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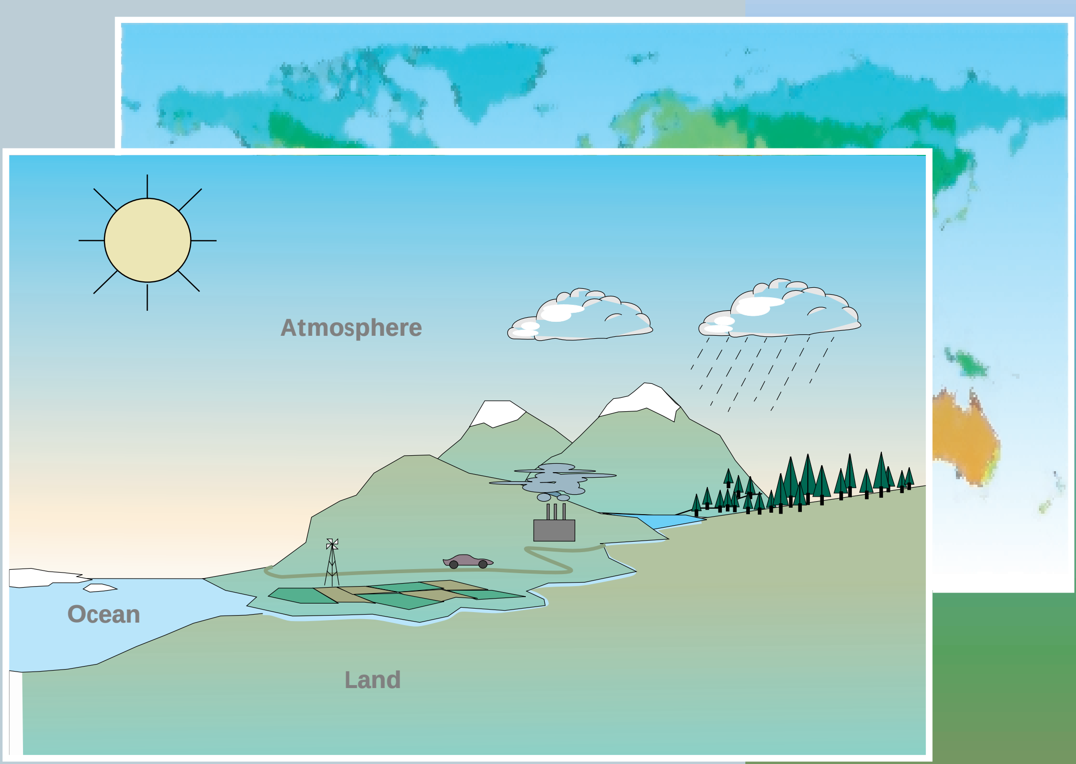

AN INTRODUCTION TO SIMPLE CLIMATE MODELS USED IN THE IPCC SECOND ASSESSMENT REPORT

IPCC Technical Paper II

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An Introduction to Simple Climate Models used in the IPCC Second Assessment Report

Edited by

John T. Houghton **L. Gylvan Meira Filho** **David J. Griggs** **Kathy Maskell**

This is a Technical Paper of the Intergovernmental Panel on Climate Change prepared in response to a request from the United Nations Framework Convention on Climate Change. The material herein has undergone expert and government review, but has not been considered by the Panel for possible acceptance or approval.

February 1997

This paper was prepared under the auspices of IPCC Working Group I, which is co-chaired by Sir John T. Houghton of the United Kingdom and Dr L. Gylvan Meira Filho of Brazil.

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Contents

Preface

This Intergovernmental Panel on Climate Change (IPCC) Technical Paper on “An Introduction to Simple Climate Models used in the IPCC Second Assessment Report” is the second paper in the IPCC Technical Paper series and was produced in response to a request made by the Subsidiary Body for Scientific and Technological Advice (SBSTA) of the Conference of the Parties (COP) to the United Nations Framework Convention on Climate Change (UNFCCC).

Technical Papers are initiated either at the request of the bodies of the COP, and agreed by the IPCC Bureau, or as decided by the IPCC. They are based on the material already in IPCC Assessment Reports and Special Reports and are written by Lead Authors chosen for the purpose. They undergo a simultaneous expert and government review, during which comments on this Paper were received from 81 reviewers from 26 countries, followed by a final government review. The Bureau of the IPCC acts in the capacity of an editorial board to ensure that review comments have been adequately addressed by the Lead Authors in the finalization of the Technical Paper.

The Bureau met in its Twelfth Session (Geneva, 3-5 February 1997) and considered the major comments received during the final government review. In the light of its observations and requests, the Lead Authors finalized the Technical Paper. The Bureau was satisfied that the agreed Procedures had been followed and authorized the release of the Paper to the SBSTA and thereafter publicly.

We owe a large debt of gratitude to the Lead Authors who gave of their time very generously and who completed the Paper at short notice and according to schedule. We thank the Co-chairmen of Working Group I of the IPCC, John Houghton and Gylvan Meira Filho who oversaw the effort, the staff of the United Kingdom Meteorological Office graphics studio who prepared the figures for publication and particularly David Griggs, Kathy Maskell and Anne Murrill from the IPCC Working Group I Technical Support Unit, for their insistence on adhering to quality and timeliness.

B. Bolin
Chairman of the IPCC

N. Sundararaman
Secretary of the IPCC

An Introduction to Simple Climate Models used in the IPCC Second Assessment Report

This paper was prepared under the auspices of IPCC Working Group I.

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SUMMARY

This Technical Paper is intended as a primer on the climate system and simple climate models (SCMs), and has two objectives: (*a*) to explain how SCMs work, the processes that are included in them, what their strengths and weaknesses are in relation to more complex models, the purposes to which they are applied, and why they have been used extensively in the Working Group I volume of the IPCC Second Assessment Report (IPCC WGI, 1996[1]); and (*b*) to fully document the procedures and assumptions used to generate the trace gas concentration, global mean temperature change, and global mean sea level rise projections presented in the SAR WGI (Section 6.3) and in the IPCC Technical Paper on Stabilization of Atmospheric Greenhouse Gases: Physical, Biological and Socio-economic Implications (IPCC TP STAB, 1997).

The major components of the climate system that are important for climatic change and its consequences, such as sea level rise, during the next century are: the atmosphere, oceans, terrestrial biosphere, glaciers and ice sheets and land surface. In order to project the impact of human perturbations on the climate system, it is necessary to calculate the effects of all the key processes operating in these climate system components and the interactions between them. These climate processes can be represented in mathematical terms based on physical laws such as the conservation of mass, momentum, and energy. However, the complexity of the system means that the calculations from these mathematical equations can be performed in practice only by using a computer. The mathematical formulation is therefore implemented in a computer program, which we refer to as a “model”. If the model includes enough of the components of the climate system to be useful for simulating the climate, it is commonly called a “climate model”. Climate system models are fundamentally different from statistical models used in some of the social sciences, which are based purely on empirical correlations and are unrelated to an underlying body of physical law.

The climate system can be represented by models of varying complexity, i.e., for any one component of the climate system a hierarchy of models can be identified. The main differences between models within a given hierarchy are:

- *The number of spatial dimensions in the model.* In a model it is necessary to represent physical quantities which vary continuously in space (e.g., temperature, humidity and wind speed) by their values at a finite number of points. The spacing between the points of the grid is the “spatial resolution”. In the most complex models of the atmosphere and ocean used to study climate (referred to as atmosphere-ocean general circulation models, or AOGCMs), such quantities are represented by a three-dimensional (longitude-latitude-height) grid with typical horizontal resolutions of several hundred kilometres. Simpler climate models may represent these physical quantities as averages over one or more spatial dimensions. Instead of, for instance, a three-dimensional grid, one might use a two-dimensional (latitude-height) grid, with each point being an average over all longitudes at a given latitude and height.

- *The extent to which physical processes are explicitly represented.* Even the most complex climate models used to project climate over the next century (AOGCMs) have a typical resolution of hundreds of kilometres in the horizontal. Many important elements of the climate system (e.g., clouds, land surface) have scales that are much smaller than this in reality. Detailed models at high resolution are available for such processes by themselves, but these are computationally too expensive to be included in a climate model. Instead, the climate model has to represent the effect of these sub-grid scale processes on the climate system at its coarse grid scale. A formulation of the effect of a small-scale process on the large-scale is called a “parametrization” (SAR WGI: Section 1.6.1). When the dimensionality of the model is reduced as described above, more processes have to be parametrized.

- *The level at which empirical parametrizations are involved.* All models rely on parametrization to represent those processes which are not explicitly represented by the model grids. The important difference between models of varying resolution and dimensionality, therefore, is the level at which parametrizations are introduced, not the need for parametrization. However, even in three dimensional AOGCMs, the large-scale behaviour of the model and the nature of processes that are explicitly computed (e.g., winds and ocean currents) can be strongly influenced by the way in which sub-grid scale processes are parametrized.

- *The computational cost of running the model*. SCMs are computationally more efficient than more complex models and can therefore be used to investigate future climate change in response to a large number of different scenarios of future greenhouse gas emissions. Such scenario analysis would be impractical with AOGCMs.

Climate models may also vary in their comprehensiveness i.e., in the number of climate components that are represented. For example, a climate model may try to model only the atmosphere, while a more comprehensive model might include the atmosphere (and atmospheric chemistry), the oceans and the terrestrial and marine biospheres.

In this report, we use the term “simple climate model” (SCM) to refer to the simplified models used in the SAR WGI (Sections 6.3, 7.5.2 and 7.5.3) to provide projections of global mean temperature and sea level change response to the IS92 emissions scenarios and the carbon dioxide (CO_2) stabilization

[1] Hereafter referred to as the SAR WG1.

profiles. The SCMs contain modules that calculate: (*a*) the concentrations of greenhouse gases for given future emissions; (*b*) the radiative forcing resulting from the computed greenhouse gas concentrations and aerosol precursor emissions; (*c*) the global mean temperature response to the computed radiative forcing; and (*d*) the sea level rise due to thermal expansion of sea water and the response of glaciers and ice sheets. These steps are briefly elaborated upon below.

Emissions to Concentrations

The calculation of future concentrations of greenhouse gases from given emissions entails modelling the processes that transform and remove the different gases from the atmosphere. For example, future concentrations of CO_2 were calculated in SAR WGI using models of the carbon cycle which include representations of the exchanges of CO_2 between the atmosphere and the oceans and terrestrial biosphere. Other greenhouse gases, rather than being exchanged between different reservoirs, are destroyed through chemical reactions. Concentrations can be derived from emissions using quite simple equations in SCMs once the atmospheric lifetimes of the gases are determined from more complex two- and three-dimensional atmospheric chemistry models.

Concentrations to Global Mean Radiative Forcing

Given the concentrations of globally uniform greenhouse gases, the direct global mean radiative forcing can be computed using simple formulae which provide a close fit to the results of detailed radiative transfer calculations. In the case of tropospheric ozone, the picture is complicated by the fact that this gas is produced from emissions of precursor gases through chemical reactions and its concentration is highly variable in space and time. In this case, concentrations are not directly computed and the radiative forcing is assumed to change based on simple linkages to other gases as a proxy for the full chemistry. Similarly, the radiative forcing due to depletion of stratospheric ozone is directly computed based on a simple relationship to emissions of chlorine and bromine containing chemicals, which has been calibrated based on the results of detailed models. Finally, the amount of aerosol in the lower atmosphere responds essentially instantaneously to changes in emissions because of the short lifetime of aerosols, so specification of an emission scenario amounts to specifying a concentration scenario. Hence, in the SCMs used in SAR WGI, global aerosol emissions are directly linked to global mean radiative forcing (both the direct and indirect components) using the results of three dimensional atmospheric general circulation models (AGCMs) which attempt to represent explicitly the processes determining the amount, distribution, and properties of aerosols in the atmosphere, and the resulting global mean forcing. These processes are poorly understood and the resultant forcings highly uncertain.

Global Mean Radiative Forcing to Global Mean Temperature

Given a scenario of global mean radiative forcing, the next step is to compute the resultant time-varying ("transient") climatic response. This depends both on the climate sensitivity and on the rate of absorption of heat by the oceans. The climate sensitivity is a measure of the global surface temperature change for a given radiative forcing and encompasses the complexity of processes responsible for the way the climate system responds to a radiative forcing, including feedback processes involving, for example, clouds, sea ice and water vapour.

The response of the SCM, for a given scenario of future greenhouse gas and aerosol precursor emissions, is governed by the climate sensitivity and a small number of parameters which control the uptake of heat by the oceans. The climate sensitivity can be estimated by four independent methods: (*a*) from simulations with three-dimensional AGCMs; (*b*) from direct observations, at the relevant temporal and spatial scales, of the key processes that determine radiative damping to space and hence climate sensitivity; (*c*) from reconstructions of radiative forcing and climate response of ancient (palaeo-) climates; and (*d*) from comparisons of ocean/climate model runs with historical global temperature records.

The climate module of the SCM only provides information about global mean temperature. For information about regional climate change, changes in other variables (e.g., precipitation), and changes in variability and extremes, three-dimensional AOGCMs are required.

Global Mean Temperature to Global Mean Sea Level Rise

Global mean sea level rise in SCMs is computed based on contributions from: (*a*) the thermal expansion of sea water, which depends on the evolving profile of temperature change in the ocean; and (*b*) glaciers, small ice-caps and ice sheets, the contributions of which are computed using simple models of these components that are driven by the global mean temperature change as computed by the SCM.

The single largest source of uncertainty in projections of future, time-dependent global mean temperature change is the equilibrium climate sensitivity, which is expected to fall within 1.5 to 4.5˚C for a CO_2 doubling. SCMs assume that the global mean temperature response to a radiative forcing perturbation depends only on the global mean value of the perturbation, and that the climate sensitivity is the same irrespective of the magnitude or direction of the radiative forcing. The dependence of climate sensitivity on the magnitude, direction, and nature of the forcing is thought to be small, in most cases, compared to the underlying uncertainty in the climate sensitivity itself (a factor of three).

The equilibrium climate sensitivity is also the single most important source of uncertainty for projections of global mean sea level rise, although the variation of temperature change with depth in the ocean and the response of glaciers and ice sheets are also important sources of uncertainty. With regard to the build-up of carbon dioxide in the atmosphere, the largest uncertainties involve interactions between the terrestrial biosphere and climate. The uncertainties in the estimated build-up of atmospheric CO_2 are thought to be small for projections spanning two to three decades, but are substantially larger for longer projections.

Both simple and complex models have important roles to play in enhancing our understanding of the range of possible future climatic changes, their impacts, and interactive effects. The more complex models are especially suited for studying those fundamental processes which are resolved by complex models but not by simple models. They also have the potential to provide credible projections of regional scale changes in climatic means and variability. Simple models can be formulated to replicate the global scale average behaviour of complex models and can be calibrated to global scale observations. Due to their computational efficiency and conceptual clarity, simple models are useful for global change scenario development and analysis, and for investigating the interactive effect of subsystem properties. The use of AOGCMs for the simulation of regional, time-varying climatic change, and the use of SCMs for more extensive sensitivity and scenario analysis, are both dictated by pragmatic considerations involving computer resources and the level of detail appropriate when coupling various components together. A long-term goal of Earth system science is the development of increasingly sophisticated coupled models of the climate system.

All climate system models used in the SAR WGI have been tested for their ability to reproduce key features of the existing climate, as well as historical and palaeo-climatic changes. While the validity of these models cannot be proven for future conditions, their ability to recover a variety of observed features of the atmosphere/ocean/biosphere system and observed changes during the recent past supports their use for projections of future climatic change.

However, many uncertainties remain regarding the modelling of the climate system. There is considerable uncertainty about the changes that might occur in some climate system processes, such as those involving clouds, in an altered climate. The effect of aerosols on the radiation balance of the climate is also not well known. Difficult-to-predict changes in the ocean circulation could have a significant effect on both regional and global climatic changes. Unexpected changes in the flow of carbon between the atmosphere and terrestrial biosphere and/or the oceans could occur. Nevertheless, research continues to improve our basic understanding of important processes and their representation in models.

1. INTRODUCTION

1.1 Aims

This Technical Paper is intended as a primer on the climate system and SCMs, and has two objectives: (*a*) to explain how SCMs work, the processes that are included in them, what their strengths and weaknesses are in relation to more complex models, the purposes to which they are applied, and why they have been used extensively in the SAR WGI; and (*b*) to fully document the procedures and assumptions used to generate the trace gas concentration, global mean temperature change, and global mean sea level rise projections presented in the SAR WGI (Section 6.3) and in IPCC TP STAB (1997).

1.2 Climate Models as Tools for Scientific and Policy Analysis

Understanding the climate system is a problem of great intrinsic scientific interest. Our growing understanding of interactions between the atmosphere, oceans, biosphere, cryosphere and land surface is revolutionizing the Earth sciences. Moreover, in recent years, a sense of urgency has infused research on modelling the climate system. The prospect of human activities altering atmospheric composition, affecting climate globally and regionally, and ultimately affecting human economies and natural ecosystems, has stimulated the development of models of the climate system.

Clearly, it is important to have useful and credible tools for policy analysis before the climate itself changes. Thus, climate system models employed by researchers contributing to the SAR WGI are motivated, at least in part, by the desire to make timely predictions of anthropogenic climatic impacts from greenhouse gas and aerosol emissions across the chain of causality from emissions to impacts.

An important concept in climate system modelling is the notion of a hierarchy of models of differing levels of complexity, dimensionality and spatial resolution, each of which may be optimum for answering different questions. It is not meaningful to judge one level as being better or worse than another, independent of the context of analysis.

Ideally, one seeks a balance whereby each component of the climate system is represented at an appropriate level of detail. How to do this is the modeller's "art". There is no methodological crank to turn, although some overall principles are clear; for example, it would be an inefficient use of computer resources to couple a detailed model for some part of the system with little effect on the particular area of concern to one with crudely represented physical processes that dominates the model output. Einstein once quipped that, "everything should be as simple as possible, but no simpler". Generations of modellers have agonized over what "no simpler" means. This has been a particularly important issue for assessments of anthropogenic climate change conducted by the IPCC.

The most general computer models for climate change employed by the IPCC are the coupled AOGCMs (see Section 3.1), which solve the equations of the atmosphere and oceans approximately by breaking their domains up into volumetric grids, or boxes, each of which is assigned an average value for properties like velocity, temperature, humidity (atmosphere) and salt (oceans). The size of the box is the models' spatial resolution. The smaller the box, the higher the resolution. An assumption of research involving general circulation models (GCMs) is that the realism of climate simulations will improve as the resolution increases.

In practice, computing limitations do not allow models of high enough resolution to resolve important sub-grid processes. Phenomena occurring over length scales smaller than those of the most highly resolved GCMs, and that cannot be ignored, include cloud formation and cloud interactions with atmospheric radiation; sulphate aerosol dynamics and light scattering; ocean plumes and boundary layers; sub-grid turbulent eddies in both the atmosphere and oceans; atmosphere/biosphere exchanges of mass, energy and momentum; terrestrial biosphere growth, decay and species interactions; and marine biosphere ecosystem dynamics — to cite a few examples. Mismatches between the scale of these processes and computationally — realizable grid scales in global models is a well-known problem of Earth system science.

To account for sub-grid climate processes, the approach has been to "parametrize" — that is, to use empirical or semi-empirical relations to approximate net (or area-averaged) effects at the resolution scale of the model (see Section 3 for further discussion). It is important to stress that all climate system models contain empirical parametrizations and that no model derives its results entirely from first principles. The main conceptual difference between simple and complex models is the hierarchical level at which the empiricism enters.

It is essential, for example, to account for the heat and carbon that enter the oceans as the climate warms from the greenhouse effect of CO_2 emitted by fossil fuel burning. The internal mixing and transport in the oceans of this energy and mass invading at the air-sea interface are key processes that must be represented in any model used to project future CO_2, climate and sea level variations. The rate at which heat and dissolved carbon penetrate the thermocline (roughly the first kilometre of ocean depth) controls how much global warming is realized for a given radiative forcing, and how much CO_2 remains in the atmosphere. In principle, these processes could be computed by AOGCMs, but AOGCMs are presently too time-consuming to run on computers for a wide range of emission scenarios. For this reason, the global mean CO_2, temperature, and sea level projections for the IS92 emission scenarios and the CO_2

stabilization calculations presented in the SAR WGI, and similar calculations in IPCC TP STAB (1997), were carried out with simple models.

The choice of the most appropriate level of parametrization for climate system modelling is a qualitative judgement based on the best scientific knowledge and computer limitations. Consider the one-dimensional upwelling-diffusion ocean introduced by Hoffert, *et al.* (1980, 1981) and subsequently developed by many other researchers (Section 3.1), used to parametrize the world's oceans in several IPCC carbon cycle, climate and sea level calculations. In this paradigm, the three-dimensional world oceans are replaced by a single horizontally-averaged column in which carbon concentration and temperature vary with depth. The column exchanges mass and energy at its top with a well-mixed ocean surface layer; at its bottom, the column is fed by cold water from a downwelling polar sea. This one-dimensional paradigm works well at simulating historical climate and carbon cycle variations. To simplify further by replacing the column with a single well-mixed box or a purely diffusive ocean would make it too simple. A well-mixed box cannot account for the fact that the mixing time of the oceans is long compared to the rates at which carbon emissions and radiative forcing at the surface are changing. The result would be incorrect rates of heat and mass uptake over time. Things are already "as simple as possible" with a one-dimensional upwelling-diffusion ocean, so we stop there.

Another frequently asked question is: "how do we know if model predictions are credible"? Science today recognizes that there is no way to prove the absolute truth of any hypothesis or model, since it is always possible that a different explanation might account for the same observations. In this sense, even the most well-established physical laws are "conditional". Rather, the test should be whether a theory or model is false. The more independent challenges that a theory or model passes successfully, the more confidence one can have in it. Indeed, the testability of a conjecture has become a necessary condition for it to be considered in the domain of science. As Sir Karl Raimund Popper, philosopher of science and developer of the doctrine of falsifiability, put it, "Our belief in any particular natural law cannot have a safer basis than our unsuccessful critical attempts to refute it" (Popper, 1969).

The application of the falsifiability rule can be seen in the values of the climate sensitivity (Section 2.3), equivalent to the equilibrium temperature change for a CO_2 doubling, estimated by the SAR WGI to lie, most probably, in the range of 1.5 to 4.5˚C (SAR WGI: Technical Summary, Section D.2). Climate sensitivity is computed in AGCMs based on a combination of physical laws and sub-grid scale model parametrizations, but is directly specified as an input in simple ocean/climate models. At least four independent methods have been used to estimate the climate sensitivity: (*a*) from simulations with three-dimensional AGCMs (Cess, *et al.*, 1989); (*b*) from direct observations, at the relevant temporal and spatial scales, of the key processes that determine radiative damping to space and hence climate sensitivity (e.g., Soden and Fu, 1995); (*c*) from reconstructions of radiative forcing and climate response of ancient (palaeo-) climates (Hoffert and Covey, 1992); and (*d*) from comparisons of ocean/climate model runs with historical global temperature records (see Section 4.2 and Figure 10). Each method has unique disadvantages and uncertainties. However, all of these independent methods give results that are consistent with the SAR WGI range 1.5 to 4.5˚C, and are inconsistent with values substantially lower or higher.

Finally, simple climate system models appear to have the drawback of dealing only with global or zonal averages, whereas regional variations of temperature and precipitation change are needed to complete the link in integrated assessments from emissions to impacts. Again, in practice, many present-day integrated assessments are conducted with models whose core transient climate calculations are done with simple ocean/climate models using regional distributions of temperature and precipitation (typically produced by AOGCMs) that have been scaled to the global mean temperature change (Santer, *et al.*, 1990; Hulme, *et al.*, 1995).

The foregoing considerations are meant to explain the rationale underlying the use of simplified models of the climate system in the SAR, and do not suggest that a particular modelling methodology or level of complexity is inherently superior for climate system analysis for all time. Indeed, the consensus of the climate modelling community is that detailed three-dimensionally resolved models of atmosphere and ocean dynamics, and correspondingly highly resolved models of the Earth's terrestrial and marine biota, are the long-term goals of Earth system science. These modelling efforts need to proceed in parallel with, and mutually reinforce, the more idealized models of the climate system used in work relating to scenario analysis and climate policy, as the IPCC process evolves.

2. CLIMATE AND THE CLIMATE SYSTEM

Climate is usually defined as the "average weather", or more rigorously, as the statistical description of the weather in terms of the mean and variability of relevant quantities over periods of several decades (typically three decades as defined by WMO). These quantities are most often surface variables such as temperature, precipitation, and wind, but in a wider sense the "climate" is the description of the state of the climate system.

The climate system consists of the following major components: (*a*) the atmosphere; (*b*) the oceans; (*c*) the terrestrial and marine biospheres; (*d*) the cryosphere (sea ice, seasonal snow cover, mountain glaciers and continental scale ice sheets); and (*e*) the land surface. These components interact with each other, and through this collective interaction, determine the Earth's surface climate. These interactions occur through flows of energy in various forms, through exchanges of water, through flows of various other radiatively important trace gases, including CO_2 (carbon dioxide) and CH_4 (methane), and through the cycling of nutrients. The climate system is powered by the input of solar energy, which is balanced by the emission of infrared ("heat") energy back to space. Solar energy is the ultimate driving force for the motion of the atmosphere and ocean, the fluxes of heat and water, and of biological activity. Figure 1 presents a schematic picture of the climate system, showing some of the key interactions between the various components and the component properties which can change (see SAR WGI: Section 1.1).

The components of the climate system influence global and regional climate in a number of distinct ways: (*a*) by influencing the composition of the Earth's atmosphere, thereby modulating the absorption and transmission of solar energy and the emission of infrared energy back to space; (*b*) through alterations in surface properties and in the amount and nature of cloud cover, which have both regional and global effects on climate; and (*c*) by redistributing heat horizontally and vertically from one region to another through atmospheric motions and ocean currents.

In the natural state, the various flows between the climate system components are usually very close to being exactly balanced when averaged over periods of one to several decades. For example, prior to the industrial revolution, the uptake of CO_2 by photosynthesis was almost exactly balanced by its release through decay of plant and soil matter, as evidenced by the near constancy of the atmospheric CO_2 concentration for several millennia prior to about 1800 (see IPCC 1994 Report[2]: Chapter 1). However, from one year to the next there can be modest imbalances which fluctuate in sign, due to the natural

[2] IPCC (1995), hereafter referred to as IPCC94.

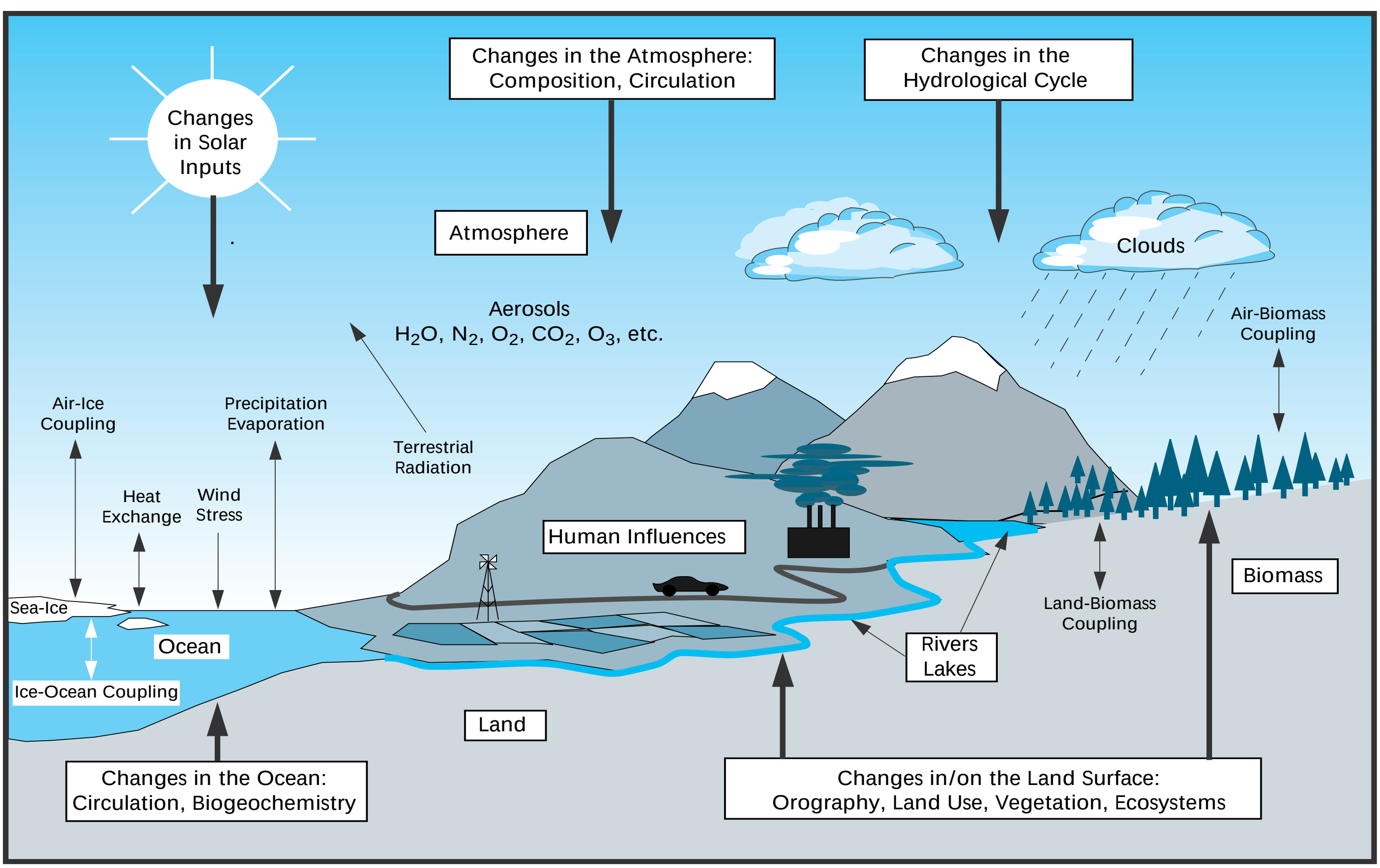

Figure 1. Schematic overview of the components of the global climate system that are relevant to climatic changes on the century time-scale (bold), their processes and interactions (thin arrows) and some elements that may change (bold arrows) (reproduced from SAR WGI, Figure 1.1).

variability of the climate system. Humans are affecting the operation of climate processes, and hence the natural balance of the climate system, through persistent regional to global scale alterations in the composition of the Earth's atmosphere and in the properties of the land surface.

2.1 Human Perturbations to the Composition of the Atmosphere

Humans are altering the concentration of greenhouse gases and aerosols, both of which influence, and are influenced by, climate. The greenhouse gases reduce the net loss of infrared heat to space, while having little impact on the absorption of solar radiation, thereby causing the surface temperature to be warmer than it would be otherwise and producing the so-called greenhouse effect (see SAR WGI: Sections 1.2.2 and 1.3.1). Aerosols, on the other hand, are important largely because of their impact on solar radiation, and have a predominantly cooling effect (see SAR WGI: Section 1.3.2).

Some greenhouse gases occur naturally but are influenced either directly or indirectly by human activity, whereas others are purely anthropogenic. The main naturally-occurring greenhouse gases are water vapour (H_2O), carbon dioxide (CO_2), ozone (O_3), methane (CH_4), and nitrous oxide (N_2O). The main groups of purely anthropogenic greenhouse gases are the CFCs, HCFCs, and HFCs (collectively known as halocarbons), and fully fluorinated species such as sulphur hexafluoride (SF_6) (see SAR WGI: Chapter 2).

Water vapour is the strongest contributor to the natural greenhouse effect, but it is the most directly linked to climate and therefore least directly controlled by human activity. This is because evaporation is strongly dependent on surface temperature, and because water vapour cycles through the atmosphere quite rapidly, about once every eight days on average. Concentrations of the other greenhouse gases, in contrast, are strongly and directly influenced by emissions associated with the combustion of fossil fuels, by forestry and most agricultural activities, and by the production and use of various chemicals.

With the exception of ozone, all of the greenhouse gases that are directly influenced by human emissions are well mixed within the atmosphere, so that their concentration is almost the same everywhere and is independent of where emissions occur. Ozone also differs from the other greenhouse gases in that it is not directly emitted into the atmosphere; rather, it is produced through photochemical reactions involving other substances — referred to as "precursors" — which are directly emitted. With regard to removal processes, all of the non-water vapour greenhouse gases except CO_2 are removed largely by either chemical or photochemical reactions within the atmosphere. Carbon dioxide, in contrast, continuously cycles between a number of "reservoirs" or temporary storage depots (the atmosphere, land plants, soils, ocean water and ocean sediments). The sources of natural greenhouse gases, and the removal processes of all greenhouse gases, are themselves influenced by climate (see SAR WGI: Sections 1.2 and 2.2).

Aerosols are suspensions of small particles in the air which influence climate primarily through their role in reflecting a portion of the incoming solar energy back to space (a direct effect) and in regulating to some extent the amount and optical properties of clouds (an indirect effect). Aerosols also absorb infrared radiation to some extent. Aerosols are produced both naturally and through human activity; natural aerosols include sea salt, dust, and volcanic aerosols, while anthropogenic aerosols are produced from burning of biomass and fossil fuels, among other sources. Some aerosols, such as dust, are directly emitted into the atmosphere. The majority of aerosols, however, are not directly emitted but, like tropospheric O_3, are produced through chemical transformation of precursor gases. All tropospheric aerosols have a short lifespan in the atmosphere due to the fact that they are rapidly washed out with rain. For this reason, and because emission source strength varies strongly from one region to another, the amount of aerosols in the atmosphere varies considerably from one region to another. The nature, amount and distribution of atmospheric aerosols are themselves influenced by climate (see SAR WGI: Sections 2.3 and 2.4).

2.2 Cloud, Surface and Dynamical Interactions

Apart from the composition of the Earth's atmosphere, a number of processes involving clouds, surface properties, and atmospheric and oceanic motions are also important to regional and global scale climate.

2.2.1 Clouds

The amount, location, height, lifespan, and optical properties of clouds exert important controls on the Earth's climate, and changes in these properties might play an important role in climatic change. The radiative impact of a given change in cloud properties, cloud amount, or cloud height depends on the location and time of year and day when the changes occur. Such changes in clouds as do occur will depend on the three-dimensional temperature and moisture fields and on atmospheric dynamical processes (i.e., those related to winds). For these reasons, three-dimensional models with high spatial resolution and a diurnal cycle hold the only prospect of correctly simulating the net effect on climate of cloud changes. However, most key cloud processes occur at scales well below the resolution of global models, so that simple area-average representations ("parametrizations") of cloud processes are required, thereby introducing the potential for substantial error in the simulated cloud changes (see SAR WGI: Sections 4.2 and 5.3.1.1.4 and Section 3 of this paper).

2.2.2 Land surface

The physical characteristics of the land surface, including the vegetation cover, have a strong effect on the absorption of solar energy and on the fluxes of heat, water vapour and momentum between the surface and atmosphere. These fluxes at any given location strongly influence the local surface climate and have effects on the atmosphere which, in some cases, extend globally. Of particular importance are changes in the extent of highly reflective ice and snow cover; as climate warms, the area of ice and snow will decrease, leading to greater absorption of solar energy and further warming. However, concurrent changes in cloud cover induced by the changes in ice and snow extent complicate the picture considerably. Correct simulation of land-surface changes and their net effect requires models with high spatial and temporal resolution on account of potential interactions with clouds and because of the spatial heterogeneity of the surface (see SAR WGI: Sections 1.4.3 and 4.4). On a time-scale of decades to centuries, changes in the vegetative cover and soil properties will also alter the exchanges of heat, moisture and momentum between the surface and atmosphere, as well as the sources and sinks of a number of greenhouse gases.

2.2.3 Oceans

The oceans play a number of important roles in the climate system and in climatic change. First, they are a major storehouse of carbon, and have played an important role in absorbing a portion of the anthropogenic CO_2 emitted up to the present. This role will continue to some extent in the future. Second, ocean currents transport substantial amounts of heat, thereby exerting a strong influence on regional climates. Changes in oceanic heat transport could significantly affect regional climatic changes, possibly causing some regions to cool temporarily and others to warm by considerably more than the global mean as the global climate warms. Third, the absorption and downward mixing of heat by the oceans considerably slows down the rate of surface warming. This reduces those impacts which depend on the rate of climatic change, but also implies that, until some time after greenhouse gas concentrations have been stabilized, there will be an irreversible commitment to more climatic change than has already occurred. Ocean currents and the rate of absorption of heat by the oceans depend on wind patterns and the exchange of heat and freshwater (through precipitation and evaporation) between the ocean and the atmosphere. At high latitudes, the presence of sea ice has a very strong effect on these exchanges, so the satisfactory simulation of sea ice is of considerable importance (see SAR WGI: Sections 1.4.2, 4.3, and 6.2; and SAR WGI: Chapter 10).

2.2.4 Atmospheric Motions

Atmospheric motions (winds) are important for transporting heat and moisture and moderating temperatures in both polar and equatorial regions. Atmospheric motions exert a strong control over the formation, nature and lifespan of clouds, thereby providing a direct coupling to both solar and infrared radiation budgets. Atmospheric heat transport and changes therein will also influence the response of sea ice and land snow cover to global mean temperature changes, thereby providing another link to the Earth's overall radiative balance. Changes in atmospheric winds, or in evaporation and precipitation due in part to changes in atmospheric winds, could also lead to significant and possibly abrupt changes in the oceans' circulation (see SAR WGI: Sections 4.2, 4.3, and 6.2).

2.3 Radiative Forcing, Feedbacks and Climate Sensitivity

The temperature of the Earth tends to adjust itself such that there is a balance between the absorption of energy from the Sun and the emission of infrared radiation from the surface-atmosphere system. If, for example, there were to be an excess of absorbed solar energy over emitted infrared radiation (as occurs with the addition of greenhouse gases to the atmosphere), temperatures would increase but, in so doing, the emission of infrared radiation to space would increase. This would reduce the initial imbalance, and eventually a new balance would be achieved, but at a new, warmer temperature (see SAR WGI: Sections 1.2 and 1.3.1).

2.3.1 Radiative Forcing

Anthropogenic greenhouse gases and aerosols affect the climate system by altering the balance between absorbed solar radiation and emitted infrared radiation, as discussed in the SAR WGI (Section 2.4). The imbalance is quantified as the "radiative forcing", which is defined as the change in net downward radiation (combined solar and infrared) at the tropopause when, for example, greenhouse gas or aerosol amounts are altered, after allowing for the adjustment of stratospheric temperatures only. The surface climate responds to the initial change in net radiation at the tropopause rather than at the surface itself or at the top of the atmosphere because the surface and troposphere are tightly coupled through heat exchanges, and respond as a unit to the combined heating perturbation. The adjustment of the stratosphere is included in the radiative forcing because the stratosphere responds quickly and independently from the surface-troposphere system. Non-anthropogenic radiative forcings relevant at the decade to century time-scales include variations in solar luminosity and volcanic eruptions, the latter producing reflective sulphate aerosols which are effective for several years if injected into the stratosphere.

The radiative forcing for a CO_2 doubling is 4.0-4.5 W m^{-2} before adjustment of stratospheric temperatures (Cess, *et al.*, 1993); allowing for stratospheric adjustment reduces the forcing by about 0.5 W m^{-2} to 3.5-4.0 W m^{-2}. If temperature were the only climatic variable to change in response to this radiative forcing, then the climate would have to warm by 1.2˚C in order to restore radiative balance. However, this very change in temperature would cause other atmospheric and surface properties to change which would lead to further alterations in the energy balance and would require further temperature changes through a series of feedback processes, which are discussed in the following section and in SAR WGI (Technical Summary, Section D).

2.3.2 Fast and Slow Feedbacks

A feedback is a process whereby an initial change in some variable ("A") leads to a change in another variable ("B") which then produces further changes in the initial variable. A positive feedback is such that the change in B leads to further changes in A in the same direction as the original change, thereby tending to amplify the initial change. A negative feedback, on the other hand, acts to diminish the initial change. Among the feedbacks which have to be considered in the calculation of global mean climatic change are the following: (*a*) *Water vapour amount:* in a warmer climate the atmospheric concentration of water vapour will increase. Since water vapour is a greenhouse gas, this represents a positive feedback; (*b*) *Clouds*: changes in clouds are difficult to calculate reliably, as noted in Section 2.2.1. Clouds have a strong radiative effect, and are, therefore, likely to produce a noticeable feedback. This feedback depends on changes in the amount, altitude and characteristics of the clouds, as well as on the reflectivity of the underlying surface, so even the sign of the feedback is uncertain; (*c*) *Areal extent of ice and snow*: a reduction in the area of sea ice and seasonal snow cover on land as climate warms will reduce the surface reflectivity, thereby tending to produce greater warming (a positive feedback). As noted in Section 2.2.2, however, concurrent changes in cloud cover complicate the picture considerably; (*d*) *Vegetation:* changes in the distribution of different biomes or in the nature of vegetation within a given biome can also lead to changes in the surface reflectivity, thereby exerting a feedback effect on climatic change; (*e*) *The carbon cycle:* the effect of climate on the terrestrial biosphere and the oceans is likely to alter the sources and sinks of CO_2 and CH_4, leading to changes in their atmospheric concentrations and hence causing a radiative feedback (see SAR WGI: Sections 1.4, 2.1, 4.2, and 4.4; and Chapters 9 and 10).

Of these feedbacks, those involving water vapour and clouds respond essentially instantaneously to climatic change, while those involving sea ice and snow respond within a few years. We therefore refer to these as "fast" feedbacks. Some vegetation and carbon cycle processes are relevant on a time-scale of decades, whereas others not listed above, such as a reduction in the area of continental ice sheets, dissolution of carbonate sediments in the ocean and enhanced chemical weathering on land (the latter two of which tend to reduce the atmospheric CO_2 concentration), require hundreds to thousands of years to unfold. These are referred to as "slow" feedbacks.

2.3.3 Climate Sensitivity: Definition

The term "climate sensitivity" refers to the steady-state increase in the global annual mean surface air temperature associated with a given global mean radiative forcing. It is standard practice to include only the fast feedback processes, including changes in water vapour, in the calculation of climate sensitivity, but to exclude possible induced changes in the concentrations of other greenhouse gases (as well as other slow feedback processes).

As noted above (in the introduction to Section 2.3), the temperature of the Earth tries to adjust itself such that there is a balance between absorbed solar radiation and emitted infrared radiation. If there is an energy surplus, temperatures will tend to increase, thereby increasing the emission of infrared radiation to space. The more strongly that infrared emission to space increases with temperature (that is, the stronger the *radiative damping*), the smaller the temperature increase required to re-establish zero net energy balance and the smaller the climate sensitivity. Changes in the albedo (reflectivity) of the atmosphere-surface system also contribute (positively or negatively) to the radiative damping. The fast feedback processes, thus, affect climate sensitivity by affecting the ease with which excess heat can be radiated to space — that is, by altering the radiative damping.

It is common practice to use CO_2 doubling as a benchmark for comparing climate model sensitivities. As reported in the SAR WGI (Technical Summary, Section D.2), the climate sensitivity for a CO_2 doubling is expected to fall between 1.5 and 4.5˚C. To the extent that the global mean temperature response depends only on the global mean forcing, any combination of greenhouse gas, solar luminosity and aerosol forcings which give the same net forcing as for a doubling of CO_2, will produce the same global mean temperature response in steady state. To the extent that the climate sensitivity is constant, the steady-state temperature response will vary in proportion to the net forcing. However, as discussed below, both of these conditions are only rough approximations.

2.3.4 Climate Sensitivity: Constancy and Independence

Given the many non-linearities associated with the fast feedback processes, which determine the climate sensitivity as defined above, one might expect that the climate sensitivity will depend both on the magnitude of the forcing and on the vertical, latitudinal and seasonal distribution of the forcing. However, experiments with a variety of models indicate that, for forcings up to the magnitude that could be experienced during the next century, the climate sensitivity is approximately constant (that is, the global mean surface temperature response is roughly proportional to the global mean forcing). Also, for a number of different forcings, the climate sensitivity is largely independent of the specific combination of factors producing a given global mean forcing. In particular, the global mean temperature response to a mixture of greenhouse gas increases is within about 10 per cent of the response to a CO_2 increase alone having the same global mean forcing as for the mixture of gases (IPCC94: Sections 4.1.1 and 4.8; and SAR WGI: Section 6.2.1.1).

On the other hand, the rough proportionality between global mean forcing and global mean temperature response established for well-mixed gases and solar luminosity variations can break down for cases involving very large and spatially or seasonally heterogeneous forcings (such as those due to variations in the Earth's orbit, which occur over periods of tens of thousands of years), or in which particularly strong interactions between the forcing and clouds occur. This appears to be the case for changes in tropospheric O_3 and in tropospheric aerosols, both of which produce much stronger spatial variations in the radiative forcing than for changes in well-mixed gases, and which have a decidedly different vertical pattern of forcing (IPCC94: Sections 4.1.1 and 4.8).

In spite of the possibility that the global mean climate sensitivity to aerosol and tropospheric O_3 changes is different from that for changes in other greenhouse gases, the SCMs used in the SAR WGI (Section 6.3) are such that the same sensitivity is assumed for all of these forcings. However, the climatic response to a given aerosol increase depends on both the climate sensitivity to aerosol increases and on the aerosol forcing, the latter being highly uncertain (ranging from -0.2 W m^{-2} to -2.3 W m^{-2}; see SAR WGI: Section 2.4.2). Thus, the uncertainty in climatic change due to possible differences in the climate response to increases in aerosols and in well-mixed greenhouse gases is, at present, overwhelmed by the uncertainty in the aerosol forcing itself.

2.3.5 Regional Climate Response

Irrespective of the extent to which the global mean temperature response depends only on the net global mean forcing, different combinations of forcings involving O_3, aerosols and well-mixed greenhouse gases will produce substantially different climatic changes in any given region. This is especially true for increases in tropospheric aerosols, where regional cooling can occur in the midst of global mean warming, and to a lesser extent for stratospheric and tropospheric O_3 changes (SAR WGI: Chapter 6). Thus, the climatic change in a given region associated with a given global mean forcing depends on the specific forcings involved when combining aerosol and ozone forcings with those of well-mixed greenhouse gases, even if the global mean temperature response is roughly the same. Furthermore, when large net negative forcings occur at the regional scale due to the effects of aerosols, the cooling effects will not be restricted to the immediate regions where aerosols occur, due to the effects of heat transport by winds and ocean currents.

There will also be strong regional variations in the climatic response to greenhouse gas increases even in the case of well-mixed gases, such as CO_2 and CH_4, whose forcing is relatively uniform from one region to the next. This is due to spatial variations in the nature and strength of various feedback processes (such as those involving snow cover, sea ice and clouds) and in atmospheric winds and ocean currents, which can be expected to change in response to overall changes in the global climate (see SAR WGI: Chapter 6).

3. SIMULATING CLIMATIC CHANGE

In order to project the impact of human perturbations on the climate system, it is necessary to calculate the effects of all the key processes operating in the climate system. These processes can be represented in mathematical terms, but the complexity of the system means that the calculations can only be performed in practice using a computer. The mathematical formulation is therefore implemented in a computer program, which we refer to as a "model". If the model includes enough of the components of the climate system to be useful for simulating the climate, it is commonly called a "climate model".

A climate model which explicitly included all our current understanding of the climate system would be too complex to run on any existing computer. For practical purposes, some compromises have to be made. The basic question is: in how much detail should the components and processes of the climate system be represented? If the representation is simplified, fewer calculations are needed and the model can be run faster or on a less powerful computer.

The most detailed model of a particular process is one which is based on fundamental physical principles which we believe to be invariant. Such a model would be applicable to any climate. In order to represent the process in a way which can be used in a climate model, additional, simplifying assumptions have to be introduced. In some cases, empirically–derived relationships are included. When this is necessary, the range of the validity of the model will inevitably become more limited. As far as possible, climate models make use of basic physical principles or of simplifications which introduce minimal uncertainty. This is necessary because the conditions of a changed climate may be quite different from current conditions, so relationships derived empirically or statistically for the current climate will not necessarily hold (SAR WGI: Section 1.6).

In the most complex climate models, physical quantities which vary continuously in three dimensions are represented by their values at a finite number of points arranged in a three-dimensional grid. This is clearly necessary because we can do only a finite number of calculations. The spacing between the points of the grid is the "spatial resolution". The finer the resolution, the larger the number of points, and the more calculations there are to be done. Hence, the resolution is limited by the computing resources available. The typical resolution that can be used in a climate model is hundreds of kilometres in the horizontal. Many important elements of the climate system (e.g., clouds, land surface variations) have scales much smaller than this. Detailed models at high resolution are available for such processes by themselves, but these are computationally too expensive to be included in a climate model. Instead, the climate model has to represent the effect of these sub-grid scale processes on the climate system at its coarse grid scale. A formulation of the effect of a small-scale process on the large-scale is called a "parametrization" (SAR WGI: Section 1.6.1). All climate models use parametrization to some extent.

Another kind of simplification used in climate models is to average over a spatial dimension. Thus, instead of a three-dimensional longitude-latitude-height grid, one might use a two-dimensional latitude-height grid, with each point being an average over all longitudes at that latitude and height. When the dimensionality is reduced, more processes have to be parametrized.

In the following sub-sections, we briefly outline the major types of models that have been developed for each of the major steps involved in simulating the climate and sea level response to anthropogenic emissions. This provides a context for the specific simple climate models that have been used by the IPCC and which are described in Section 4.

3.1 A Hierarchy of Atmosphere and Ocean Climate Models

Some of the main types of models for the atmospheric and oceanic components of the climate system are as follows:

One-dimensional radiative-convective atmospheric models. These models are globally (horizontally) averaged but contain many layers within the atmosphere. They treat processes related to the transfer of solar and infrared radiation within the atmosphere in considerable detail, and are particularly useful for computing the radiative forcing associated with changes in the atmosphere's composition. The change in atmospheric water vapour amount as climate changes must be prescribed (based on observations), but the impact on radiation associated with a given change in water vapour can be accurately computed. Radiative-convective models thus provide one means for determining one of the key feedbacks which are important to climate sensitivity through a combination of observations and well-established physical processes.

One-dimensional upwelling-diffusion ocean models. The atmosphere is treated as a single well-mixed box that exchanges heat with the underlying ocean and land surface. The absorption of solar radiation by the atmosphere and surface depends on the specified surface reflectivity and atmospheric transmissivity and reflectivity. The emission of infrared radiation to space is a linearly increasing function of atmospheric temperature in this model, with the constant of proportionality serving as the infrared radiative damping. The ocean is treated as a one-dimensional column which represents a horizontal average over the real ocean, excluding the limited regions where deep water forms and sinks to the ocean bottom, which are treated separately. Figure 2 illustrates this model. The sinking in polar regions is represented by the pipe to the side of the column. This sinking and the compensating upwelling within the column represent the global scale thermohaline circulation. This model is used primarily to study the role of the oceans in the surface temperature response to changes in radiative forcing.

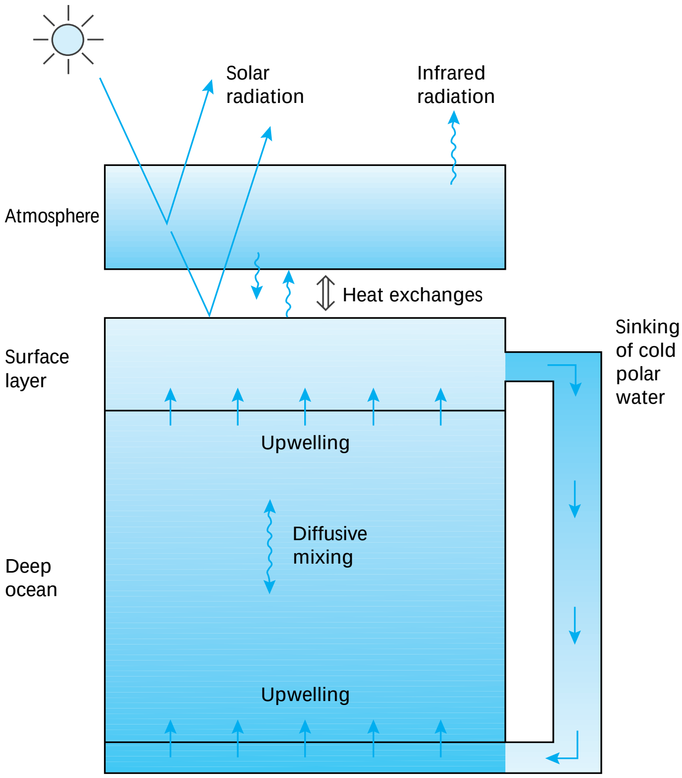

Figure 2. Illustration of the upwelling-diffusion climate model, consisting of a single atmospheric box, a surface layer representing both land and the ocean mixed-layer, and a deep ocean. Solar and infrared radiative transfers, air-sea heat exchange, and deep ocean mixing by diffusion and thermohaline overturning are all represented in this model and are indicated in the figure (based on Harvey and Schneider, (1985)).

One-dimensional energy balance models. In these models, the only dimension that is represented is the variation with latitude; the atmosphere is averaged vertically and in the east-west direction, and is often combined with the surface to form a single layer. The multiple processes of north-south heat transport by the atmosphere and oceans are usually represented as diffusion, while infrared emission to space is represented in the same way as in the upwelling-diffusion model. These models have provided a number of useful insights concerning the interaction of horizontal heat transport feedbacks and high latitude feedbacks involving ice and snow.

Two-dimensional atmosphere and ocean models. Several different two-dimensional (latitude-height or latitude-depth) models of the atmosphere and oceans have been developed (e.g., Peng and Arking (1982), for the atmosphere; Wright and Stocker (1991), for the ocean). The two-dimensional models permit a more physically based computation of horizontal heat transport than in one-dimensional energy balance models. In some two-dimensional ocean models (e.g., Wright and Stocker, 1991) the intensity of the thermohaline overturning is determined by the model itself, while in others (e.g., de Wolde, *et al.*, 1995) it is prescribed, as in the one-dimensional upwelling-diffusion model. The one-dimensional energy balance atmosphere-surface climate model has also been coupled to a two-dimensional ocean model (Harvey, 1992; de Wolde, *et al.*, 1995, and Bintanja, 1995). It is relatively easy to run separate two-dimensional ocean models for each of the Atlantic, Pacific, and Indian Ocean basins, with a connection at their southern boundaries (representing the Antarctic Ocean) and interaction with a single, zonally-averaged atmosphere.

Three-dimensional atmosphere and ocean general circulation models. The most complex atmosphere and ocean models are the three-dimensional AGCMs and ocean general circulation models (OGCMs), both of which are extensively reviewed in the SAR WGI (Chapter 5). These models divide the atmosphere or ocean into a horizontal grid with a typical resolution of 2-4° latitude by 2-4° longitude in the latest models, and typically 10 to 20 layers in the vertical. They directly simulate winds, ocean currents, and many other variables and processes characterizing the atmosphere and oceans. Both AGCMs and OGCMs have been used extensively in a stand-alone mode, with prescribed ocean surface temperatures and sea ice in the case of AGCMs and with prescribed surface temperatures and salinities, or the corresponding heat and freshwater fluxes, in the case of OGCMs. An AOGCM consists of an AGCM coupled to an OGCM, with information about the state of the atmosphere and ocean adjacent to, or at the sea surface, used to compute exchanges of heat, moisture and momentum between the two components**.**

AOGCMs compute radiative transfer through the atmosphere (explicitly modelling clouds, water vapour and other atmospheric components), snow and sea ice, surface fluxes, transport of heat and water by the atmosphere and ocean, as well as the uptake of heat by the oceans (which delays and modifies the initial surface temperature response but contributes to sea level rise through expansion of ocean water as it warms). Thus, coupled AOGCMs explicitly compute the fast feedback processes, whose interactive effect determines climate sensitivity. Because of computational constraints, however, the majority of these processes are parametrized to some extent (see SAR WGI, Sections 4.2 and 4.3, concerning processes in atmospheric and oceanic GCMs, respectively). More detailed representations are either not practical or have not been developed for use in a global model. Some parametrizations inevitably include constants which have been tuned to observations of the current climate. AOGCMs attempt to explicitly represent a large number of processes, while simpler models represent these processes by a small number of adjustable parameters.

3.2 Models of the Carbon Cycle

The carbon cycle is an integral part of the climate system, and governs the build-up of atmospheric CO_2 in response to human emissions. The key processes that need to be accurately simulated are photosynthesis and respiration on land, and the net exchange of CO_2 between the ocean and atmosphere. Because

CO_2 is chemically inert in the atmosphere and of rather uniform concentration, natural changes in atmospheric CO_2 concentration depend only on the global sum of the photosynthesis, respiration, and air-sea flows. However, each of these flows exhibits substantial variation in time and space and depends on a number of poorly understood sub-processes (SAR WGI: Chapters 9 and 10; IPCC WGII, 1996[3]: Chapter A). For example, long-term changes in both the photosynthetic and respiration flows of carbon between the terrestrial biosphere and the atmosphere are modulated by processes involving soil nutrients and micro-organisms, while the air-sea flow is modulated by a number of processes that affect the concentration of CO_2 in surface water. These include vertical mixing of total dissolved carbon and the net sinking of particulate organic matter and carbonate material into the deep ocean, which is driven in part by surface biological productivity. The latter, in turn, will be influenced by vertical mixing of nutrients and changes in temperature. Changes in ocean circulation will, thus, influence the air-sea exchange of CO_2 by altering the exchange of total dissolved carbon between the surface layer and deep ocean, and by altering biological productivity by changing the flow of nutrients from the deep ocean to surface layer.

The one-dimensional upwelling-diffusion model can be used as the oceanic part of the carbon cycle (Hoffert, *et al.,* 1981; Piehler and Bach, 1992). The global mean atmosphere-ocean exchange of CO_2, the vertical mixing of total dissolved carbon by thermohaline overturning and diffusion, and the sinking of particulate material produced by biological activity can all be represented in this model. A two-dimensional ocean model has been used as the oceanic component of the global carbon cycle (Stocker, *et al.*, 1994). Finally, OGCMs can be used as the oceanic component of the global carbon cycle, in which the model-computed ocean currents and other mixing processes are used, in combination with simple representations of biological processes and air-sea exchange (e.g., Bacastow and Maier-Reimer, 1990; Najjar, *et al.*, 1992). At the time of the SAR, CO_2 uptake calculations using three-dimensional models had been published only for standalone OGCMs, in which the circulation field and surface temperatures were fixed. In a coupled simulation, changes in both of these variables, in response to increasing greenhouse gas concentrations, would alter the subsequent uptake of CO_2 to some extent (see SAR WGI: Chapter 10).

A variety of globally aggregated box models of the terrestrial biosphere have also been developed and used, in conjunction with simple models of the oceanic part of the carbon cycle, to project future atmospheric CO_2 concentration. The commonly used global box models are quantitatively compared in Harvey (1989). Because the terrestrial biosphere is globally aggregated in the SCMs used in the SAR WGI, it is not possible to simulate separate responses in different latitude zones (e.g., net release of carbon through temperature effects at high latitudes, net uptake of carbon in the tropics due to CO_2 fertilization), as obtained in regionally resolved models (van Minnen, *et al.*, 1996). Rather, only a global mean response is simulated. Since regional responses vary non-linearly with temperature and atmospheric CO_2 concentration, the use of globally aggregated models undoubtedly introduces errors.

[3] Hereafter referred to as SAR WGII.

The role of the terrestrial biosphere in global climatic change has been simulated using relatively simple models of vegetation on a global grid with a resolution as fine as 0.5˚ latitude x 0.5˚ longitude (SAR WGI: Section 9.4). Such grid-point models simulate the distribution of potential rather than actual vegetation; to simulate the latter requires taking into account human disturbances and soil properties These models have been used to evaluate the impact on net ecosystem productivity of higher atmospheric CO_2 (which tends to stimulate photosynthesis and improve the efficiency of water use by plants)[4] and warmer temperatures (which can increase or decrease photosynthesis and increase decay processes). These models distinguish, as a minimum, standing biomass from soil organic matter. The more sophisticated varieties track the flows of both carbon and nitrogen (taken to be the limiting nutrient), and include feedbacks between nitrogen and the rates of both photosynthesis and decay of soil carbon (e.g., Rastetter, *et al.,* 1991, 1992; Melillo, *et al.,* 1993).

Grid point models of the terrestrial biosphere have been used to assess the effect on the net biosphere-atmosphere CO_2 flux of hypothetical (or GCM-generated) changes in temperature and/or atmospheric CO_2 concentration, but generally without allowing for shifts in the ecosystem type at a given grid point as climate changes. More advanced ecosystem models are being developed and tested that link biome models (which predict changing ecosystem types) with ecophysiological models (which predict carbon fluxes) (SAR WGI: Section 9.4). Simulations with these and earlier models demonstrate the potential importance of feedbacks involving the nutrient cycle and indicate the potential magnitude of climate-induced changes in terrestrial biosphere-atmosphere CO_2 fluxes. However, individual models still differ considerably in their responses (VEMAP Members, 1995). As with models of the oceanic part of the carbon cycle, such simulations have yet to be carried out interactively with coupled AOGCMs. These models also have not yet been combined with ocean carbon uptake OGCMs.

Rather detailed models of the marine biosphere, involving a number of species and interactions, have also been developed and applied to specific sites or regions (e.g., Gregg and Walsh, 1992; Sarmiento, *et al.*, 1993; Antoine and Morel, 1995).

3.3 Models of Atmospheric Chemistry and Aerosols

Atmospheric chemistry is central to the distribution and amount of ozone in the atmosphere because chemical reactions are

[4] The stimulatory effect of higher atmospheric CO_2 concentration on photosynthesis is referred to as CO_2 “fertilization”, and is discussed further in the accompanying box overleaf.

Ecosystem responses to changes in atmospheric CO_2 concentration and climate

Plants in terrestrial ecosystems occupy a central role in the terrestrial carbon cycle, as they take in atmospheric CO_2 during photosynthesis and store carbon as biomass. Photosynthetic rates are influenced by plant type, ambient CO_2 concentrations and temperature, and are often constrained by nutrients and moisture availability. Higher ambient CO_2 levels could enhance plant growth through the CO_2-fertilization effect and through an increase in water use efficiency. The response of plants to higher atmospheric CO_2 concentrations depends, in part, on the particular photosynthetic pathway used (i.e., whether the plants are C_3 or C_4 — see SAR WGI: Section 9.2). This introduces significant regional differences in the response of plants to higher CO_2. Climatic change further influences plant growth through several pathways. Warmer temperature can either increase or decrease the rate of photosynthesis but will tend to increase the rates of plant respiration, which returns carbon to the atmosphere. The decay of dead biomass (predominantly in soils) also releases CO_2 to the atmosphere; such soil respiration is a function of soil type, soil temperature, moisture and nutrient availability. Thus, changes in ambient CO_2 concentrations and climate influence ecosystem productivity in a highly non-linear and complex fashion. These ecophysiological processes are discussed in detail in the SAR WGI (Chapter 9) and SAR WGII (Chapter A).

Additional changes in ecosystem productivity are caused by changes in land-use practices, nitrogen fertilization (both fertilizer applications and nitrogen deposition through air pollution), and irrigation. However, most SCMs generally only consider deforestation, the most obvious land-use change, which has led to large and immediate changes in global carbon storage.

The relationships used in SCMs to compute the response of the terrestrial biosphere to changes in atmospheric CO_2 concentrations and climate emphasise mainly enhanced plant growth under changed conditions. The parametrizations used are largely based on short-term glasshouse experiments with responsive plant species under ideal conditions, and do not consider the complex non-linear and interactive effects, systemic feedbacks, and changes in land-use. Simulations with such models suggest that the biosphere will increase its carbon uptake under future conditions. In real ecosystems, the response could be quite different. The complexity and heterogeneity of terrestrial ecosystems and their responses thus make it difficult and dangerous to extrapolate from current conditions far into the future.

responsible for both the production and removal of ozone (O_3). The dominant chemical reactions and sensitivities are significantly different for the stratosphere and troposphere. These processes can only be adequately modelled with three-dimensional atmospheric models (in the case of the troposphere) or with two-dimensional (latitude-height) models (in the case of the stratosphere). Atmospheric chemistry is also critical to the removal of methane (CH_4) from the atmosphere and, to a lesser extent, all other greenhouse gases except water vapour (H_2O) and CO_2. In the case of CH_4, a change in its concentration affects its own removal rate and, hence, subsequent concentration changes. An accurate simulation of changes in the removal rate of CH_4 requires specification of the concurrent concentrations of other reactive species, in particular nitrogen oxides (NO_x), carbon monoxide (CO) and the volatile organic compounds (VOCs); and use of a model with latitudinal and vertical resolution. However, simple globally averaged models of chemistry-climate interactions have been developed. These models treat the global CH_4-CO-OH cycle in a manner which takes into account the effects of the heterogeneity of the chemical and transport processes, and provide estimates of future global or hemispheric mean changes in the chemistry of the Earth's atmosphere. Some of the models also simulate halocarbon concentrations and the resulting atmospheric chlorine concentration, as well as radiative effects due to halocarbons (Prather, *et al.*, 1992). An even simpler approach is to treat the atmosphere as a single well-mixed box, but to account for the effects of atmospheric chemistry by making the CH_4 lifetime depend on CH_4 concentration in a way that roughly mimics the behaviour of the above-mentioned globally averaged models or of models with explicit spatial resolution.

Atmospheric chemistry is also central to the distribution and radiative properties of aerosols, although chemistry is only part of what is required in order to simulate the effects of aerosols on climate. The key processes that need to be represented are the emissions of aerosols or aerosol precursors; atmospheric transport, mixing, and chemical and physical transformation; and removal processes (dry deposition, rain out and wash out). Since part of the effect of aerosols on climate arises because they serve as cloud condensation nuclei (leading to rain out), it is important to be able to represent the relationship between changes in the aerosol mass input to the atmosphere and, ultimately, the radiative properties of clouds. Establishing the link between aerosol emissions and cloud properties, however, involves several poorly understood steps and is highly uncertain.

Atmospheric O_3 and CH_4 chemistry is being incorporated into AGCMs for climate simulation purposes. Geographically-distributed sulphur aerosol emissions have been used as the input to AGCMs and, in combination with representations of aerosol chemical and physical processes, have been used to compute the geographical distribution of sulphur aerosol mass and the direct (cloud-free) effects on radiative forcing. Simple models have, on the other hand, considered the direct and indirect effects of aerosols originating from both industrial and biomass sources.

3.4 Models of Ice Sheets

High resolution (20 km x 20 km horizontal grid), two- and three-dimensional models of the polar ice sheets have been developed and used to assess the impact on global mean sea level of various idealized scenarios for temperature and precipitation changes over the ice sheets (e.g., Huybrechts and Oerlemans, 1990; Huybrechts, *et al.*, 1991). AGCM output has also recently been used to drive a three-dimensional model of the East Antarctic ice sheet (Verbitsky and Saltzman, 1995), but has not yet been used to assess the possible contribution of changes in mountain glaciers to future sea level rise. Output from high resolution ice sheet models can be used to develop simple relationships in which the contribution of ice sheet changes to future sea level is scaled with changes in global mean temperature.

3.5 Computation of Sea Level Rise

Sea level rise is an important output of climate, glacier and ice sheet models, but it differs from other climate system model outputs in that it is not involved in any feedbacks. That is, sea level rise itself will not affect the subsequent changes of climate to any significant degree. Furthermore, the energy involved in melting Antarctic or Greenland ice sheets and albedo effects due to changes in their area, are small compared to the forcings. Thus, it does not matter whether sea level rise is computed alongside climate model computations, or as a separate operation using climate model results. The components of sea level rise are (*a*) the thermal expansion of the ocean, which is computed from the evolving profile of ocean warming as given by a coupled atmosphere-ocean climate model; (*b*) the contribution from mountain glaciers and ice-caps; and (*c*) the contribution from the Greenland and Antarctic ice sheets (SAR WGI: Chapter 7). The latter two components require either globally-averaged or regionally-distributed temperature change results from coupled atmosphere-ocean climate models, which are used to drive glacier/ice-cap and ice sheet models.

3.6 Utilization of Simple and Complex Models

As indicated above, a wide range of models exists for most of the components of the climate system. In the remainder of this Technical Paper, we shall use the term “simple climate model” (SCM) to refer primarily to the upwelling-diffusion climate and ocean carbon cycle models, since such models were used extensively in the SAR WGI for the computation of CO_2 build-up, temperature change and sea level rise. We shall use the term “complex model” to refer to the atmospheric and ocean GCMs, whether run as stand-alone models or as coupled models. In reality, there is a continuous variation in both the complexity and comprehensiveness of climate system models. Figure 3 compares the models described above in terms of comprehensiveness and complexity (Integrated Assessment Models, also shown in Figure 3, are discussed in Section 3.6.3). By complexity, we mean the level of detail with which the individual model components are treated, while by comprehensiveness we mean the number of components included.

It should be noted that none of the models cited above represents the most complex model available. For example, very high resolution models of clouds, with a grid spacing of tens of meters and covering several tens of square kilometres, have been developed although even these include approximations of microphysical processes. Similarly, highly detailed models of plant photosynthesis and respiration have been developed and calibrated against measurements on individual leaves. Thus, even the most complex models used for simulating global scale climatic change are simplified in several important respects, and ultimately depend on parametrizations of processes that they cannot explicitly represent.

The essential common features of the models used for climate projection in the SAR WGI are that they can calculate the response of surface temperature to radiative forcing, and that they include the ocean, because of its dominant influence on the rate of climatic change. The essential difference between simple and complex models is the degree of simplification, or the level at which parametrization is introduced. Simple linked models have been used to go from emissions of a suite of gases to concentrations, climatic change, and sea level rise. Figure 4 illustrates the structure of such calculations using SCMs as done for the SAR WGI and in the IPCC Technical Paper on Stabilization of Atmospheric Greenhouse Gases (IPCC TP STAB, 1997).

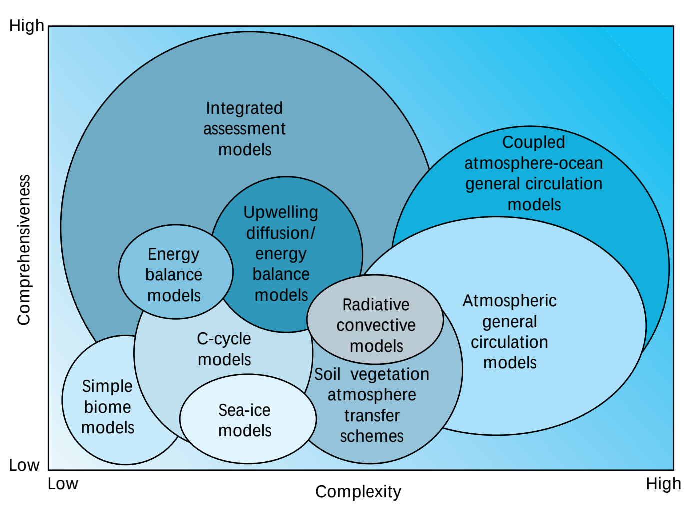

Figure 3. Schematic comparison of different climate models, and climate model components, in terms of comprehensiveness (vertical axis) and complexity (horizontal axis). Comprehensiveness refers to the number of components or processes included in the model, while complexity refers to the detail with which those components are treated.

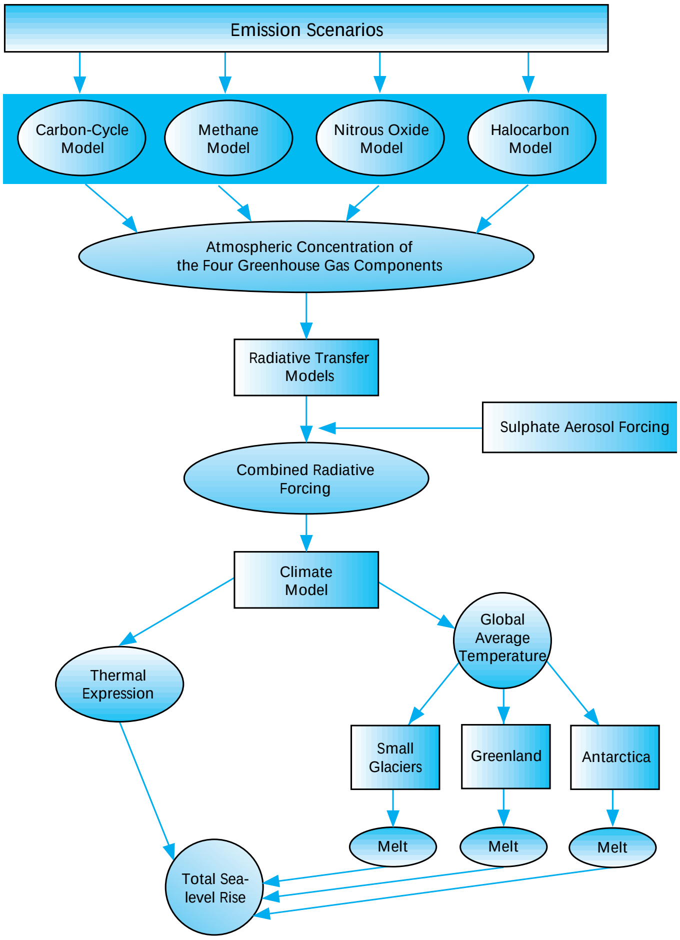

Figure 4. Steps involved in calculating greenhouse gas and aerosol concentration changes, climatic change, and sea level rise using simple climate models.

An important test of any model is its ability to replicate observations at the scale of the model resolution. Since the same world ocean is involved in the uptake of both anthropogenic CO_2 and heat, a properly formulated model should be able to simulate simultaneously both the pre-industrial (essentially steady state) profile of temperature and carbon variables with depth in the ocean, as well as the observed changes over time in atmospheric temperature, CO_2 concentration, and carbon isotope ratios. Figure 5 illustrates the ability of the one-dimensional model of Jain, *et al.* (1995) to simulate simultaneously the observed vertical ocean profiles of total carbon and carbon isotopes and the observed variation of CO_2 concentration and the ^{13}C and ^{14}C amounts from the industrial revolution to the present[5]. Despite the agreement shown in Figure 5, major uncertainties in the carbon cycle do remain, and there is the potential for significant errors in future projections of CO_2 build-up.

[5] For the sake of clarity in the presentation of the results, the effects of nuclear bomb testing (which injected large amounts of ^{14}C into the stratosphere) have not been including here. However, global cycle models are able to simulate the estimated observed oceanic uptake of ^{14}C following nuclear bomb testing, as shown, for example, in Jain, *et al.*, (1995).

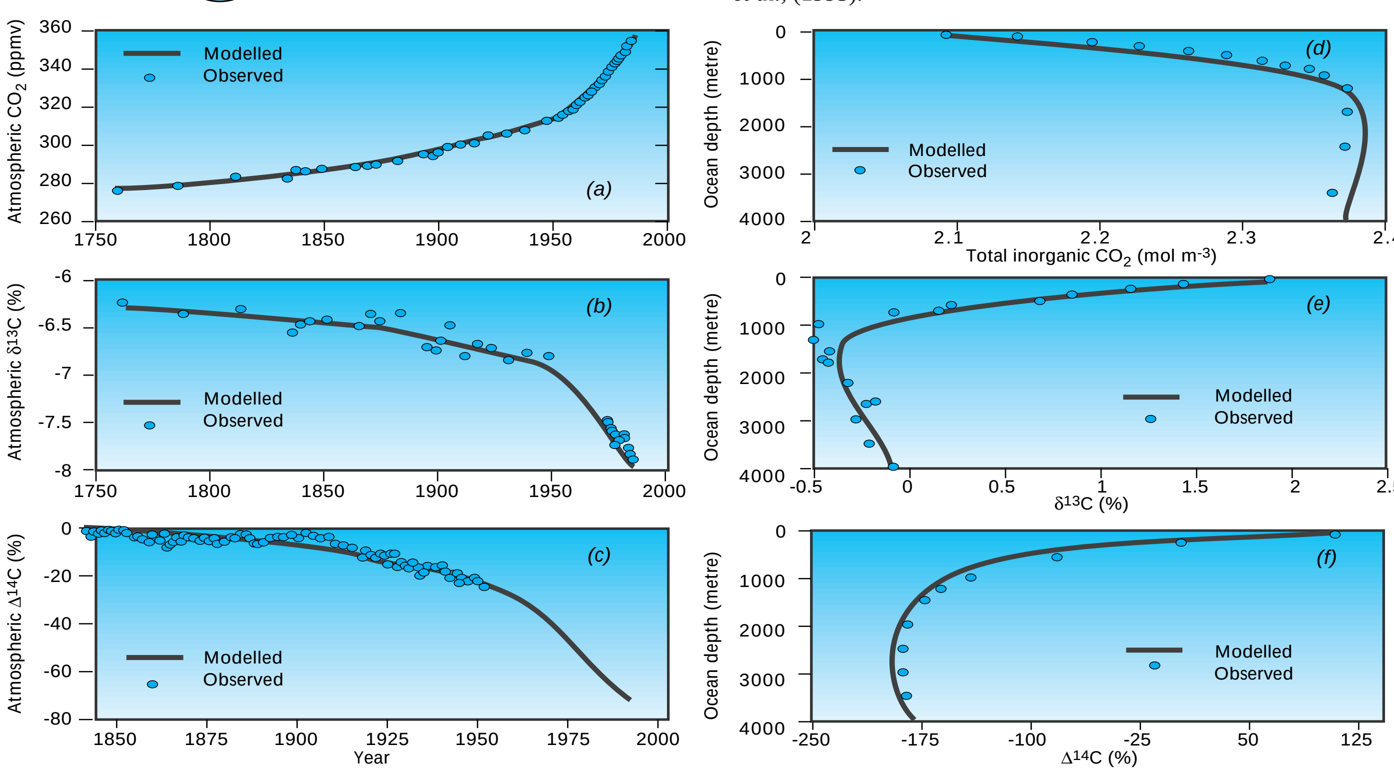

Figure 5. Comparison of observed and model-simulated historical variation of (*a*) atmospheric CO_2 concentration; (*b*) atmospheric $\delta^{13}C$ (a measure of the ratio of ^{13}C to total C ratio); (*c*) atmospheric $\Delta^{14}C$ (a measure of the ratio of ^{14}C to total C ratio); and observed and model-simulated vertical profiles of (*d*) total dissolved carbon (*e*) $\delta^{13}C$, and (*f*) $\Delta^{14}C$ as simulated by the carbon cycle model of Jain, *et. al.* (1995).

This table compares simple and complex models with reference to the different uses to which they can be put (see text for discussion and clarification).

Simple Models	*Complex Models*
Generally produce zonally- or globally-averaged results, and only for temperature and temperature changes, not for other variables such as rainfall.	Simulate the past and present geographical variation of temperature, as well as other variables of climatic interest such as rainfall, evaporation, soil moisture, cloudiness, and winds; and provide credible continental scale changes of at least some of these variables.
Cannot simulate possible changes in climatic variability as output consists of the climate change signal only.	Have the potential to simulate changes in important modes of interannual variability (e.g., *El Niño)* as well as mean values.
The effects of physical processes are approximated based on globally- or zonally-averaged computations with low temporal resolution.	Many physical processes are directly simulated, necessitating the use of a short time-step but allowing resolution of the diurnal cycle.
Climate sensitivity and other subsystem properties must be specified based on the results of complex models or observations. These properties can be readily altered for purposes of sensitivity testing.	Climate sensitivity and other subsystem properties are computed based on a combination of physical laws and sub-grid scale model parametrizations.
Sufficiently fast that multiple scenarios can be simulated, and that runs with a wide range of parameter values can be executed. Can be initialized in a steady state at little computational cost.	Computational cost strongly limits the number of cases that can be investigated and the ability to initialize in a steady state.
Useful for sensitivity studies involving the interaction of large-scale climate system components.	Useful for studying those fundamental processes which can be resolved by the model.
Analysis is easy because simple models include relatively few processes. Interpretation of simple model results may give insights into the behaviour of more complex models.	Model behaviour is the result of many interacting processes, as in the real world. Studies with complex models indicate what processes need to be included in simple models and, in some cases, how they can be parametrized.
One-dimensional models cannot simulate climatic surprises, for example sudden ocean circulation changes. Two-dimensional ocean models can give some insight into such changes.	AOGCMs can simulate major changes in ocean circulation but the timing and nature of such changes may not yet be reliable.

3.6.1 Comparison of Simple and Complex Models

Both simple and complex models have important but different roles to play in projecting future climatic change due to human activities. The table summarizes the principal differences between simple and complex models.

The key processes that determine climate sensitivity and the longer term feedbacks involving the terrestrial and marine biosphere depend on regionally-distributed and regionally-heterogeneous processes, and require three-dimensional models if they are to be reliably simulated. Complex models are also needed for the simulation of regional climatic change and of variability on short time-scales; for identifying which processes need to be included in simple models (namely, those in which the effects of small-scale variability do not average out); and for studying those fundamental physical processes which can be resolved by global scale, three-dimensional models but not simpler models (such as the role of localized oceanic convection in the large-scale ocean circulation, or the interaction between winds and large-scale heating patterns in the atmosphere). Complex models provide scenarios of time-evolving regional climatic change, as well as diurnal and seasonal patterns of climatic change and changes in extremes and variability at many time-scales. They, therefore, can be used in the interpretation of observed regional scale climatic changes. On the other hand, complex models are computationally costly, are sometimes difficult to understand, and require high resolution data inputs, which in some cases simply do not exist. They produce outputs which contain substantial temporal and spatial variability (sometime referred to as “noise”); this makes analysis of their results a complicated task, as is the case for the real climate system.

Simple models represent only the most critical processes. Consequently, they are relatively easy to understand and inexpensive to run, so that multiple diagnostic tests can be executed. They are useful mainly for exploring global scale questions. The upwelling-diffusion model, for example, has been used to investigate the role of the oceans in delaying the climatic response to increasing greenhouse gas concentrations and the role of ocean mixing-climate feedbacks in modifying the transient response (e.g., Hoffert, *et al.*, 1980; Harvey and Schneider, 1985; Morantine and Watts, 1990), in exploring the importance of natural variability in observed global mean temperature variations during the past century (Wigley and Raper, 1990; Schlesinger and Ramankutty, 1995), in setting constraints on the magnitude of the global mean aerosol cooling effect (Wigley, 1989), and in assessing the relative roles of greenhouse gases, aerosols, and solar variability in explaining global mean temperature variations during the past century (Kelly and Wigley, 1992; Schlesinger and Ramankutty, 1992). The climate sensitivity in simple models is a prescribed parameter and is held constant for a given simulation. In complex models, the climate sensitivity is determined as a consequence of the explicitly computed processes and sub-grid scale parametrizations in the model, and is free to vary as the climate itself changes.

The sub-components of simple models can be constrained to replicate the overall behaviour of the more complex model sub-components. For example, the climate sensitivity of simple models can be made to equal that of any particular AGCM or AOGCM by altering a single model parameter whose value implicitly accounts for the net, global mean effect of all the fast feedback processes which influence heat loss to space (on the other hand they cannot say, *a priori*, what that value should be). Similarly, the vertical diffusion coefficient and the upwelling velocity can be readily altered such that the oceanic uptake of heat (and associated sea level rise) closely matches that of any given OGCM. Globally-aggregated biosphere models can be adjusted to replicate the sensitivity to atmospheric CO_2 and temperature changes obtained by regionally-distributed models. This allows the simple models to emulate the behaviour of the more detailed, regionally-resolved models.

Another consequence of the different computational demands of simple and complex models relates to initialization. Ideally, one should begin a simulation with anthropogenic forcing starting from a steady-state (or “balanced”) climate, so that the simulated changes are due to the applied perturbation and not a consequence of the starting state. However, since the spin-up of coupled AOGCMs to a steady state requires thousands of simulated years, some anthropogenic forcing experiments using coupled AOGCMs have been started with the model in a non-equilibrium state. In such cases, “control” run projections with no imposed forcing yield a slowly changing or drifting climate. In order to determine the anthropogenic component of future change in such cases, one procedure is to subtract the control run climatic change from that of the perturbed run, on the assumption that the drift in the perturbed run is the same as in the control run, and that the climatic change and drift add linearly. This problem clearly complicates the experimental design and could also affect the occurrence of abrupt ocean circulation changes. Simple one- and two-dimensional models, in contrast, can always be spun up to a steady state prior to applying an anthropogenic perturbation because of the low computational cost involved. In reality, natural variability exists in the atmosphere-ocean system, so that an exact equilibrium has never existed. However, the magnitude of such variability at the century time-scale is expected to be much smaller than human-induced climatic change over the next century.

One-dimensional models are clearly incapable of anticipating climatic “surprises”, resulting from major changes in ocean circulation for example, although they can be used to assess the implications of such surprises. Complex AOGCMs have the potential to project such major changes in ocean circulation, although they are not yet crafted sufficiently well to do this reliably. Multi-basin versions of two-dimensional ocean models (i.e., Stocker and Wright, 1991) which have been calibrated to simulate the observed climate and ocean circulation can also provide insights into the conditions under which major ocean circulation changes could occur.

3.6.2 Data Limitations of Biosphere Models

Spatially-detailed terrestrial biosphere models are highly dependent on data-sets of land cover, land use, terrain, climate and soil characteristics. The quality of the existing data sets is currently low due to classification problems, data availability and poor temporal and spatial coverage (SAR WGII: Section 2.5.3). The marine biosphere has, in some ways, a less complex spatial heterogeneity than the terrestrial biosphere and is therefore simpler to model. Nevertheless, the available data on spatial heterogeneity in the biosphere limits the use of spatially-explicit models and adds to their uncertainty in both input variables, parameter settings and results. Although some of the spatially explicit models are included in climate models (e.g., Goldewijk, *et al.*, 1994), they are still mainly research tools to assess responses of the biosphere more comprehensively. Simple, globally aggregated models of the terrestrial and marine biosphere are currently more frequently used tools for the analyses of alternative scenarios involving the biosphere. These models have been calibrated against global scale observations but cannot simulate the detailed response of the biosphere. In the long run, regionally resolved models will have to be used.

3.6.3 Policy Development

SCMs are ideal for exploring the global scale consequences of alternative emission scenarios and for investigating the interactive effect of specific assumptions concerning the behaviour of individual subsystem components. Climate sensitivity and other key parameters (such as ocean mixing coefficients, biosphere feedbacks and ice-melt parameters) can be directly specified in simple models, and many sensitivity tests can be performed for each of a wide variety of emission scenarios. For these reasons, simple models were used extensively in the SAR WGI to explore the impact of alternative emissions scenarios of CO_2 and other gases on global mean surface temperature change and sea level rise (SAR WGI: Sections 6.3, 7.5.2, and 7.5.3).

Relatively simple climate and carbon cycle models have also been used as one of the core components of Integrated Assessment Models (IAMs), which are based on the integration of models that simulate the most critical processes of the climate system (human emissions, biosphere, oceans and atmosphere), and are used to explore the impacts of diverse emissions scenarios generated by alternative energy sources, different land-use changes, pollution control, and population policies. Although the climate component of such models is globally- (e.g., Wigley and Raper, 1995) or zonally-aggregated (as in de Haan, *et al.*, 1994), they have been linked to a number of regionally resolved sub-models spanning a wide range of human activities and impacts. One of the more advanced IAM is the IMAGE 2 model, which is described in Alcamo (1994). This model calculates emissions of different greenhouse gases from energy and land use; computes atmospheric concentrations by accounting for atmospheric chemistry and carbon uptake by the oceans and biosphere; and computes changes in climate and sea level as well as impacts on ecosystems and agriculture. These calculations allow for a transient determination of driving forces (including changed policies), climatic change, and its impacts. The policy relevance of such models lies in the comprehensiveness of simulations of many components in the climate system (see Figure 3).

The premise behind using simple models for policy analysis, with their focus on global scale changes, is that preventative responses to the risk of climatic change might be a collective response based on global scale aggregated impacts and risks, rather than on the local impacts and risks for any given nation undertaking a response. On the other hand, regionally-resolved models are needed, in conjunction with sector- and region-specific impact assessment tools, in order to translate global scale changes into specific impacts and hence to determine the globally-aggregated risk associated with a given magnitude and distribution of global scale change.

4. SIMPLE CLIMATE MODELS USED IN THE IPCC SECOND ASSESSMENT REPORT

In this section, we provide details concerning the specific SCMs and modules, and the associated assumptions, used in the SAR WGI. We begin with the computation of radiative forcing from emission scenarios, followed by the projection of global mean temperature change, and finally, the projection of future changes in sea level (as illustrated in Figure 4).

4.1 The Biogeochemical Component of a Simple Climate Model: Turning Emissions into Radiative Forcing

The following subsections describe the methods used in the SCM simulations described in the SAR WGI to compute the perturbation in greenhouse gas concentrations (SAR WGI: Sections 2.1 and 6.3), and the radiative forcings associated with perturbations in greenhouse gas and aerosol amounts (SAR WGI: Section 6.3). The quantitative relationships used are summarized in Appendices 1 and 2.

4.1.1 Treatment of Well-Mixed Gases with Well-Defined Lifetimes

The rate of removal from the atmosphere of nitrous oxide (N_2O) and the halocarbons is, to a first approximation, linearly proportional to the amount of gas in the atmosphere. That is, a fixed fraction of the amount of gas present at the start of a given year is removed per year, so that if the concentration of the gas doubles, for example, the mass removal rate doubles. These gases also have long lifetimes in the atmosphere relative to the time required for complete wind mixing to occur, so they are of relatively uniform concentration. As a result, the atmosphere can be regarded as a single, well-mixed box. The most important parameter is the average lifespan of a molecule of gas in the atmosphere, τ, which provides the link between concentration and rate of removal. Figure 6 illustrates the treatment of these gases. The numerical values of τ as adopted in the SAR WGI (Section 6.3) are summarized in Appendix 1; since the main removal process for most gases occurs through chemical reactions in the atmosphere, we use the term τ_{atm} in Appendix 1.

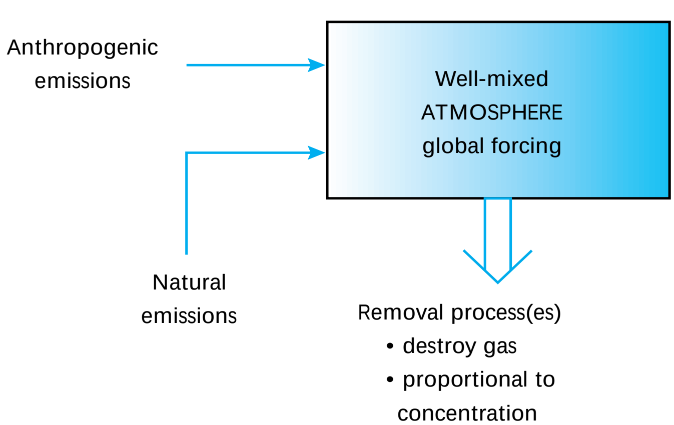

Figure 6. Schematic illustration of the treatment of well-mixed gases (CH_4, N_2O, halocarbons) in simple climate models. The removal rate is linearly proportional to concentration in the case of N_2O and halocarbons, but varies non-linearly with atmospheric concentration in the case of CH_4.

Methane (CH_4) is somewhat more complicated in that τ depends on the concentration itself. Nevertheless, the atmosphere can still be treated as a single well-mixed box as far as CH_4 is concerned, and concentration changes can be computed if the CH_4 lifetime is updated during the course of the computations. Thus, Figure 6 can also be applied to CH_4 as long as it is understood that the lifetime varies with the concentration itself, so that the removal rate now varies non-linearly with the concentration. As noted in Section 3.3, the dependence of the CH_4 lifetime on CH_4 concentration is affected by the concurrent concentrations of NO_x, CO and VOCs in the atmosphere, which vary significantly between regions. Emissions of these gases are also likely to change significantly over time, but, for purposes of computing changes in CH_4 removal rate time in SAR WGI (Section 6.3), these emissions were assumed to be constant. This feedback is based on calculations using three-dimensional models, as discussed by Osborn and Wigley (1994). The currently estimated CH_4 lifetime is given in Appendix 1.

In addition to removal by chemical reactions in the atmosphere, CH_4 is also absorbed by soils, a process that is also accounted for in the SAR WGI (Section 6.3) projections of global mean temperature and sea level. If soil absorption was the only removal process, the average lifespan of methane in the atmosphere would be about 150 years. We denote this lifespan by the term τ_{soil} in Appendix 1.

4.1.2 Treatment of Carbon Dioxide

Unlike the gases discussed in the preceding section, CO_2 does not have a well-defined lifetime. This is due to the multiplicity and complexity of processes involved in the removal of CO_2 from the atmosphere (as discussed in Section 3.2). Figure 7 illustrates the carbon cycle components and flows that have been included in the simple carbon cycle models used in SAR WGI (Sections 2.1 and 6.3). In two of the simple models used in the SAR WGI — those of Jain, *et al.*, (1995) and Siegenthaler and Joos (1992) — ocean chemistry and vertical mixing processes are explicitly computed using the one dimensional upwelling-diffusion model or a variant of it. In the third model used in the SAR WGI — that of Wigley (1991) — a reasonably accurate mathematical representation of the uptake of carbon by an OGCM, which was first employed by Harvey (1988), is used.

These three carbon cycle models are such that, when driven by anthropogenic fossil fuel emissions, the simulated build-up of

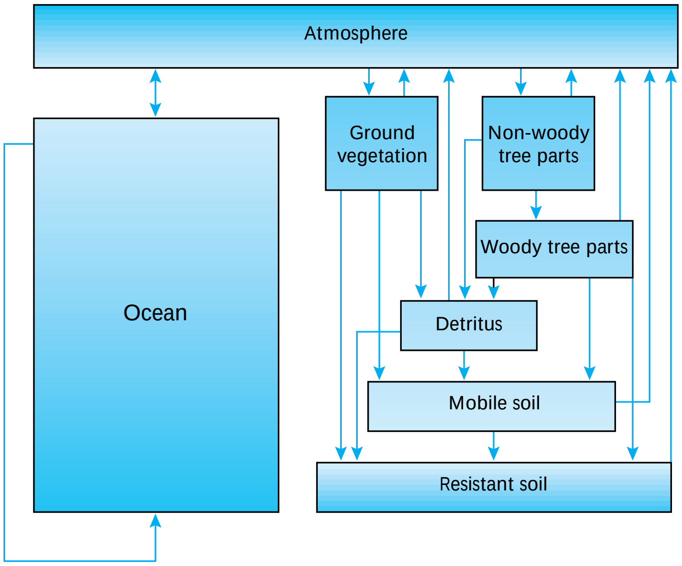

Figure 7. Components of the carbon cycle and the flows of carbon that are commonly included in simple models. The oceanic component can be formulated as an upwelling-diffusion model, or can be represented by a mathematical function (known, formally, as a convolution integral) which can be used to closely replicate the behaviour of other models, including OGCMs, used as part of the carbon cycle.

atmospheric CO_2 since the industrial revolution is close to that observed. Hence, when net emissions due to deforestation and forest regrowth are added (estimated to be 1.1±1.0 GtC/yr averaged over the 1980s; see SAR WGI: Table 2.1), additional sinks are required in order to avoid too large a CO_2 build-up. One way to create such a sink, which is the method used in the SAR WGI calculations (Sections 2.1 and 6.3), is to specify a CO_2 fertilization effect on the terrestrial biosphere. The larger the assumed past land-use emissions, the greater the required fertilization effect. If this effect is then extrapolated in some way (not necessarily linearly) into the future, the projected future CO_2 concentration will be lower the greater the assumed past and present land-use emissions (given that land-use emissions will eventually fall). The long-term validity of this extrapolation is highly uncertain (SAR WGI: Sections 2.1.1 and 9.2.3.2; and SAR WGII: Section A.2.3)

As noted in SAR WGI (Section 2.1) and in IPCC94 (Chapter 1), there are other mechanisms besides a CO_2 fertilization effect through which the carbon cycle could be balanced in the presence of net land-use emissions. For example, nitrogen fertilization of portions of the terrestrial biosphere as a by-product of NO_x emissions could be causing an extra 0.2-1.0 GtC/yr of carbon to be taken up (SAR WGI: Sections 2.1.1 and 9.2.3.4). Climatic changes during recent decades could also be causing the terrestrial biosphere to absorb a significant amount of carbon (SAR WGI: Sections 2.1.1 and 9.2.3.1). To the extent that these mechanisms have been operative, the CO_2 fertilization effect is weaker; to the extent that they do not increase as quickly as a CO_2 fertilization effect, extrapolation of an overestimated CO_2 fertilization effect will lead to projected atmospheric CO_2 concentrations that are too small.

A number of other processes that could influence future atmospheric CO_2 concentrations have also been neglected in projections of global mean temperature and sea level (SAR WGI: Section 6.3) and in the CO_2 stabilization calculations (SAR WGI: Section 2.1). In particular, no account has been taken of the potential for accelerated respiration of biomass and soil carbon due to warmer temperatures (leading to a potentially large release of CO_2), release of carbon to the atmosphere due to die back of forests if climatic zones shift too rapidly, or the impact of warmer ocean temperatures and changes in ocean circulation on the oceanic uptake of CO_2 (potentially leading to either a small release or additional absorption of CO_2). Until the relative importance of alternative mechanisms for absorbing anthropogenic CO_2 is better known, quantification of the uncertainties in future atmospheric CO_2 projections will remain difficult (see SAR WGI: Chapters 9 and 10 for a discussion of the potential impact of these processes on the carbon cycle).

4.1.3 Treatment of Gases not Directly Emitted

Tropospheric ozone is produced indirectly through chemical reactions involving CH_4, CO, NO_x, and VOCs, which have both natural and anthropogenic sources. Proper computation of tropospheric ozone build-up requires three-dimensional atmospheric chemistry/transport models. Since the relationships between CO, NO_x, VOCs and tropospheric O_3 build-up are uncertain, and the adequacy of current three-dimensional models is questionable, only the increase in tropospheric O_3 associated with increasing CH_4 concentration has been included in the SAR WGI (Section 6.3) projections of global mean temperature and sea level beyond 1990. This forcing is assumed to be directly proportional to the increase in methane concentration, with a value of 0.08 W m^{-2} in 1990. Up to 1990, tropospheric ozone radiative forcing associated with emissions other than CH_4 is also included. This forcing is assumed to have been proportional to fossil fuel emissions and to have reached a value of 0.32 W m^{-2} by 1990, and is then held constant. The total forcing in 1990 due to changes in tropospheric ozone has an uncertainty of at least ±50 per cent (see IPCC94: Section 4.3.6).

Problems also remain with regard to stratospheric models, which still cannot fully explain observed stratospheric O_3 losses. In the SAR WGI projections of global mean temperature and sea level (Section 6.3), stratospheric O_3 loss is assumed to vary with the tropospheric chlorine loading to the power 1.7, plus a bromine loading term weighted relative to chlorine by a factor of about 40 at present. The forcing associated with stratospheric ozone loss is then assumed to be directly proportional to the ozone loss, leading to the relationship between forcing and chlorine and bromine loading given in Appendix 2. This relationship was calibrated by comparing the computed global mean forcing due to stratospheric ozone changes with detailed radiative transfer calculations based on the observed ozone loss over the period 1979 to 1990 (Ramaswamy, *et al.*, 1992). The total direct halocarbon forcing in 1990 calculated using the expression in Appendix 2 is 0.27 W m^{-2}, and 0.1 W m^{-2} when stratospheric O_3

depletion is taken into account. The 1990 halocarbon forcing has relatively low uncertainty (± 20 per cent), while the uncertainty in the forcing associated with stratospheric O_3 depletion is at least ±50 per cent (see SAR WGI: Sections 2.4.1.1 and 2.4.1.2). The change in stratospheric ozone in the future, implied by use of the forcing-effective chlorine loading relationship given in Appendix 2, agrees very well with that computed by complex models.

4.1.4 Treatment of Aerosols

The global mean concentrations of three kinds of aerosols have increased through human activity by a sufficiently large amount to have important effects on climate: sulphate (SO_4) aerosols, which are produced from the oxidation of sulphur-containing precursors and which are emitted through the combustion of coal and oil and from smelting of certain metals; soot (black carbon) aerosols, directly released from combustion of coal, oil, and biomass; and organic aerosols (other than soot), released from the combustion of biomass or produced from chemical transformation of VOCs (IPCC94: Chapter 3). Dust aerosols from land-surface changes might also have noticeable climatic impacts (SAR WGI: Sections 2.3 and 2.4)

As discussed in Section 3.3, the processes determining the amount, distribution, and properties of aerosols in the atmosphere can be simulated, and the global mean forcing computed, only by using three-dimensional AGCMs. When using SCMs, one must therefore use results from AGCMs to establish a quantitative link directly between present global emissions and present global mean forcing. Because the atmospheric aerosol burden responds essentially instantaneously to changes in emissions, specification of an emission scenario amounts to specifying a concentration scenario. In the SAR WGI (Section 6.3), the relationship between emissions and atmospheric aerosol loading is assumed to be linear. Although this is not exactly true, the error so introduced is overwhelmed by uncertainties in the link between atmospheric aerosol loading and global mean radiative forcing. In practice, atmospheric aerosol loading is not explicitly computed; rather, global emissions are directly linked to global mean forcing using the results of AGCMs (as discussed below in Section 4.1.5).

For sulphur, two emission scenarios were considered in the SAR WGI (Section 6.3): one in which anthropogenic emissions are held constant at the 1990 level after 1990, and one in which the emissions of SO_2 are as specified in the IS92*a* scenario (IPCC, 1992: Table A3.12). In the latter case, total anthropogenic sulphur emissions will increase from 75 TgS in 1990 to 147 TgS in 2100. Dust aerosols are neglected in the SAR WGI (Section 6.3) projections of global mean temperature and sea level, while the radiative forcing associated with organic aerosols from biomass burning is assumed to scale with gross deforestation up to 1990 (when the forcing is assumed to have been -0.2 W m^{-2}), then is held constant.

4.1.5 Calculating Radiative Forcing From Concentrations

Given the concentrations of globally uniform greenhouse gases, the direct radiative forcing can be computed by using simple formulae which provide a close fit to the results of detailed radiative transfer calculations. In the case of CH_4, indirect forcings also arise through the formation of stratospheric water vapour from oxidation of CH_4, and through effects on tropospheric O_3. In the SAR WGI (Section 6.3), the stratospheric water vapour forcing is assumed to vary directly with the CH_4 forcing, while the tropospheric O_3 forcing due to CH_4 emission is assumed to vary linearly with the increase in CH_4 concentration (see Appendix 2).

The forcing associated with both stratospheric and tropospheric O_3 changes varies substantially regionally, since the O_3 changes themselves exhibit strong regional variation (IPCC94: Section 2.6; SAR WGI: Section 2.2). It is assumed in the SAR WGI (Section 6.3) that the global mean climatic response is proportional to the global mean forcing, which in turn is assumed to be directly related to the change in global mean concentration. As noted in the SAR WGI (Section 2.2), changes in stratospheric O_3 provoke further radiative forcings through induced changes in tropospheric chemistry, and this indirect forcing could be two to three times the direct forcing. Due to uncertainties in the magnitudes of these potential effects, they have been neglected in the SAR WGI projections of global mean temperature and sea level. As noted in Section 2.3.4, the assumption that the relationship between global mean temperature response and global mean forcing is the same for O_3 as for CO_2 might introduce further error. However, this error is at present overwhelmed by the large (factor of two to three) uncertainty in the forcings due to both tropospheric and stratospheric O_3 changes.

As discussed in Section 4.1.4, the global mean aerosol forcing in the models used in the SAR WGI (Section 6.3) is based on the ratio of present-day global emissions to present-day forcing, as computed from an AGCM for a limited number of aerosol distributions. Since atmospheric aerosol concentrations vary directly and immediately with emissions, this contains an implicit relationship between concentration and forcing. The direct component of the forcing is assumed to vary linearly with concentration and hence with emissions, while the indirect forcing is assumed to increase more slowly than emissions, based on our understanding of the key physical mechanisms involved. Both the direct and indirect global mean forcings by sulphate aerosols are highly uncertain (SAR WGI: Section 2.4.2 and 6.3.2); in the SAR WGI projections of global mean temperature and sea level, these forcings are assumed to have been -0.3 W m^{-2} (out of an uncertainty range of -0.2 to -0.8 W m^{-2}) and -0.8 W m^{-2} (out of an uncertainty range of 0.0 to -1.5 W m^{-2}), respectively, with the indirect forcing varying with the logarithm of concentration and thus of emission (see Appendix 3). Thus, as sulphate aerosol loading increases the indirect forcing becomes smaller relative to the direct forcing.

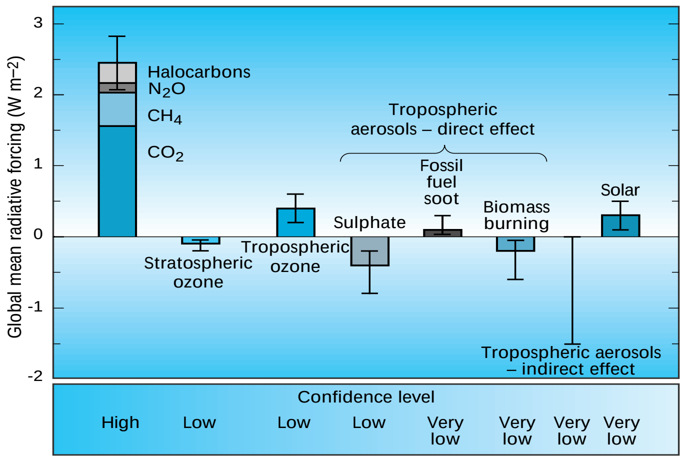

Figure 8. Estimated global mean radiative forcing (W m^{-2}) and associated uncertainty due to changes in greenhouse gas concentrations and aerosols from pre-industrial times to the present (1992) and in solar output from 1850 to the present (reproduced from SAR WGI: Figure 2.16).

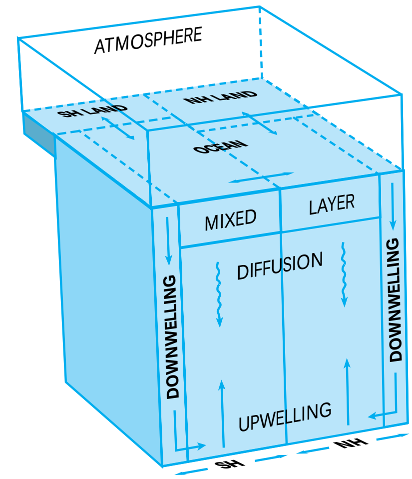

Figure 9. Illustration of a variant of the one-dimensional upwelling-diffusion model having separate land and sea boxes within each hemisphere, and separate polar sinking and upwelling in each hemisphere. This variant was used in the SAR WGI (Section 6.3 and 7.5.2).

As an indication of the relative importance of the different forcings, Figure 8 shows the forcings for 1990 and the associated uncertainty range as given by the SAR WGI (Section 2.4).

4.2 Translating Radiative Forcing into Global Mean Temperature Change

Given a scenario of global mean radiative forcing, the next step is to compute the resultant transient (time-varying) climatic change. This depends on both the climate sensitivity and on the rate of absorption of heat by the oceans. For the projections of global mean temperature (and sea level) change resulting from the IS92 emissions scenarios presented in SAR WGI (Section 6.3 and 7.5.2), a variant of the one-dimensional upwelling-diffusion model (described in Section 3.1) was used. This variant consists essentially of two one-dimensional upwelling-diffusion models strapped together, one for the northern hemisphere (NH) and one for the southern hemisphere (SH), and distinguishes between land and sea. It is illustrated in Figure 9. The original version of this variant is described in Wigley and Raper (1993), although it had been modified for the SAR WGI to include different climate sensitivities for land and ocean and a variable upwelling rate (see Raper and Cubasch, 1996 and SAR WGI: Section 6.3.1). A limited number of sea level cases was also presented (in SAR WGI: Section 7.5.3) using the two-dimensional ocean and one-dimensional atmospheric model of de Wolde, *et al.*, (1995) and Bintanja (1995), which was also introduced in Section 3.1.

There are four key parameters in the upwelling-diffusion model (and the variant shown in Figure 9): (*a*) the infrared radiative damping factor, which governs the change in infrared emission to space with temperature. This factor includes the effect of feedbacks involving water vapour, atmospheric temperature structure, and clouds, which are explicitly computed in more complex models. Because the infrared radiative damping to space is a key determinant of climate sensitivity, the model climate sensitivity can be readily altered — to match observational constraints or the results of other models — by changing the value of this factor; (*b*) the intensity of the thermohaline circulation, which consists of water sinking in polar regions (at a temperature which is prescribed in the model) and upwelling throughout the rest of the ocean; (*c*) the strength of vertical ocean mixing by turbulent eddies, which is represented as a diffusion process; and (*d*) the ratio of warming in the polar regions (which are not explicitly represented in the model) to the global mean surface layer warming, which determines the change in temperature of water in the sinking branch of the thermohaline circulation.

The other model used in the SAR WGI for climatic change projections (other than coupled AOGCMs) is the atmosphere-ocean climate model of de Wolde, *et al.*, (1995) and Bintanja (1995). The oceanic part of this model is a two-dimensional upwelling diffusion model, in that it contains both vertical heat diffusion and the thermohaline overturning (as in the one-dimensional upwelling-diffusion model). This model has horizontal resolution and includes parametrizations of north-south heat transport, as well as simple representations of sea ice and land snow cover. The ratio of polar to global mean surface warming is not directly specified in this model, but is determined by changes in north-south heat transport, ice and snow distribution, and vertical heat fluxes. The climate sensitivity also is not directly specified, but arises from the interaction of a number of different model processes. As in the one-dimensional upwelling-diffusion model, the intensity of the ocean thermohaline overturning and the value of the vertical diffusion coefficient must be directly specified.

Diffusive mixing produces a downward heat flux (from the warm surface to cooler sub-surface water). The thermohaline overturning, in contrast, produces an upward heat flux because it entails sinking of cold polar water and the upwelling of less cold water elsewhere. This shall be referred to here as the advective heat flux. In steady state, the net heat flux between the surface and deep ocean is zero (that is, the diffusive and advective heat fluxes exactly cancel).

As the surface and atmosphere warm in response to a radiative heating perturbation, the downward diffusive heat flux increases, which tends to slow down the subsequent rate of surface warming. The upward advective heat flux can increase or decrease as the climate warms, depending on the rate of warming of the downwelling source water in polar regions relative to the global mean surface layer and on changes in the sinking flux/upwelling velocity. The greater the specified (or computed) polar warming relative to the mean warming, the slower the mean surface temperature response to a heating perturbation. Similarly, variations in the upwelling velocity as a function of time or as a function of surface warming can be imposed in both the one-dimensional and two-dimensional upwelling-diffusion models, based on the variation in upwelling observed in coupled AOGCM experiments. A reduction in the upwelling velocity in response to surface warming tends to slow the surface temperature response, since this reduces the net heat flux toward the surface layer. Conversely, a strengthening of the thermohaline overturning will accelerate the surface temperature response, and can even cause a temporary overshoot of the equilibrium response (see Harvey and Schneider, 1985; and Harvey, 1994).

A third, minor, feedback that can be imposed in both the one-dimensional and two-dimensional upwelling-diffusion models is between the vertical diffusion coefficient and the vertical temperature gradient. It is expected that an increase in the temperature gradient (associated with greater initial warming at the surface) will lead to a weaker diffusion coefficient, which in turn will permit a slightly faster surface warming. However, this feedback was not included for the SAR WGI projections; rather, the diffusion coefficient is assumed to be constant both in the vertical and with time.

It should be stressed that neither alteration in the polar/global mean surface warming ratio in the one-dimensional upwelling-diffusion model, nor feedback between surface temperature and the thermohaline overturning or vertical diffusion coefficient, has any effect on the steady-state surface temperature response to an external forcing change[6]. This is because, in steady state, there is no net heat flux to or from the deep ocean, and the global mean surface-atmosphere steady-state temperature response is governed by radiative damping to space. However, these three factors do strongly influence the rate of approach to steady state, as noted above. Furthermore, each of these factors strongly influences the steady-state deep ocean temperature. Thus, the greater the polar sea warming, the greater the mean deep ocean warming. An increase in thermohaline overturning intensity results in a smaller deep ocean warming, while a reduction in overturning intensity leads to greater deep ocean warming. Finally, a reduction in the vertical diffusion coefficient will lead to smaller deep ocean warming. These differences in deep ocean warming can lead to dramatic differences in the thermal expansion component of global mean sea level rise associated with a given surface warming (see also Section 5).

It is assumed in both models that the global mean temperature response to a radiative forcing perturbation depends only on the global mean value of the perturbation, and that the climate sensitivity is the same irrespective of the magnitude or direction of the radiative forcing. As discussed in Section 2.3.4, the dependence of climate sensitivity on the magnitude, direction, and nature of the forcing is thought to be small, in most cases, compared to the underlying uncertainty in the sensitivity itself (a factor of three).

The two most important uncertainties in projections of future global mean temperature change are the climate sensitivity and the aerosol forcing, which partly offsets the heating due to increasing greenhouse gas concentrations. Figures 10*a* and *b* (SAR WGI: Figure 8.4) illustrate the impact of alternative assumptions concerning climate sensitivity and aerosol forcing, as computed using a one-dimensional upwelling-diffusion model. Comparison with Figure 10*c* shows that solar variability may also be an important contributor to past observed global mean changes, and its incorporation improves the agreement between model and global mean observations. The effect of uncertainties in the climate sensitivity and aerosol forcing for future climatic change is illustrated in Figure 11 for the central IPCC (1992) emission scenario, IS92a. The figure shows temperature changes over 1990 to 2100 for climate sensitivities of 1.5, 2.5 and 4.5˚C, for the changing aerosol (full lines) and constant aerosol (dashed lines) cases. The central sensitivity value gives a warming of 2.0˚C (changing aerosols) to 2.4˚C (constant aerosols). The range in warming due to uncertainty in the climate sensitivity is large, and aerosol-related uncertainties are larger for higher sensitivities.

Consistency Between Biogeochemical and Energy Balance Model Components

An ideal, fully integrated model, at any level of complexity, should have both chemical (e.g., CO_2) and climate (e.g., temperature, sea level) outputs that are derived simultaneously using the same physics, where appropriate. At the simple model level, consistency between the carbon cycle and energy balance components requires, as a minimum, that the

[6] In the case of the two-dimensional upwelling diffusion model, the global mean temperature response will depend slightly on the imposed variation in the thermohaline overtuning, since such changes will modify the north-south heat transport and lead to somewhat different changes in the amount of ice and snow than if the thermohaline overturning is held fixed.

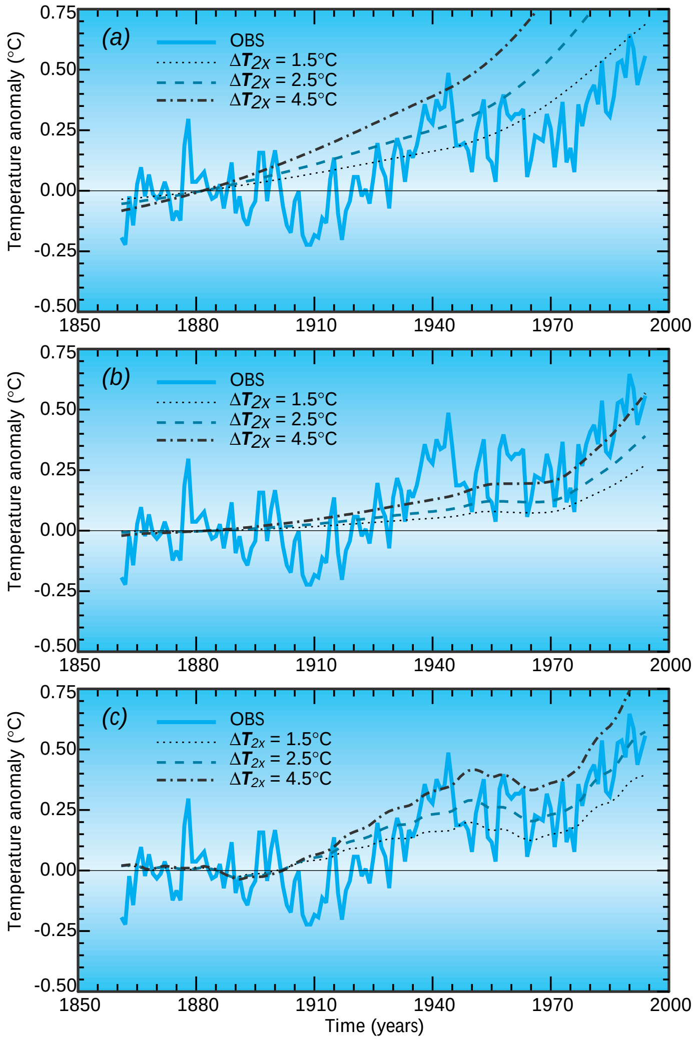

Figure 10. Observed changes in global mean temperature from 1861 to 1994 compared with those simulated using an upwelling diffusion-energy balance model. The model was run first with forcing due to (*a*) greenhouse gases alone; (*b*) greenhouse gases and aerosols; and (*c*) greenhouse gases, aerosols and an estimate of solar irradiance changes. The global mean greenhouse forcing in 1990 in all cases was 2.3 W m^{-2} out of an uncertainty range of 2.0 to 2.8 W m^{-2}, the global mean aerosol forcing in 1990 was -1.3 W m^{-2} out of an uncertainty range of -0.2 to -2.3 W m^{-2}, and the solar forcing over the period 1861 to 1990 was 0.4 W m^{-2} out of an uncertainty range of 0.1 to 0.5 W m^{-2}. Simulations were carried out with climate sensitivities of 1.5, 2.5 and 4.5°C (reproduced from SAR WGI: Figure 8.4).

same ocean model be used to advect and diffuse heat as is used to advect and diffuse total dissolved carbon and other chemical tracers used in the oceanic part of the carbon cycle. None of the models used in the SAR WGI incorporates this level of integration. For example, the global mean temperature and sea level results reported in SAR WGI (Sections 6.3, 7.5.2 and 7.5.3) were based on separate simple carbon cycle and climate models. The integration of these two components could be important in cases where there are substantial changes in the intensity of the thermohaline circulation (i.e., the upwelling rate), since this would alter both the thermal response and the rate of oceanic carbon uptake. In the SAR WGI, the effect of upwelling changes on the thermal response only was considered. However, the impact of upwelling changes on carbon uptake might be comparatively small, based on OGCM experiments reported by Bacastow and Maier-Reimer (1990).

4.3 Calculating Sea Level Change

Global warming is expected to cause changes in the ocean volume through thermal expansion caused by the flux of heat into the oceans, through the melting of glaciers and ice-caps, and through changes in the volume of the Greenland and Antarctic ice sheets (see Figure 4). In the SAR WGI (Section 7.5.2), the primary set of sea level rise projections was generated using the one-dimensional upwelling-diffusion model described in Section 4.2 to compute the thermal expansion component of sea level rise. The global mean surface air temperature change from this model was used to drive a conceptually simple model of glaciers and small ice-caps which takes into account the fact that there is a distribution of glacier altitudes and characteristics today (Wigley and Raper, 1995). A variety of assumptions concerning the linkage between changes in global mean temperature and the Greenland and Antarctic ice sheets was considered. An alternative set of projections was also generated using the two-dimensional upwelling-diffusion model (also described in Section 4.2) combined with more detailed calculations of the response of Antarctic and Greenland ice-caps (SAR WGI: Section 7.5.3). The resultant sea level

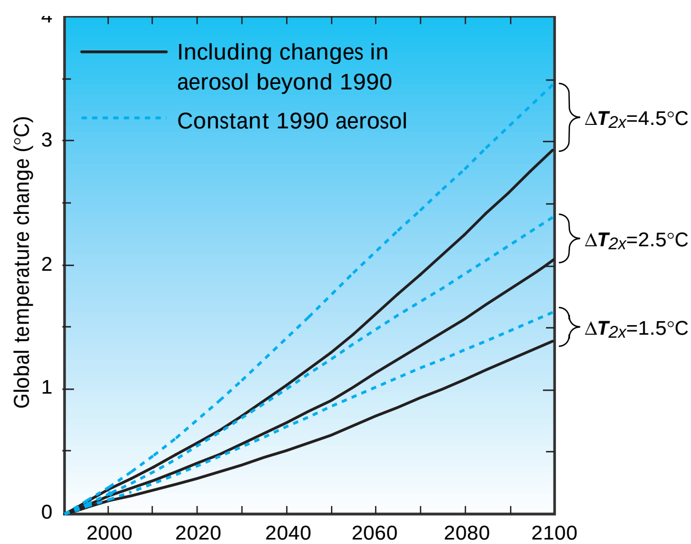

Figure 11. Global mean temperature change from 1990 as projected by the one-dimensional upwelling-diffusion model described in Section 4.2 for emission scenario IS92a, for climate sensitivities of 1.5, 2.5 and 4.5˚C and with aerosol emissions increasing (solid lines) or constant after 1990 (dashed lines). Reproduced from SAR WGI (Figure 6.20).

changes in both cases are global mean values; to project regional sea level changes requires taking into account vertical land movement and changes in ocean currents and winds. Computation of the latter two effects requires the use of a coupled AOGCM, as in Gregory (1993).

In the following subsections, the methods used to compute sea level changes using the one-dimensional and two-dimensional upwelling-diffusion models, and the glacier and ice-cap models to which they are coupled, are briefly described.

4.3.1 Calculations Starting From the One-Dimensional Upwelling-Diffusion Model

The thermal expansion component of sea level rise is computed from the variation of globally-averaged ocean temperature change with depth. The most important model parameter controlling thermal expansion over the next one hundred years is the climate model sensitivity, which strongly influences the heat flux into the ocean. The ratio of polar to mean surface layer warming and the change in the thermohaline overturning intensity are also important to sea level rise, as discussed in Section 4.2, particularly on longer time-scales. For the one-dimensional model calculations presented in the SAR WGI, it was assumed that the polar source regions for downwelling water warm by 20 per cent of the global mean surface layer warming, and that the thermohaline overturning weakens slightly as the climate warms (as in some coupled AOGCMs). The resultant thermal expansion component of sea level rise, associated with the surface temperature response curves of Figure 11 with changing aerosols, is 20, 28 and 40 cm for climate model sensitivities of 1.5, 2.5 and 4.5˚C, respectively.

For the calculation of the land-based ice contribution to sea level rise, the ice masses were divided into three groups: the glaciers and ice-caps, the Greenland ice sheet, and the Antarctic ice sheet.

For the glaciers and ice-caps, a simple model which relates glacier volume to temperature change was used (Wigley and Raper, 1995). There are three important parameters in this model: (*a*) the initial (1880) global ice volume, which was assumed to be 30 cm sea level equivalent; (*b*) the minimum temperature increase which, if it were maintained, would cause a given glacier to eventually disappear; and (*c*) the glacier response time. Because there is a distribution of critical temperature warmings and glacier response times in nature, a distribution of minimum temperature increases required for disappearance of a glacier, and of glacier response times, is assumed in the calculations. As the simulated global mean temperature increases, greater melting of glaciers within the model distribution occurs. The ranges of glacier response times and warmings required for eventual disappearance of small glaciers are themselves uncertain, so different sets of assumptions have been adopted and are listed in Appendix 3. The assumptions listed as “high” in Appendix 3 will give a relatively large contribution to sea level rise, while those listed as “low” will give a relatively small contribution to sea level rise.

The assumed initial glacier and ice-cap volume is important because it sets an upper limit to the sea level rise from this source. However, the correct value of this parameter is controversial; a value of 50±10 cm is given in Table 7.1 of the SAR WGI. The difference between this range and the value adopted for the SCM sea level projections (30 cm) reflects the difficulty in estimating this parameter. The initial ice volume and other parameter values were chosen so as to match, as the central value, the estimated contribution to sea level rise during the period 1900-1961 of 1.6 cm sea level, equivalent. Estimates of the past contribution to sea level rise of glaciers and ice-caps based on direct observations over the last century are uncertain by a factor of two. There are many reasons for this uncertainty, including: (*a*) different time periods used in the analysis; (*b*) differences in the total estimated glacier areas; (*c*) incomplete climatic data from the glaciated regions; (*d*) crude approximations to dynamic feedbacks; and (*e*) neglect of refreezing of meltwater and of iceberg calving. The central value used here of 1.6 cm sea level equivalent over 1900-1961 is at the low end of the range of the estimates of 0.35 mm/yr with uncertainty of at least ± 0.1 mm/yr, over 1890-1990, given in the SAR WGI (Section 7.3.2.2). The estimated contribution of glaciers and ice-caps to sea level rise for 1990 to 2100, when climate sensitivities of 1.5, 2.5 and 4.5˚C are combined with the low, medium, and high ice parameters of Appendix 3, respectively, are 7, 16 and 25 cm, respectively (again using the temperature response curves of Figure 11 with changing aerosols).

The response time of the Greenland and Antarctic ice sheets is long compared to the time-scale considered here, so, for simplicity, the areas of the ice sheets are assumed to be constant and effects related to ice flow are neglected. However, the uncertainties even in the present mass balance of the ice sheets are large. The SAR WGI (Section 7.3.3.2) concludes that an imbalance between accumulation and losses of the ice sheets of up to 25 per cent cannot be detected by current methods using currently available data.

For modelling purposes, the mass balance of both ice sheets is divided into two components (Wigley and Raper, 1993). The first represents the gain or loss of ice due to the initial state of the ice sheet, and has units mm/yr sea level rise. If the ice sheet was initially in equilibrium with the climate in 1880 (the initial time), this component would be zero, but if it was not in equilibrium but still reacting to a previous temperature change, then it would be non-zero. This component is denoted by the symbol ΔBo in Appendix 3, where the values used for the low, medium, and high sea level rise cases are given.

The second component is assumed to be linearly dependent on the temperature change relative to the initial state, and has units mm/yr/˚C sea level rise. The values used are given in Appendix 3 and are based on estimates of the sensitivity of the ice sheets to a 1˚C climatic warming as computed by the two- and three-dimensional ice sheet models that are directly used for the calculations with the two-dimensional upwelling-diffusion model (SAR WGI: Section 7.3.3.3; and Section 4.3.2, below). For Antarctica, the temperature dependent term is assumed to have two sensitivities: a sensitivity value for the mass balance (which is negative), and a second sensitivity that represents the influence of a possible instability of the West Antarctic ice sheet. Given our present knowledge, it is clear that, while the West Antarctic ice sheet has had a very dynamic history, estimating the likelihood of a collapse during the next century is not yet possible (SAR WGI: Section 7.5.5). A small value (based on MacAyeal, 1992) is included in the model, however, to acknowledge the possibility of a contribution from this source.

For the period up to 1990, the ice sheet changes are driven by the model-computed, global mean surface temperature change. For the future, however, a temperature warming of 1.5 times the global mean warming since 1990 is used to drive further changes in the Greenland ice sheet. The factor of 1.5 is based on the summer regional warming response over Greenland as obtained by coupled AOGCMs. The computed contribution to sea level rise from 1990 to 2100 are 1, 6 and 14 cm for Greenland and -9, -1 and 8 cm for Antarctica, when climate sensitivities of 1.5, 2.5 and 4.5˚C are combined with the low, medium, and high ice sheet parameters, respectively.

When the individual contributions described above are concatenated together in such a way as to maximise the range in overall sea level rise (that is, when the “low” contribution from one component is combined with the “low” contribution from another, and similarly for the “high” contributions), the modelled sea level rise from 1880 to 1990 is 2-19 cm if the warming over this period is 0.5˚C, with a central estimate of around 10 cm. In Table 7.7 of SAR WGI, a range of -19 cm to 37 cm is given based on a synthesis of model results and observations. The range given here is designed to be less than that of the SAR WGI Table 7.7 because, as the high or low limits from various factors are concatenated together, the probabilities associated with the limits of the resulting range become very small. The range of 2 to 19 cm reported here can be compared with the 10 to 25 cm range based on tidal gauge data, which is also given in Table 7.7 of SAR WGI. While modelled and tidal gauge ranges overlap, there is still a problem in reconciling the past changes, which emphasises the uncertainties in projections for the future.

Figure 12 shows the net result of the above individual contributions to sea level for the period 1990-2100 for the temperature response curves of Figure 11. As in Figure 11, results are shown for the two aerosols cases of Section 4.1.4. The combination of low, medium, and high ice melt parameters with the low, medium, and high climate sensitivities, respectively, gives total sea level rises of 20, 49, and 86 cm, respectively, for the case with increasing aerosol emissions, and 23, 55, and 96 cm for the case with constant aerosol emissions. Figure 13 shows the contributions of the individual components to sea level rise for the medium ice melt parameters and medium (2.5˚C) climate sensitivity.

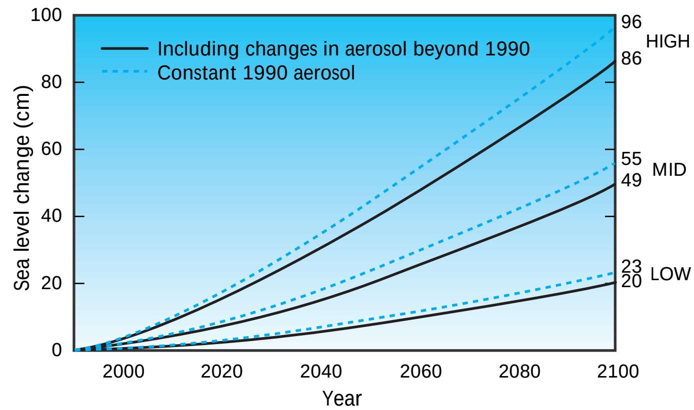

Figure 12. Global mean sea level changes based on the one-dimensional upwelling-diffusion model described in Section 4.3.1 for the same cases as shown in Figure 11 (reproduced from SAR WGI, Figure 7.7).

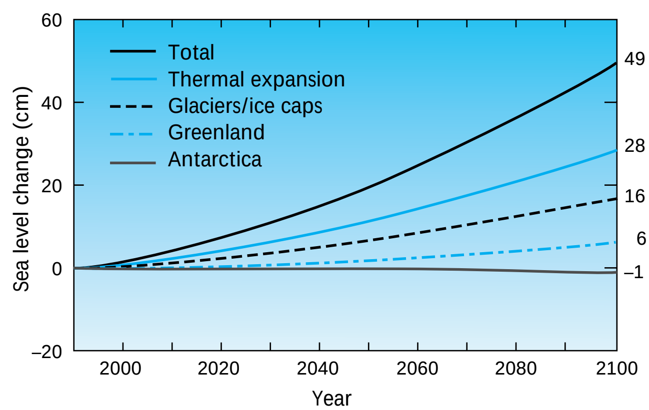

Figure 13. The individual contributions to the “MID” sea level rise case shown in Figure 11 (reproduced from SAR WGI, Figure 7.8).

4.3.2 Calculations Starting From the Two-Dimensional Upwelling-Diffusion Model

The second set of sea level rise calculations used in the SAR WGI (Section 7.5.3) is also based on the summation of separate contributions from ocean thermal expansion, melting of glaciers and ice-caps, and changes in the Greenland and Antarctic ice sheets. However, the procedures used to compute the contributions from these components differ in several important ways from those described above.

The thermal expansion component is computed using a two-dimensional upwelling-diffusion model (de Wolde, *et al.*, 1995), applied separately to the Atlantic, Pacific, and Indian Ocean basins and coupled to a zonally (east-west) averaged atmospheric model (Bintanja, 1995). Besides computing the thermal expansion component of sea level rise, this coupled atmosphere-ocean model calculates latitudinally and seasonally varying changes in surface air temperature. These changes in turn are used as input to glacier, ice-cap and ice sheet models.

Studies of well-observed glaciers indicate that glaciers in a wetter climate are more sensitive to changes in air temperature than glaciers in dry regions. This arises because the area-elevation distribution is different and the albedo feedback is more effective for glaciers with high precipitation snowfall. For the calculation of the glacier response to climatic change, all glaciers and small ice-caps on Earth have, therefore, been placed in one of 100 regions, each characterized by the present-day precipitation rate and glacierized area. For each region, the sensitivity of the glacier mass balance to changes in temperature depends on the mean annual precipitation (see Oerlemans and Fortuin, 1992). Model calculations start in 1990, although at present most glaciers are not in equilibrium. To account for the observed present-day thinning of several glaciers, projections of the contribution of glaciers and ice-caps to sea level change include a constant long-term trend of 0.5 mm/yr sea level rise, which is consistent with observations.

The sea level contributions of the Greenland and Antarctic ice sheets are estimated using dynamic ice flow models. In the case of Greenland, a two-dimensional (latitude-longitude) model with a horizontal resolution of 20 x 20 km is used (Cadee, 1992), while a three-dimensional model of the Antarctic ice sheet with 20 km horizontal resolution and 14 layers is used (Huybrechts, 1992; Huybrechts and Oerlemans, 1990). Both ice sheet models are forced with the zonally-averaged temperature changes produced by the coupled atmosphere-ocean climate model. In the case of Greenland, the accumulation rate is held constant at the observationally based estimate for the present (Ohmura and Reeh, 1991), and changes in the rate of melting are computed using a simple surface energy balance model (van de Wal and Oerlemans, 1994). Model calculations start in 1990, at which time the Greenland ice sheet is assumed to have been in a state of equilibrium. In the case of Antarctica, a combination of observations and theory suggests that the accumulation rate should increase with increasing temperature, in proportion to the increase in the ability of air over Antarctica to hold moisture. The accumulation rate over Antarctica is therefore derived from present-day estimated observed values and is subsequently increased in proportion to the increase in atmospheric saturation water vapour pressure over Antarctica as the climate warms. Ablation (ice melting) in Antarctica is of minor importance. The initial state of the ice sheet was obtained by integrating the ice sheet model over the last two glacial cycles (spanning more then 200 000 years). Although this exercise indicates that there is a long-term negative mass balance at present, this is not included in projections of the Antarctic contribution to sea level rise because of the large uncertainty in the result; instead, projections of the Antarctic contribution to sea level change are calculated as the difference between runs with and without anthropogenic greenhouse gas and aerosol forcings.

As is the case for the calculations presented in Section 4.3.1, a wide range of model input parameters is possible, giving a wide range of sea level results. However, the middle or "best" estimate values obtained here differ significantly from the middle results shown in Section 4.3.1. Results obtained here are shown in Figure 14, and should be compared with the corresponding results in Figure 13. The largest difference is in the thermal expansion contribution to sea level, followed by the difference in the Antarctic contribution. Although the reasons for these differences were not entirely resolved at the time of publication of the SAR WGI, several differences in model features were identified (SAR WGI: Section 7.5.3.2). The differences likely to be important for the thermal expansion component of sea level rise include the meridional resolution of the two-dimensional model, the different model formulations of heat exchange between atmosphere and oceans, the absence of sea ice in the upwelling-diffusion model, different climate sensitivities (2.5˚C for the one-dimensional model middle case, 2.2˚C for the two-dimensional model, the latter not being adjustable), and the way in which the thermohaline circulation is represented. In the case of the Antarctic contribution, different temperature perturbations are used to force the ice sheet and smaller ice sheet sensitivities are used for the results presented in Section 4.3.1.

4.3.3 Uncertainties in Sea Level Projections

Uncertainties in the thermal expansion component of global mean sea level change are linked with those of surface temperature change itself, because thermal expansion is computed from the variation of ocean temperature change with depth. For climate model sensitivities ranging from 1.5 to 4.5˚C, the uncertainty in thermal expansion is about a factor of two over the next century. The main uncertainties in deriving the land ice contribution to sea level rise from global mean temperature

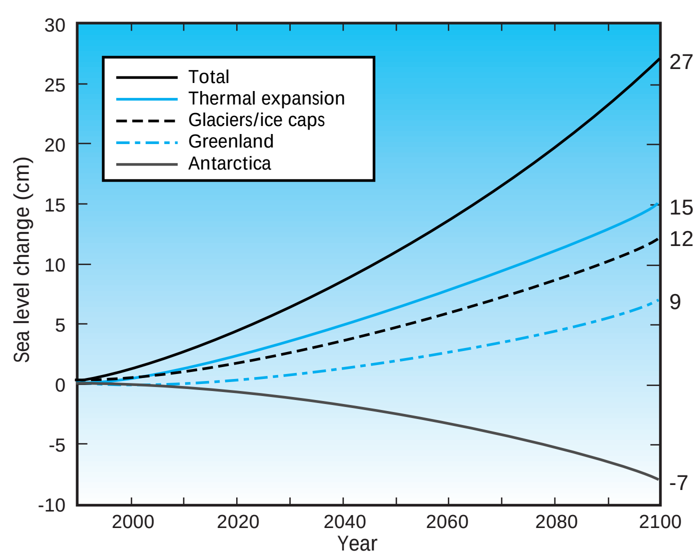

Figure 14. The individual contributions to the "MID" sea level rise case for emission scenario IS92*a* as computed using the two-dimensional upwelling-diffusion model described in Section 4.3.2 (reproduced from SAR WGI, Figure 7.11).

change are: the regional distribution of the temperature changes; the initial volume of the glaciers and ice-caps, and their sensitivity to increases in temperature; and the initial state of balance of the Greenland and Antarctic ice sheets and their sensitivity to temperature changes. Uncertainties in sea level rise cannot, therefore, be separated from uncertainties in global mean temperature change. However, changes in accumulation will also affect the volume of land-based ice. For the glaciers and ice-caps and for the Greenland ice sheet, accumulation has been assumed constant, where for the Antarctic ice sheet, accumulation is assumed to increase as temperature increases. Figures 11 and 12 express the uncertainty in temperature and sea level rise.

5. COMPARISON OF SURFACE TEMPERATURE CHANGES AND OCEAN THERMAL EXPANSION AS SIMULATED BY AOGCMs AND SCMs

Simple climate models have been, and will continue to be, used for analysis of the global scale implications of alternative emissions scenarios or of alternative assumptions concerning the properties of individual model components. It is, therefore, pertinent to compare the global mean temperature and sea level projections as simulated by one- and two-dimensional upwelling-diffusion models on the one hand, and AOGCMs on the other hand.

Figure 15 compares the change in global mean surface air temperature as simulated by several different AOGCMs with that of the one-dimensional upwelling-diffusion model with a CO_2 doubling climate sensitivity of 2.5˚C and that of the two-dimensional climate model (whose sensitivity is fixed at 2.2˚C). The spread in the AOGCM results can be largely explained by the differences in the model climate sensitivity, which varies from 2.1 to 4.6˚C. Note that the interannual variability in the AOGCM response is absent in the SCM response, which increases smoothly but is otherwise similar to the AOGCM response. Comparison of Figure 15 with Figure 11 illustrates the ability of upwelling-diffusion models to span the results of most AOGCMs when a range of values for the climate sensitivity is used.

A further illustration of the comparability of AOGCM and SCM time-dependent behaviour is given in Figure 16, which compares the global mean temperature change for the Geophysical Fluid Dynamics Laboratory (GFDL) AOGCM and the upwelling-diffusion climate model when both models are driven by various rates of increase in atmospheric CO_2 concentration (see SAR WGI: Section 6.3.1). To ensure a valid comparison, the SCM climate sensitivity was set at the GFDL model value of 3.7˚C. All other parameter values remained unchanged. The value of the land/ocean sensitivity differential (1.3), chosen on the basis of other GCM results (Raper, *et al.*, 1996), is similar to that for the GFDL model. The thermohaline circulation in the SCM was made to vary with surface warming in a manner that closely approximated the variation in the GFDL model (Manabe and Stouffer, 1994). The surface temperature responses are seen to agree well over a wide range of forcings.

As a final example of the ability of the one-dimensional upwelling-diffusion model to replicate AOGCM results, both the global mean temperature change and the ocean thermal expansion obtained for the 2xCO_2 and 4xCO_2 stabilization simulations of Manabe and Stouffer (1994) are compared

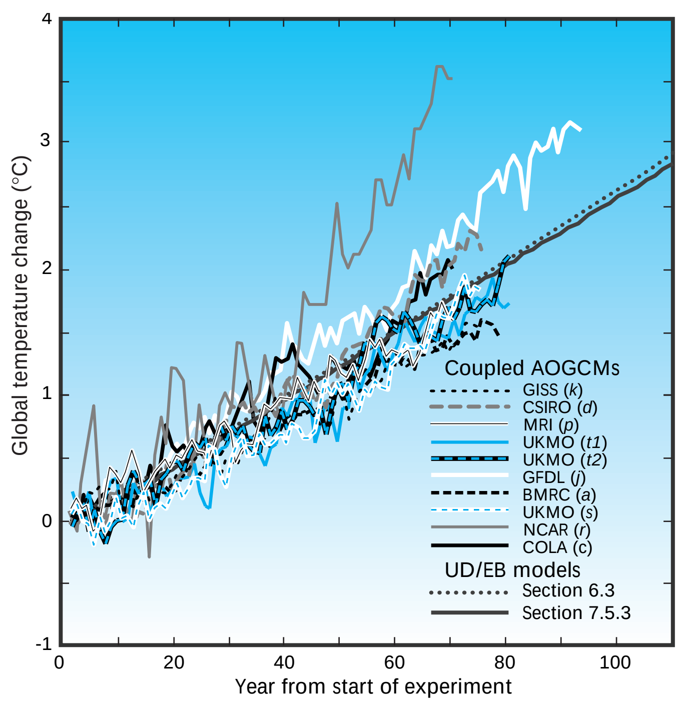

Figure 15. Comparison of global mean surface air temperature change as simulated by several different AOGCMs (with climate sensitivity varying from 2.1 to 4.6˚C), the one-dimensional upwelling-diffusion climate model (climate sensitivity of 2.5˚C), and the two-dimensional upwelling-diffusion model (climate sensitivity of 2.2˚C), in each case driven by a 1 per cent per year (compounded) CO_2 concentration increase (reproduced from SAR WGI, Figure 6.4).

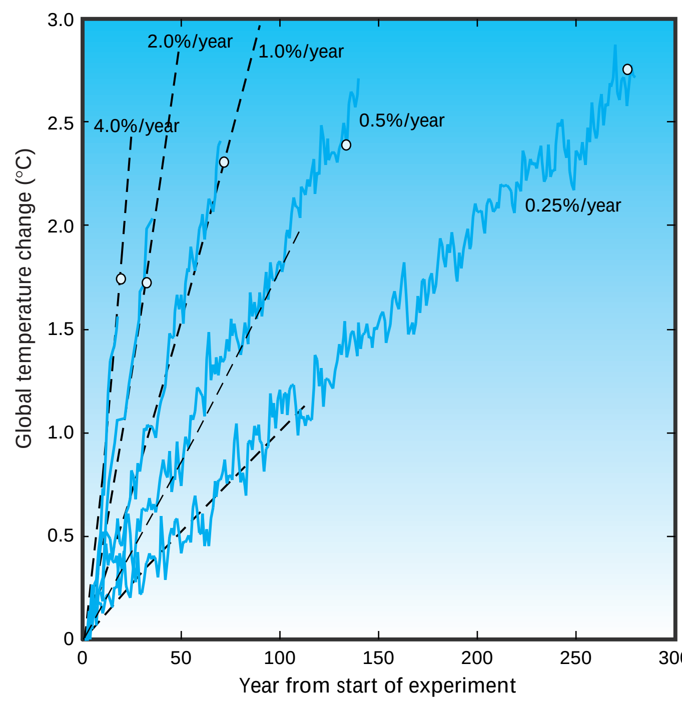

Figure 16. Global mean surface air temperature increase as computed by the GFDL AOGCM (solid lines) and the one-dimensional upwelling-diffusion climate model with a CO_2 doubling sensitivity of 3.7˚C. Results are shown for cases in which the atmospheric CO_2 concentration increases by 0.25, 0.5, 1, 2 and 4 per cent per year (reproduced from SAR WGI, Figure 6.13).

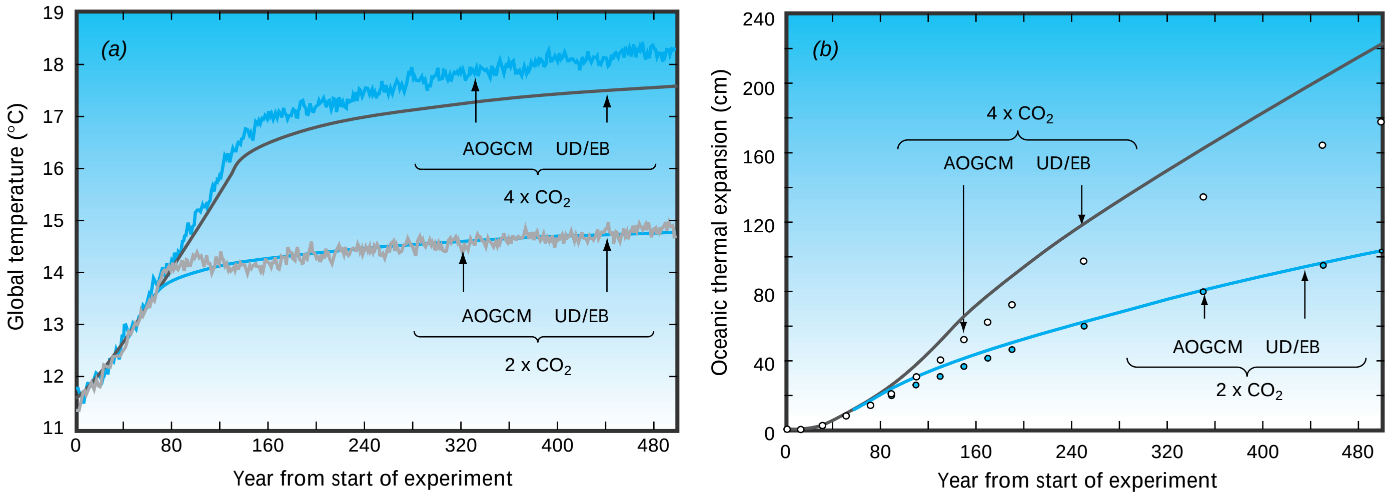

Figure 17. Comparison between the AOGCM results of Manabe and Stouffer (1994) and the one-dimensional upwelling-diffusion model for cases in which the atmospheric CO_2 concentration increases by 1 per cent per year (compounded) until the concentration has either doubled or quadrupled and is then stabilized (*a*) global mean surface air temperature; (*b*) sea level rise due to ocean thermal expansion (reproduced from SAR WGI, Figure 6.17).

with the one-dimensional model results in Figure 17. For the $2xCO_2$ case, the agreement is excellent for both the global mean temperature and the thermal expansion results. For the $4xCO_2$ case, the one-dimensional model gives lower warming and higher expansion, implying that the flux of heat into the deeper layers of the ocean is greater than in the AOGCM.

Other comparisons have been carried out by Raper and Cubasch (1996) using the Max Plank Institute (MPI) AOGCM described in Cubasch, *et al.* (1992). When the climate sensitivity is adjusted to give reasonably good agreement between the two models for temperature, the thermal expansion estimate from the SCM is greater than that in the AOGCM. This again implies that there is a greater flux of heat into the ocean in the one-dimensional model than in this particular AOGCM. The reason for this was not resolved at the time of publication of the SAR WGI. Note that when the heat flux into the ocean is larger (smaller) the surface temperature change is smaller (larger) and the thermal expansion is larger (smaller) (Harvey, 1994).

To summarize, it is possible to replicate the behaviour of a wide range of complex AOGCMs with SCMs. Of even greater importance than the ability to replicate the behaviour of any one AOGCM is the ability of SCMs to span the range of results that are obtained with different AOGCMs. Thus, SCMs provide a convenient and computationally fast tool for use in scenario and sensitivity analyses, in which a large number of model runs is required to cover the different emissions scenarios and to span the uncertainties inherent in AOGCMs.

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Appendix 1

Summary of methods used to compute concentrations of greenhouse gases in the SAR WGI (Chapter 2 and Section 6.3) and the IPCC Technical Paper on Stabilization of Atmospheric Greenhouse Gases (IPCC TP STAB, 1997).

Constituent	***Method for Computing Concentration***
CO_2	Concentration depends on the net flows between a number of carbon reservoirs that are represented within the models
CH_4	One box model, $dC/dt = ßE - C\,(1/\tau_{atm} + 1/\tau_{soil})$ τ_{atm} is a function of methane concentration and emissions of CO, NOx, and VOCs*. τ_{atm} = 9.08 years in 1990 and τ_{soil} = 150 years
N_2O	One box model, $dC/dt = ßE - C/\tau_{atm}$ τ_{atm} is fixed at a value of 120 years
CFC-11	Same as for N_2O, with τ_{atm} = 50 years
CFC-12	Same as for N_2O, with τ_{atm} = 102 years
HCFC22	Same as for N_2O, with τ_{atm} = 13.3 years
HCFC134a	Same as for N_2O, with τ_{atm} = 14 years
Other halocarbons	Treated explicitly as for CFC-11, gas by gas
Stratospheric Water Vapour	Concentration not explicitly specified[+]
Tropospheric Ozone	Concentration not explicitly specified[+]
Stratospheric Ozone	Concentration not explicitly specified[+]
Sulphate Aerosols	Concentration not explicitly specified[+]
Biomass Burning Aerosols	Concentration not explicitly specified[+]

In the above, C represents the atmospheric concentration of the corresponding gas, E the mass emission rate per year, $ß$ a factor that converts from mass to concentration, and τ_{atm} the mean lifespan of a molecule of the constituent in the atmosphere when accounting for chemical removal. In the case of methane, an additional removal process is through absorption by soils, and τ_{soil} is the mean lifetime a methane molecule would have if absorption by soils were the only removal process.

*VOCs = volatile organic compounds

[+]The radiative forcing is directly computed from emissions or from the concentration of some other gas, as indicated in Appendix 2.

Appendix 2

Functional dependence of radiative forcing on greenhouse gases and aerosols used in the SAR WGI (Section 6.3) and in IPCC TP STAB (1997). As discussed in the text, some of the forcing terms, as well as the natural sulphur emissions and anthropogenic sulphur emissions in 1990, are subject to considerable uncertainty. ΔQ_{CH_4}-pure is the methane forcing before correction for overlap with N_2O. $C(t)$ and $e(t)$ refer to concentrations and anthropogenic emissions of the gas in question at time t, while C_0 is the pre-industrial concentration. Sulphate aerosol indirect forcing depends on the natural sulphur emission, e_{nat}, which was assumed in the SAR WGI to be 42 TgS/yr, a higher value than currently accepted. Using a lower value leads to a slightly lower future indirect forcing (e.g., by 0.02 W m^{-2} averaged over 1990-2100 for emission scenario IS92*a*).

Constituent	*Method for Computing Radiative Forcing (W m^{-2})*
CO_2	$\Delta Q = 4.37 \ln(C(t)/C_o)/\ln(2)$*
CH_4	$\Delta Q = 0.036(\sqrt{C(t)}-\sqrt{C_o})$-(correction for overlap with N_2O)$^+$, where C and C_o are in ppbv and C_o=700 ppbv
N_2O	$\Delta Q = 0.14(\sqrt{C(t)}-\sqrt{C_o})$-(correction for overlap with CH_4)$^+$, where C and C_o are in ppbv and C_o=280 ppbv
CFC-11	$\Delta Q = 0.000\,22\, C(t)$ (where C is in pptv)
CFC-12	$\Delta Q = 0.000\,28\, C(t)$ (where C is in pptv)
HCFC22	$\Delta Q = 0.000\,189\, C(t)$ (where C is in pptv)
HCFC134a	$\Delta Q = 0.000\,169\, C(t)$ (where C is in pptv)
Other halocarbons	Treated explicitly (ΔQ varies with C), gas by gas
Stratospheric Water Vapour	$\Delta Q = 0.05\, \Delta Q_{CH_4}$-pure
Tropospheric Ozone	$\Delta Q = 8.62 \times 10^{-5}\, \Delta Q_{CH_4}$ for O_3 formation due to CH_4 build-up ΔQ associated with O_3 formation due to emissions of other gases ramps up to an assumed 1990 value of 0.32 W m^{-2}, then is held constant due to uncertainties
Loss of stratospheric ozone†	$\Delta Q = -[0.000\,552 \sum(\{NCl_i C_i\}^{1.7}) + 3.048 \sum(NBr_i C_i)]/1\,000$ where C_i is the concentration (pptv) of chlorine- or bromine-containing gas i, NCl_i and NBr_i are the numbers of chlorine or bromine atoms in gas i, and the summation is over all gases considered, (NBr_i = 1 for the two halons considered)
Sulphate Aerosols, Direct Forcing	$\Delta Q = e(t)/e_{1990}\, \Delta Q_{dir,1990}$, where $\Delta Q_{dir,1990}$ = -0.3 W m^{-2} and e_{1990}=69 TgS/yr
Sulphate Aerosols, Indirect Forcing	$\Delta Q = \frac{\log(1+e(t)/e_{nat})}{\log(1+e_{1990}/e_{nat})}\Delta Q_{indir,\,1990}$ where $\Delta Q_{indir,1990}$ = -0.8 W m^{-2} and e_{nat}=42 TgS/yr
Biomass Burning Aerosols	ΔQ = ramps to -0.2 W m^{-2} in 1990, and is held constant thereafter

* In the SAR WGI, the forcing is written as $6.3\ln(C(t)/C_0)$. The form used here is somewhat more transparent in that the coefficient in front of $\ln(C(t)/C_0)$ is equal to the forcing that is assumed for a CO_2 doubling. The forcing of 4.37 W m^{-2} that had been used in the SAR WGI and IPCC TP STAB (1997) is about 0.5 W m^{-2} too large. Since, for most results presented in the SAR WGI and IPCC TP STAB (1997), the climate response to a CO_2 doubling is directly specified, this error will not affect the results except to the extent that the warming effect of non-CO_2 gases will be slightly too small relative to the warming effect of CO_2.

+ See First IPCC Assessment Report (IPCC, 1990), Table 2.2 for details concerning the overlap term.

† The climate forcing due to loss of stratospheric ozone does not include effects of ozone loss on tropospheric chemistry.

Appendix 3

Parameter values for the ice-melt module described in the text, and used to obtain the low, medium and high sea level rise estimates for this Technical Paper and IPCC TP STAB (1997).

	Glaciers and Ice-caps		*Greenland*		*Antarctica*		
Case	τ (yr)	ΔT^* (˚C)	ΔB_o (mm/yr)	ß (mm/yr/˚C)	ΔB_o (mm/yr)	$ß_1$ (mm/yr/˚C)	$ß_2$ (mm/yr/˚C)
High	35-65	0.6-2.5	0.0	0.5	0.6	-0.15	0.2
Medium	70-130	0.7-3.0	0.0	0.3	0.1	-0.30	0.1
Low	105-195	0.9-4.5	0.0	0.1	-0.4	-0.45	0.0

τ is the range of glacier and ice-cap response times.

ΔT^* is a range of minimum temperatures for eventual disappearance of glaciers and ice-caps.

ΔB_o is the rise in sea level caused by the initial imbalance of the Greenland or Antarctic ice sheet.

ß and $ß_1$ are sensitivities of the mass balance (in terms of sea level rise) to global mean temperature changes.

$ß_2$ is the sensitivity of the areal mean Antarctic mass balance (in terms of sea level rise) to changes in temperature through possible instability of the West Antarctic ice sheet.

Appendix 4

GLOSSARY OF TERMS

Aerosol
A collection of airborne particles. The term has also come to be associated, erroneously, with the propellant used in “aerosol sprays”.

Biomass
The total weight or volume of organisms in a given area or volume.

Biome
A naturally occurring community of flora and fauna (or the region occupied by such a community) adapted to the particular conditions in which they occur (e.g., tundra).

Capital stocks
The accumulation of machines and structures that are available to an economy at any point in time to produce goods or render services. These activities usually require a quantity of energy that is determined largely by the rate at which that machine or structure is used.

Carbon cycle
The term used to describe the exchange of carbon (in various forms, e.g., as carbon dioxide) between the atmosphere, ocean, terrestrial biosphere and geological deposits.

Carbonaceous aerosol(s)
Aerosol(s) *(q.v.)* containing carbon.

Climate
Climate is usually defined as the “average weather”, or more rigorously, as the statistical description of the weather in terms of the mean and variability of relevant quantities over periods of several decades (typically three decades as defined by WMO). These quantities are most often surface variables such as temperature, precipitation, and wind, but in a wider sense the “climate” is the description of the state of the climate system.

Climate change (FCCC usage)
A change of climate which is attributed directly or indirectly to human activity that alters the composition of the global atmosphere and which is in addition to natural climate variability observed over comparable time periods.

Climate change (IPCC usage)
Climate change as referred to in the observational record of climate occurs because of internal changes within the climate system or in the interaction between its components, or because of changes in external forcing either for natural reasons or because of human activities. It is generally not possible clearly to make attribution between these causes. Projections of future climate change reported by IPCC generally consider only the influence on climate of anthropogenic increases in greenhouse gases and other human-related factors.

Climate sensitivity
In IPCC reports, climate sensitivity usually refers to the long-term (equilibrium) change in global mean surface temperature following a doubling of atmospheric CO_2 (or equivalent CO_2) concentration. More generally, it refers to the equilibrium change in surface air temperature following a unit change in radiative forcing ($^\circ C/W\ m^{-2}$).

Cloud condensation nuclei
Airborne particles that serve as an initial site for the condensation of liquid water and which can lead to the formation of cloud droplets.

CO_2 fertilization
The enhancement of plant growth as a result of elevated atmospheric CO_2 concentration.

Cryosphere
All global snow, ice and permafrost.

Damage function
The relation between changes in the climate and reductions in economic activity relative to the rate that would be possible in an unaltered climate.

Discount rate
The annual rate at which the effect of future events are reduced so as to be comparable to the effect of present events.

Diurnal temperature range
The difference between maximum and minimum temperature over a period of 24 hours.

Eddy mixing
Mixing due to small scale turbulent processes (eddies). Such processes cannot be explicitly resolved by even the finest resolution Atmosphere-Ocean General Circulation Models currently in use and so their effects must be related to the larger scale conditions.

Equilibrium response
The steady state response of the climate system (or a climate model) to an imposed radiative forcing.

Equivalent CO_2
The concentration of CO_2 that would cause the same amount of radiative forcing as the given mixture of CO_2 and other greenhouse gases.

External impacts/externalities
Impacts generated by climate change (or some other environmental change) that cannot be evaluated by a competitive market because of a lack of information and or the inability to act on that information.

Falsifiability rule
Science today recognizes that there is no way to prove the absolute truth of any hypothesis or model, since it is always possible that a different explanation might account for the same observations. In this sense, even the most well established physical laws are "conditional". Hence, with scientific methodology it is never possible to prove conclusively that a hypothesis is true, it is only possible to prove that it is false.

Feedback
When one variable in a system triggers changes in a second variable that in turn ultimately affects the original variable; a positive feedback intensifies the effect, and a negative feedback reduces the effect.

Flux adjustment
To avoid the problem of a coupled atmosphere-ocean general circulation model drifting into some unrealistic climatic state (e.g., excessively warm temperatures in the tropical Pacific ocean), adjustment terms can be applied to the fluxes of heat and precipitation (and sometimes the surface stresses resulting from the effect of the wind on the ocean surface) before being imposed on the model ocean.

Fossil fuel reserves
The quantity of a fossil fuel that is known to exist, based on geological and engineering evidence, and that can be recovered under current economic conditions and operating capabilities.

Fossil fuel resources
The quantity of fossil fuel that is thought to exist and that may be recoverable based on an explicit scenario for future economic conditions and operating capabilities.

GDP
Gross Domestic Product. The value of all goods and services produced (or consumed) within a nation's borders.

Greenhouse gas
A gas that absorbs radiation at specific wavelengths within the spectrum of radiation (infrared radiation) emitted by the Earth's surface and by clouds. The gas in turn emits infrared radiation from a level where the temperature is colder than the surface. The net effect is a local trapping of part of the absorbed energy and a tendency to warm the planetary surface. Water vapour (H_2O), carbon dioxide (CO_2), nitrous oxide (N_2O), methane (CH_4) and ozone (O_3) are the primary greenhouse gases in the Earth's atmosphere.

Halocarbons
Compounds containing either chlorine, bromine or fluorine and carbon. Such compounds can act as powerful greenhouse gases *(q.v.)* in the atmosphere. The chlorine and bromine containing halocarbons are also involved in the depletion of the ozone layer.

Infrared radiation
Radiation emitted by the Earth's surface, the atmosphere and by clouds. Also known as terrestrial and long-wave radiation. Infrared radiation has a distinctive spectrum (i.e., range of wavelengths) governed by the temperature of the Earth-atmosphere system. The spectrum of infrared radiation is practically distinct from that of solar *(q.v.)* or short-wave radiation because of the difference in temperature between the Sun and the Earth-atmosphere system.

Integrated assessment
A method of analysis that combines results and models from the physical, biological, economic and social sciences, and the interactions between these components, in a consistent framework, to project the consequences of climate change and the policy responses to it.

Lifetime
In general, lifetime denotes the average length of time that an atom or molecule spends in a given reservoir, such as the atmosphere or oceans. It is not to be confused with the response time of a perturbation in concentration. CO_2 has no single lifetime.

Marginal cost
The cost on one additional unit of effort. In terms of reducing emissions, it represents the cost of reducing emissions by one more unit.

Marine biosphere
A collective term for all living marine organisms.

Market damages
The value of damages generated by climate change (or some other environmental change) and evaluated based on information available to and usable by a competitive market.

Mitigation marginal cost function
The relation between the total quantity of emissions reduced and the marginal cost of the last unit reduced. The marginal cost of mitigation generally increases with the total quantity of emissions reduced.

Nitrogen fertilization
Enhancement of plant growth through the deposition of nitrogen compounds. In IPCC reports, this typically refers to fertilization from anthropogenic sources of nitrogen such as, man-made fertilizers and nitrogen oxides released from burning of fossil fuels.

"No-regrets" mitigation options
"No-regrets" mitigation options are those whose benefits, such as reduced energy costs and reduced emissions of local/regional pollutants, equal or exceed their cost to society, excluding the benefits of climate change mitigation. They are sometimes known as "measures worth doing anyway".

Non-market damages
Damages generated by climate change (or some other environmental change) and that cannot be evaluated by a competitive market because of a lack of information and/or the inability to act on that information.

Optimal control rate
The rate of intervention at which the net present value of the marginal costs of the intervention, equals the net present value of the marginal benefits of the intervention.

Parametrize (parametrization)
In climate modelling, this term refers to the technique of representing processes that cannot be explicitly resolved at the resolution of the model (sub-grid scale processes) by relationships between the area averaged effect of such sub-grid scale processes and the larger scale flow.

Photosynthesis
The metabolic process by which plants take CO_2 from the air (or water) to build plant material, releasing O_2 in the process.

Portfolio analysis
The mix of actions available to policy makers to reduce emissions or adapt to climate change.

Precautionary principal
Avoiding a solution that is irreversible, because the assumptions on which the solution is based may prove incorrect, in favour of a seemingly inferior solution that can be reversed.

Radiative damping
An imposed positive radiative forcing *(q.v.)* on the Earth-atmosphere system (e.g., through the addition of greenhouse gases) represents an energy surplus. The temperature of the surface and lower atmosphere will then increase and in turn increase the amount of infrared radiation being emitted to space, thus a new energy balance will be established. The amount that emissions of infrared radiation to space increase for a given increase in temperature is known as the radiative damping.

Radiative forcing
A simple measure of the importance of a potential climate change mechanism. Radiative forcing is the perturbation to the energy balance of the Earth-atmosphere system (in W m^{-2}) following, for example, a change in the concentration of carbon dioxide or a change in the output of the Sun; the climate system responds to the radiative forcing so as to re-establish the energy balance. A positive radiative forcing tends to warm the surface and a negative radiative forcing tends to cool the surface. The radiative forcing is normally quoted as a global and annual mean value. A more precise definition of radiative forcing, as used in IPCC reports, is the perturbation of the energy balance of the surface-troposphere system, after allowing for the stratosphere to re-adjust to a state of global mean radiative equilibrium (see Chapter 4 of IPCC94). Sometimes called "climate forcing".

Respiration
The metabolic process by which organisms meet their internal energy needs and release CO_2.

Soil moisture
Water stored in or at the continental surface and available for evaporation. In IPCC (1990) a single store (or "bucket") was commonly used in climate models. Today's models which incorporate canopy and soil processes view soil moisture as the amount held in excess of plant "wilting point".

Solar luminosity
A measure of the brightness of (i.e., the amount of solar radiation *(q.v.)* being emitted by) the Sun.

Solar radiation
Radiation emitted by the Sun. Also known as short-wave radiation. Solar radiation has a distinctive spectrum (i.e., range of wavelengths) governed by the temperature of the Sun. The spectrum of solar radiation is practically distinct from that of infrared *(q.v.)* or terrestrial radiation because of the difference in temperature between the Sun and the Earth-atmosphere system.

Spatial scales
Continental 10 - 100 million square kilometres (km^2).
Regional 100 thousand - 10 million km^2.
Local less than 100 thousand km^2.

Spin-up
"Spin-up" is a technique used to initialize an AOGCM. At present it is not possible to diagnose accurately the state of the coupled atmosphere-ocean system and therefore it is not possible to prescribe observed starting conditions for an experiment with an AOGCM. Instead, the atmosphere and ocean components of the model are run separately, forced with "observed" boundary conditions, followed perhaps by a further period of "spin-up" when the atmosphere and ocean are coupled together, until the AOGCM is near to a steady state.

Stratosphere
The highly stratified and stable region of the atmosphere above the troposphere *(q.v.)* extending from about 10 km to about 50 km.

Sustainable development
Sustainable development is development that meets the needs of the present without compromising the ability of future generations to meet their own needs.

Terrestrial biosphere
A collective term for all living organisms on land.

Thermocline
The region in the world's ocean, typically at a depth of 1 km, where temperature decreases rapidly with depth and which marks the boundary between the surface and deep ocean.

Thermohaline circulation
Large-scale density-driven circulation in the oceans, driven by differences in temperature and salinity.

Transient climate response
The time-dependent response of the climate system (or a climate model) to a time-varying change of forcing.

Tropopause
The boundary between the troposphere *(q.v.)* and the stratosphere *(q.v.)*.

Troposphere
The lowest part of the atmosphere from the surface to about 10 km in altitude in mid-latitudes (ranging from about 9 km in high latitudes to about 16 km in the tropics on average) where clouds and "weather" phenomena occur. The troposphere is defined as the region where temperatures generally decrease with height.

Turn-over time
The ratio between the mass of a reservoir (e.g., the mass of N_2O in the atmosphere) and the rate of removal from that reservoir (e.g., for N_2O, the rate of destruction by sunlight in the stratosphere *(q.v.)*).

Volatile Organic Compounds (VOCs)
Any one of several organic compounds which are released to the atmosphere by plants or through vaporization of oil products, and which are chemically reactive and are involved in the chemistry of tropospheric ozone production. Methane, while strictly falling within the definition of a VOC, is usually considered separately.

Wet/dry deposition
The removal of a substance from the atmosphere either through being washed out as rain falls (wet deposition) or through direct deposition on a surface (dry deposition).

WGII LESS scenario
Scenarios developed for the SAR WGII to assess low CO_2-emitting supply systems for the world. The scenarios are referred to as LESS: Low-Emissions Supply System.

"When" and "where" flexibility
The ability to choose the time (when) or location (where) of a mitigation option or adaptation scheme in order to reduce the costs associated with climate change.

Appendix 5

ACRONYMS AND ABBREVIATIONS

AGCM	Atmosphere General Circulation Model
AOGCM	Atmosphere-Ocean General Circulation Model
CFCs	Chloro-flouro-carbons
COP-2	Second Conference of the Parties to the UN/FCCC
GDP	Gross Domestic Product
GFDL	Geographical Fluid Dynamics Laboratory
HCFCs	Hydro-chloro-fluoro-carbons
HFCs	Hyro-fluoro-carbons
IAM	Integrated Assessment Model
IIASA	International Institute for Applied Systems Analysis
IMAGE	Intergated Model to Assess the Greenhouse Effect
IPCC	Intergovernmental Panel on Climate Change
IS92	IPCC Emissions Scenarios defined in IPCC (1992)
OECD	Organization for Economic Cooperation and Development
OGCM	Ocean General Circulation Model
R&D	Research and Development
S Profiles	The CO_2 concentration profiles leading to stabilization defined in the 1994 IPCC Report (IPCC, 1995)
SAR	IPCC Second Assessment Report
SBSTA	Subsidiary Body of the UN/FCCC for Scientific and Technological Advice
SCM	Simple Climate Model
SPM	Summary for Policymakers
TPs	IPCC Technical Papers
UN	United Nations
UNFCCC	United Nations Framework Convention on Climate Change
UV	Ultraviolet
VEMAP	Vegetation/Ecosystem Modelling and Analysis Project
VOCs	Volatile Organic Compounds
WEC	World Energy Council
WGI, II & III	IPCC Working Groups I, II and II
WMO	World Meteorological Organization
WRE Profiles	The CO_2 concentration profiles leading to stabilization defined by Wigley, *et al.* (1996)

Chemical symbols

Br	Atomic bromine
CFC-11	$CFCl_3$, or equivalently CCl_3F (trichlorofluoromethane)
CFC-12	CF_2Cl_2, or equivalently CCl_2F_2 (dichlorodifluoromethene)
CH_4	Methane
Cl	Atomic chlorine
CO	Carbon monoxide
CO_2	Carbon dioxide
HCFC-134a	CH_2FCF_3
HCFC-22	CF_2HCl (chlorodifluoromethane)
N_2O	Nitrous oxide
NO_x	The sum of NO & NO_2
O_3	Ozone
OH	Hydroxyl
S	Atomic sulphur
SO_2	Sulphur dioxide
SO_4^{2-}	Sulphate ion

Appendix 6

UNITS

SI (Système Internationale) Units

Physical Quantity	Name of Unit	Symbol
length	metre	m
mass	kilogram	kg
time	second	s
thermodynamic temperature	kelvin	K
amount of substance	mole	mol

Fraction	Prefix	Symbol	Multiple	Prefix	Symbol
10^{-1}	deci	d	10	deca	da
10^{-2}	centi	c	10^{2}	hecto	h
10^{-3}	milli	m	10^{3}	kilo	k
10^{-6}	micro	μ	10^{6}	mega	M
10^{-9}	nano	n	10^{9}	giga	G
10^{-12}	pico	p	10^{12}	tera	T
10^{-15}	femto	f	10^{15}	peta	P
10^{-18}	atto	a			

Special Names and Symbols for Certain SI-derived Units

Physical Quantity	Name of SI Unit	Symbol for SI Unit	Definition of Unit
force	newton	N	kg m s^{-2}
pressure	pascal	Pa	kg $m^{-1}s^{-2}$ (=N m^{-2})
energy	joule	J	kg m^{2} s^{-2}
power	watt	W	kg m^{2} s^{-3} (= Js^{-1})
frequency	hertz	Hz	s^{-1} (cycles per second)

Decimal Fractions and Multiples of SI Units Having Special Names

Physical Quantity	Name of Unit	Symbol for Unit	Definition of Unit
length	ångstrom	Å	10^{-10} m = 10^{-8} cm
length	micron	μm	10^{-6} m
area	hectare	ha	10^{4} m^{2}
force	dyne	dyn	10^{5} N
pressure	bar	bar	10^{5} N m^{-2} = 10^{5} Pa
pressure	millibar	mb	10^{2} N m^{-2} = 1 Pa
weight	ton	t	10^{3} kg

Non-SI Units

°C	degrees Celsius (0°C = 273 K approximately) Temperature differences are also given in °C (=K) rather than the more correct form of "Celsius degrees"
ppmv	parts per million (10^{6}) by volume
ppbv	parts per billion (10^{9}) by volume
pptv	parts per trillion (10^{12}) by volume
bp	(years) before present
kpb	thousands of years before present
mbp	millions of years before present

The units of mass adopted in this report are generally those which have come into common usage, and have deliberately not been harmonized, e.g.,

kt	kilotonnes
GtC	gigatonnes of carbon (1 GtC = 3.7 Gt carbon dioxide)
PgC	petagrams of carbon (1PgC = 1 GtC)
MtN	megatonnes of nitrogen
TgC	teragrams of carbon (1TgC = 1 MtC)
TgN	teragrams of nitrogen
TgS	teragrams of sulphur

Appendix 7

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List of IPCC outputs

I. IPCC FIRST ASSESSMENT REPORT (1990)

a) **CLIMATE CHANGE — The IPCC Scientific Assessment.** The 1990 report of the IPCC Scientific Assessment Working Group *(also in Chinese, French, Russian and Spanish).*

b) **CLIMATE CHANGE — The IPCC Impacts Assessment.** The 1990 report of the IPCC Impacts Assessment Working Group *(also in Chinese, French, Russian and Spanish).*

c) **CLIMATE CHANGE — The IPCC Response Strategies.** The 1990 report of the IPCC Response Strategies Working Group *(also in Chinese, French, Russian and Spanish).*

d) **Overview and Policymaker Summaries, 1990.**

Emissions Scenarios (prepared by the IPCC Response Strategies Working Group), 1990.

Assessment of the Vulnerability of Coastal Areas to Sea Level Rise — A Common Methodology, 1991.

II. IPCC SUPPLEMENT (1992)

a) **CLIMATE CHANGE 1992 — The Supplementary Report to the IPCC Scientific Assessment.** The 1992 report of the IPCC Scientific Assessment Working Group.

b) **CLIMATE CHANGE 1992 — The Supplementary Report to the IPCC Impacts Assessment.** The 1990 report of the IPCC Impacts Assessment Working Group.

CLIMATE CHANGE: The IPCC 1990 and 1992 Assessments — IPCC First Assessment Report Overview and Policymaker Summaries, and 1992 IPCC Supplement *(also in Chinese, French, Russian and Spanish).*

Global Climate Change and the Rising Challenge of the Sea. Coastal Zone Management Subgroup of the IPCC Response Strategies Working Group, 1992.

Report of the IPCC Country Study Workshop, 1992.

Preliminary Guidelines for Assessing Impacts of Climate Change, 1992.

III. IPCC SPECIAL REPORT, 1994

a) **IPCC Guidelines for National Greenhouse Gas Inventories** (3 volumes), 1994 *(also in French, Russian and Spanish).*

b) **IPCC Technical Guidelines for Assessing Climate Change Impacts and Adaptations,** 1994 *(also in Arabic, Chinese, French, Russian and Spanish).*

c) **CLIMATE CHANGE 1994 — Radiative Forcing of Climate Change and An Evaluation of the IPCC IS92 Emission Scenarios.**

IV. IPCC SECOND ASSESSMENT REPORT, 1995

a) **CLIMATE CHANGE 1995 — The Science of Climate Change.** (including Summary for Policymakers). Report of IPCC Working Group I, 1995.

b) **CLIMATE CHANGE 1995 — Scientific-Technical Analyses of Impacts, Adaptations and Mitigation of Climate Change.** (including Summary for Policymakers). Report of IPCC Working Group II, 1995.

c) **CLIMATE CHANGE 1995 — The Economic and Social Dimensions of Climate Change.** (including Summary for Policymakers). Report of IPCC Working Group III, 1995.

d) **The IPCC Second Assessment Synthesis of Scientific-Technical Information Relevant to Interpreting Article 2 of the UN Framework Convention on Climate Change,** 1995.

(Please note: the IPCC Synthesis and the three Summaries for Policymakers have been published in a single volume and are also available in Arabic, Chinese, French, Russian and Spanish).

IV. IPCC TECHNICAL PAPERS

Technologies, Policies and Measures for Mitigating Climate Change — IPCC Technical Paper 1.
(also in French and Spanish)

An Introduction to Simple Climate Models used in the IPCC Second Assessment Report — IPCC Technical Paper 2.
(also in French and Spanish)

Stabilization of Atmospheric Greenhouse Gases: Physical, Biological and Socio-economic Implications — IPCC Technical Paper 3.
(also in French and Spanish)