceans will store or give up in a climate change scenario. It is essential to measure the current systems in the deep ocean as well as in the upper ocean (to depths of several hundred meters) to describe the heat transports within the ocean which affect sea-surface temperature. Measurements in the deep ocean will make it possible to monitor the Atlantic thermohaline circulation which is a key factor for identifying decadal-scale shifts in the ocean circulation. Measurements of temperature in the deeper ocean may provide a detection of climate change less obscured by seasonal cycle and shorter period variability. For example Figure 6.2 shows a warming trend between 1957 and 1992 that has been observed in subsurface north Atlantic waters at 24°N and is most pronounced between depths of 0.7 and 2.5 km with values up to 0.5°C.

In limited areas, comprehensive ocean-measurement networks exist. An example is the network to support the research, detection, and prediction of El Niño, the Tropical Ocean Atmosphere (TAO) array in the tropical Pacific Ocean. This network of moored instrument systems provides temperature and current measurements in the upper levels of the ocean across the Pacific between 10°S and 10°N. However, that network does not involve deep ocean water. For decadal-scale variability, long observational records will be required in both the Pacific and Atlantic ocean areas. In the Atlantic Ocean area, the decadal-scale variability involves deep water conditions which need to be thoroughly measured.





Sea-level observations are needed for several reasons. First, spatial sea-level variations relate to ocean currents and provide information useful for understanding ocean dynamics. Spatial sea-level variations can be measured from satellites. Second, sea-level changes can have potentially deleterious impacts on coastal regions. Long term *in-situ* monitoring is needed to isolate changes due to climate change from many other factors that can affect sea level in order to make projections for future impacts.

6.3.2.3 Surface hydrology

Routine long-term soil moisture observations are woefully scarce over the world. Only a few countries such as the Russian Federation have observations that extend over many decades. Understanding climate change processes over land areas will require such information.

6.3.2.4 Surface land cover

Remote sensing from satellites together with calibration from local surface observations is providing an important monitoring of land surface changes. This information helps to define changes in the land surface forcing of the atmosphere and the response of the terrestrial biosphere to climate variations. It is important to maintain this observing activity for monitoring and studying climate change.

6.3.2.5 Cryosphere

Satellite observations are important for monitoring snow and ice cover over land, and sea-ice extent over the oceans. However, many of the records to date are undocumented with regard to processing procedures and are of short duration. It will be important to give more attention to data quality and continuity for climate monitoring purposes.

6.3.3 OBSERVATIONS FOR CLIMATE RESPONSES

6.3.3.1 Surface temperature

The surface temperature of the Earth is a key descriptor for climate. Surface station reports are important for providing the observational data as satellites are incapable of measuring the details of temperature in the surface boundary layer (the lowest several meters in the atmosphere). There are numerous deficiencies in these observational records. Over land, the density of reporting stations varies by continent. Africa, Central America, and South America have large areas which are not adequately covered. Figure 6.3, which presents mean maximum temperatures reported by surface stations, shows the uneven spacing in station distribution. To compound the problem, the number of reporting stations is decreasing even in those areas that are already sparsely represented. As shown in Figure 6.4, the decrease was more than 10 per cent from 1989 to 1994.

There are many factors which introduce non-uniform spatial and temporal biases into land surface temperature data. These include changes in instrumentation, instrument shelters, station location and time of observations. Changing an observation site from a city to a nearby airport can cause systematic changes in the temperature due to the ‘heat island’ effect of the urban area. In many cases, the principles for long-term climate prediction presented in Section 6.2 were not observed. A common omission has been the overlapping of measurements between old and new observing systems when the observing system is changed. Many of the temperature biases introduced by these changes are of magnitudes similar to those expected with climate change. Thus, it is important to define these biases and remove them from the climate record.

Sea-surface temperatures have been measured extensively over large parts of the oceans from ships that voluntarily make temperature measurements (voluntary observing ships and ships of opportunity), as well as from former ship weather stations, and from buoy systems. Ships of opportunity and voluntary observing ships have provided considerable data for more than 100 years in some parts of the world. However, the method of measurement has changed over the years from sampling with buckets of water drawn from the ocean to sampling of water coming into the engine intake. Efforts have been made to adjust data to a common frame of reference. However, information is lacking about the details of the measurement so it has been difficult to make precise corrections to historical data. A good review is provided by Parker, *et al.*, in Reference no. 2 [pp. 429-470].

Satellites provide virtually complete coverage of the oceans for sea-surface temperature measurement. It is a challenge to combine this data with that obtained by *in-situ* measurements because the satellite is measuring temperatures at a very thin surface layer (‘skin’ temperatures) whereas *in-situ* measurements are for a deeper surface layer. In addition satellite measurements are affected by atmospheric turbidity which requires adjustments to the data. The recent volcanic eruption of Mt. Pinatubo caused biases as large as 1°C in the satellite temperature data.

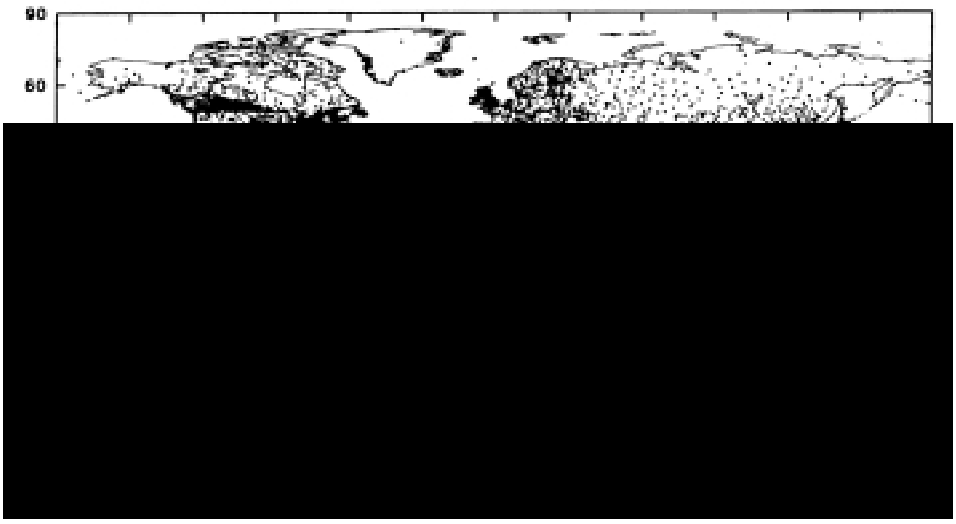
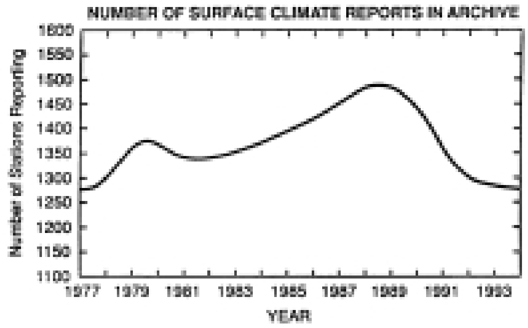


Figure 6.3 — Climatological stations with 1961-90 normals data for mean maximum temperature. All station data has been received by the Climate Research Unit at the University of East Anglia from National Meteorological Agencies. A total of 6 632 locations are shown. [from page 418, Reference no. 2, with permission of Kluwer Academic Publishers].

Figure 6.4 — Number of 'CLIMAT' messages (containing monthly temperature summaries) received at the UK Met. Office through the Global Telecommunications Systems for the period of 1977-1993. [from page 72, Reference no. 2, with permission of Kluwer Academic Publishers].



6.3.3.2 Precipitation

Rainfall is a very important climate parameter for all forms of life on land. Nevertheless, its measurement is very inadequate. Precipitation has much local-scale variability which makes it difficult to find measurements representative for an area. Some of this variability is due to topographic effect. Rain gauge measurements are quite sensitive to rain gauge design, to wind conditions and to whether the precipitation is rain or snow. For snow conditions gauges can seriously underestimate precipitation amounts. Changes in rain gauge design and observing practices which have been common throughout the world result in a major task for producing homogeneous climate data. Figure 6.5 shows a sampling of changes that have been made throughout the world with estimates on the biases introduced into the data.

Satellite and radar sensing systems provide additional information on precipitation, and this information should become better quantified with time. However, these data are available only for short times and the radar data has limitations in the observations of diurnal cycle variability. It will be essential to calibrate and combine gauge and remotely-sensed data to get long-term climate records. The gauge data will remain important as calibration for the remote-sensed data.

For reasons mentioned above, there is considerable uncertainty in climate records of precipitation. Nevertheless, there are first estimates of precipitation trends as reported by IPCC [pp. 152-156, Reference no. 3]. For example, analysis of data suggests an increase in precipitation in high northern latitudes during this century and an increase of winter precipitation in northern mid-latitudes. It will require considerable effort to obtain climate records for precipitation that will be useful for identifying trends due to climate change.

6.4 STRATEGIES FOR IMPROVING LONG-TERM CLIMATE MONITORING

A number of steps are being taken to improve long-term climate monitoring. A few are briefly presented here.

6.4.1 DATA RECOVERY AND RECALIBRATION ('REHABILITATION')

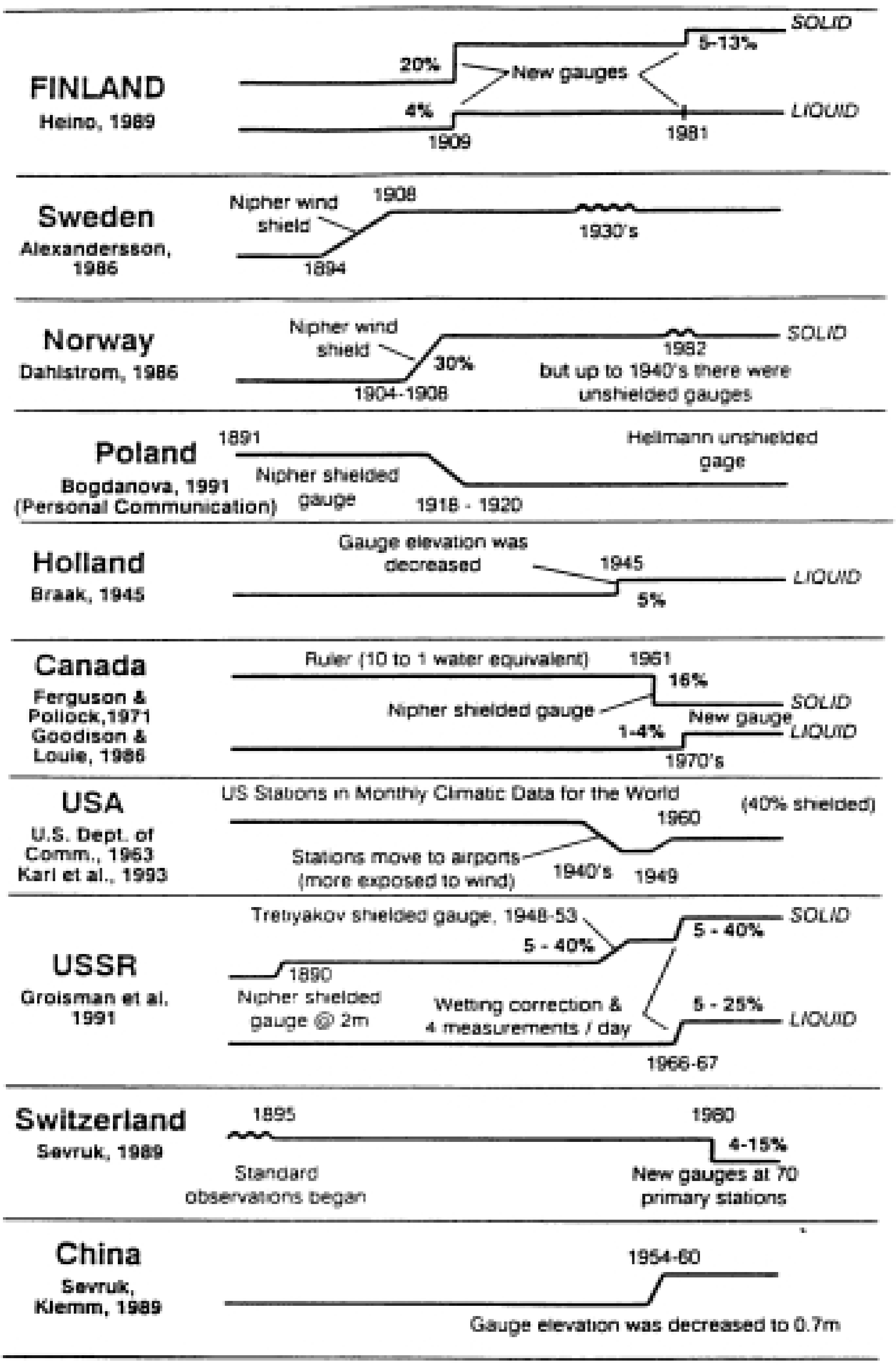
There are ongoing efforts to gather data from archives throughout the world and to render it into a form that allows it to be easily accessed and incorporated into the climate data files. Digitizing hand written records is an important aspect of this work. For all such data it is also necessary to study existing metadata (documentary records about site changes, exposure details, etc.) to correct for artificial biases in the data. WMO has its Data Rescue programme (DARE), to help in this area.

6.4.2 REANALYSIS

Construction of analysis (spatial distribution maps) from data is important for describing and understanding processes in the climate system. The process of analysis also helps to fill in where observational data is missing and to provide the information database needed for numerical model studies.

Numerical models themselves are important components of the analysis process. Adjustments in the patterns that evolve in numerical models as observational information is introduced achieves a 'four dimensional data assimilation' whereby best-estimate and internally consistent initial analyses for prediction models are produced from observational data. This has been the practice since the advent of operational Numerical Weather Prediction (NWP) in the 1950s. The improvement of NWP models has improved the quality of the analysis fields representing observations.

Figure 6.5 — Time-varying biases of precipitation measurement for various countries (from Karl et al., *1993*. [from page 77, Reference no. 2, with permission of Kluwer Academic Publishers].



Changes of the model formulation over time have led to variations in the quality of the analysis and have introduced artificial time-varying biases into the data. In order to obtain an observation analysis record that has a uniform (homogeneous) frame of reference in time, reanalysis has been conducted using the same numerical prediction model. For the purposes of NWP, a reanalysis has been carried out to obtain a consistent dataset of atmospheric observations over a significant portion of the world from around 1957.

General reanalysis is currently being done by two programmes. In the United States, the Environmental Modelling Center of the National Centers for Environmental Prediction (NCEP) together with the National Center for Atmospheric Research (NCAR) are using a forecast model that became operational in 1994 to reanalyse initial conditions for daily data at many levels in the atmosphere. Their plan is to go back to 1957. In Europe, the European Centre for Medium-Range Weather Forecasts (ECMWF) is also conducting a reanalysis of daily data at many atmospheric levels using one of their recent comprehensive forecast models. Their reanalysis period starts in 1979. These reanalysis projects are of central interest to the World Climate Research Programme (WCRP). The reanalysis data will greatly facilitate long-term climate monitoring. Further reanalysis information is given in WMO (1997).

6.4.3 INCREASING THE NUMBER OF MEASUREMENTS

The increase of *in-situ* measurements for greenhouse gases, aerosols and ozone is considered a relatively low-cost enhancement of great value for observations directly related to climate change. In particular, increases in the number of places where flask measurements for CO_2 concentration are obtained are needed to provide validation points for future satellite measurements. More *in-situ* measurements of aerosol chemical and physical characteristics would greatly benefit determination of its radiative properties. An increased number of balloon-borne measurements of ozone and water vapour in the stratosphere would provide important validation data for future satellite measurement.

6.4.4 NEW MEASUREMENTS SYSTEMS

Satellites will play a key role in new measurements for climate monitoring. Many new systems are being developed for satellites to improve measurements of the radiation budget, clouds, trace gases, surface changes on Earth and so forth. Careful attention will have to be given to calibration of the data and overlapping measurements between different satellite systems.

The Earth Observing System (EOS) programme of the U.S. National Aeronautics and Space Administration (NASA) is a good example of the advancement in remote sensing from satellites. The EOS programme is developing over 20 new satellite systems to obtain or improve measurements of many climate system parameters (NASA, 1995). The list includes: solar radiation (total irradiation and ultraviolet component only); atmospheric water vapour vertical distribution; images of the land surface, water, ice and clouds; Earth radiation budget; topography of the sea surface and ice sheets; flux of trace gases (including carbon dioxide) at the air-sea interface; global aerosol distribution; cloud properties such as optical thickness; cloud heights; planetary boundary layer heights; global distributions of numerous trace gases and aerosols in the upper troposphere, stratosphere and mesosphere; location and radiant energy of lightning flashes; precipitation rate; cloud water; sea-surface temperature; soil moisture; angular solar reflectance functions from the top of the atmosphere, clouds and the Earth's surface; biological processes (chlorophyll concentration, vegetation productivity, etc); tropospheric pollution; and sea-surface winds.

Innovative approaches for temperature measurements are being considered. One idea is to monitor global temperature changes by measuring the Earth's electrical fields. Overall electric potential variations with height are related to the number and intensity of thunderstorms in the world. To the extent that thunderstorm activity relates to temperature, measurement of electric potential would provide a measure of temperature conditions. Another idea is that the measurement of temperatures at great depths (like 600 m) below the surface of the Earth could provide signals for long-term temperature changes. This is already being done in the ocean as shown in Figure 6.4. Finally, the travel time for acoustic waves in the ocean over great distances may provide useful information on temperature changes in the deep ocean since the sound speed depends, in part, on temperature.

CHAPTER 7

MODELLING, DETECTION AND ATTRIBUTION OF RECENT AND FUTURE CLIMATE CHANGE

7.1 INTRODUCTION

The projections for future climate change are continually updated as climate models are improved observational records are expanded, understanding of the climate system is improved and estimates of the climate forcing due to human activity are refined. All of these areas involve complex considerations and uncertainties. A large number of research groups throughout the world are focused on improving the projections for future climate through the coordination efforts of the Intergovernmental Panel on Climate Change (IPCC), the World Climate Research Programme (WCRP) and other substantive programme activities of the World Meteorological Organization (WMO) and the International Council of Scientific Unions (ICSU).

The results published at any one time will be superseded by new results. The differences between successive conclusions may be confusing to those outside the community working on climate change. One must appreciate that the changes in conclusions are incremental and part of a coordinated approach to finding answers.

Intercomparisons among many model experiments involving long-term simulations are required to establish a meaningful understanding of their characteristics, sensitivities and uncertainties for any given specification of human climate-forcing impact. As a result, only a limited number of forcing scenarios can be modelled and it takes time for the gains made in climate-change projections to become apparent as models and forcing specifications are improved.

This chapter provides a perspective on the scope and type of conclusions that are being reached. The material presented is based primarily on the material in chapters 6 and 8 of Reference no. 3. Current work and conclusions are overviewed here with full recognition that the details will be different in the future. Some updates from chapter 9 of the 2001 IPCC report (Reference no. 7) are also included.

This chapter includes discussion of recent climate change because that is the direct antecedent for calibration and understanding of projections for the future. It first presents modelling results and then moves to the detection (analysis of observations) and attribution (understanding the causes) for climate change. Attribution is the key factor for isolating human-produced effects from natural changes in the climate system.

7.2 MODEL RESULTS FOR CLIMATE CHANGE

7.2.1 RECENT CLIMATE CHANGE

Figure 3.5 in Chapter 3 identifies the primary anthropogenic impacts on the climate system up to the present. Initial model simulation studies considered only the carbon dioxide component. More recent model simulation studies have included the sulphate aerosol component. Note that in the description of these experiments, the carbon dioxide concentration used may be an equivalent concentration to represent all of the greenhouse gases.

Model simulations for recent global mean temperature changes in the last 45 and 120 years have already been shown in Chapters 4 and 5, respectively. The predicted global mean temperature increases when the observed increases in both carbon dioxide and sulphate aerosol are included in the model (see Figure 5.5). The sulphate aerosol itself leads to a cooling effect which partially offsets the warming due to carbon dioxide. Earlier model studies which had not included sulphate aerosols gave larger values for predicted warming. The general upward trend appears quite realistic when compared with observations. The observed temperature record shows much more variability within the 1860-1990 period than is simulated in the model. However, the model result has been smoothed over time. Yearly mean model values would show similar variability.

Source	*DJF**	*MAM**	*JJA**	*SON**	*Year*
Simulated, increase in equivalent CO_2 since 1900	–0.46	–0.35	–0.08	–0.34	–0.29
Simulated, aerosol forcing and equivalent CO_2 increase since 1990	–0.43	–0.27	–0.16	–0.32	–0.27
Observations of recent change (1981 to 1990 mean less 1951 to 1981 mean)	–0.28	–0.17	–0.19	–0.36	–0.19

* DJF = December, January, February; MAM = March, April, May; JJA = June, July, August; SON = September, October, November.

Table 7.1 — Changes in diurnal range of 1.5 m temperature averaged over seasons and the annual cycle. The simulated and observed values are averaged over the regions where observations are available. The observed data are from Horton (1995) and the simulations from Mitchell et al. *(1995). The changes in greenhouse gas forcing (represented by an equivalent increase in CO_2) and direct sulphate aerosol forcing are those estimated to have occurred since 1900. Note that the observed changes are available only over the latter half of this period and are the difference between the mean for 1981 to 1990 and the mean for 1951 to 1980.* [from page 295, Reference no. 3].

Diurnal temperature range decreases are predicted by theory when the greenhouse gases and sulphate aerosols are increased. Model results show a decrease in diurnal temperature range since 1900 over Northern Hemisphere continental areas as observed increases in carbon dioxide and sulphate aerosols are included in the model. The observations also show decreases in diurnal temperature range (see Table 7.1). However, other processes also affect diurnal temperature ranges such as cloudiness and surface evaporation. The relative importance of all these effects is still uncertain.

In summary, model simulations generally have suggested that the human production of carbon dioxide and sulphate aerosols, starting with the industrial era, has already caused climate change in terms of global warming. Recent numerical model simulations have given estimates for the global-mean warming that range from 0° to +1.6°C, as shown in Table 7.2. These results imply that such changes are currently ongoing and that the atmosphere and the climate system currently are not in equilibrium. As reported in the 2001 IPCC report (Reference no. 7) recent studies further confirm this conclusion.

This pre-existing change and current lack of equilibrium must be considered when designing model experiments for future climate change that start from present conditions. The year 1990 has often been used as the initial time for future climate-change simulations. If the climate model is defined as being in equilibrium before starting the future climate-change experiment, the initial rate of simulated temperature increase will be erroneously suppressed as the model develops the radiative imbalance conditions. This is called the ‘cold start’

Study	*Sensitivity to doubling CO_2 (°C)*	*Direct aerosol forcing (Wm^{-2})*	*Temperature response of equilibrium due to aerosols (°C)*	*Temperature response of equilibrium to combined aerosol and CO_2 forcing since 1900 (°C)*
S1 Roeckner *et al.* (1995)	2.8	-0.7	–0.9*	0.5
S2 Taylor and Penner (1994)	5.2#	–0.9	–0.9	0.6+
S3 Mitchell *et al.* (1995a)	5.2	–0.6	–0.8	1.6
S4 Le Treut *et al.* (1995)	3.9	–0.3 (direct) –0.8 (indirect)	–1.6$	0.0$

Table 7.2 — Equilibrium global mean response to the increase in greenhouse gases and sulphate aerosol concentrations over the 20th century. The experimental designs in the four studies differ, so some of the entries are derived under specific assumptions defined below. [from page 293, Reference no. 3].

S1 and S3 use the aerosol distribution of Langer and Rodhe (1991) and represent aerosols as an increase in surface albedo. S2 derives the sulphate loading from a coupled atmosphere sulphur cycle model and includes an explicit radiative scattering treatment of aerosols.

- * Assuming 40 per cent increase of CO_2 gives 50 per cent of the warming due to doubling, and subtracting this value from the combined forcing experiment in the final column.
- # Assuming 25 per cent increase in CO_2 gives 29 per cent of the warming due to doubling.
- + Using a 25 per cent increase in CO_2, whereas S1 and S3 use a 40 per cent increase to allow for changes in all greenhouse gases.
- $ The forcing used includes both the estimated direct and indirect forcing. In the last column, CO_2 was increased by 25 per cent. Substantially higher sensitivities are found if a colder simulation is used. Although the global mean temperature change in S4 is zero, the model gives a cooling in the Northern Hemisphere and a warming in the Southern Hemisphere.

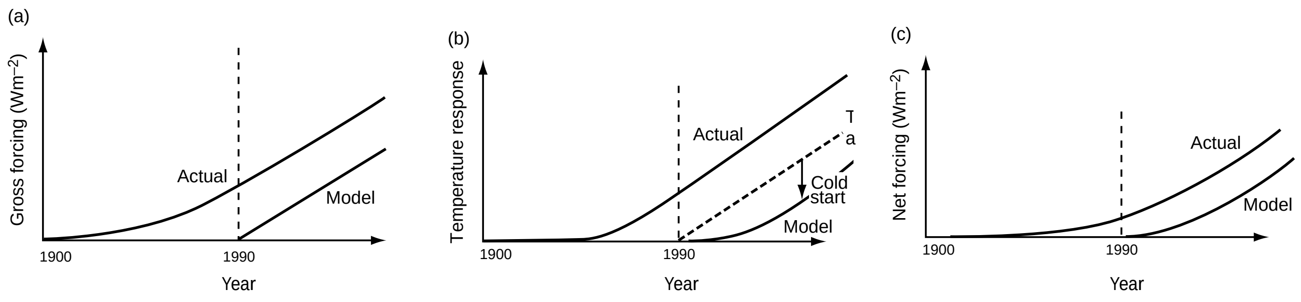


Figure 7.1 — Schematic diagrams of radiative forcing and temperature response, showing the effect of neglecting the effect of past forcing (the 'cold start' problem). Panel (a) shows forcing due to increases in CO_2, with a rate of increase rising gradually to 1990 as observed, and maintained at one per cent/yr thereafter (left curve) and from a one per cent/yr increase starting abruptly in 1990, as in idealized experiments (right curve). Panel (b) shows temperature response to the forcing in (a). The upper curve, which is the response in the case with the gradual initial increase as observed, has been transposed vertically to zero at 1990 (dashed curve) to highlight the initial slow response in the case of an abrupt increase used in idealized experiments (lower curve). The difference between the two curves is known as 'the cold start' and is an artefact of the experimental design. Panel (c) shows the net forcing (which allows for the increased loss of radiation to space as the model warms) in the case with a gradual start to the forcing. The lower curve shows the net forcing (upper curve) in the idealized case. Note the net heating at 1990 which maintains the warming of the ocean in the upper curve in (b). To heat 300m depth of water by 0.3°C/decade (typical of the AOGCM experiments in climate change assessments) requires a net heating of 1.5 Wm^{-2} (cf. 4 Wm^{-2} for a doubling of CO_2) which takes several decades to build up with a 1 per cent/yr increase in CO_2. [from page 313, Reference no. 3].

problem. It is estimated that the lag effect on predicted temperature changes will last for several decades, a time scale determined by the adjustment time of the oceans. Rates of temperature change would be underestimated during this time as shown in Figure 7.1.

7.2.2 FUTURE CLIMATE CHANGE

A range of climate model predictions have been made. The largest group dealt with greenhouse-gas impacts alone. In the description of these experiments, as noted before, reference may be made to the changing concentration of carbon dioxide alone, or to an equivalent carbon-dioxide concentration, that is calculated to represent the effect of all the greenhouse gas changes attributable to human activity. Experiments have been carried out to examine the effects of doubling equivalent carbon dioxide concentrations either all at once or gradually with rates of increase varying from 0.25 per cent per year to 4 per cent per year. Note that the current observed rate of increase of carbon dioxide itself is about 0.5 per cent per year, and the overall increase from the pre-industrial period to 1990 has been 26 per cent (see Table 3.2). A current rate of 1.0 per cent per year

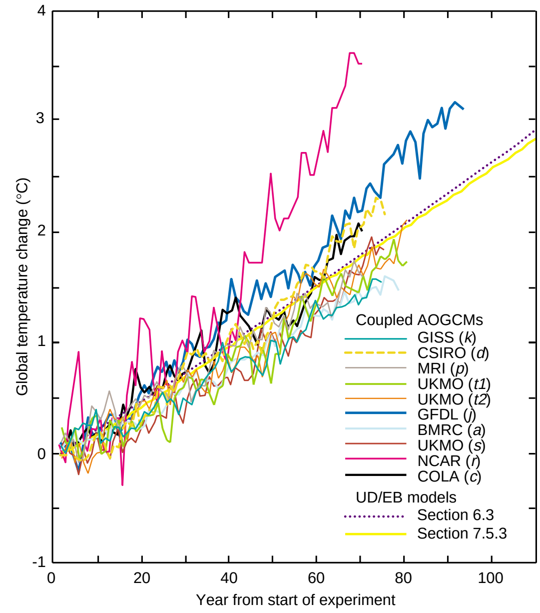
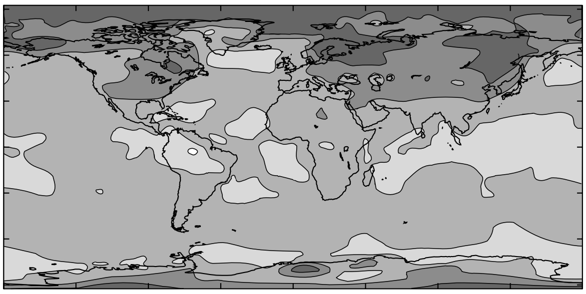


Fig. 7.2 — Comparison between several AOGCM simulations (climate sensitivities between 2.1 and 4.6°C) and two versions of the simpler UD/EB-type models (with climate sensitivities of 2.5°C and 2.2°C). All models were forced with one per cent/yr (compound) increase of atmospheric CO_2 concentration from equilibrium or near-equilibrium in 1990. [from page 300, Reference no. 3].



for equivalent carbon dioxide is close to reality. More recently, experiments have included sulphate aerosols and other greenhouse gases more explicitly. Other experiments have looked at factors of two and four for carbon dioxide increases. Projections have been generally made to about the year 2100, although some studies have gone to the year 2500. Many predictions have started with 1990 conditions; others have gone back to pre-industrial time, eliminating the 'cold start' problem.

7.2.2.1 Mean conditions

Results from experiments with fully-coupled ocean-atmosphere climate models have shown large differences in the simulations for climate change impacts even for globally-averaged surface temperature. A comparison was made among ten models for the case where (equivalent) carbon dioxide concentration was increased at the rate of one per cent per year starting from 1990 values and the model conditions were initially in equilibrium. This rate of increase gives a doubling of CO_2 in about 70 years.

Results for global surface temperature are shown in Figure 7.2. Also shown are results from the simpler type of climate model labelled (UD/EB) discussed in chapter 4. The warming after 70 years ranges from 1.5° to 3.8°C. This variability of model response has led to a calibration descriptor for climate models called 'climate sensitivity.' Climate sensitivity is defined as the increase in the equilibrium value of global mean surface temperature produced by the model with a doubling of carbon dioxide concentration. The climate sensitivity value is larger than the values shown in Figure 7.2 after 70 years because the 70-year value is not an equilibrium value. It would be necessary to run the model for many years with the CO_2 concentration held constant at its doubled value in order to get the equilibrium value of temperature.

Differences between the model results are even more dramatic for spatial patterns and local/regional temperature change values. Figure 7.3 shows a comparison of annual-mean temperature spatial patterns between two of the models shown in Figure 7.2 at the time the CO_2 reaches a doubled value. Note that the increases tend to be larger over land than over water and larger at higher latitudes in both models. However, differences in the details between the models can be large as is seen over northern Africa.

Some experiments for CO_2 concentration doubling and with sulphate aerosol effects included have simulated mean temperature decreases in regional areas in both summer and winter seasons. These areas included parts of China and the United States where aerosol concentrations were high. However, the uncertainty in results for regional areas is large, as shown by the range of values in the model intercomparisons in Figure 5.6 and Figure 7.4. Differences in predicted changes of seasonal mean temperatures ranged up to 5°C. For the same simulations, predictions of seasonal mean precipitation changes differred by up to two mm/day. In some cases, different models even predicted changes of opposite sign.

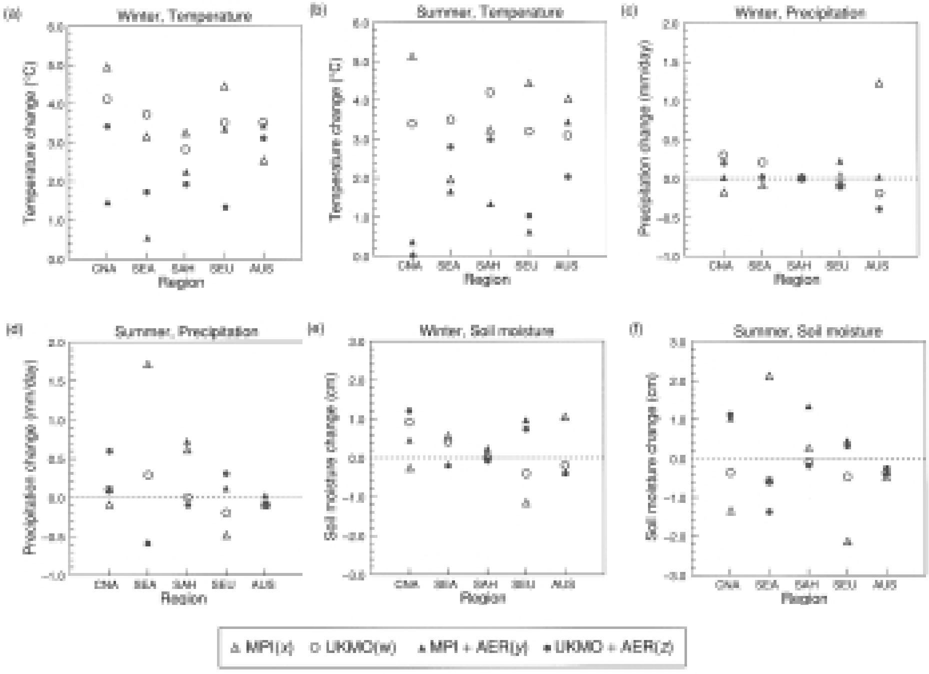


Figure 7.4 — Simulated regional changes from 1880-1889 to 2040-2049 (experiments x,y) or from pre-industrial to 2030-2050 (experiments w,z). Experiments x and w include greenhouse gas forcing only, whereas y and z also include direct sulphate aerosol effects. The x, y, w and z indicators are in the model list at the bottom of the figure. (a) Temperature (December to February); (b) Temperature (June to August); (c) Precipitation (December to February); (d) Precipitation (June to August); (e) Soil moisture (December to February); (f) Soil moisture (June to August). CNA=Central North America; SEA=South East Asia; SAH=Sahel; SEU=Southern Europe; AUS=Australia. [from page 306, Reference no. 3].

Seasonal soil moisture predictions were also in varying directions with differences as large as three cm.

Note that the model simulations discussed in the previous paragraph cover different time periods and have different CO_2 concentration variations than do the model simulations discussed in the two preceding paragraphs. This does not change the overall characteristics discussed here.

Climate-change assessments have been made with the climate models using standardized projections for the changes in anthropogenic forcing of the climate system. For the 1995 IPCC reports six emission scenarios established in the 1992 IPCC report were used in the model studies. The assumptions for these scenarios (referenced IS92a-f) are described in Table 7.3. The scenarios cover the period from 1990 to 2100 and provide examples of low (IS92c), medium (IS92a), and high (IS92e) impacts.

A new set of 40 emission scenarios (referenced SRES for ‘Special Report on Emission Standards’) was established in 2000 (Nakićenović *et al.*, 2000) to define future projections for anthropogenic emissions. Of these, 35 scenarios which contain data on the full range of gases required for climate modelling, define the full set of scenarios used for climate-change projections. A subset of these was used in studies for the 2001 IPCC report. These scenarios generally define the range of possibilities in terms of demographic, economic and technological development. A group of six ‘marker’ or ‘illustrative’ scenarios from this set serves to describe the primary range of uncertainty for future projections. These six scenarios are highlighted in Table 7.4 (from Chapter 9, Reference no. 7). Basically the A categories have larger emission outcomes than the B categories. The three subsets of the A1 category (A1FI, A1T, and A1B) refer to fossil fuel intensive, non-fossil fuel sources, and reliance not on only one energy source cases, respectively. The projections for total radiative forcing out to 2100 for the six SRES ‘marker’ categories along with three of the scenarios examined in the IPCC 1992 studies (i.e. IS92c, IS92a, and IS92e for low, medium and high impacts, respectively) are shown in Figure 7.5.

Note that the IPCC Working Group II, which examined impacts, adaptations, and mitigations of climate change, has adapted revised anthropogenic forcing scenarios based on anticipated new efficiencies for anthropogenic energy production systems that will cut back on increases in carbon dioxide production. See the 1995 IPCC Working Group II report (pp. 47-50, Reference

Table 7.3 — Summary of Assumptions in the Six IPCC 1992 Alternative Scenarios†. (from page 12, IPCC, 1992a).

Scenario	*Population*	*Economic growth*	*Energy supplies††*	*Other†††*	*CFCs*
IS92a	World Bank 1991 11.3 B by 2100.	1990-2025: 2.9 per cent. 1990-2100: 2.3 per cent.	12.000 EJ conventional oil. 13.000 EJ natural gas. Solar costs fall to $0.075/kWh. 191 EJ of biofuels available at $70/barrel.	Legally enacted and internationally agreed controls on SO_x, NO_x and NMVOC emissions.	Partial compliance with Montreal Protocol. Technological transfer results in gradual phase out of CFCs, also in non-signatory countries by 2075.
IS92b	World Bank 1991 11.3 B by 2100.	1990-2025: 2.9 per cent. 1990-2100: 2.3 per cent.	Same as 'a'.	Same as 'a' plus commitments by many OECD countries to stabilize or reduce CO_2 emissions.	Global compliance with scheduled phase out of Montreal Protocol.
IS92c	UN Medium Low Case 6.4 B by 2100.	1990-2025: 2.0 per cent. 1990-2100: 1.2 per cent.	8.000 EJ conventional oil. 7.300 EJ natural gas. Nuclear costs decline by 0.4 per cent annually.	Same as 'a.'	Same as 'a'.
IS92d	UN Medium Low Case 6.4 B by 2100.	1990-2025: 2.7 per cent. 1990-2100: 2.0 per cent.	Oil and gas same as 'c'. Solar costs fall to $0.065/kWh. 272 EJ of biofuels available at $50/barrel.	Emissions controls extended worldwide for CO, NO_x, NMVOC and SO_x. Halt deforestation. Capture and use of emissions from coal mining and gas production and use.	CFC production phase out by 1997 for industrialized countries. Phase out of HCFCs.
IS92e	World Bank 1991 11.3 B by 2100.	1990-2025: 3.5 per cent. 1990-2100: 3.0 per cent.	18.400 EJ conventional oil. Gas same as 'a'. Phase out nuclear by 2075.	Emissions controls (30 per cent pollution surcharge on fossil energy).	Same as 'd'.
IS92f	UN Medium High Case 17.6 B by 2100.	Same as 'a'.	Oil and gas same as 'e.' Solar costs fall to $0.083/kWh. Nuclear cost increase to $0.09/kWh.	Same as 'a'.	Same as 'a'.

† The assumption for the 1990 Scenario A are described in IPCC (1990) Annex A. pp. 331-339.
†† All scenarios assume coal resources up to 197 000 EJ. Up to 15 per cent of this resource is assumed to be available as $1.30/gigajoule at the mine.
††† Tropical deforestation rates (for closed and open forest) begin from an average rate of 17.0 million hectares/year (FAO, 1991) for 1981-1990, then increase with population until constrained by availability of land not legally protected. IS91d assumes an eventual halt of deforestation for reasons other than climate. Above-ground carbon density per hectare varies with forest type from 16 to 117 tons C/hectare, with soil C ranging from 68 to 100 T C/ha. However, only a portion of carbon is released over time with land conversion, depending on type of land conversion.

A1. The A1 storyline and scenario family describe a future world of very rapid economic growth, global population that peaks in mid-century and declines thereafter, and the rapid introduction of new and more efficient technologies. Major underlying themes are convergence among regions, capacity building and increased cultural and social interactions, with a substantial reduction in regional differences in per capita income. The A1 scenario family develops into three groups that describe alternative directions of technological change in the energy system. The three A1 groups are distinguished by their technological emphasis: fossil intensive (A1FI), non-fossil energy sources (A1T), or a balance across all sources (A1B) (where balanced is defined as not relying too heavily on one particular energy source, on the assumption that similar improvement rates apply to all energy supply and end-use technologies).

A2. The A2 storyline and scenario family describe a very heterogeneous world. The underlying theme is self-reliance and preservation of local identities. Fertility patterns across regions converge very slowly, which results in continuously increasing population. Economic development is primarily regionally oriented and per capita economic growth and technological change are more fragmented and slower than in other storylines.

B1. The B1 storyline and scenario family describe a convergent world with the same global population, that peaks in mid-century and declines thereafter, as in the A1 storyline, but with rapid change in economic structures toward a service and information economy, with reductions in material intensity and the introduction of clean and resource-efficient technologies. The emphasis is on global solutions to economic, social and environmental sustainability, including improved equity, but without additional climate initiatives.

B2. The B2 storyline and scenario family describe a world in which the emphasis is on local solutions to economic, social and environmental sustainability. It is a world with continuously increasing global population, at a rate lower than A2, intermediate levels of economic development, and less rapid and more diverse technological change than in the B1 and A1 storylines. While the scenario is also oriented towards environmental protection and social equity, it focuses on local and regional levels.

Table 7.4 — The Emissions Scenarios of the Special Report on Emissions Scenarios (SRES). (from page 554, Reference no. 7).

no. 4) for a description of the low CO_2-emitting energy supply system (LESS) scenarios.

The uncertainty in climate-model predictions for climate change is large. The sources for this uncertainty include the uncertainty in the forcing to be prescribed for the system (as shown in Figure 7.5), and uncertainty due to model formulation (as demonstrated by the range in climate sensitivities in the model intercomparisons shown above in Figure 7.2). Note that the uncertainty in the forcing specification is even larger than that due to the model formulation.

As a result of these uncertainties, it has been necessary to examine a wide range of forcing scenarios. A simple reference climate model of the UD/EB type described in Chapter 4 has been used to make an overall initial analysis for the many cases to determine general characteristics of climate change impacts because of limitations in the computer resources needed to use the comprehensive climate models for so many experiments. The comprehensive climate

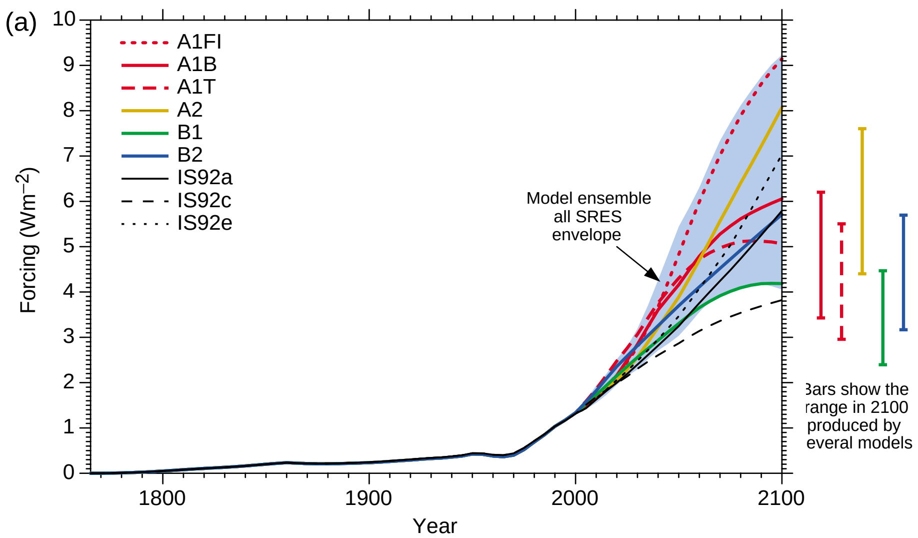


Figure 7.5 — Estimated historical anthropogenic radiative forcing (W m^{-2}) followed by projections for future forcing based on the six SRES marker scenarios and three of the scenarios from the 1992 IPCC (IS92a, IS92c, IS92e). Also shown is the total range of variation (shaded area) for the 35 SRES scenarios that could be used for climate modeling studies. (from Figure 9.13a on page 554 in Reference no. 7).

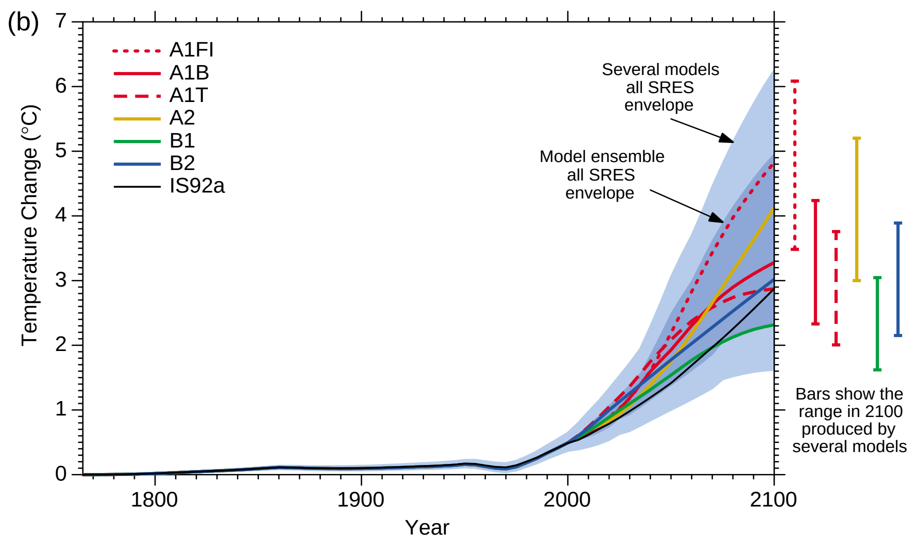


Figure 7.6 — Simple model results for the estimated historical anthropogenic global-mean surface temperature change (°C) followed by projections for future changes based on the six SRES marker scenarios and middle-range of the scenarios from the 1992 IPCC (IS92a) using a simple climate model tuned to seven ocean-atmosphere climate models. Also shown is the total range of variation (shaded area) for all SRES. (from Figure 9.13b on page 554 in Reference no. 7).

models are used to examine the details of climate change for selected cases. The UD/EB model was calibrated using the comprehensive climate models and set to represent a model with 'climate sensitivity' of 2.5°C. The global-mean surface temperature change predictions out to 2100 for the scenarios used in Figure 7.5 (except for IS92a and IS92e) are shown in Figure 7.6. These figures also show the full range of variability 'envelopes' for the full set of 35 SRES scenarios described above.

The UD/EB-type model has also been used to examine other possible impacts due to projected future increases in the greenhouse gases. As an example, possible impacts on the thermohaline circulation in the oceans were studied in experiments that required 1000-year model simulations (see Stocker and Schmittner, 1997). Use of the UD/EB-type model made it possible to perform the set of very long simulations needed for the study. Stocker and Schmittner showed an example where the thermohaline circulation change depended on the rate of increase of greenhouse-gas concentration in the atmosphere. A slow rate of increase to a final enhanced value caused the thermohaline circulation to be reduced. A rapid rate of increase to the same final enhanced value resulted in a total cessation of the thermohaline circulation.

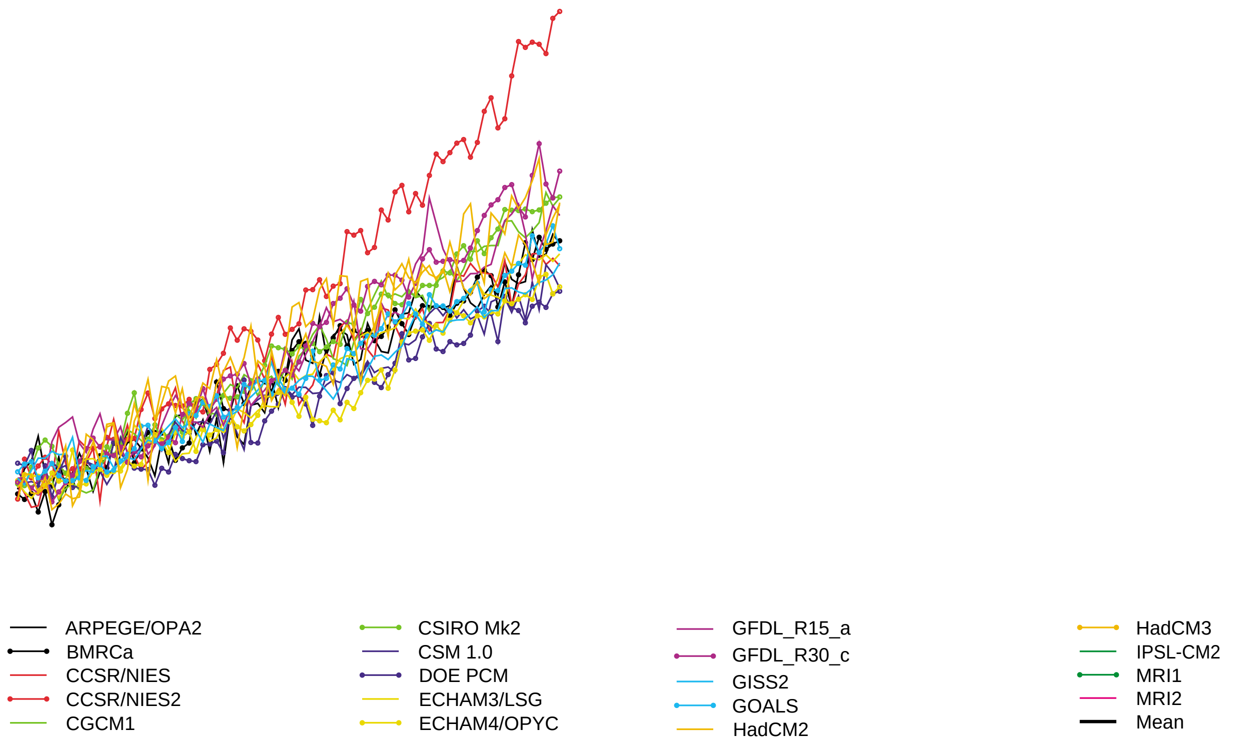
Studies with more sophisticated models reported in the 2001 IPCC report (Reference no. 7) still show considerable variation among models for the same forcing scenario, a doubling of CO_2 in 70 years. See Figure 7.7 for a comparison of the results from 19 models for both temperature and precipitation changes. The change in global mean temperature ranges from +1.1 to +3.1°C and the change in global mean precipitation ranges from -0.2 to +5.6 per cent. This variability is similar to that reported in the 1996 IPCC studies shown in Figure 7.2 before.

7.2.2.2 Variability

There have been climate-change analyses for many specific aspects of the climate system other than mean conditions. Studies have generally been with individual climate model simulations and not the whole group of climate models. Thus, results are quite preliminary.

The decrease of diurnal variability over continental areas has already been discussed. For longer-term variability, it is sufficient to reproduce the essence of the summary comments from page 330 in the 1995 IPCC report (Reference No. 3).

(i) Experiments with different model configurations indicate that zonal-mean, mid-latitude, intermonthly temperature variability may be reproduced, but there are no consistent results in regard to changes in persistent anomalies called 'blocks,' one of the contributors to intermonthly variability. Generalization of climate change impacts from these experiments is difficult since different definitions of blocking have been used and different models show different changes of geographical patterns of blocking.



(ii) One study shows that intermonthly and interannual variability differences between GCMs with a simple mixed-layer (ocean model) and those coupled to a full ocean model are larger than changes in either type of model due to increased CO_2 alone, pointing to the importance of using a model with some representation of ENSO-like phenomena (mixed layer ocean models cannot represent ENSO processes).

(iii) ENSO-like variability in Sea-Surface Temperatures (SSTs) found in several Atmospheric-Ocean General Circulation Models (AOGCMs) still exists with increased CO_2 conditions. The variability shows either little change or a slight decrease in the eastern tropical Pacific Ocean.

(iv) Precipitation variability associated with ENSO events increased in the climate change simulations, especially over the tropical continents. It was suggested that this could be associated with the mean increase of tropical SSTs.

(v) Several models indicate enhanced interannual variability of area-averaged summer rainfall in the South Asian monsoon.

(vi) In a number of simulations decadal and longer time-scale variability obscured the signal of climate change in the rates of global warming and patterns of zonal mean temperature change. In one case multi-century variability in ENSO phenomena was as large as the mean change caused by CO_2 increase. This indicates that natural climate variability on long-time scales will continue to be problematic for CO_2 climate change analysis and detection.

7.2.2.3 CHANGES IN EXTREME EVENTS

Extreme event statistics are an important aspect of climate. Predicting such events requires predicting changes in the probability distributions. Changes in variability will strongly affect the occurrence of extreme events and, in some cases, even more than would result from changes in mean values. For instance, decreasing diurnal variability would be expected to decrease the probability of extreme temperature events (hot and cold). Global climate models do not have the resolution needed to predict most extreme events. Techniques such as statistical downscaling and high resolution regional climate models are needed. These techniques were discussed earlier. A summary of some observed and modelled changes listed in the 2001 IPCC report (Reference no.7) are shown in Table 7.5.

Confidence in observed changes (latter half of the 20th century)	**Changes in Phenomenon**	**Confidence in projected changes** (during the 21st century)
Likely	Higher maximum temperatures and more hot days [a] over nearly all land areas	Very likely
Very likely	Higher minimum temperatures, fewer cold days and frost days over nearly all land areas	Very likely
Very likely	Reduced diurnal temperature range over most land areas	Very likely
Likely, over many areas	Increase of heat index [b] over land areas	Very likely, over most areas
Likely, over many Northern Hemisphere middle to high latitude land areas	More intense precipitation events [c]	Very likely, over many areas
Likely, in a few areas	Increased summer continental drying and associated risk of drought	Likely, over most mid-latitude continental interiors. (Lack of consistent projections in other areas)
Not observed in the few analyses available	Increase in tropical cyclone peak wind intensities [d]	Likely, over some areas
Insufficient data for assessment	Increase in tropical cyclone mean and peak precipitation intensities [d]	Likely, over some areas

[a] Hot days refers to a day whose maximum temperature reaches or exceeds some temperature that is considered a critical threshold for impacts on human and natural systems. Actual thresholds vary regionally, but typical values include 32°C, 35°C or 40°C.

[b] Heat index refers to a combination of temperature and humidity that measures effects on human comfort.

[c] For other areas, there are either insufficient data or conflicting analyses.

[d] Past and future changes in tropical cyclone location and frequency are uncertain.

Table 7.5 — Estimates of confidence in observed and projected changes in extreme weather and climate events. (from page 575, Reference no. 7).

Some of the examples below from the 1995 IPCC report (Reference no. 3) show where changes in extreme weather events were inferred from overall weather or wind pattern changes in the experiments with CO_2 global warming. As in the previous section, these examples are intended to illustrate areas of interest and not areas where conclusions of high certainty have been reached.

7.2.2.3.1 Wind

One source of extreme wind events in the middle latitudes is synoptic-scale storms. The occurrence and intensity of such storms relates to baroclinic field intensity and moisture supply. Changes in middle-latitude storm intensity and tracks have been examined in some climate-change simulations. There has been some evidence in the models of storm tracks being displaced poleward and storms being more intense as global warming occurred.

Tropical cyclones are the primary cause of extreme wind events (as well as heavy rain events) in tropical and subtropical regions. Some global climate models can simulate aspects of tropical cyclone occurrence. However, analysis of global warming simulations has not shown any clear-cut patterns of changes in frequency, area of occurrence, time of occurrence, mean intensity, or maximum intensity of tropical cyclones. Recall that the observations for Atlantic tropical storm activity have shown no systematic change in the past 50 years (see Figure 2.13).

The 1995 IPCC report goes on to highlight the following problems for clarifying climate change impacts on tropical cyclones:

(i) Tropical cyclones cannot be simulated adequately in present general circulation climate models.
(ii) Some aspects of ENSO are not simulated well in general circulation climate models.
(iii) Other large-scale changes in the atmospheric general circulation which could affect tropical cyclones such as jet stream activity cannot yet be discounted.
(iv) Natural variability of tropical storms is very large, so small trends are likely to be lost in the noise.

7.2.2.3.2 Temperature

The effects of doubling the concentration of CO_2 in climate models have been analysed in terms of changes in daily maximum and minimum temperatures. Changes of up to 10°C were found in regions over land areas. The larger changes were related to modelled alterations in snow cover, soil moisture, and cloud cover.

An analysis of climate-model simulations for Victoria, Australia, in a global warming experiment, clearly documented the large change in occurrences of extremes associated with a small change in the mean value. In a low-warming scenario where the mean temperature increased about 0.5°C in that area, there was a 25 per cent increase in summertime days with temperatures over 35°C and a 25 per cent decrease in wintertime days when the temperature went below 0°C. The probability that five consecutive days would exhibit such extreme high or low temperature conditions also showed notable changes.

7.2.2.3.3 Precipitation

A warmer climate is expected to have a more active hydrological cycle as increased evaporation generally leads to higher water vapour content in the atmosphere. It is believed that this would lead to increases in rainfall, including extreme rainfall events. Recent model studies for doubled CO_2 cases have shown an increase in the intensity of single precipitation events along with an overall precipitation increase with temperature. In general, the resolution in the climate models is not good enough to represent the atmospheric convective elements that actually cause heavy rains. Model experiments for doubled CO_2 cases have shown both decreases and increases in rainfall for areas that normally have much rain. In some cases, the predicted rainfall increased while, at the same time, the number of days with rain events decreased.

In one experiment where the mean precipitation decreased by 22 per cent in a southern Europe area, the frequency of occurrence for 30-day dry spells more than doubled in the summer.

It will take considerably more research and model experimentation to clarify the expectations for mean and extreme precipitation associated with climate change.

7.2.3 REDUCING MODEL UNCERTAINTIES AND IMPROVING CLIMATE CHANGE ESTIMATES

The discussion in Chapter 4 on the aspects of current climate models which contribute the most uncertainty to model simulations provides the basis for developing the model needed to improve climate change assessments. The 1995 IPCC report [pp. 345-348, Reference no. 3] discussed nine areas considered to be most important for improving global climate models; these are summarized below.

(*a*) Cloud modeling

It is important to improve the parameterization of cloud formation and dissipation, as radiative energy transfer is very sensitive to cloud cover. This will require improvements in microphysical parameterization to better represent the ice phase in cloud, particle size distribution and the type of precipitation (rain or snow). Particle size distribution and ice phase components are all important for determining the radiative properties of clouds, particularly for solar radiation. The proper simulation of snow is important for impacts on surface albedo and ice field growth. Furthermore, improvement is needed in parameterization for cloud scale dynamics to include deep convection and turbulence effects. Additional observational information will be necessary for this work.

(*b*) Ocean component

Improvement of resolution is essential to improving the ocean component. It is felt that the horizontal resolution needs to be reduced to much less than a 1° latitude-longitude grid to resolve the smaller-scale eddies that influence the circulation with even smaller grid spacing in tropical areas. Currently, some ocean general circulation models have grid spacing as small as 1/6° in both latitude and longitude. Improving the thermohaline circulation simulation is necessary for representation of the dynamics of the full ocean and long time-scale interactions provided by the oceans in the climate system. Observation of the temperature, salinity and motions in the deeper ocean areas will be required for this modelling improvement work. Finally, parameterization for sub-grid scale processes needs to be improved.

(*c*) Flux adjustments

To maintain appropriate balances in the coupled ocean-atmosphere system, it is desirable to avoid the need for using flux adjustments at the ocean-atmosphere interface in the climate models. The use of a flux adjustment makes it more difficult to interpret variability in the model simulations; many of the major modelling centers no longer use it, because the realism of their climate models has improved.

(*d*) Longer periods for simulations

More general availability of ensembles of 100- to 1000-year (or even longer) simulations will help to calibrate and validate climate models using past-climate variations. It is expected that computer capacity will continue to grow to make this possible.

(*e*) Sea ice component

Sea-ice modelling should include motion effects to properly represent dynamic and thermodynamic feedbacks to the ocean and atmosphere.

(*f*) Land surface processes

A fully interactive land-surface component should be incorporated into the global climate models used for climate-change assessments particularly for land areas. This will require model formulations that represent land-surface structure (land-surface type) and functions (processes within the land-surface features).

(*g*) Radiation computation

It is necessary to improve the radiation computational scheme so that, in particular, the water vapor and aerosol effects on solar radiation are represented better.

(*h*) Global carbon cycle

The oceanic and land-surface components of the carbon cycle should be incorporated into the climate models. This incorporation depends on other improvements in the oceanic and land-surface components mentioned above. In the ocean, deep circulation effects are important. It will be necessary to obtain better observations for the deep ocean currents and carbon chemical components.

(*i*) Tropospheric chemistry

The radiative effects of tropospheric sulphates, dust, ozone and other greenhouse gases (CFCs, methane, and nitrous oxide) should be incorporated into the climate model. This will require modelling the appropriate chemical processes for these substances and predicting the size distribution for the aerosol particles.

7.3 DETECTION AND ATTRIBUTION FOR CAUSES OF RECENT CLIMATE CHANGE

7.3.1 INTRODUCTION

A key challenge for climate change assessment is to determine how much of the observed changes in climate are, in fact, due to the effects of human activity (anthropogenic factors) and to describe these aspects. It is important to make the climate change issue understood in order to gain public interest for taking appropriate responses to limit climate change and to deal with impacts on society. Meteorologists need to be able to give a clear response to the question: “has climate change already occurred?”.

It is recognized that humans have already had significant impact on environmental conditions, particularly since the beginning of the industrial era. The discussion about climate change has included changes up to the present as well as those anticipated in the future. Considerable attention is now being given to showing that climate change is already in progress. The task has two parts: first to isolate signals due to anthropogenic influences from the background ‘noise’ of natural variability (detection), and second to define the specific causes of these effects (attribution).

The studies used to detect climate change and to attribute its causes have become increasingly sophisticated. The approaches, from simplest to most complex, may be described by four ‘stages’ listed below:

Stage 1: Examine global or hemisphere-mean values of a single atmospheric descriptor, (most commonly used has been annual-mean surface temperature);

Stage 2: Examine spatial patterns of a single atmospheric descriptor, (again, most commonly used has been surface temperature averaged over a season or a year, although three dimensional temperature patterns have been used);

Stage 3: Same as Stage 2 but the data (model or observed) have been filtered in space and/or in time to make the anthropogenic component signal more detectable; and, Stage 4: Simultaneously use more than one atmospheric descriptor in the analysis made for any of the first three stages.

Model results are often used as signals for the attribution of the data. It is possible to isolate the causes for model signals by model-sensitivity studies where hypothesized causal factors are varied one at a time. Results can be extrapolated from observed conditions where there is correspondence of model signals to observed change signals.

7.3.2 RECENT PROGRESS

Recent progress in detection and attribution studies has been made possible by a number of advances. First, more realistic climate simulations have become available from improved climate models. Anthropogenic forcing factors such as for sulphate aerosols have been added. The first experiments had carbon dioxide or general greenhouse-gas forcing only. Second, more and longer control simulations have provided more reliable statistics for natural variability. Third, more sophisticated statistical analysis techniques have been developed.

Nevertheless, there are many challenges and uncertainties that must be dealt with for the detection and attribution of climate change. Correlations of climate variations with natural forcing factors may appear sufficient to explain all of the observed variability and trends. For instance, Friis-Christensen and Lassen (1991), found a remarkable correlation between the length of the solar sunspot cycle and Northern Hemisphere land temperature anomalies from 1860 to 1985 (see Figure 7.8).

Subsequent study has suggested that although solar impacts are important in climate variations, they do not explain most of the recent warming trend (Kelly and Wigley, 1992). Uncertainties about natural variability exist because of limitations in the observational data as discussed in Chapter 6. Instrument data with sufficient quality to make variability estimates for the atmosphere go back as far as 150 years, but the coverage over the globe during the early years was quite limited; it still has deficiencies. Paleoclimatological data are difficult to interpret and very incomplete in coverage of space and time. Very long climate-model control-case simulations have helped to fill in the gaps for natural variability. The uncertainties of climate model simulations for climate change signals have already been discussed.

Discussion of several recent studies can help to illustrate accomplishments in detection and attribution. Many studies have been made at the Stage-1 level using globally-averaged surface temperature observations. The likelihood that observed trends in the observational record for global-mean surface temperature could result from natural variability has been evaluated. In this analysis it was necessary to use variability statistics from climate models because observational data for variability was insufficient.

The observed global-mean temperature trend for the past 100 years was compared with values expected from linear trends of natural variability characteristics defined by three climate models. Linear trends were fitted to a number of different assembled time-series segments (overlapping ‘chunks’) taken from multi-century model simulations made by these models with no anthropogenic radiative forcing changes. This procedure produced a statistical distribution of possible trends that could be found in the model-simulation data for time scales

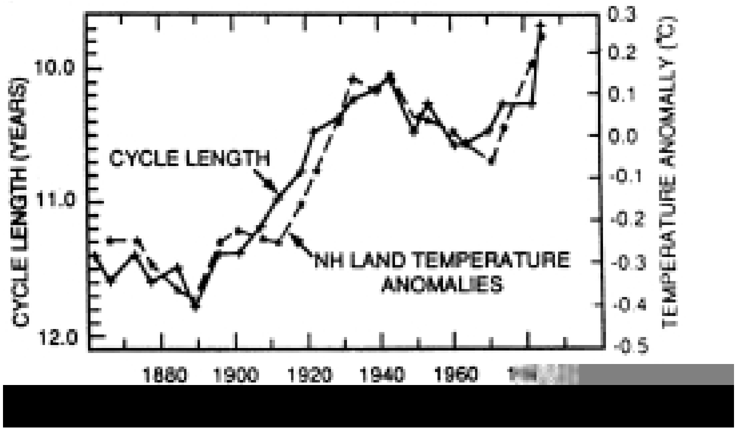


Figure 7.8 — Variations in solar-cycle length and Northern Hemisphere temperature anomalies. The two plotted variables parallel each other quite remarkably (from Friis-Christensen and Lassen, 1991, with permission). [from page 186, Hoyt and Schatten, 1997, with permission of Oxford University]

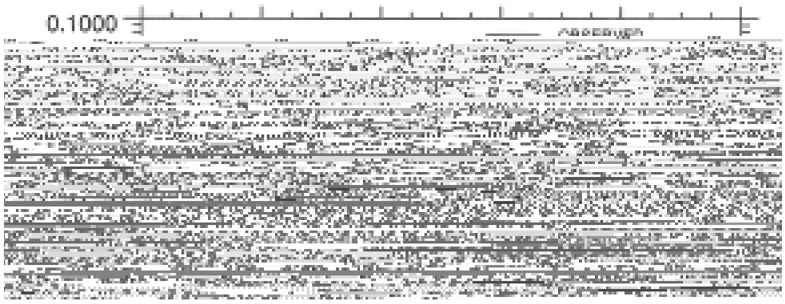


Figure 7.9 — Significance of observed changes in global mean, annually averaged near-surface temperature. The solid line gives the magnitudes of the observed temperature trend (°C/year) over the recent record — i.e. over 10 years (1984 to 1993), 20 years (1974 to 1993), etc. to 100 years. Observed data are from Jones and Briffa (1992). Model results (dashed lines) are from three AOGCM control integrations: the GFDL control run (Stouffer et al.*, 1994), the first 600 years of the 1000-year ECHAM-1/LSG control integration (Hasselmann et al., 1995), and the first 310 years of the UKMO control run (Mitchell et al., 1995). Linear trends were fitted to overlapping 'chunks' of the model temperature series, thus allowing sampling distributions of trends to be generated for the same 10- to 100-year time scales for which observed temperature trends were estimated. The 95th percentiles of these distributions are plotted with dashed lines for each model control run and each trend length.* [from page 423, Reference no. 3].

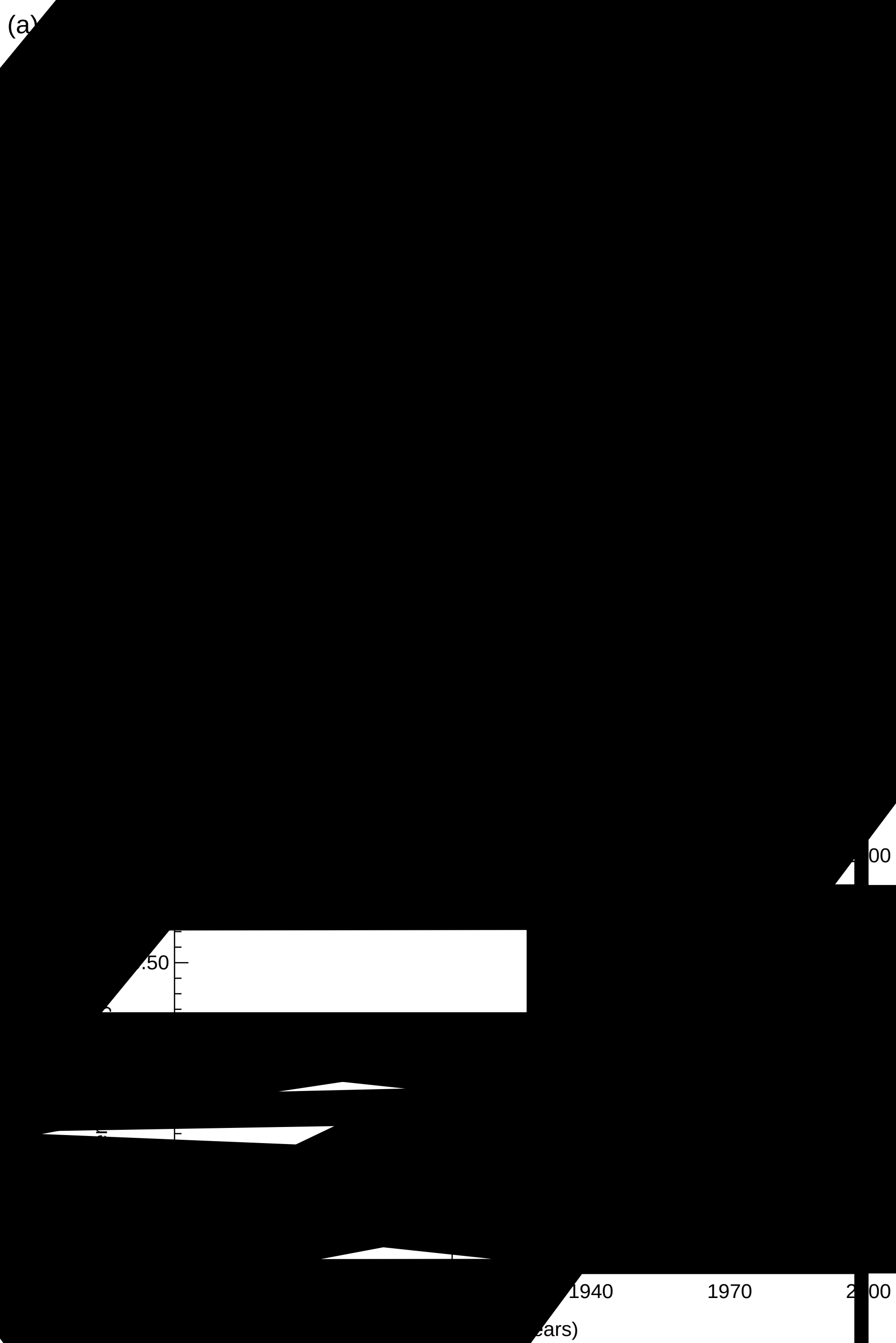
ranging from 10 to 100 years. The observed temperature trend for various lengths of record ending at the present time was then compared with the distribution of trends found in the model data for the same length of record (see Figure 7.9). The results show that *for any record longer than 20 years, the observed warming trend is higher than the level that might be expected from natural variability alone, for more than 95 per cent of the natural variability possibilities*. This, of course, is not certain proof of climate change, but the results indicate probability is low that natural processes alone could account for the observed trend. The conclusion may be exaggerated because the model variability an underestimate of natural variability in the real climate system. The italicized part in the sentence above is an answer that can be given to the question: 'has climate change already occurred?'

Sensitivity of climate variation to forcing can provide information on how anthropogenic forcing may account for observed climate variations. Again with attention to globally-average surface temperature, studies with simple climate models have identified aspects of observed variations that can be attributed to human forcing effects. Figure 7.10 shows the matching between observed and modelled temperature obtained by adding anthropogenic greenhouse gases and aerosol forcing to a model system. In this case, an upwelling diffusion-energy balance model (UD/EB) was used. The adjustment factor is the 'climate sensitivity' of the model. Results suggest that there is some correspondence of overall trends in the observations to those attained in the model.

In the Stage 2 level of analysis, spatial patterns of change are evaluated. This represents a more rigorous analysis and one that can provide more confidence in the attribution of the observed changes to human-produced climate change. The patterns are called 'fingerprints.' They may involve the total value of the variable or structures associated with variable magnitudes that have certain internal coherency such as Empirical Orthogonal Functions (EOFs).

An example of one fingerprint is the vertical structure for temperature in the atmosphere. Enhanced greenhouse forcing, such as that due to increased carbon dioxide, causes a warming in the troposphere and a cooling in the stratosphere. Forcing due to solar radiation increase would be expected to increase temperature at all levels. Figure 7.11 shows the vertical pattern of temperature changes due to changes in anthropogenic forcing from the pre-industrial age to the present in both observations and model simulations. The observed pattern of change has a vertical structure qualitatively similar to that expected from an enhanced greenhouse effect and is not of the form expected for solar radiation increase. Specifically, a common pattern of stratospheric cooling and tropospheric warming is evident in the observations and in both model experiments. In the model data, this pattern primarily reflects the direct radiative effect of changes in atmospheric CO_2. Temperature changes in both the observations and in the experiment with combined CO_2+aerosol forcing also show a common pattern of hemispherically

Figure 7.10 — Observed changes in global mean temperature over 1861 to 1994 compared with those simulated using [redacted] upwelling diffusion-en[redacted] balance climate mod[redacted] model was run first wi[redacted] due to greenhouse [redacted] (a), then with gr[redacted] and aerosols (b), [redacted] greenhouse [redacted] an estim[redacted]



asy[redacted]osphere, with reduced warming i[redacted] the N[redacted] is absent in the CO_2-only case.

Ea[redacted] fingerprint. To the extent that finge[redacted] prints are [redacted]tify causes for patterns of change seen i[redacted] both observ[redacted]ations. For example, horizontal fingerprin[redacted] for surface-te[redacted]ange expected with greenhouse-gas enhanc[redacted] ment would hav[redacted] over polar and continental areas such as show[redacted] in Figure 7.3. In co[redacted]orizontal fingerprint for aerosol effects would mo[redacted] likely have large am[redacted] over industrialized continental areas and resemble the aerosol concentrat[redacted] variations shown in Figure 3.3.

Analysis of the correspondence of observed or modelled patterns of climate change with the fingerprints defined for each forcing component can be done by statistical methods. An example is the EOF approach where the statistically ‘most dominant’ one or two component structures (EOFs) are identified for the patterns

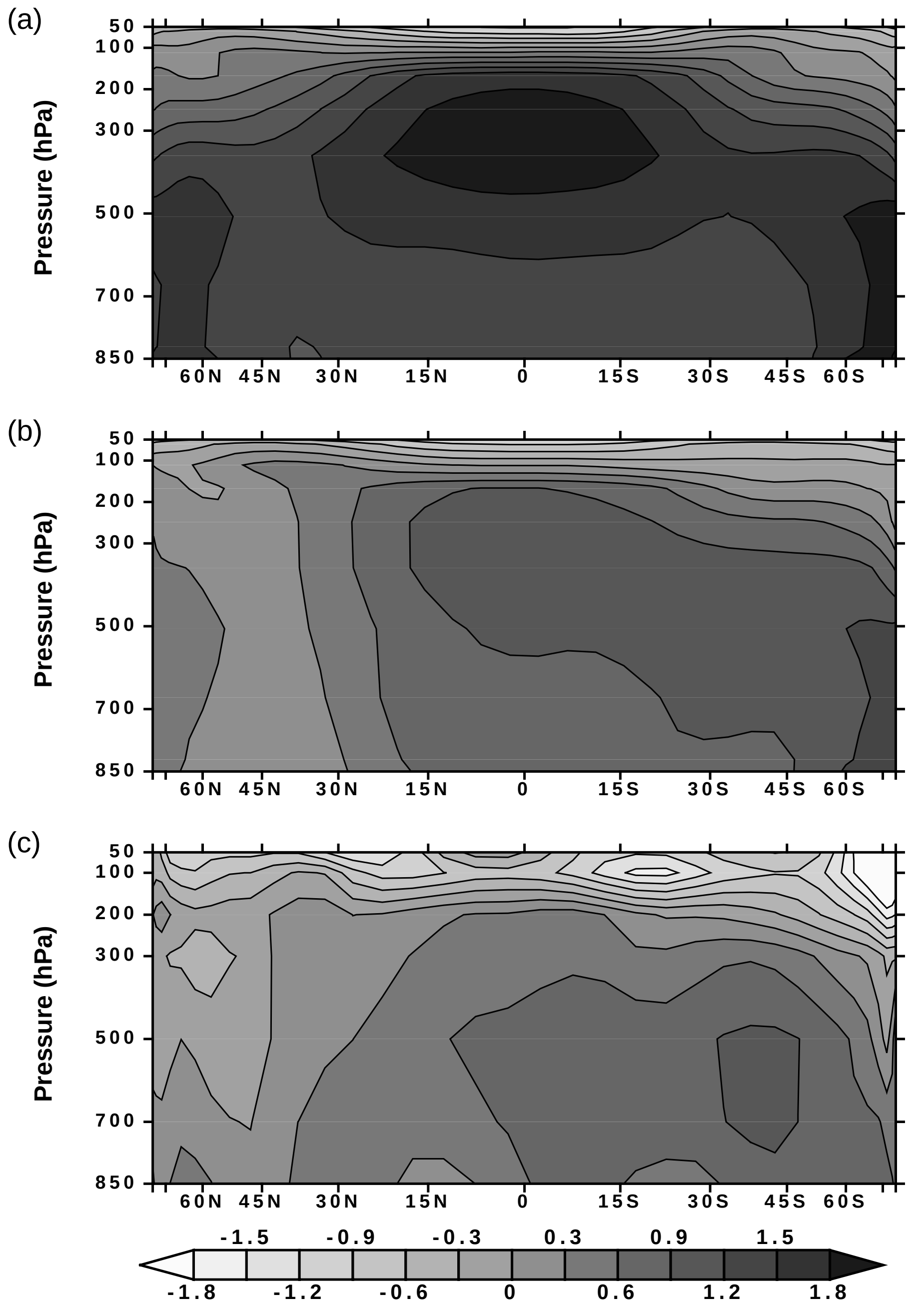


Figure 7.11 — Modelled and observed changes in the zonal-mean, annual-average temperature structure of the atmosphere (°C). (a) Model results are from equilibrium response experiments performed by Taylor and Penner (1994), in which an AGCM with a mixed-layer ocean was coupled to a tropospheric chemistry model and forced with present-day atmospheric concentration of CO_2 (b) and by the combined effects of present-day CO_2 levels and sulphur emissions (c). Model changes are expressed relative to a control run with pre-industrial levels of CO_2 and no anthropogenic sulphur emissions. Observed changes (c) are radiosonde-based temperature measurements from the data set by Oort and Liu (1993) and are expressed as total least-squares linear trends over the 25-year period extending from May 1963 to April 1988 (i.e., °C/25 years). Dark shading at and above the 150 hPa pressure level highlights regions of cooling. Dark shading generally below the 200 hPa level highlights warming except for the region at 300 hPa and 60°N which has cooling. For further details refer to Santer et al. *(1995).* [from page 428, Reference no. 3].

and fingerprints along with an amplitude factor for each. Then correspondence of the overall patterns can be measured by examining the amplitude factors.

In conclusion, the overall assessment of the scientific community involved with the 1995 Intergovernmental Panel on Climate Change report concerning the existence of climate change was that: '*the balance of evidence suggests a discernible human influence on global climate*'. The detection of and attribution for climate change will continue to receive attention from many scientists. One can expect many new results with more definitive conclusions in forthcoming years. The subsequent IPCC assessment of 2001 (Reference no. 7) had a stronger conclusion about the human influence on global climate stating that: '*there is new and stronger evidence that most of the warming over the last 50 years is attributable to human activities*' (Summary for Policy Makers in Reference no. 7).

CHAPTER 8

POTENTIAL IMPACTS OF CLIMATE CHANGE

8.1 INTRODUCTION

Climate change is expected to have an impact on a wide range of ecological and socio-economic areas including human health. It has been considered important to investigate these impacts at the same time as climate change itself is assessed. In its 1995 second assessment, IPCC provided an extensive report on potential impacts of climate change (Reference no. 4).

Impact assessment is still in a preliminary stage as it is difficult to quantify and most studies have been quite limited in scope. Analyses generally have used simple assumptions about climate-change conditions and have considered only limited aspects of the complex interactive stress factors. The overall impact on a system depends on both sensitivity to the climate-condition changes and the adaptability and compensating factors that the system itself possesses. In many cases actual impacts will depend on regional climates for which the estimates of change are far more uncertain than for global-mean conditions.

Some examples of potential impacts are presented here. These were chosen to focus on areas in which operational meteorologists may be directly involved in discussions with government agencies or citizens.

8.2 TERRESTRIAL ECOSYSTEMS

There are a number of ways in which climate change will affect terrestrial ecosystems. They are discussed briefly below and followed by a focused discussion of several specific ecosystems.

Terrestrial ecosystems depend directly on temperature and precipitation climatology as shown in Figure 1.16. Changes in temperature and precipitation, including their extremes and seasonal or daily variations, will influence the distribution of biomes in the world, including those in agriculture. Rising temperatures alone would be expected to foster the poleward migration of biome species. For example, a warming could be expected to improve options for agriculture in subarctic regions.

Changes in climate will influence other determinants of the ecosystem condition such as disease, pest cycles and the incidence of fires. The increase in atmospheric carbon dioxide concentrations is expected to increase the primary productivity of plants, i.e. make them grow faster. This could change the balance among plants competing for the same space.

The largest adverse impacts on the ecosystem are anticipated to be from direct human activity itself. The clearing of land for agriculture and urbanization and the segmentation of ecosystems will be important factors. More unfavourable impacts are likely in tropical and subtropical developing countries compared to developed countries because of population pressures and lack of resources for adaptation and mitigation responses to climate change impacts.

8.2.1 AGRICULTURE (PLANT CROPS)

Climate change will affect crops in a number of ways. These include the growth process of crop plants and those of insects, weeds and diseases as discussed above.

Growth will be affected by changes in CO_2 concentration as well as those in temperature, moisture supply and severe weather. CO_2 increases alone are expected to increase the productivity of annual crops, particularly those that may be limited by existing concentrations of carbon dioxide (i.e. most crops including wheat, rice, barley, cassava and potato, and most trees). For these crops, increases on the order of 30 per cent would be expected for a doubling in CO_2 if there were no other changes in conditions. The increases would be less for plants that have a special CO_2-concentrating mechanism (such crops as maize, millet, sugar cane, sorghum, and many tropical grasses).

Temperature and moisture supply have a dominant influence on crop growth. Each crop type has an optimal temperature range for growth. Moisture supply is critical throughout the growth period of the crop. The moisture supply depends on both precipitation and evaporation (evapotranspiration from the plant). The diurnal and day-to-day variability for temperature and moisture supply along with mean values are important climatic factors for plants because of the negative impact of extremes.

Very preliminary impact assessments have been made for a selection of crops using simple estimates of climate change. The estimates were derived from climate-model simulations, historical data, or just an outright specification of a temperature change. The impacts show extreme variability, often ranging from increases to decreases of yields for a given crop. Some examples are shown for areas in Africa, south Asia, Latin America, and western Europe in Tables 8.1 to 8.4. One may find comparable summaries for all other major regions of the Earth in chapter 13 in the 1995 IPCC Working Group II report (Reference no. 4). The climate models referenced are briefly described earlier in Chapter 4.

It is clear from these tables that there is great uncertainty in the quantitative impact assessments of climate change on agricultural crops. It is strongly suggested that crop yields and productivity will vary a great deal from one region to another. Some areas will see improvement in yields, others will see decreases. Agricultural patterns will change. It is likely that overall impacts will be significant. The overall effects for a region will depend on many factors in that region such as irrigation options, current agricultural infrastructure, adaptations in farming practices and so forth. It will take serious planning efforts to deal most constructively with the changes.

8.2.2 FORESTS

Overall regional impacts of climate change for forests are briefly summarized on p. 26 of Reference no. 4. As the projected temperature increases are smaller in tropical latitudes, tropical forests will be affected less than those in other latitudes. However, climate change in terms of the amount and seasonality of rainfall could have larger impacts. Even so, other human impacts will likely affect tropical forests more than does climate change. Temperate forests will be impacted by temperature, precipitation and CO_2 changes differently from region to region. However, the negative aspects of such changes on temperate forests will be minimized by reforestation and forest management programmes, since most temperate forests are located in developed countries. Boreal forests will be most affected by climate change as warming is expected

Table 8.1 — Selected crop studies for Africa and the Middle East for climate-change scenarios from climate models, observations, and prescribed changes. [from page 438, Reference no. 4].

Study	*Scenario*	*Geographic scope*	*Crop(s)*	*Yield impact in per cent*	*Other comments*
Eid. 1994	GISS, GFDL, UKMO	Egypt	Wheat Maize	-75 to -18 -65 to +6	w/CO_2 effect; also temperature and precipitation sensitivity; adaptation would require heat-resistant variety development.
Schulze *et al*., 1993	+2°C (1)	South Africa	Biomass Maize	decrease increase	Mapped results, not summarized as average change for entire region.
Muchena, 1994	GISS, GFDL, UKMO	Zimbabwe	Maize	-40 to -10	w/CO_2 effect; also temperature and precipitation sensitivity; adaptation (fertilizer and irrigation) unable to fully offset yield loss.
Downing, 1992	+2/+4°C, ± 20 per cent precipitation	Zimbabwe Senegal Kenya	Maize Millet Maize	-17 to -5 -70 to -63 decrease	Food availability estimated to decline in Zimbabwe; carrying capacity fell 11 to 38 per cent in Senegal; overall increase for all crops in Kenya with zonal shifts.
Akong'a *et al*., 1988 broader socio-	Historical droughts, sensitivity		Kenya livestock	Maize, of drought	negative effects Considered economic impacts, small-holder impacts, a policy implications.
Sivakumar, 1993	1945-1964 vs. 1965-1988	Niger West Africa	Growing season	reduced 5-20 days	Crop variety development, timely climate information seen as important adaptation strategies.

Study	*Scenario*	*Geographic scope*	*Crop(s)*	*Yield impact in per cent*	*Other comments*
Rosenzweig and Iglesias (eds.), 1994[1]	GCMs	Pakistan	Wheat	-61 to +67	UKMO, GFDL, GISS, and +2°C, +4°C, and ± 20% precipitation range is over sites and GCM scenarios with direct CO_2 effect; scenarios w/o CO_2 and w/ adaptation also were considered; CO_2 effect important in offsetting losses of climate-only effects; adaptation unable to mitigate all losses.
		India	Wheat	-50 to +30	
		Bangladesh	Rice	-6 to +8	
		Thailand	Rice	-17 to +6	
		Philippines	Rice	-21 to +12	
Qureshi and Hobbie, 1994	average of five GCMS	Bangladesh	Rice	+10	GCMs included UKMO, GFDLQ, CSIRO9, CCC and BMRC; GCM results scaled to represent 2010; includes CO_2 effect.
		India	Wheat	decrease	
		Indonesia	Rice	-3	
			Soyabean	-20	
			Maize	-40	
		Pakistan	Wheat	-60 to -10	
		Philippines	Rice	decrease	
		Sri Lanka	Rice	-6	
			Soyabean	-3 to +1	
			Coarse grain	decrease	
			Coconut	decrease	
Parry *et al.*, (eds), 1992	GISS	Indonesia	Rice	approx. -4	Low estimates consider adaptation; also estimated overall loss of farmer income ranging from $10 to $130 annually.
			Soyabean	-10 to increase	
			Maize	-65 to -25	
		Malaysia	Rice	-22 to -12	Maize yield affected by reduced radiation (increased clouds); variation in yield increases; range is across seasons.
			Maize	-20 to -10	
			Oil palm	increase	
			Rubber	-15	
Matthews *et al.*, 1994a, 1994b	three GCMs	Thailand sites	Rice	-5 to +8	Range across GISS, GFDL, and UKMO GCM scenarios and crop models; included direct CO_2 effect; varietal adaptation was shown to be capable of ameliorating the detrimental effects of a temperature increase in currently high-temperature environments.
		India	Rice	-3 to +28	
		Bangladesh		-9 to +14	
		Indonesia		+6 to +23	
		Malaysia		+2 to +27	
		Myanmar		-14 to +22	
		Philippines		-14 to +14	
		Thailand		-12 to +9	

1 Country studies were by Qureshi and Iglesias, 1994; Rao and Sinha, 1994; Karim *et al.*, 1994; Tongyai, 1994; and Escaño and Buendia, 1994, for Pakistan, India, Bangladesh, Thailand, and the Philippines, respectively.

Table 8.2 — Selected crop studies for south and south-east Asia for climate change scenarios from climate models. [from page 439, Reference no. 4].

to be largest at high latitudes. Increased fire and pest outbreaks will negatively impact the southern regions of boreal forests, whereas the increased temperature and moisture supply in the northern regions of the boreal forests will enhance the forest, which is expected to advance northward into the tundra.

Forest fires are a matter of special interest. Forest fires occur commonly in seasonally dry forest areas due to human and natural causes. They play an important role in the ecosystem dynamics in tropical, temperate and boreal zones. Handling them is an important part of forest management. In some cases of major fires there can be extreme danger to human life and property locally, as well as region-wide health effects from air pollution. Temperature increases due to climate change are expected to increase drought conditions, which would lead to more favourable conditions for forest fires in seasonally dry areas. This would be of particular concern to areas that do not have integrated fire, pest and disease management.

8.2.3 DESERTS, LAND DEGRADATION, AND DESERTIFICATION

Roughly 30 per cent of the earth's land surface is desert or semi-desert; as shown by the dry climate areas (BS and BW classifications) in Figure 1.20. These are areas where the lack of moisture is a serious impediment to the growth of plants. Adjacent to these areas and elsewhere are regions, estimated at 17 per cent of the total earth land surface, where

Study	Scenario	Geographic scope	Crop(s)	Yield impact in per cent	Other comments
Baethgen, 1992, 1994	GISS, GFDL, UKMO[1]	Uruguay	Barley Wheat	-40 to -30 -30	w/ and w/o CO_2; with adaptation, losses were 15 to 35 per cent; results indicate increased variability.
Baethgen and Magrin, 1994	UKMO	Argentina Uruguay	Wheat	-10 to -5	w/CO_2; high response to CO_2, high response to precipitation.
Siquera *et al.*, 1994; Siquera, 1992	GISS, GFDL UKMO[1]	Brazil	Wheat Maize Soybean	-50 to -15 -25 to -2 -10 to +40	w/CO_2; w/o adaptation; adaptation scenarios did not fully compensate for yield losses; regional variation in response.
Liverman et al., 1991, 1994	GISS, GFDL, UKMO[1]	Mexico	Maize	-61 to -6	w/CO_2; adaptation only partly mitigated losses.
Downing, 1992	+3°C. 25 per cent precip.	Norte Chico Chile	Wheat Maize Potatoes Grapes	decrease increase increase decrease	The area is especially difficult to assess because of the large range of climates within a small area.
Sala and Paruelo, 1992, 1994	GISS, GFDL, UKMO[1]	Argentina	Maize	-36 to -17	w/ and w/o CO_2; better adapted varieties could mitigate most losses.

[1] These studies also considered yield sensitivity to +2 and +4°C and -20 and +20 per cent change in precipitation.

Table 8.3 — Selected crop studies for Latin America for climate-change scenarios from climate models and prescribed changes. [from page 444, Reference no. 4].

'desertification' ascribed to human activity is occurring. This process is an ecological degradation which causes economically-productive land to become less productive, more desert-like and incapable of continuing to sustain an existing community of people. Currently about one sixth of the earth's population lives in regions where such desertification is occurring.

Desertification involves a number of factors. Soil that is cultivated for agriculture can be eroded by water flow and wind. Salinization of soils can occur near coastlines due to sea-level rises. Inland salinization can occur due to salt accumulation related to erosion, seepage and wind deposition. Overall significant causes of desertification can be traced to overcultivation, overstocking, fuel and wood collection, salinization and urbanization.

It is difficult to define the effects of global climate change in these arid areas, especially where direct effects due to human activity are already taking place. Generally climate-change temperature increases would be expected to increase stresses on plants. This would tend to make desert conditions more severe and to accentuate desertification processes. The increase in CO_2 would be expected to reduce plant transpiration and to increase the water use efficiency of plants.

Climate model projections for changes in precipitation are very uncertain. Impacts of precipitation changes depend critically on changes in distribution throughout the year and in extreme events, aspects which are not reliably handled by the models. Most model simulations to date do not suggest significantly wetter conditions in arid regions.

In summary, local environmental change due to human activity (desertification) is significant in many arid areas and is independent of global climate change. It is not expected that climate change will offset the desertification process, but rather that climate-change-related factors, such as increased drought conditions resulting from rising temperatures, will increase the vulnerability of land to desertification. Many arid and semi-desert areas are in developing countries where the negative impacts would be most severe.

8.3 FRESHWATER RESOURCES MANAGEMENT

Analysis of climate-change impacts on freshwater resources must include their effects on both water supply and water demand.

A rough analysis of water supply can be based on the runoff of surface water, which depends on the difference between precipitation over a river catchment

Study	*Scenario*	*Geographic scope*	*Crop(s)*	*Yield impact in per cent*	*Other comments*
Oleson *et al.*, 1993	*	Northern Europe	Cauliflower	Increase	Quality affected by temperature; longer season.
Goudriaan and Unsworth, 1990	+3°C	Northern Europe	Maize (fodder)	Increase	Shift to grain production possible.
Squire and Unsworth, 1988	+3°C	Northern Europe	Wheat	Increase	
Kettunen *et al.*, 1988	GCMs	Finland	Potential yield	+10 to +20	Range is across GISS and UKMO GCMs.
Rötter and van Diepen, 1994	+2°C (winter), +1.5°C summer	Rhine area	Cereals, sugar beet, potato, grass	+10 to +30	Also +10 per cent winter precipitation; includes direct effect of CO_2; range is across crop. agroclimatic zone, and soil type; decreased evapotranspiration (1 to 12 per cent), except for grass.
U.K. Dept. of Environment, 1991	GCMs +1, +2°C	U.K.	Grain, horticulture	Increase or level increase	Increased pest damage; lower risk of crop failure.
Wheeler *et al.*, 1993	*	U.K.	Lettuce	Level	Quality affected; more crops per per season possible.
Semonov *et al.*, 1993	*	U.K./France	Wheat	Increase or decrease	Yield varies by region; UKMO scenario negative; includes adaptation and CO_2.
Delécolle *et al.*, 1994	GCMs** +2, +4°C	France	Wheat, maize	Increase or level	Northward shift; w/adaptation, w/CO_2; GISS, GFDL and UKMO GCMs.
Iglesias and Minguez, 1993	GCMs**	Spain	Maize	-30 to -8	w/adaptation, w/CO_2; irrigation efficiency loss; see also Minguez and Iglesias, 1994.
Santer, 1985	+4°C	Italy/Greece	Biomass	-5 to +36	Scenarios included -10 per cent precipitation.
Bindi *et al.*, 1993	+2, +4°C and *	Italy	Winter wheat	Not estimated	Crop growth duration decreases; adaptation (using slower developing varieties) possible.

* Climate scenarios included GISS, GFDL and UKMO and time-dependent scenarios, using GCM methodology, based on emission scenarios proposed by the IPCC in 1990. Composite scenarios for temperature and precipitation were based on seven GCMs and scaled by the global-mean temperature changes associated with the IPCC 1990 emissions scenarios for the years 2010, 2030, and 2050 (Barrow, 1993).

** These studies also considered yield sensitivity to +2 and +4°C and -20 and +20 per cent change in precipitation.

Table 8.4 — Selected crop studies for Western Europe for climate-change scenarios from climate models and prescribed changes. [from page 445, Reference no. 4].

and evaporation over the same area. This approach would neglect the effects due to groundwater recharge and salinization due to sea-level rise and so forth. As discussed earlier, climate-change estimates for precipitation and evaporation from climate models have considerable uncertainty. It is useful to describe the water supply in terms of the amount available per person so that population impacts are included but separated from climate-change impacts.

An impact analysis was made for the year 2050 using the results of three climate-model simulations for climate change. Results for selected countries are shown in Table 8.5. The values for the three quantities shown are for: current (1990) fresh water availability in m^3/yr/person; availability in the year 2050 assuming current water amounts but with the increased population expected; and the range of water availability in the year 2050 based on three transient climate-model scenarios. Comparison of the second and third quantities shows the predicted impact of climate change on water availability.

As can be seen in Table 8.5, there is a wide range of estimates for the water availability with climate-change conditions. This is due to the significant variations in precipitation and evaporation among the three climate models. Predictions for any given country range from increases to decreases in most cases. Clearly there is great uncertainty in the results. However, overall the number of countries with shortages in water supply is expected to increase by the year 2050. This is due to population increases which are not offset by precipitation increases (actually precipitation minus evaporation increases) even for the most favourable of the three model predictions.

A value of 1000 m^3/yr/person has been used as the baseline for minimum water requirement. Based on this number, only two of the 21 countries listed in Table 8.5 had a water shortage in 1990, eight of these countries would have a water shortage in 2050 due to population increase alone, and anywhere from seven to eleven would have a water shortage in 2050 based on both population and climate change projections.

Freshwater demands are expected to increase with the warming associated with climate change. However, this impact will be combined with population growth, economic factors, and changes in agricultural, industrial and domestic practices, acting at the same time. Irrigation for agriculture is the biggest user for water extracted from the natural reservoirs (rivers, lakes and ground water). Temperature increases are expected to increase irrigation requirements more than precipitation enhancement would decrease such requirements. Impacts of climate change on municipal uses for domestic and industrial purposes are not well defined.

8.4 SEA-LEVEL RISE

Observations have detected sea-level rises over the last 100 years of between eight and 31 cm (cf. Figure 2.14). The overall global mean value is estimated to be between 10 and 25 cm. This is related to the observed net melting of glaciers and ice fields and the net warming of the surface air temperature. Temperature increases in the ocean lead to change in sea level due to thermal expansion. Thermal expansion and ice melting are estimated to be of comparable importance for the sea-level changes. There are local variations in the rate of sea-level change due to geological factors of ‘post-glacial rebound’ (rising of earth surface as the weight of an overlying ice mass is removed) and other earth crust movements.

Table 8.5 — Water availability (m^3/yr/person) in 2050 for the present climatic conditions and for three transient climate model scenarios (GFDL, UKMO, MPI) compared with the present. The range in simulated values for the three climate models is shown in the right hand column. [from page 478, Reference no. 4].

Country	*Present Climate (1990)*	*Present Climate (2050)*	*Scenario Range (2050)*
China	2,500	1,630	1,550–1,780
Cyprus	1,280	820	620–850
France	4,110	3,620	2,510–2,970
Haiti	1,700	650	280–840
India	1,930	1,050	1,060–1,420
Japan	3,210	3,060	2,940–3,470
Kenya	640	170	210–250
Madagascar	3,330	710	480–730
Mexico	4,270	2,100	1,740–2,010
Peru	1,860	880	690–1,020
Poland	1,470	1,250	980–1,860
Saudi Arabia	310	80	30–140
South Africa	1,320	540	150–500
Spain	3,310	3,090	1,820–2,200
Sri Lanka	2,500	1,520	1,440–4,900
Thailand	3,380	2,220	590–3,070
Togo	3,400	900	550–880
Turkey	3,070	1,240	700–1,910
Ukraine	4,050	3,480	2,830–3,990
United Kingdom	2,650	2,430	2,190–2,520
Vietnam	6,880	2,970	2,680–3,140

Climate-change warming is expected to cause a sea-level rise of between 20 and 100 cm by the year 2100 as ocean warming and ice melting occurs at a faster rate than in the past 100 years. Figure 8.1 shows the wide range in the estimates which have been made. This wide range results from great uncertainty about how ice melt will proceed in polar regions. The rate of sea-level rise is projected to be anywhere from the same to five times greater than that experienced in the last 100 years.

The rise in sea level will have important impacts on coastal zones and small islands. Coastal erosion, flooding, saltwater intrusion and sedimentation changes will be factors that impact on human settlements, agriculture, freshwater supply and quality, fisheries, and human health. Millions of people will be affected in countries such as Bangladesh, Benin, China, Egypt, India, Japan, The Netherlands and Nigeria. In some, such as the Marshall Islands, entire countries will be affected. Table 8.6 shows estimates for impacts for a 100 cm sea-level rise.

8.5 STORMS

Changes in the frequency, intensity and location of storms may be an important aspect of climate-change impacts over many parts of the world. Tropical cyclones, i.e. typhoons and hurricanes, affect many areas in and adjacent to the Indian, Pacific and Atlantic Oceans, primarily in the Northern Hemisphere. Damage caused by flooding and high winds can be devastating for land areas. In the middle latitudes, high winds and precipitation associated with extratropical cyclones can have major impacts even at locations distant from the oceans. Some comments on the effects due to climate change were made in Chapter 6 as part of the discussion on changes in extreme events.

Both tropical cyclones and the severe weather aspects of extratropical cyclones are of regional scale (mesoscale) and are difficult to represent in global-climate models. Assessments of modification of these systems due to climate change up to now have been based primarily on inferences from changes in large-scale conditions.

The ENSO cycle in the tropical Pacific Ocean area is known to affect both tropical and extratropical cyclones. In particular the warm El Niño phase of ENSO is associated with enhanced extratropical storminess in the western and southern United States. Accordingly, it is important to identify the climate-change impact on ENSO. Global climate models have been able to represent ENSO, but further experimentation is needed to isolate climate change impacts.

8.6 HUMAN HEALTH

8.6.1 INTRODUCTION

It is anticipated that global climate change will have a wide-ranging and net adverse impact on human health, including increased loss of life. This will be due both to direct causes such as the increased severity of heat waves and to indirect effects such as changes in local food productivity and in the range of diseases transmitted by organisms in air or water. Direct health impacts will also result from concurrent environmental changes such as in the concentration of toxic and carcinogenic air pollutants. Figure 8.2 summarizes some of the major direct and indirect impacts.

Many impacts will result from disturbances in ecological systems. Such changes could bring disease conditions to a human population that was formerly outside the range of such conditions. As an example, atmospheric warming could lead to mosquitoes reaching higher levels in mountainous areas introducing malaria to people living there. Populations in developing countries may require additional resources to deal effectively with such changes in disease patterns.

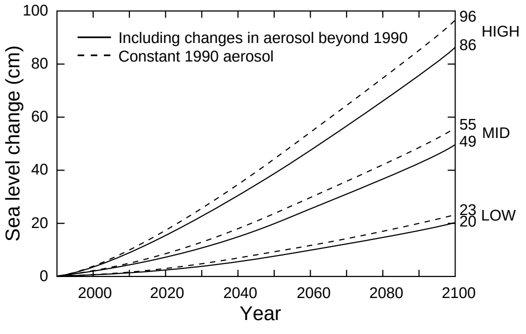


Figure 8.1 — Scenario IS92a sea-level rise from 1990 to 2100 for high, medium and low ice-melt parameter specifications. (See Chapter 7 in Reference no. 3 for more details.) [from page 296 in Reference no. 4].

	People affected		Capital value at loss		Land at loss		Wetland at loss	Adaptation/ Protection Costs	
	no. of people (1000s)	% Total	Million US$1	% GNP	km^2	% Total	km^2	Million US$1	% GNP
Antigua[2], (Cambers, 1994)	38	50	-	-	5	1.0	3	71	0.32
Argentina (Dennis *et al*.,1995a)	-	-	>5000[7]	>5	3400	0.1	1 100	>1 800	>0.02
Bangladesh (Huq *et al*.,1995: Bangladesh Government, 1993)	71 000	60	-	-	25 000	17.5	5 800	>1 000[10]	>0.06
Belize (Pernetta and Elder, 1993)	70	35	-	-	1 900	8.4	-	-	-
Benin[3] (Adam, 1995)	1350	25	118	12	230	0.2	85	>400[10]	>0.41
China (Bilan, 1993; Han *et al*., 1993)	72 000	7	-	-	35 000	-	-	-	-
Egypt (Delft Hydraulics *et al*., 1992)	4 700	9	59 000	204	5 800	1.0	-	13 100[11]	0.45
Guyana (Kahn and Sturm, 1993)	600	80	4 000	1 115	2 400	1.1	500	200	0.26
India (Pachauri, 1994)	7 100[6]	1	-	-	5 800	0.4	-	-	-
Japan (Mimura *et al*., 1993)	15 400	15	849 000	72	2 300	0.6	-	>156 000	>0.12
Kiribati[2] (Woodroffe and McLean, 1992)	9	100	2	8	4	12.5	-	3	0.10
Malaysia (Midun and Lee. 1995)	-	-	-	-	7 000	2.1	6 000	-	-
Marshall Islands[2] (Holthus *et al*., 1992)	20	100	160	324	9	80	-	>360	>7.04
Mauritius[4] (Jogoo, 1994)	3	<1	-	-	5	0.3	-	-	-
The Netherlands (Peerbolte *et al*., 1991)	10 000	67	186 000	69	2 165	5.9	642	12 300	0.05
Nigeria (French *et al*., 1995)	3 200[6]	4	17 000[7]	52	18 600	2.0	16 000	>1 400	>0.04
Poland (Pluijm *et al*., 1992)	240	1	22 000	24	1 700	0.5	36	1 400	0.02
Senegal (Dennis *et al*., 1995b)	110[6]	>1	>500[7]	>12	6 100	3.1	6 000	>1 000	>0.21
St. Kitts-Nevis[2] (Cambers, 1994)	-	-	-	-	1	1.4	1	50	2.65
Tonga[2] (Fifita *et al*., 1994)	30	47	-	-	7	2.9	-	-	-
United States (Titus *et al*., 1991)	-	-	-	-	31 600[8]	0.3	17 000	>156 000	>0.03
Uruguay[5] (Volonté and Nicholls, 1995)	13[6]	<1	1 700[7]	26	96	0.1	23	>1 000	>0.12
Venezuela (Volonté and Arismendi, 1995)	56[6]	<1	330[7]	1	5 700	0.6	5 600	>1 600	>0.03

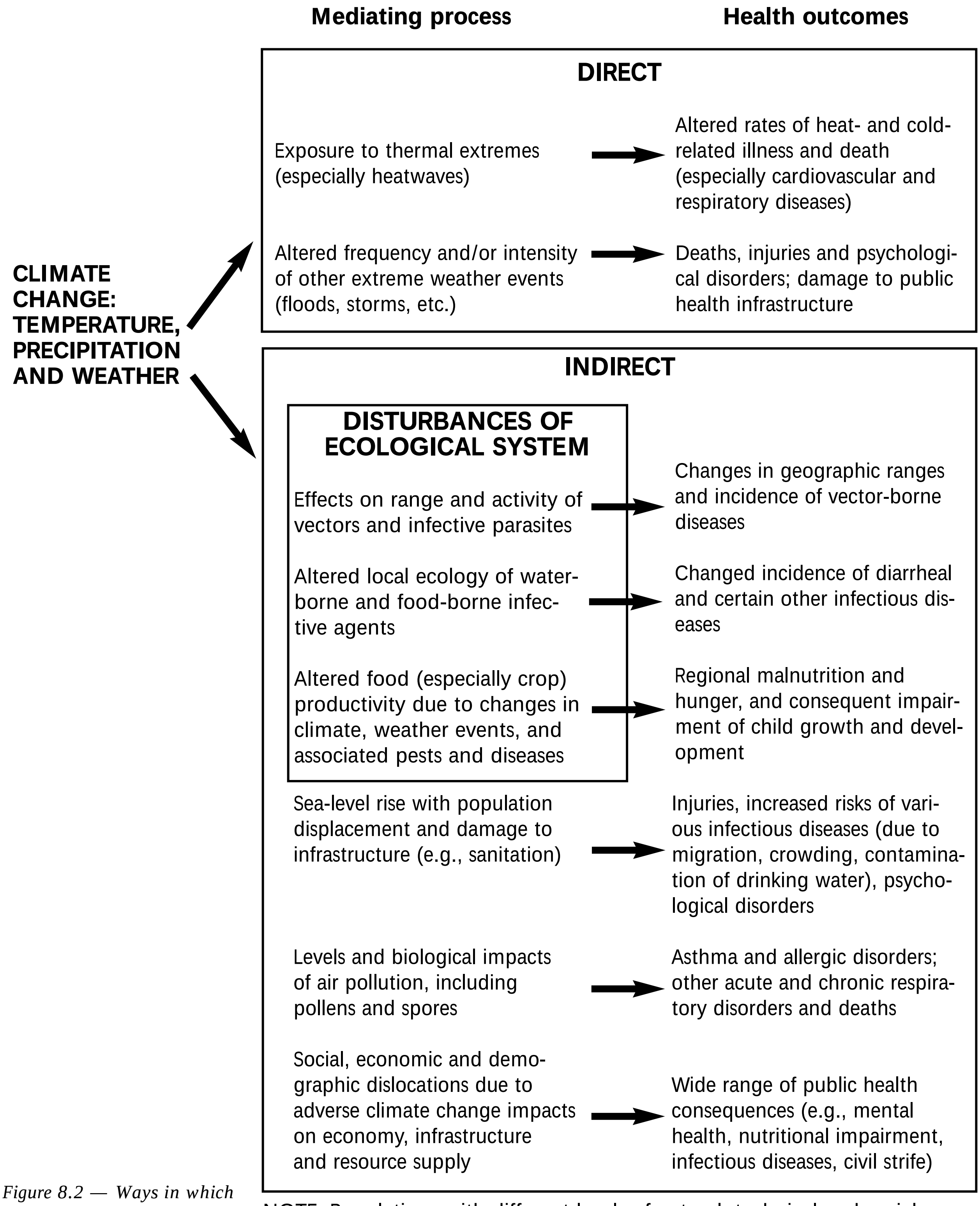
Table 8.6 — Synthesized results of country case studies. Results are for existing development and a one metre rise in sea level. People affected, capital value at loss, land at loss, and wetland at loss assume no measures (i.e. no human response), whereas adaptation assumes protection except in areas with low population density. All costs have been adjusted to 1990 US Dollars (adapted from Nicholls, 1995). [from page 308, Reference no. 4].

1 Costs have been adjusted to reflect 1990 US Dollars.
2 Minimum estimates-incomplete national coverage.
3 Precise year for financial values not given are assumed to be 1992 US$.
4 Results are linearly interpolated from results for a two metre sea-level rise scenario.
5 See also review in Nicholls and Leatherman (1995a).
6 Minimum estimates—number reflects estimated people displaced.
7 Minimum estimates—capital value at loss does not include ports.
8 Best estimate is that 20 000 kM^2 of dry land are lost, but about 5.400 kM^2 are converted to coastal wetlands.
9 Adaptation only provides protection against a one-in-20 year event.
10 Adaptation costs are linearly extrapolated from a 0.5-m sea-level rise scenario.
11 Adaptation costs include 30-year development scenarios.

A range of possible health impacts to climate change are summarized here. Research has yet to quantify such impacts and it may be difficult to isolate them from overall trends in world health (except for stratospheric ozone reduction). Nevertheless, it is important to be aware of the possibilities because of the potential impacts on whole communities or populations.

8.6.2 POTENTIAL DIRECT EFFECTS

Global warming is expected to increase the frequency of extremely hot days as discussed in Chapter 6. Recognizing that extremely hot spells are related to increases in mortality rates, it is logical to conclude that mortality due to excessively hot conditions will increase with global warming. Estimates of changes in heat-related mortality rates for selected large cities using temperatures predicted by global climate models show that by the year 2020 the rate will double and nearly quadruple by the year 2050. These results are based on current mortality rate increases due to hot weather. They neglect other factors that may operate in a hot spell such as increased air pollution and mitigation programmes that might be implemented with global warming.



NOTE: Populations with different levels of natural, technical and social resources would differ in their vulnerability to climate-induced health impacts.

Figure 8.2 — Ways in which climate change can affect human health. (from page 565 of Reference no. 4).

On the other hand, global warming is also expected to reduce the occurrence of cold conditions in parts of the world. In such areas mortality due to cold conditions, including accompanying respiratory diseases, is expected to decrease. On balance for the world it is felt that the sensitivity of death rates to hotter summers may be greater than that for the warmer winters. Thus, the overall temperature impact is negative. Considerably more research, especially in developing countries, is needed to quantify the impacts of temperature on health.

Other weather extremes such as droughts, floods and high winds also cause adverse health effects including deaths for humans. Understanding the impact of these factors due to climate change will require defining the changes in such weather extremes.

8.6.3 POTENTIAL INDIRECT EFFECTS

8.6.3.1 Vector-borne diseases

It is expected that climate change will have a significant impact on vector-borne diseases particularly in tropical and subtropical countries. 'Vector-borne' refers to the processes by which the infective agent of the disease is transmitted by a living organism (the vector) such as a mosquito, tsetse fly, bug or tick. The climate change may directly affect the living patterns of the vector which in turn will affect the spread of the disease among humans as a secondary or indirect effect

Vectors such as mosquitoes are extremely sensitive to temperature. For instance, the anopheline mosquito species, which transmits malaria, normally does not survive if the mean winter temperature is below 16-18°C, or if nighttime temperatures in summer are sufficiently low. Furthermore, the temperature at which the mosquito can incubate the malaria parasite has a lower limit, e.g., 18°C and 14°C for the *Plasmodium falciparum* and *Plasmodium vivax* parasites, respectively. If climate change increases temperature, then the mosquito can become an active vector for malaria at higher levels in mountainous country and at latitudes more poleward of the tropical areas. This can expand the malaria area to include populations which were not previously exposed and which would lack naturally-acquired immunity.

A large number of vector-borne diseases are active in the tropical and subtropical areas as listed in Table 8.7. The number of people likely to be exposed is enormous. Alteration (expansion) of the risk areas due to climate change will affect a large number of additional persons. The diseases considered to be most likely to be involved are malaria, schistosomiasis, onchocerciasis, dengue fever and yellow fever.

8.6.3.2 Water-borne and food-borne diseases

Alteration of water-borne and food-borne infectious diseases is another major potential indirect impact of climate change. Diarrheal diseases such as cholera and dysentery are spread by untreated water and water systems infiltrated by run-off water. Climate change would have an impact to the extent that it changes or increases flooding situations and by providing warmer environments for bacterial development. The rise in sea level due to climate change would be expected to increase coastal flooding and to degrade sewage disposal systems.

8.6.3.3 Agricultural productivity and food supplies

Climate-change impacts such as increased drought, desertification, increased severe weather and increased coastal flooding could have serious impacts on food supply, especially in developing countries. This could have negative impacts on nutrition and health of the population.

8.6.3.4 Air pollution

Air pollution effects may be increased by climate change. This is in addition to the human-produced air pollution which is a factor in producing the climate change in the first place. Climate changes may impact levels of pollen and other biotic allergens from birch trees, grasses, oilseed rape crops and ragweed. These effects would be further enhanced by pollutants, such as ozone, that are generated by fossil fuel combustion and the action of solar radiation.

8.6.4 STRATOSPHERIC OZONE DEPLETION AND INCREASED EARTH-SURFACE ULTRAVIOLET RADIATION

A serious health hazard has arisen from reduction of ozone in the stratosphere due to human-produced CFC and other chlorine- and bromine-containing gases. Ozone in the stratosphere helps to protect life at the Earth's surface by absorbing most of the harmful ultraviolet radiation in the sunlight. It has been estimated that the ozone concentration in the middle and high latitudes has, on average, been reduced by 10 per cent in the past ten years. The increase of ultraviolet radiation causes skin cancer, eye cataracts and damage to the local immune system in the skin. The suppression of the immune system leads to increased susceptibility to infectious diseases.

Disease	*Vector*	*Population at risk (million)*[1]	*Number of people currently infected or new cases per year*	*Present distribution*	*Likelihood of altered distribution with climate change*
Malaria	Mosquito	2400[2]	300-500 million	Tropics/Subtropics	+++
Schistosomiasis	Water snail	600	200 million	Tropics/Subtropics	++
Lymphatic filariasis	Mosquito	1094[3]	117 million	Tropics/Subtropics	+
African Trypanosomiasis (Sleeping sickness)	Tsetse fly	55[4]	250 000-300 000 cases per year	Tropical Africa	+
Dracunculiasis (Guinea worm)	Crustacean (Copepod)	100[5]	100 000 per year	South Asia/ Arabian Peninsula/ Central-West Africa	?
Leishmaniasis	Phlebotomine Sand fly	350	12 million infected. 500 000 new cases per year[6]	Asia/Southern Europe/Africa/ Americas	+
Onchocerciasis (River blindness)	Black fly	123	17.5 million	Africa/Latin America	++
American Trypanosomiasis (Chagas' disease)	Triatomine bug	100[7]	18 million	Central and South America	+
Dengue	Mosquito	1800	10-30 million per year	All Tropical Countries	++
Yellow fever	Mosquito	450	<5 000 cases per year	Tropical South America and Africa	++

Table 8.7 — Major tropical vector-borne diseases and the likelihood of change of their distribution with climate change. [from page 572, Reference no. 4].

+ = likely. ++ = very likely. +++ = highly likely. ? = unknown.

1 Top three entries are population-prorated projections (based on 1989 estimates).
2 WHO, 1995b.
3 Michael and Bundy, 1995.
4 WHO, 1994a.
5 Ranque, personal communication.
6 Annual incidence of visceral leishmaniasis; annual incidence of cutaneous leishmaniasis is between one and one and a half million cases/yr (PAHO, 1994).
7 WHO, 1995c.

The estimated increase in skin cancer is of major concern. Effects are more notable at high latitudes than in tropical regions because the depletion of ozone has been greater at high latitudes. As an example, it is estimated that if the current latitude-dependent reduction in ozone is maintained for the next few decades the incidence of the skin cancer, basal cell carcinoma, will increase by a factor of one to two per cent at very low latitudes, three to five per cent for the 15-25° latitudes, eight-12 per cent at the 35-45° latitudes, and 13-15 per cent at the 55-65° latitudes (Madronich and de Gruijl, 1993).

CONCLUDING REMARKS

The broad scope of these lecture notes should make it clear that there is a wide range of topics related to the study and understanding of climate change. It requires collaboration among persons from many different professions to answer the questions about the magnitude of climate change and the effects it is likely to have on our way of life.

The lecture notes were designed to promote sufficient understanding of the science and technology of climate change so that the student will be equipped to critically evaluate reports concerning climate change and to explain material to government representatives and lay persons. The material also gives background information to those who wish to pursue further study. It should be emphasized again that data and assessments of climate change are being continually updated. Thus, it is necessary to seek out current reports and publications to keep up to date on the topic.

The lecture notes have given some examples of the many types of impact that climate change would have on the human community, and on plant and animal communities. Some of the impacts may be considered to be favourable, but a large number are unfavourable and would require actions to reduce their negative effects. Climate change clearly implies many other kinds of changes. The world community needs to be alert and provide assistance to those in the world who lack the resources to respond to the negative impacts of climate change.

Some people still doubt that climate change has occurred despite statistical evidence at the 95 per cent significance level. Despite their doubts relating to climate change, environmental impacts of humans on the atmosphere, oceans, and biosphere are recognized. The net result of these impacts has already been an overall degradation of our environment. It is important for the meteorological community to do its part to raise awareness of projected human effects on world climate.

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