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PREFACE

The aim of the Environment Program is to provide the knowledge required for the development of policies aimed at ensuring environmental security. We recognize that the environmental issues cannot be treated in isolation if we are to achieve our goal. Environmental issues are closely linked with global concerns regarding increasing population, political and military security, technological and economic change, and humanitarian and social questions. Activities in the Program are therefore focussed on environmental problem areas which possess urgent needs for concise and realistic policy actions aimed at both reducing the stresses on the environment and implementing adjustment strategies. One of two themes in the Program is derived from expected global climate change caused by increasing atmospheric concentration of radiatively active gases, and its consequences for managed and natural ecosystems, with particular emphasis on agriculture, forestry and water resources.

The following paper is aimed directly at the questions concerning our major source of information on future climate change, that is, climate described by general circulation models (GCMs) of the atmosphere. Each of the present suite of GCMs, used in exploring climate response, is designed to correctly characterize different aspects of atmospheric dynamics, and hence, none of them will produce the same estimated daily temperature or precipitation patterns. Perhaps more important, none of the GCMs were developed to assess climate response to radiatively-active gases. Hence, none of them are more than coincidentally suited for the task, and all have very serious deficiencies for the purpose. For these reasons, Dr. Harrison's discussion of the most prominant GCMs used in climate change assessment, and her comparison of their output characteristics is a critical document for our progress on climate impacts research in the Environment Program. Harrison's paper fills a void in the literature, allowing the biologists, hydrologists, land planners, and agronomists involved in this research to understand the nature, strengths and weaknesses of predictions of climate response to increasing greenhouse gas concentrations.

Prof. Bo R. Doos, Leader
Environment Program

AN INTRODUCTION TO GENERAL CIRCULATION MODELLING EXPERIMENTS WITH RAISED CO_2

Sandy P. Harrison

INTRODUCTION

The possible effects of raised CO_2 and other greenhouse gases on climate have been investigated using general circulation models (GCMs) by several modelling groups including the UK Meteorological Office (UKMO), the US National Center for Atmospheric Research at Boulder (NCAR), the Goddard Institute for Space Studies, New York (GISS), the Geophysical Fluid Dynamic Laboratory of NOAA, at Princeton (GFDL), and Oregon State University, at Corvallis (OSU). The results of these model experiments are increasingly being used by specialists in other disciplines to assess the potential impacts of the greenhouse effect on other earth systems, including natural vegetation, crops and water resources.

Although the various GCMs share many common characteristics, they also differ in many ways, and the differences quite substantially affect the results of "greenhouse" simulations. Furthermore, the models are continually being updated and refined in key areas of uncertainty such as the representation of the ocean and land surface, and of clouds. It is therefore important that scientists concerned with impact assessments are aware of the variety of model formulations and the kinds of effects that different formulations will have on the reliability of model experiments.

The aim of this paper is to describe the basic structure of existing GCMs, the processes they simulate and how these are represented, with emphasis on differences between models that may be important for simulations of the greenhouse effect. Tables 1 and 2 summarise the characteristics of each model at the time of each set of raised CO_2 experiments. The paper does not attempt to describe the results of the simulations, which have been reviewed by Manabe (1983), Schlesinger (1984), Schlesinger and Mitchell (1985), Schlesinger and Mitchell (1987), Mitchell (1988), Schlesinger (1988), Mitchell (1989) and Schlesinger (1989).

GENERAL CIRCULATION MODELS

General circulation models (GCMs) are numerical simulation models that represent the physics of the atmospheric circulation mathematically. They simulate the dynamics of the three-dimensional structure of the atmosphere, coupled with the surface water and energy balances (e.g. Henderson-Sellers and McGuffie, 1987; Washington and Parkinson, 1986).

There are two basic types of GCM: **grid-point** and **spectral** models. These two types differ in the techniques they use to represent the horizontal structure of model variables (Simmons and Bengtsson, 1984). In grid-point or finite-difference models, variables are represented at a large number of grid points obtained by dividing the Earth's surface into a regular rectangular grid. Several different types of grid are used, the primary difference being whether the grid is regular in longitude or physical distance. The resolution of grid-point models is generally in the range of 2-5^{o} of latitude and longitude. In spectral models, the variables are represented in terms of truncated expansions of spherical harmonics. The spatial resolution of spectral models is determined by the level and type of truncation (Gordon and Stern, 1972; Henderson-Sellers and McGuffie, 1987). Spectral models have certain advantages over grid-point models: they are less subject to numerical instability; they place fewer limitations on the length of the time step; and, in general, they are computationally more efficient (Manabe et al., 1979b; Girard and Jarraud, 1982).

GCMs differ in the way they treat terrestrial geography. The earliest models (**sectorial** models) had a limited computational domain, corresponding to a sector of the globe, and used an idealised distribution of land and sea. Such models are useful exploratory tools but cannot be used to simulate regional climatic changes. Later models (**global** models) have a global computational domain and incorporate so-called realistic geography: that is, the distribution of land and sea corresponds to the real world distribution. In practice, the distribution of land, ice and sea in most models is made on a gridbox basis, that is each gridbox is assigned to the category which covers the largest area of the gridbox. The resulting distribution of land, ice and water is therefore simplified. For example, in many of the models the Mediterranean and the Baltic are isolated bodies of water with no outlet to the Atlantic, and islands such as New Zealand are omitted because they are too small. The GISS model is the only one which assigns fractional values of land, ice and water to an individual gridbox. Climatic variables (e.g. temperature) are calculated separately for each category of surface type and an area-weighted average value is then calculated for the entire gridbox.

In sectorial models with idealised geography, the land area was assumed to be flat. The global models with realistic geography incorporate surface topography at a scale appropriate to the model resolution. Since elevation is represented as the average over a whole gridbox, the resulting topography is highly smoothed. To illustrate this: the highest model elevations in the western United States are typically about 1800m in comparison with actual maximum elevations of ca. 3660m. Relatively small mountain systems such as the Alps may not appear.

The models are driven (or **forced**) by incoming solar radiation (insolation). In the simplest formulations, annual average solar radiation is used. A slightly more advanced method, allowing the effects of seasonal differences in insolation to be approximated, is to force the model with a monthly or seasonal average. For example, some of the simulations made with the NCAR CCM were forced by mean January and mean July values for solar radiation.

The average intensity of solar radiation as it enters the top of the atmosphere is known as the solar constant. According to satellite measurements, the solar constant is 1366-1367 W/m^2. Different models use different values of the solar constant. In some cases this is because they use a pre-satellite estimate (e.g. OSU which uses the old "best" estimate of 1354 W/m^2), in other cases the solar constant is altered as a simple way to tune the model to give a good simulation of the present climate (e.g. GFDL uses a high value of 1443.7 W/m^2 in some runs for this purpose). Such differences between models are unimportant, provided inter-model comparisons are made on simulated **anomalies**, i.e. differences between climatic variables as simulated under perturbed (e.g. high CO_2) and normal (control) conditions.

The intensity of solar radiation is least when the earth is furthest away from the sun (aphelion) in July and is nearly 7% greater when the earth is closest to the sun (perihelion) in January. Models with a true seasonal cycle (e.g. UKMO, NCAR M88a, GISS Model II, GFDL G15, OSU 2LM) vary the intensity of solar radiation through the year in a realistic fashion. The most advanced models incorporate not only the seasonal variations but also the diurnal variations in solar radiation (e.g. UKMO, GISS Model II).

CLIMATE FEEDBACKS

The response of the climate to changes in external forcing (boundary conditions) is affected by a number of positive and negative feedbacks. The most important feedbacks are related to the behaviour of the ocean, clouds, sea ice, and surface hydrology. The various models, and indeed versions of individual models, deal with the processes related to each of these feedback mechanisms in substantially different ways. The processes are sufficiently complex that a full physical simulation is impossible; instead, the processes have to be **parametrised** - represented approximately by simplified equations. The parametrisation of surface processes and clouds remains one of the key areas of uncertainty in climate modelling.

Ocean Treatment

Oceans are wet (and therefore act as an unlimited source of moisture for the hydrological cycle), they store heat, they advect heat from the near-surface layers down to the deep ocean, and they also advect heat (through e.g. currents) horizontally. The earliest attempt to incorporate oceans into GCM experiments was by prescribing SSTs from modern climatological data. These **prescribed swamp ocean** models act as a moisture source, but cannot respond to atmospheric temperature changes. **Energy-balance swamp ocean** models calculate SSTs using an energy-balance approach. Swamp oceans are wet, a characteristic which affects both surface temperatures and means that they act as a moisture source, but they do not store or advect heat. They are run with an atmospheric model forced by some kind of average value for solar radiation.

Simple **slab** or **mixed-layer** oceans (e.g. UKMO $11LM_a$) represent an improvement in the treatment of ocean behaviour because, in addition to acting as a moisture source, they store heat. Slab oceans may have a fixed or a seasonally-varying depth, prescribed from climatological data to correspond to the isothermal mixed-layer of the upper part of the ocean. Slab ocean models can usefully be run with a full seasonal cycle and thus simulate the effects of heat storage on the cycle of sea-surface temperatures through the year. In the simplest slab ocean formulations there is no attempt to incorporate either horizontal heat advection or heat advection into the deep ocean, with the result that simulated SSTs are generally too high in equatorial latitudes and too low towards the poles. More advanced slab ocean models (e.g. UKMO $11LM_b$) attempt to parametrise the 2-dimensional dynamical behaviour of the ocean (e.g. currents) through prescribed heat convergence (or divergence). In other words, a pattern of heat transport is included that more closely simulates the present-day pattern of sea-surface temperatures.

Finally, there are fully computed ocean models (OGCMs) which can be coupled to atmospheric GCMs (e.g. NCAR WM89, GFDL COAM, OSU CGCM). OGCMs incorporate both heat storage and 3-dimensional dynamics (Bryan, 1989). The simplest method of linking OGCMs with atmospheric GCMs is to run them simultaneously such that there is a continuous two-way feedback between the atmosphere and the ocean (Washington and Chervin, 1980). Such models (e.g. NCAR WM89, OSU CGCM) are said to be **synchronously coupled.** However, synchronously coupled atmosphere-ocean models are very demanding of computer time. Capitalising on the fact that the ocean responds much more slowly than the atmosphere, various more complicated methods of linking the ocean and atmospheric components have been developed in order to reduce computing time (e.g. Manabe and Bryan, 1969; Bryan and Lewis, 1979; Manabe et al., 1979a). **Asynchronously-coupled** ocean-atmosphere models (e.g. GFDL COAM) are run sequentially, with the time-averaged output of one component used to force the other component. Asynchronous coupling is appropriate for model experiments concerned with the long-term average response to changes in CO_2 concentration, but investigations of the evolution of climate through time can only be made using synchronously coupled models (Washington and Chervin, 1980).

Cloud treatment

Clouds affect how much of the incoming solar radiation actually reaches the earth's surface. Low clouds (within 2 km of the earth's surface, cf. Slingo, 1990) reflect back incoming radiation, and thus have a cooling effect on surface temperatures. High clouds also reflect back incoming solar radiation, but they reflect less than the ground surface or low clouds in the infrared part of the spectrum. Their net effect is therefore to raise surface temperatures. The climate response to raised CO_2 is likely to be particularly sensitive to the treatment of high cloud.

Early versions of the various models used a **prescribed cloud** distribution, using climatological data on the zonal and vertical distribution of clouds. Different models used somewhat different climatological data for prescribing cloud distribution. The same prescribed distribution was used both for the control and raised CO_2 simulations. The raised CO_2 simulations therefore make no attempt to incorporate the effects that might occur because of changes in the distribution of clouds in a warmer world.

The development of schemes to predict cloud distribution made it possible to incorporate the effects of cloud feedbacks (**interactive clouds**). In the simplest predictive schemes, cloud cover is a function of relative humidity (e.g. UKMO $11LM_b$, GISS model II, GFDL VC). A number of more complex formulations, including an explicit cloud water variable, have been developed by the UKMO group. It is clear that climate sensitivity to raised CO_2 varies according to the cloud predictive scheme used (Mitchell et al., 1989), but the processes involved are complex and it is not clear what is the best way to predict cloud behaviour.

Sea-ice treatment

The role of sea ice in the climate system is discussed by e.g. Hibler (1984), Semtner (1984), van Ypersele (1989). Sea ice has a higher albedo than sea water (0.6-0.7 for bare ice, and up to 0.9 for snow-covered ice) and thus increases the amount of solar energy reflected back into space. Sea ice formation can also lead to a marked reduction in surface temperatures because sea ice is relatively isolated from the ocean heat reservoir. Sea ice acts as a barrier to exchanges between the atmosphere and ocean (e.g. of sensible heat, momentum, water vapour, CO_2). Finally, the expulsion of salt during sea ice formation increases the density of the sea water beneath the ice, destabilises the water column, and is thought to play an important role in deep ocean circulation.

Sea ice formation and melting are governed by heat fluxes between the atmosphere, the ice and the ocean. In the earliest model experiments the extent of sea ice was prescribed from climatological data. In model experiments with a energy-balance swamp ocean, sea ice exists whenever the SST is below the value at which sea water freezes. In experiments with a slab ocean, the existence and thickness of sea ice is determined by an energy budget scheme which includes accumulation through snowfall and sea water freezing, and destruction through ice melting and sublimation.

The NCAR and GFDL coupled atmosphere-ocean models incorporate a simple thermodynamic sea-ice model, based on Semtner (1976), which predicts sea ice formation and extent. The OSU CGCM uses a slightly more advanced thermodynamic model (Parkinson and Washington, 1979) which allows for the horizontal inhomogeneities of sea ice (e.g. the presence of open water areas or leads).

Sea ice moves in response to winds and currents, and since it is not a rigid material may also deform under pressure. This dynamic behaviour controls ice distribution and thickness, and can thus affect the heat exchange between the atmosphere and the ocean. Although dynamic sea-ice models are being developed that incorporate these complexities (e.g. Hibler, 1979, 1988; Semtner, 1987), they have not been incorporated into coupled atmosphere-ocean models.

Soil moisture treatment

The storage of moisture in the soil affects heat and moisture fluxes. Moist soil loses energy through latent heat flux and remains cool, but dry ground warms until it is hot enough for energy to be lost through sensible heat flux. As a result, surface temperatures are considerably higher over dry ground. The soil also acts as a source of moisture for precipitation: rainfall is increased locally in areas with high soil moisture stores (Manabe, 1975; Charney et al., 1977; Shukla and Mintz, 1982; Rind, 1982). Changes in soil moisture storage are therefore likely to have important feedback effects on climate.

The treatment of soil moisture storage in most GCMs is relatively simple: the soil is represented as a **bucket** with a fixed water-holding capacity, most commonly the equivalent of 15 cm of water. The simulated moisture content of the soil is increased by rainfall, condensation and snowmelt, and decreased by evaporation. When the bucket is full, that is when the soil is saturated, additional moisture inputs are lost to the system as **runoff**. This use of the term runoff is not equivalent to its standard use in hydrology, since there is no mechanism to transport excess moisture horizontally over the land surface. Evaporation is assumed to occur at potential rate when the amount of water in the soil bucket reaches a specified level, which varies between 33-75% of maximum water-holding capacity in different models. The ratio of evaporation to potential evaporation is generally treated as a linear function of soil moisture content up to this specified level.

In the real world, runoff is also generated when the rate of moisture delivery is greater than the rate at which the soil can absorb it (the infiltration rate). The infiltration rate is partly dependent on soil type. Some models (e.g. the UKMO $11LM_c$ model) parametrise this process by increasing runoff non-linearly as a function of precipitation rate, and by assigning different rates of increase to different soil types. Moisture can also be lost from the soil through gravitational drainage out of the root zone. The rate of gravitational drainage is again dependent on soil type. Gravitational drainage is included in some model representations of the soil, and is parametrised as a non-linear function of soil moisture content and soil type.

In reality, soil water-holding capacity varies with soil type, structure, layering and depth. Furthermore, the distribution of moisture within the soil is an important determinant of the ease with which water can be removed, either through evaporation from the surface or by gravitational drainage. The treatment of surface hydrology in GISS Model II attempts to include these complexities. In the GISS model, the soil is treated as a two-layer bucket in which each layer can have a different water-holding capacity and water-holding capacity is varied geographically. The moisture content of the top layer is increased by rainfall, condensation, snowmelt and upward diffusion from the lower layer, and decreased by evaporation and gravitational drainage into the lower layer. The moisture content of the lower layer is increased by gravitational drainage from above and decreased by upward diffusion into the upper layer. Unfortunately, in the implementation of this soil model many of the advantages of the advanced conceptualisation of soil processes are lost. The treatment of soil moisture storage and surface hydrology remains one of the weaker parts of current climate models.

Considerably more complex and realistic treatments of fluxes between the land surface and atmosphere, including water vapour flux, are included in such models as the Simple Biosphere (SiB) model (Sellers et al., 1986, 1988) and the biosphere-atmosphere transfer scheme (BATS) (Dickinson et al., 1986; Wilson et al., 1987). These models are designed for incorporation into GCMs and for investigating the effects of land-surface characteristics (such as vegetation structure) on the atmosphere. However, these more complex models are not included in any of the raised CO_2 simulations discussed here.

EQUILIBRIUM AND TRANSIENT GCM EXPERIMENTS

When a system is perturbed, changes occur until the system is once more in equilibrium with the external conditions (forcing). Most GCM simulations of the effects of raised CO_2 have simulated this **equilibrium response** to an instantaneous or **step-function increase** in CO_2. In general, the models have simulated the response to a large increase in CO_2 (e.g. a doubling or quadrupling) in order to optimise the chance of distinguishing the effects of the CO_2 increase from the natural day-to-day and year-to-year variability that is a feature of both the real atmosphere and of GCM simulations.

The speed with which equilibrium is re-established after a step-function change in external conditions depends on the magnitude of the change and the response times of the components of the system. The climate system is complex, and its various components respond at different rates. The atmosphere itself responds relatively quickly (weeks), while the greater heat capacity of the oceans means that they respond more slowly. It is thought that the state of the mixed-layer of the ocean may lag radiation changes by several decades (e.g. Hasselmann, 1979; Thompson and Schneider, 1979; Hoffert et al., 1980) and the full 3-dimensional system of ocean currents may have even greater inertia. Finally, large continental ice sheets take thousands of years to build or melt (Imbrie, 1985).

The time-dependent or **transient response** of the climate system on a time-scale of decades to centuries can be investigated using coupled ocean-atmosphere models. The transient response to a step-function increase in CO_2 concentration has been investigated using the NCAR WM89 model, the GFDL $COAM_b$ model, and the OSU CGCM. These simulations do not allow for possible ice-sheet reduction, but they do explicitly include the dynamics of the ocean circulation which may act to delay the full effects of the change.

However, CO_2 is actually increasing continuously. It has been suggested that the transient and equilibrium responses to changing CO_2 concentrations may not be very different if the change in small and occurs slowly, but could be significantly different when the change is large and/or fast (Schneider and Thompson, 1981; Harvey, 1989). The transient response to a large step-function increase in CO_2 concentration may be significantly different from the transient response to gradually increasing CO_2. The NCAR WM89 model has been used to investigate the transient response to a **continuous increase** in CO_2 concentration.

Model runs forced by continuously changing CO_2 and other greenhouse gases have also been made by the GISS group (e.g. GISS IId). However, although these experiments were described as transient runs (because they are subject to so-called "transient forcing", i.e. a continuous change in the radiation regime), they were made with a slab ocean and thus could not simulate the transient response of the ocean circulation.

RAISED CO_2 EXPERIMENTS

UKMO model

The first set of raised CO_2 experiments (Mitchell, 1983) was made with the 5LM version of the UKMO grid-point model, as described by Slingo (1982) and developed from the original formulation of Corby et al. (1977). The model has 5 vertical layers and a quasi-uniform horizontal grid with a grid length of approximately 330km. The model takes into account both the diurnal and the seasonal variations in solar radiation. Cloud behaviour is prescribed, using zonal mean cloud amounts derived from seasonal climatological data sets. Sea surface temperatures (SSTs) and sea-ice extent are prescribed from climatological data sets. The albedos of sea, sea ice and permanent snow are fixed, the albedo of snow-free land varies as a prescribed function of latitude (following Corby et al., 1977), and the albedo of snow covered land is a function of snow depth. The treatment of surface hydrology is relatively simple: the moisture content of the soil is increased by rainfall, condensation and snowmelt, and decreased by evaporation. The maximum amount of soil moisture storage is 20cm, and "runoff" occurs when the soil is saturated. The ratio of evaporation to potential evaporation is a linear function of soil moisture content, such that evaporation occurs at the potential rate when soil moisture is equal to or greater than 10cm.

The control experiment (Mitchell, 1983) was run for 1192 days, with "normal" CO_2 and modern SSTs and sea-ice extent. The results are averaged over the last 3 years (1095 days) of the simulation.

The first experiment ($2xCO_2$) was run for just over one year, with doubled CO_2 and modern SSTs and sea-ice extents. The results are averaged over the last year (365 days) of the simulation. This experiment estimates the short term response to an instantaneous increase in CO_2 in the absence of cloud or ocean feedbacks. Since vertical mixing between the mixed layer and the deep ocean could delay the oceanic response, this experiment provides a useful lower limit to possible climatic changes due to doubling CO_2.

The second experiment (C2S2) was initialised from day 153 (late October) values of the $2xCO_2$ experiment, and run for 855 days with doubled CO_2 and SSTs increased everywhere by 2K. The results are averaged over the final two years. The experiment estimates the possible effects of oceanic feedbacks on the response to CO_2 doubling. It is assumed that the CO_2 effect on the ocean is sufficiently small to be considered as a perturbation of the basic global circulation. The estimate of 2K is a plausible value for the temperature increase based on the range of estimates from single column radiative-convective equilibrium models.

The third experiment ($10xCO_2$) was run for just over one year, with decupled CO_2 and modern SSTs and sea-ice extents. The results are averaged over the last year of the simulation. The experiment improves the chance of detecting local responses raised CO_2 concentrations by enhancing the response to relative to the level of the model's natural variability. The response to increasing CO_2 is thought to vary logarithmically with CO_2 concentration, so the changes in this simulation should be 3.3 times those in the $2xCO_2$ experiment.

Mitchell and Lupton (1984) used the 5LM version of the UKMO model to examine the response to quadrupled CO_2 levels.

This experiment (C4SL) was run for 1134 days from day 148 (22 October) of the control run, with quadrupled CO_2 and prescribed SSTs which were changed by different amounts depending on latitude. The SST increments were chosen, on the basis of previous experiments, such that there was no net change in surface heating at each latitude. This constraint requires that the implied zonally-averaged meridional advection of heat by the ocean is unchanged. The extent of sea-ice is reduced compared to previous experiments (e.g. C2S2). The results are averaged over the final 3 years (1095 days) of the simulation.

Note that in order to compare the results from C4SL with previous experiments, Mitchell and Lupton (1984) refer to a hypothetical experiment (C4S4). The "results" from this experiment are simply double those of C2S2.

Wilson and Mitchell (1987a) used the 5LM model to examine the consequences of raised CO_2 concentrations for the climate of Europe. The control run ($control_{ext}$) was the same as that described in Mitchell (1983), except that it was extended to 1464 days (4 yr). The results were averaged over the last 1095 days (3 yr) of the simulation. The experimental run ($C4SL_{ext}$) was the same as the $4xCO_2$ experiment of Mitchell and Lupton (1984), except that it was extended to 1316 days. The results were again averaged over the last 3 years (1095 days) of the simulation.

Mitchell et al. (1987) describe raised CO_2 experiments made with an higher resolution version of the UKMO model, the 11LM version. The model has 11 vertical layers and a regular grid of 2.5^o latitude by 3.75^o longitude. The model takes into account both the diurnal and the seasonal variations in solar radiation. Cloud behaviour is prescribed, using zonal mean cloud amounts derived from seasonal climatological data sets. Sea surface temperatures (SSTs) and sea-ice extent are prescribed from climatological data sets. The albedos of sea, sea ice, permanent snow, land and snow-covered land are all fixed. The treatment of surface hydrology is the same as in the 5LM version, except that full evaporation takes place when the soil moisture content is 5cm, and runoff when it is 15cm.

The 11LM version has been described by Slingo (1985a, 1985b) and a comparison of the two versions, using a 220km quasi-uniform grid, was made by Mitchell and Bolton (1983). The large-scale responses of the two models are apparently similar, but there are important differences at a regional scale (e.g. in precipitation). The climatology of the 11LM version is somewhat more similar to climatological observations, particularly with regard to precipitation.

The control experiment (Mitchell et al., 1987) was run for 8 years starting with real data for 25th July 1979, with "normal" CO_2 and modern SSTs and sea-ice extent. Seasonal values were averaged over the 8 years while annual mean values were apparently computed from the last 2 years of the run.

The experiment (2C2S) was initialised from 1 March of the 2nd year of the control and run for 3 years, with doubled CO_2 and SSTs increased everywhere by 2K. Seasonal values were averaged over the 3 years while annual mean quantities were computed from the last 2 years of the experiment.

Subsequent raised CO_2 experiments (Wilson and Mitchell, 1987b; Mitchell and Warrilow, 1987; Mitchell et al., 1989) have been made using slightly modified versions of the 11LM model, coupled to a simple "slab" ocean with a 50m fixed-depth oceanic mixed layer and an energy-balance sea-ice model. The 2-dimensional dynamic behaviour of the ocean (e.g. heat advection by currents) is simulated through the prescription of oceanic heat convergence.

The version used by Wilson and Mitchell (1987b) has 11 vertical layers and a regular grid of 5^{o} latitude by 7.5^{o} longitude. Cloud behaviour is predicted, using a scheme where cloud cover is a function of relative humidity (RH scheme). The albedos of sea, sea ice, and permanent snow are fixed; land albedo varies geographically according to vegetation cover and soil type (based on data in Wilson and Henderson-Sellers, 1985); the albedo of snow-covered land varies with snow depth.

The control experiment (Wilson and Mitchell, 1987b) was run for 20 years, with a CO_2 level of 323 ppmv. The results are averaged over the last 15 years of the simulation.

The experiment ($2xCO_2$) was started from the end of year 7 of the control simulation and run for 38 years, with doubled CO_2 (646 ppmv). The results are averaged over the last 15 years of the simulation. Although the control run had reached equilibrium after 10 years, after 25 years the experiment had only reached 90% of the final equilibrium level. This experiment estimates the response to an instantaneous increase in CO_2 taking into account possible cloud and ocean feedbacks.

The sensitivity of the response to raised CO_2 to the model's representation of land surface hydrology was investigated by Mitchell and Warrilow (1987). They used the version of the 11LM-slab ocean model described in Wilson and Mitchell (1987b), with a regular grid of 5^o latitude by 7.5^o longitude, cloud behaviour predicted according to the RH scheme, and geographically varying land albedo. However, the surface hydrology was altered such that "runoff" is generated when the infiltration capacity of the soil is exceeded as well as when the soil is saturated, and water can be removed from the soil by gravitational drainage from the root zone in addition to evaporation. Runoff increases non-linearly with precipitation, the rate of increase being more rapid with convective precipitation, and also increases from coarse to fine soils. Drainage from the root zone increases non-linearly with soil moisture content, and decreases from coarse to fine soils. The root depth, which affects the partitioning between runoff and evaporation, is set to 1m (considered typical for woodland vegetation).

Three control simulations and three doubled CO_2 experiments were run with this model($control_{cly}$, $control_{med}$, $control_{snd}$ etc.), using runoff and drainage parameters appropriate to clay, medium and sandy soils respectively. A further paired set of runs ($control_{frz}$ and $2xCO_{2frz}$) with a medium soil was made in which the infiltration capacity of frozen soil was set to zero, so that runoff occurred immmediately with all rain or snow-melt events. All the simulations were run for 6 years. The results are averaged over the last 5 years of each simulation.

The sensitivity of the response to raised CO_2 to the model's parametrization of cloud processes was investigated by Mitchell et al. (1989). They used the version of the 11LM-slab ocean model described in Mitchell and Warrilow (1987), with a regular grid of 5^o latitude by 7.5^o longitude, geographically varying land albedo, and the more complex surface hydrology formulations. However, a series of different schemes were used to predict cloud behaviour. Cloud treatment was changed by including an explicit cloud-water variable for all but deep convective cloud, to incorporate the effects of changes in the state of cloud water (CW scheme). In the first version of this scheme, ice-cloud particles form as a function of temperature once the temperature falls below 0^o C and are assumed to start falling at once with a fixed fall speed of 1m s^{-1}. In a second version of the scheme (CWH) the ice fall speed is parametrized in terms of ice-water content. A third version of the scheme (CWRP) modifies the CW formulation so that cloud radiative properties vary with cloud water content.

Control (320 ppmv) and doubled CO_2 (640 ppmv) simulations were made with each of these cloud schemes. The results are averaged over the last 5 years of each simulation. The results are compared to the runs made using the RH cloud scheme presented by Mitchell and Warrilow (1987). Note that there appears to be a slight discrepancy between the CO_2 concentrations used in the two sets of experiments.

NCAR Community Climate Model

The NCAR Community Climate Model (CCM) is a spectral model with so-called realistic geography that evolved from the Australian spectral model described by Bourke et al. (1977) and McAvaney et al. (1978).

The first set of raised CO_2 experiments (Washington and Meehl, 1983) were made with a version of the model described by Pitcher et al. (1983). The model has 9 vertical layers and a horizontal resolution of 4.4^o latitude by 7.5^o longitude (i.e. 40 by 48 grid points). The model is driven by annual average solar forcing, with a value of 1370 W m^{-2} for the solar constant. Cloud behaviour is prescribed, from climatological data sets. SSTs and sea-ice extent are calculated using a simple energy-balance swamp ocean. The albedos of sea, sea ice, permanent snow, desert and non-desert land are fixed. There is no allowance for the gradual attenuation of snow or sea-ice cover, so albedo of a grid cell only changes when the whole cell is free of snow or ice. This constraint is likely to reduce the model sensitivity at high latitudes to increases in CO_2. The treatment of surface hydrology follows Washington and Williamson (1977): the moisture content of the soil is increased by rainfall and snowmelt, and decreased by evaporation. The maximum amount of soil moisture storage is 15cm, and "runoff" occurs when the soil is saturated. The ratio of evaporation to potential evaporation is a linear function of soil moisture, such that evaporation occurs at the potential rate when soil moisture is equal or greater than 11.25cm.

The control experiment (Washington and Meehl, 1983) was run for 600 days, with modern CO_2 and SSTs and sea-ice extent. The results are averaged over the last 360 days of the simulation.

The 2xCO_2 and 4xCO_2 experiments were also run for 600 days, with doubled and quadrupled CO_2 respectively. The results were averaged over the last 360 days of the simulation.

A second set of experiments (Washington and Meehl, 1983) was made with a version of the model with an interactive cloud scheme, as described by Ramanathan et al. (1983). In this version, cloud formation occurs when relative humidity is greater than 80%, and convective clouds are formed when the vertical gradient of the equivalent potential temperature is less than zero.

The control experiment (Washington and Meehl, 1983) was run for 670 days, with modern CO_2 and SSTs and sea-ice extent. The results are averaged over the last 360 days of the simulation.

The $2xCO_2$ and $4xCO_2$ experiments were also run for 670 days, with doubled and quadrupled CO_2 respectively. The results were averaged over the last 360 days of the simulation.

Washington and Meehl (1986) used the same version of the model, with interactive clouds and a simple energy balance swamp ocean, to investigate the sensitivity of CO_2 response to sea-ice and snow albedo/melting parametrization. In the original experiments, described in Washington and Meehl (1983), sea ice forms at -1.8° C and always has an albedo of 0.7, while the albedo of snow was set at 0.8 for the shortwave and 0.55 for the longwave part of the solar spectrum. In the revised formulation (SSIA), sea ice and snow albedos vary with temperature, such that the albedo of sea ice is 0.35 if the surface temperature is > -10° C and 0.7 if the temperature is < or equal to -10° C, and similarly snow albedo is set to 0.4 if the surface temperature is > -10° C and 0.8 if the surface temperature is < or equal to -10° C.

The effects of ice-albedo feedback on the response to CO_2 warming should be greater when the initial conditions are colder, and there is therefore more snow and sea ice. In order to test this, Washington and Meehl (1986) ran an additional set of experiments (SSIA+DSC) in which the surface temperature of the control run was lowered by reducing the solar constant by 2% to 1343 W m^{-2}.

The SSIA control and doubled CO_2 runs were all started at the end of the original 670-day control simulation described in Washington and Meehl (1983) and run for an additional 1040 days. The results are averaged of the last 360 days of the simulations. The SSIA+DSC control and doubled CO_2 runs were also started at the end of the original 670-day control simulation and run for 680 days. The results are again averaged over the last 360 days of the simulations.

Washington and Meehl (1984) describe raised CO_2 experiments with the NCAR CCM, coupled to a simple slab ocean with a 50m fixed-depth oceanic mixed layer. SSTs are determined by a simple energy balance and seasonal heat storage. There is thus no attempt to simulate ocean heat transport. The initial formation of sea ice depends on SST, and its subsequent growth and melting on a simple energy-balance model. The atmospheric model has 9 vertical layers and a horizontal resolution of 4.5^o latitude by 7.5^o longitude (i.e. 40 by 48 grid points), and takes into account seasonal variations in solar radiation but does not include a diurnal cycle (Meehl and Washington, 1988). Cloud behaviour is predicted, according to the scheme of Ramanathan et al. (1983). The albedos of sea ice, snow-covered surfaces, desert and non-desert land are fixed; sea albedo varies as a function of solar zenith angle. The treatment of surface hydrology follows Washington and Williamson (1977).

The control experiment was started from the end of the annual mean solar forcing experiment described by Washington and Meehl (1983), then run through 12 solar cycles each lasting ca 40.6 days (1st phase), then 4 solar cycles each lasting ca 121.7 days (2nd phase), and finally run for 11 solar cycles with the standard 365-day length. The results are averaged over the last 3 years of the simulation.

The 2xCO_2 experiment was started at the beginning of 2nd phase of the control run, and then run through 4 solar cycles each lasting 40.6 days and then 11 solar cycles of the standard length. Thus the total length of the experimental run was 15 years. The results are averaged over the last 3 years of the simulation.

Results from these experiments are also analysed in Meehl and Washington (1985), Bates and Meehl (1986), Meehl and Washington (1986), Dickinson et al. (1987) and Meehl and Washington (1988).

Meehl (1988) describes a series of experiments made with the atmospheric GCM described by Washington and Meehl (1984). In the first experiment (SPEC SST) the atmospheric GCM is coupled to a swamp ocean and driven by the annual cycle of observed SSTs. This experiment differs from earlier runs with a swamp ocean because it includes a seasonal cycle. The model is run for 5 years, and results are averaged over the last 3 years of the simulation. In the second and third experiments (MIX1 and MIX2), the atmospheric GCM is coupled to a simple mixed-layer slab ocean as described by Washington and Meehl (1984). MIX1 is run with modern CO_2, and MIX2 with doubled CO_2. Both experiments are run for 12 years, preceded by 2 phases with accelerated annual cycles. Results are averaged over the last 3 years of each simulation. Note that Meehl (1988) describes the MIX1 and MIX2 runs as though they are equivalent to the control and doubled CO_2 runs of Washington and Meehl (1984), but the length of the final phase with the non-accelerated annual cycle in the earlier paper is only 11 years.

The sensitivity of the climate response to raised CO_2 to the model parametrization of cloud processes has been investigated using a 12-layer version of the NCAR CCR (Slingo, 1990), driven by average solar radiation for January and July, with prescribed sea surface temperatures. The first version ($12LM_1$) has a horizontal resolution determined by rhomboidal spectral truncation at wave 15 (ca. 4.5^o latitude by 7.5^o longitude), and clouds predicted using the "standard" scheme. In the second version ($12LM_2$), the horizontal resolution is the same but the cloud prediction scheme of the ECMWF medium-range forecast GCM is used. The third version ($12LM_3$) has a higher horizontal resolution, determined by triangular spectral truncation at wave 42 (ca. 2.8^o latitude by 2.8^o longitude), and uses the ECMWF cloud prediction scheme. Experiments were run with doubled CO_2. The results were averaged over 500 days of each simulation.

Washington and Meehl (1989) and Washington (1990) describe raised CO_2 experiments with the NCAR CCM, synchronously coupled to a coarse-grid ocean general circulation model. The atmospheric model is the seasonal-cycle version of the NCAR CCM described by Washington and Meehl (1984). The atmospheric model has 9 vertical levels, a horizontal resolution of 4.5^o x 7.5^o, interactive clouds, and a simple soil bucket model. The albedos of sea ice, snow-covered surfaces, desert and non-desert land are fixed; sea albedo varies as a function of solar zenith angle. The ocean general circulation model (OGCM), which was adapted from Semtner (1974), has been described by Washington et al. (1980) and Meehl et al. (1982). It has 4 vertical layers and a horizontal resolution of 5^o latitude by 5^o longtitude. Sea-ice formation and extent are calculated using a simple thermodynamic model (Semtner, 1976).

The ocean model was started from the end of an uncoupled ocean model experiment with observed atmospheric forcing (Meehl et al., 1982). The atmospheric model was started from the 15th year of the climate simulation described by Washington and Meehl (1984). The coupled model was then run synchronously for 16 years (though with several changes in ocean diffusion parameters). Three experiments (control, $2xCO_2$, and transient CO_2) were begun at this point and each run for a further 30 years.

The control experiment (Washington and Meehl, 1989) was run with CO_2 concentration set to 330 ppmv. The doubled CO_2 experiment (Washington and Meehl, 1989) was run with CO_2 concentration set to 660 ppmv (equivalent to an instantaneous doubling of CO_2). In the transient experiment (Washington and Meehl, 1989; Washington, 1990), CO_2 concentration was started at 330 ppmv and increased linearly by 1% a year, such that the CO_2 concentration was 429 ppmv by the end of the 30 year run. The results of the various experiments are averaged over the last 5 years of the simulations.

GISS model

The GISS GCM is a grid-point model with so-called realistic geography. The basic structure and development of the model are described by Hansen et al. (1983).

The first set of raised CO_2 experiments (Hansen et al., 1984) were made with a slightly modified version of Model II (Hansen et al., 1983). The model has 9 vertical layers and a horizontal resolution of 8^{o} latitude by 10^{o} longitude. Each grid cell has appropriate fractions of land, ocean and sea ice. The model takes into account both the diurnal and the seasonal variations in solar radiation. The value of the solar constant is taken as 1367 W m^{-2}. Cloud cover and height are computed, using a relative humidity type scheme. In the documented version of Model II, SSTs ands sea ice extent are prescribed from climatological data sets. In the modified version used for the raised CO_2 experiments, SSTs and sea ice extent are computed using a simple mixed layer ocean with a maximum depth of 65m (Hansen et al., 1984). Ocean heat transport is obtained from the divergence of heat implied by energy conservation at each ocean grid point, using mixed layer depths specified from monthly climatological data. The heat capacity of the mixed layer is also prescribed. Ocean ice cover is computed by a simple energy-balance model, such that sea ice grows horizontally until the whole grid cell is covered and then increases in thickness. The temperature of the oceanic mixed layer is not allowed to exceed 0^{o} C until all the ice in a grid cell has melted; the excess heat is used to melt the ice. Sea ice, land ice and sea albedos are fixed. Land albedo is a function of vegetation type, with a separate value for each season in both the visible and the near IR part of the spectrum. The distribution of the 8 vegetation types (desert, tundra, grass, shrub, woodland, deciduous, evergreen and rainforest) are derived from data in Matthews (1983). Snow albedo depends on snow depth, age, masking by vegetation and the albedo of the underlying ground. Surface hydrology is characterised by a spatially-variable two layer soil bucket. The field capacity, or maximum amount of water storage, of each layer is specified according to vegetation type, such that the field capacity of the upper layer is 1 cm in desert, 3 cm for tundra, grassland, shrub, woodland, deciduous and evergreen, and 20 cm for rainforest; the field capacity of the lower layer is 1 cm in desert, 20 cm in tundra and grassland, 30 cm in shrub and woodland, and 45 cm in deciduous, evergreen and rainforest (Rind, 1984). The field capacity of each grid cell is determined by area weighting the values associated with each vegetation type over the cell. Rain falling onto the surface is divided into runoff and infiltration. Runoff is proportional to the fractional wetness of the top soil layer (that is the actual moisture content divided by field capacity) and the precipitation rate, such that there is no runoff when the top layer of the soil is dry, and runoff increases linearly as a function of soil wetness up to a maximum of half the precipitation rate, except that all the precipitation runs off when the top layer is saturated. The soil moisture content of the top layer is reduced by evaporation (a function of soil moisture content and potential evaporation, where potential evaporation is based on a drag law parametrization) and drainage into the lower soil layer. The rate of downward drainage is proportional to the field capacity of the top layer and the difference in soil wetness between the top and bottom layers. Downward drainage is effectively suppressed during the "growing season" by the assumption that upward diffusion is infinitely fast. The growing season is defined as the whole year between 30^{o}N and 30^o S, May-August north of 30^o N, November-February south of 30^o S; there is assumed to be no growing season in desert regions.

The control experiment (Hansen et al., 1984) was run from January 1st for 35 years, with a CO_2 level of 315 ppmv. The results are averaged over the last 10 years of the simulation.

The 2xCO_2 experiment (Hansen et al., 1984; Rind, 1988a, 1988b) was also run from January 1st for 35 years, with an instantaneous doubling of CO_2 to 630 ppmv. The results are averaged over the last 10 years of the simulation. Since the ocean heat transport is constrained to be the same as in the control run, there is no ocean feedback effect in this simulation.

The control simulation produces ca 15% less sea-ice than observed. In order to assess the role of sea ice in climate sensitivity to CO_2, a second set of experiments (Hansen et al., 1984; Rind, 1988) was run with a modified version of the model in which the constraint that the ocean mixed layer cannot exceed 0^o C until all the sea ice in a grid cell has melted is removed. The control run with this version gives 23% greater sea ice than observed.

The length of the alternate control run and alternate 2xCO_2 run are not specified. Results are averaged over the last 5 years of these runs (Rind, 1988a).

Rind (1987, 1988a, 1988b) used the fine-grid version of GISS model II to examine the impact of changes in the SST gradient of the climate response to raised CO_2. This version of the model has a vertical resolution of 9 layers, a horizontal resolution of 4^o latitude by 5^o longitude, and prescribed ocean behaviour. Other features of the model are as described in Hansen et al. (1983, 1984).

The control simulation (Rind, 1987) was initialised from a previous simulation of several years length and run for 3.5 years, with 315 ppmv CO_2 and prescribed SSTs. The prescribed SSTs were the average SSTs of the last 10 years of the coarse-grid control simulation (Hansen et al., 1984). The results are averaged over the last 3 years of the simulation.

The 2CO2 experiment was initialised in the same way as the control simulation and run for 3.5 years, with 630 ppmv CO_2 and prescribed SSTs. The prescribed SSTs were the average SSTs from the last 10 years of the coarse-grid $2xCO_2$ simulation (Hansen et al., 1984).

A second, and otherwise similar, raised CO_2 experiment (ALT $2xCO_2$) was made in which the prescribed SSTs were selected so as to amplify the high latitude changes and reduce the low latitude changes. This experiment resulted in about 5% less sea ice than 2CO2. The experiment was run for 3 years.

Hansen et al. (1987, 1988) describe a series of experiments designed to investigate the transient climate response to rising CO_2 levels. The experiments were made with a slightly modified version of Model IIa (Hansen et al., 1983), which differs from the previous version only in that the maximum mixed layer depth is prescribed from seasonal observations rather than being fixed at 65m.

Hansen et al. (1987) made two transient runs using different radiative forcing scenarios (Case A and Case B), thought to bracket the actual rate of change in greenhouse forcing. Both runs use documented (in the case of CO_2) or estimated (in the case of CCl_2F_2, CCl_3F, CH_4, and N_2O) changes in trace gases between 1958 and 1984 and also include documented changes in stratospheric aerosols for this period (in particular those arising from the eruptions of Mt. Agung, 1963-1965, and El Chichon, 1982-1985). Case A is based on the assumption that the growth rates of trace gases will continue to increase, but that the next few decades will be free of volcanic eruptions creating large stratospheric effects. Details of the rates of change of individual trace gases used in this run are given in Table 3. Case B makes similar assumptions to Case A, except that the increases in chlorofluorocarbons are multiplied by 1.15, and CH_4 and N_2O are also increased. The net greenhouse forcing is about 25% more than in Case A. The simulations are run from 1958 to 2001 (44 years).

Hansen et al. (1988) describe a separate set of transient experiments. The control simulation (Hansen et al., 1988) was run for 100 years, with atmospheric composition fixed at estimated 1958 values (315 ppmv for CO_2, 1400ppbv for CH_4, 292.6 ppbv for N_2O, 15.8 pptv for CCl_3F and 50.3 pptv for CCl_2F_2).

Three transient runs (Hansen et al., 1988) were made using different radiative forcing scenarios (A, B and C). All the runs use documented (in the case of CO_2) or estimated (in the case of CCl_2F_2, CCl_3F, CH_4, and N_2O) changes in trace gases between 1958 and 1984 and also include documented changes in stratospheric aerosols for this period (in particular those arising from the eruptions of Mt. Agung, 1963-1965, and El Chichon, 1982-1985). Scenario A is based on the assumption that the growth rates of trace gases will continue to increase exponentially, but that the next few decades will be free of volcanic eruptions creating large stratospheric effects. The potential effects of O_3, stratospheric H_2O and minor chlorine and fluorine compounds are approximated in this scenario by doubling the estimates for CCl_3F and CCl_2F_2. Scenario B (Hansen et al., 1988: Rind et al., 1989) is based on the assumption that the growth rates of trace gases will decrease, such that the increase in greenhouse forcing will be linear, but that the mean stratospheric aerosol optical depth will be comparable to that in the volcanically active period 1958-1985. Scenario C is based on the assumption that the growth of trace gases is reduced between 1990-2000, such that greenhouse forcing ceases to increase after 2000, but that the mean stratospheric aerosol optical depth will be comparable to that in the volcanically active period 1958-1985. Neither Scenario B nor C make any attempt to include the potential effects of O_3, stratospheric H_2O, and minor chlorine and fluorine compounds. Details of the trace gas and stratospheric aerosol changes used in each simulation are given in Table 4. All three simulations are run from 1958 to 2060 (102 years). Scenario A reaches a climate forcing equivalent to doubled CO_2 by 2030, Scenario B by 2060, but Scenario C never reaches that level.

GFDL model

The first set of raised CO_2 experiments (Manabe and Wetherald, 1975) were made with a grid-point model of limited computational domain with idealized geography, as described by Manabe (1969). The model (MW75) has 9 vertical layers and a quasi-uniform grid with a resolution of ca. 500 km. The model is driven by annual average solar focing, and diurnal variations are eliminated by using the effective mean zenith angle. Cloud behaviour is prescribed from climatological data, and is a function of latitude and height only. SSTs of the swamp ocean are calculated using a simple energy-balance swamp ocean. Land and sea albedos are a function of latitude (Manabe, 1969). Snow and sea ice albedos are fixed: when the surface temperature is below -25^{o} C both have an albedo of 0.7; when the surface temperature is above -25^{o} C, snow has an albedo of 0.45 and sea ice an albedo of 0.35. Surface hydrology is represented by a simple bucket model where the moisture content of the soil is increased by rainfall and snowmelt, and decreased by evaporation. The maximum amount of soil moisture storage is 15cm, and "runoff" occurs when the soil is saturated. The ratio of evaporation to potential evaporation is a linear function of soil moisture, such that evaporation occurs at the potential rate when soil moisture is equal or greater than 11.25cm.

In the initial raised CO_2 experiments, the computational domain is a sector with meridional boundaries 120^o apart and extending from 81.7^o N to the equator (Manabe and Wetherald, 1975). Cyclic continuity is assumed at the two meridional boundaries, but "free slip insulated walls" are placed at the equator and 81.7^o N. The sector is divided into two equal parts between the equator and 66.5^o N: one half being land and the other ocean. The whole sector between 66.5 and 81.7^o N is land.

The control simulation is started from an isothermal, dry, motionless atmosphere, and run for 800 days. A second control run, initialised from day 40 of the first control run, was made. The two runs converged towards a similar equilibrium. The control results are an average of the last 100 days of both of these runs. The same method was used to obtain results for the doubled CO_2 experiment ($2xCO_2$).

Wetherald and Manabe (1979) and Manabe and Wetherald (1980) used a modified version of the MW75 model in a second set of raised CO_2 experiments. In this version, the computational domain is extended to the pole. This is made possible by the use of a regular grid with a resolution of 4.5^o latitude by 5^o longitude (Manabe et al., 1975). Cyclical continuity is assumed at the meridional boundaries of the sector, and a free-slip insulated wall is placed at the equator. The geography is simplified, such that the sector is divided equally into land and ocean from the equator to the pole. Cloud behaviour is predicted, using a simple scheme where cloud is formed whenever condensation of water vapour is predicted. The temperature at which the values of snow and sea ice albedos are modified is changed from -25^o C (in Manabe and Wetherald, 1975) to -10^o C. Other characteristics of the model as as described in Manabe (1969) and Manabe and Wetherald (1975).

The control run (Manabe and Wetherald, 1980) is started from an isothermal, dry, motionless atmosphere and integrated for 1200 days, with modern CO_2. The results are averaged over the last 500 days of the simulation.

The $2xCO_2$ and $4xCO_2$ experiments were initialised from the end of the control simulation and run for 1200 days. The results are averaged over the last 500 days of the simulations.

Some of the results from these experiments are discussed in Wetherald and Manabe (1986).

Manabe and Stouffer (1979, 1980) describe raised CO_2 experiments made with the GFDL spectral general circulation model with a global computational domain and so-called realistic geography. The model (G15) has 9 vertical layers and a horizontal resolution determined by spectral truncation at wave number 15 (ca 4.8^o latitude by 8^o longitude; Gordon and Stern, 1982). The model is driven by the seasonal variations in solar radiation; for simplicity the diurnal variation in insolation is ignored. Cloud behaviour is prescribed from climatological data, and is zonally uniform and invarient through time. Ocean behaviour is modelled as a simple mixed layer ocean with a uniform thickness of 68m. The temperature of the mixed layer beneath sea ice is constrained to be equal to -2^o C. The model does not attempt to simulate the 2-dimensional dynamic behaviour of the ocean. Albedo varies as a function of latitude over the ocean and geographically over the land based upon Posey and Clapp (1964). The albedo of sea ice and continental snow varies between 0.6 and 0.7 as a function of latitude, with reduced values for thin or melting sea ice and thin snow. Surface hydrology is represented by a simple bucket model where the moisture content of the soil is increased by rainfall and snowmelt, and decreased by evaporation. The maximum amount of soil moisture storage is 15cm, and "runoff" occurs when the soil is saturated. The ratio of evaporation to potential evaporation is a linear function of soil moisture, such that evaporation occurs at the potential rate when soil moisture is equal or greater than 11.25cm.

The control experiment (Manabe and Stouffer, 1979) was run for 12 years, starting from an isothermal, dry, motionless atmosphere, with 300 ppmv CO_2. To save on computation time the atmospheric part of the model is run with an accelerated seasonal cycle for the first 7 years, while the final 5 years are run with a full cycle of 365 days (Manabe and Stouffer, 1980). The results are averaged over the last 3 years of the simulation.

The quadrupled CO_2 experiment (4xCO_2) was run for 14 years, with 1200 ppmv CO_2. To save on computation time the atmospheric part of the model is run with an accelerated seasonal cycle for the first 8 years, while the final 6 years are run with a full cycle of 365 days (Manabe and Stouffer, 1980). The results are averaged over the last 3 years of the simulation.

Wetherald and Manabe (1981) used a spectral model similar to that described by Manabe and Stouffer (1979, 1980) but with a limited computational domain and idealized geography in order to investigate the influence of seasonal variations on model sensitivity to raised CO_2. The computational domain is a sector with meridional boundaries 120^o apart extending from pole to pole. The geography is simplified, such that the sector is bisected into land and ocean. It is assumed that the earth's surface is flat. Cyclical continuity is assumed at the meridional boundaries.

The model (S15a) has 9 vertical layers, a horizontal resolution determined by spectral truncation at wave number 15 (ca. 4.8^{o} latitude by 8^{o} longitude; Gordon and Stern, 1982), and is driven by seasonal variations in solar radiation. It differs from the model described by Manabe and Stouffer (1979, 1980) only with respect to the computational domain and the treatment of the surface albedo of ice and snow. In Wetherald and Manabe (1981), the albedo of sea ice is assumed to be 0.7 when the surface temperature is below -10^{o} C and 0.35 when it is above -10^{o} C. Similarly, the snow albedo is assumed to be 0.70 if the surface temperature is below -10^{o} C, and 0.45 when it is above -10^{o} C.

The control simulation was run for 19 years, with 300 ppmv CO_2. The atmospheric part of the model is run with an accelerated seasonal cycle for the first 8 years, while the final 11 years are run with a full cycle of 365 days. The results are averaged over the last 4 years of the simulation.

The quadrupled CO_2 experiment ($4xCO_2$) was run for 20 years, with 1200 ppmv CO_2. In order to save computation time the atmospheric part of the model is run with an accelerated seasonal cycle for the first 9 years. The results are averaged over the last 4 years of the simulation.

A second version of the model (S15b) was constructed which was driven by annual mean insolation. The control simulation was run with 300 ppmv CO_2 and the quadrupled CO_2 simulation ($4xCO_2$) with 1200 ppmv CO_2. In order to save computation time, a 400-day integration of the atmospheric part of this model was synchronised with a 26 year integration of the mixed layer ocean model, and then both models were run together for a further 400 days. The results are averaged over the last 200 day period of each integration.

Manabe et al. (1981) analyse and compare the results of raised CO_2 simulations made with three versions of the GFDL model (S15a, G15 and G21). All three models are run with 300 ppmv CO_2 (control simulation) and 1200 ppmv CO_2 ($4xCO_2$). The S15 model was originally used and described by Wetherald and Manabe (1981). According to Manabe et al. (1981) the time integration of this model was performed over 20 years, and results are averaged over the last 4 years of each simulation. In the original reference the length of the integration is given as 19 years for the control and 20 years for the $4xCO_2$ experiment. The G15 model was originally used and described by Manabe and Stouffer (1979, 1980). According to Manabe et al. (1981) the time integration of this model was performed over 13 years, and the results are averaged over the last 3 years of each simulation. In the original references the length of the time integration is given as 12 years for the control and 14 years for the $4xCO_2$ experiment.

The G21 model is identical to the G15 model, except that it has higher horizontal resolution determined by spectral truncation at wave number 21 (ca 3.4^{o} latitude by 5.7^{o} longitude; Gordon and Stern, 1982). The runs with this model were initialised from the final equilibrium state of the G15 model and run for a further 5 years (i.e. 13 + 5 years). The results are averaged over the last 3 years of each simulation.

The sensitivity of the response to raised CO_2 to the model's parametrization of cloud behaviour was investigated by Manabe and Wetherald (1986, 1987) and Wetherald and Manabe (1986, 1988), using two different cloud schemes: one with prescribed (FC) and one with predicted (VC) clouds.

The FC version of the model is somewhat similar to G15, with a vertical resolution of 9 layers, a horizontal resolution determined by truncation at wave number 15 (ca 4.8^{o} latitude by 8^{o} longitude), incorporating a seasonal cycle of insolation, and coupled to a slab ocean. The oceanic component of the model differs from G15, however, because the depth of the mixed layer is less (50m). The treatment of snow and ice albedos is also modified, with the albedos fixed at different values above and below -10^{o} C. The albedo of the ocean varies with latitude and that of the land geographically, as in G15. Although cloud behaviour is prescribed in both G15 and FC, the prescribed distribution differs between the two models. Cloud cover differs between the northern and southern hemispheres in the FC model but is essentially symmetric in both hemispheres in the earlier G15 version of the model. This difference in cloud behaviour is responsable for the more realistic simulation of surface temperature in the southern hemisphere with the FC version of the model.

The VC version of the model differs from the FC version only with respect to cloud behaviour. VC includes a simple scheme which predicts cloud formation whenever the relative humidity exceeds 99%.

Three simulations were made using the FC model, with 300 (1X-FC or control), 600 (2X-FC) and 1200 (4X-FC) ppmv CO_2, and a further two runs were made with the VC model, with 300 (1X-VC) and 600 (2X-VC) ppmv CO2. Each simulation was started with an isothermal, dry, motionless atmosphere and an isothermal mixed layer ocean, and run for approximately 40 years. (The exact length of each simulation varied slightly depending on how long it took to reach equilibrium.) To reduce computing time, the atmosphere and ocean were coupled non-synchronously for the first 10 years of each simulation, as described by Manabe and Stouffer (1980). Results are averaged over the last 10 years of each simulation.

Bryan et al. (1982) describe preliminary results of raised CO_2 experiments with a coupled ocean-atmosphere model (COAM). The model has idealized geography with three identical continents with meridional boundaries 60o apart, separated by 3 oceans of the same width, and with a condition of mirror symmetry imposed at the equator. The land surface is flat. The atmospheric part of the model has been documented by Gordon and Stern (1982). It has 9 vertical layers, a horizontal resolution determined by spectral truncation at wave number 15 (ca. 4.8o latitude by 8o longitude; Gordon and Stern, 1982), and is driven by annual mean solar radiation. Cloudiness is prescribed from climatological data, and is zonally uniform. Albedo varies as a function of latitude over the ocean and the land. The albedos of sea ice and continental snow are apparently assigned (Manabe, 1983) according to the scheme used by Manabe and Stouffer (1979, 1980). Surface hydrology is represented by a simple bucket model where the moisture content of the soil is increased by rainfall and snowmelt, and decreased by evaporation. The maximum amount of soil moisture storage is 15cm, and "runoff" occurs when the soil is saturated. The ratio of evaporation to potential evaporation is a linear function of soil moisture, such that evaporation occurs at the potential rate when soil moisture is equal or greater than 11.25cm. The ocean model (Bryan et al., 1975) has 12 vertical levels, a horizontal resolution of approximately 4.5o latitude and 3.8o longitude, and includes a simple sea-ice model.

The atmosphere, upper ocean and deep ocean are coupled non-synchronously, to allow for the very different time scales on which they operate, using a method developed by Manabe and Bryan (1969), Bryan and Lewis (1979) and Manabe et al. (1979a). One year in the atmospheric model was taken to correspond to 110 years in the upper ocean model; one year in the upper ocean model was taken to correspond to 25 years in the deep ocean. Convergence to a climatic equilibrium was reached after an integration equivalent to 6 years in the atmosphere, 650 years in the upper ocean, and 16,000 years in the deep ocean. The equilibrium climate is then perturbed by a step-function quadrupling of CO_2 and the atmospheric and oceanic components of the model are run synchronously for a further 25 years.

Results from these experiments are also analysed by Manabe (1983), Spelman and Manabe (1984) and Bryan and Spelman (1985).

Manabe and Bryan (1985) describe a series of experiments made with the coupled atmosphere-ocean model used by Bryan et al. (1982) with six different concentrations of CO_2, namely 150 ppmv (X/2), 212 ppmv (X/sqrt(2)), 300 ppmv (1X or control), 600 ppmv (2X), 1200 ppmv (4X) and 2400 ppmv (8X). To speed up the convergence to equilibrium the ocean and atmosphere are coupled non-synchronously couple, such that the ocean is integrated over a period of 110 years while the atmosphere is integrated over 1 year. The convergence towards equilibrium in the deeper ocean is further accelerated by a device, described by Bryan (1984), which is equivalent to reducing the heat capacity of the water. For the control (1X) experiment, convergence occurs after integrating the atmospheric part of the model for 7.7 years and the ocean for 850 years. For the raised CO2 experiments, the corresponding times are 11.8 years and 1290 years respectively. The results from each simulation are averaged over the final 600 days of the atmospheric part of the model and the last 300 years of the oceanic part of the model.

The results of these experiments are also discussed by Bryan and Manabe (1988), and the results of the quadrupled CO_2 simulation (4X, 1200 ppmv CO_2) by Bryan and Manabe (1985).

Bryan et al. (1988) describe a series of runs using the coupled ocean-atmosphere model of Bryan et al. (1982), with a different idealised geometry ($COAM_b$). The model has 3 identical continents extending from near the north pole to about 45o S, and a tri-symmetric southern polar continent. The ratio of ocean to land at each latitude corresponds to the Earth's present geography. The atmospheric and ocean components of the model were asynchronously coupled to reduce computation time.

In the control ("normal" CO_2) run, the atmosphere was integrated over the equivalent of 8.2 years, the upper ocean over the equivalent of 1250 years, and the deep ocean over 34,000 years. A doubled CO2 (2xCO2) experiment was run in the same way.

The transient response to changes in CO_2 concentration was investigated by coupling the atmospheric and oceanic components of the model with no distortion of the thermal time scales. The control run was integrated for 110 years. Three doubled CO_2 runs (A, B and C) were made. Each run was initialised from a different stage of the control simulation and run for 50 years. Ensemble averages of these 3 runs were used to provide a measure of the transient response to doubled CO2.

OSU model

The OSU model is a grid-point model that evolved from the Mintz-Arakawa GCM (Arakawa et al., 1969; Gates et al., 1971; Gates, 1973, 1975a, 1975b). Simulations of January and July climates were made with a version of this model, the RAND two-level atmospheric model (Gates and Schlesinger, 1977). These model experiments formed the basis for improvements to the parametrization of various boundary layer and radiative processes in the OSU model. Schlesinger and Gates (1979, 1980) describe simulations of January and July climates made with the OSU model. A simulation driven by seasonal changes in solar forcing is described by Schlesinger and Gates (1981).

Gates et al. (1981) describe raised CO_2 experiments made with the OSU model, as described in Schlesinger and Gates (1979, 1980, 1981). The model has 2 vertical layers and a horizontal grid with a resolution of 4^{o} of latitude by 5^{o} of longitude. The model takes into account the seasonal variations in solar radiation. Cloud behaviour is predicted using a relative humidity scheme such that large-scale clouds form when the relative humidity exceeds 90%. SSTs and sea ice extent are prescribed, based on monthly climatological data from Alexander and Mobley (1976). The land surface is classified into 9 surface types (woodland, grassland and cultivated areas; forest; steppe and grassland; steppe desert; desert; tundra, mountains and artic areas; water; land ice; sea-ice), based on data from Posey and Clapp (1964). The albedo of the oceans is fixed; other surface types have an assigned snow-free albedo; this albedo is progressively modified as a function of tha amount of snow, once a critical mass of snow has covered the surface, to a maximum value for snow-covered surfaces (Schlesinger and Gates, 1981). Surface hydrology appears to be treated by means of a simple bucket model.

The control simulation (Schlesinger and Gates, 1981; Gates et al., 1981) was run for 39 months, with 322 ppmv CO2 and modern SSTs and sea ice distribution. The initial conditions for the doubled (2xCO2) and quadrupled (4xCO2) CO2 experiments were taken as the state on November 1st of the second year of the control integration. Both raised CO2 experiments (Gates et al., 1981) were run for 9 months, with 644 ppmv (2xCO2) and 1288 ppmv (4xCO2) respectively, and with modern SSTs and sea ice distribution. These experiments, then, estimate the direct radiation effects of increased CO2 in the absence of ocean feedbacks.

Schlesinger and Zhao (1989) describe raised CO_2 experiments made with the OSU atmospheric GCM, coupled to a simple slab ocean. The atmospheric component of the model is a modified version of that described by Schlesinger and Gates (1979, 1980, 1981) and documented by Ghan et al. (1982), with 2 vertical layers and a horizontal grid of 4^{o} of latitude by 5^{o} of longitude. Cloud behaviour is predicted using a modified relative humidity scheme, such that large-scale clouds form in the lower vertical layer when the relative humidity exceeds 85% and in the upper vertical layer when relative humidity exceeds 95%. The land surface is classified into 9 surface types, based on data from Posey and Clapp (1964), each of which has a characteristic snow-free albedo. The albedo of land and sea ice is fixed at 0.60, a higher value than that used in previous simulations (0.45), while the albedo of snow covered surfaces is reduced by 0.1 compared to earlier simulations (e.g. Gates et al., 1981). Surface hydrology appears to be treated by means of a simple bucket model. The oceanic component of the model (Pollard et al., 1983: model version 1) is a 60-m deep mixed-layer ocean with a simple thermodynamic sea ice model. Oceanic heat flux into the base of the sea ice is prescribed as 17 W m^{-2} in the northern hemisphere and 6 W m^{-2} in the southern hemisphere.

The control simulation (1 x CO_2) was run with a CO_2 concentration of 326 ppmv. The doubled CO_2 simulation (2 x CO_2) was run with a CO_2 concentration of 652 ppmv. Both simulations were started from the same initial conditions. The state of the atmosphere was taken as those on November 1st of year 1 of a 10-year integration that was itself initialised from an earlier model simulation. The initial ocean mixed-layer temperatures and sea-ice thicknesses were prescribed from the climatological observations of Alexander and Mobley (1976). In order to minimise computational demands, the model was run with an accelerated seasonal cycle (30.4 days) and a reduced mixed-layer depth (5m) for the first 45 years. Equilibrium was reached by the end of ca 35 years; the additional period of accelerated integration was necessary to allow for re-equilibration after correction of a coding error. The control simulation was then run without acceleration for a further 24 years, while the 2 x CO_2 experiment was run without acceleration for a further 16 years. Results are averaged over the last 10 years of each simulation.d

The effects of ocean feedbacks on the climatic response to raised CO_2 have been investigated using the OSU atmospheric model synchronously coupled to an oceanic GCM (CGCM: Schlesinger et al., 1985). The atmospheric component of the coupled model is a modified version of the model as described by Schlesinger and Gates (1979, 1980, 1981), incorporating both seasonal and diurnal variations in solar radiation, with interactive clouds (Ghan et al., 1982). The oceanic component of the coupled model is a modified version of the OGCM described by Han (1984a, 1984b, 1988) but including an Arctic Ocean. The OGCM has 6 vertical layers, a horizontal resolution of 4^o latitude by 5^o longitude, and realistic bottom topography. The model predicts currents water temperatures, and salinity. However, surface salinity is constrained to be the same as the observed salinity field, according to Levitus (1982). Sea-ice formation and extent are calculated using a simple thermodynamic model, following Semtner (1976) and Parkinson and Washington (1979). The performance of the coupled model is analysed by Gates et al. (1985) and Han et al. (1985).

The control experiment (1 x CO_2) was run for 16 years with CO_2 concentration set to 326 ppmv (Gates et al., 1985; Schlesinger et al., 1985). The raised CO_2 experiment was run for 16 years with CO_2 concentration set to 652 ppmv (Schlesinger et al., 1985). Each simulation was begun from the same initial conditions.

The state of the atmosphere was taken as that on November 1st of year 1 of a 10-year atmospheric GCM integration that was itself initialised from an earlier model simulation. The state of the ocean was taken as that on November 1st of year 9 of an 11-yr oceanic simulation with prescribed monthly atmospheric forcing, itself initialised from an early 40-yr simulation with annually-averaged forcing. The coupled model was run synchronously (from 1st November of year 0 through to 31st October of year 16) for 16 years. The control run had not reached equilibrium by the end of the 16 years (Gates et al., 1985).

Longer runs made with the same coupled atmosphere-ocean model are described by Schlesinger (1986), Schlesinger and Jiang (1988), and Schlesinger (1989). The CO_2 concentration was set to 326 ppmv in the control run (1 x CO_2) and 652 ppmv in the doubled CO_2 experiment (2 x CO_2). Both simulations were integrated for 20 years. The results are examined over the whole 20 years of the simulations.

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Table 1: Comparison of model characteristics

Type	Version	Vertical resolution	Horizontal resolution	Seasonal cycle ?	Diurnal cycle ?	Cloud behaviour	Ocean behaviour	Hydrology	Albedo
UKMO Model gridpoint	5LM	5	330km grid	Yes	Yes	prescribed	swamp ocean, prescribed SSTs and sea ice	simple soil bucket, run-off when saturated	varies by latitude on land, and with depth of snow
	$11LM_a$	11	$2.5^o x 3.75^o$	Yes	Yes	prescribed	swamp ocean, prescribed SSTs and sea ice	simple soil bucket, run-off when saturated	fixed values
	$11LM_b$	11	5^o x 7.5^o	Yes	Yes	predicted, RH scheme	slab ocean, prescribed heat convergence	simple soil bucket, run-off when saturated	varies with soil and veg type, and with snow depth
	$11LM_c$	11	5^o x 7.5^o	Yes	Yes	predicted, RH scheme	slab ocean, prescribed heat convergence	drainage below root zone, run-off when saturated or infiltration rate exceeded	varies with soil and veg type, and with snow depth
	$11LM_d$	11	5^o x 7.5^o	Yes	Yes	predicted, CW scheme	slab ocean, prescribed heat convergence	drainage below root zone, run-off when saturated or infiltration rate exceeded	varies with soil and veg type, and with snow depth
	$11LM_e$	11	5^o x 7.5^o	Yes	Yes	predicted CWH scheme	slab ocean, prescribed heat	drainage below root zone, run-	varies with soil and veg

							convergence	off when saturated or infiltration rate exceeded	type, and with snow depth
	$11LM_f$	11	5^o x 7.5^o	Yes	Yes	predicted CWRP scheme	slab ocean, prescribed heat convergence	drainage below root zone, run-off when saturated or infiltration rate exceeded	varies with soil and veg type, and with snow depth
NCAR CCM spectral	WM83a	9	4.4^ox7.5^o	No	No	prescribed	swamp ocean, energy-balance SSTs and sea ice	simple soil bucket, runoff when saturated	fixed values
	WM83b	9	4.4^ox7.5^o	No	No	predicted (RH-type scheme)	swamp ocean, energy-balance SSTs and sea ice	simple soil bucket, runoff when saturated	fixed values
	$WM83_{SSIA}$	9	4.4^ox7.5^o	No	No	predicted (RH-type scheme)	swamp ocean, energy-balance SSTs and sea ice	simple soil bucket, runoff when saturated	sea-ice and snow albedo dependent on temp. (SSIA scheme)
	$WM83_{DSC}$	9	4.4^ox7.5^o	No	No	predicted (RH-type scheme)	swamp ocean, energy-balance SSTs and sea ice	simple soil bucket, runoff when saturated	sea-ice and snow albedo dependent on temp. (SSIA scheme)
	WM84	9	4.5^ox7.5^o	Yes	No	predicted (RH-type scheme)	slab ocean, energy-balance SSTs and sea ice	simple soil bucket, runoff when saturated	all fixed except sea is a function of solar

									zenith angle
	M88a	9	$4.5^{o}x7.5^{o}$	Yes	No	predicted (RH-type scheme)	swamp ocean, prescribed SSTs and sea ice	simple soil bucket, runoff when saturated	all fixed except sea is a function of solar zenith angle
	M88b	9	$4.5^{o}x7.5^{o}$	Yes	No	predicted (RH-type scheme)	slab ocean, energy-balance SSTs and sea ice	simple soil bucket, runoff when saturated	all fixed except sea is a function of solar zenith angle
	WM89	9	$4.5^{o}x7.5^{o}$	Yes	No	predicted (RH-type scheme)	OGCM, predicted SSTs and sea ice	simple soil bucket, runoff when saturated	all fixed except sea is a function of solar zenith angle
GISS Model gridpoint	**II** IIa	9	$8^{o}x10^{o}$	Yes	Yes	predicted (RH-type scheme)	slab ocean, 65m mixed layer; heat divergence prescribed; temperature beneath ice not > 0^{o} C;	2-layer soil bucket, runoff when saturated or infiltration rate exceeded; no drainage during growing season	land varies with veg type, and snow with depth,age veg cover
	IIb	9	$8^{o}x10^{o}$	Yes	Yes	predicted (RH-type scheme)	slab ocean, 65m mixed layer; heat divergence prescribed; temperature	2-layer soil bucket, runoff when saturated or infiltration rate exceeded;	land varies with veg type, and snow with depth,age

							beneath ice unconstrained	no drainage during growing season	veg cover
	IIc	9	$4^{o}x5^{o}$	Yes	Yes	predicted (RH-type scheme)	swamp ocean, prescribed SSTs and sea ice	2-layer soil bucket, runoff when saturated or infiltration rate exceeded; no drainage during growing season	land varies with veg type, and snow with depth, age veg cover
	IId	9	$8^{o}x10^{o}$	Yes	Yes	predicted (RH-type scheme)	slab ocean, mixed layer depth & heat divergence prescribed; temperature beneath ice not > 0^{o} C;	2-layer soil bucket, runoff when saturated or infiltration rate exceeded; no drainage during growing season	land varies with veg type, and snow with depth, age veg cover
GFDL gridpoint	MW75	9	500 km	No	No	prescribed function of height & latitude	swamp ocean, energy-balance SSTs and sea ice	simple soil bucket, runoff when saturated	land and sea vary with latitude; snow and ice fixed but difft above and below -25^{o} C
	MW80	9	4.5^{o} x 5^{o}	No	No	predicted when condensat^{n} occurs	swamp ocean, energy-balance SSTs and sea ice	simple soil bucket, runoff when saturated	land and sea vary with latitude; snow and ice fixed but difft above and

									below -10^{o} C
spectral	G15	9	wave no 15 ($4.8^{o} \times 8^{o}$)	Yes	No	prescribed as zonally uniform	slab ocean, 68m fixed mixed layer, -2^{o} C below sea ice	simple soil bucket, runoff when saturated	sea, snow & sea ice vary with latitude, snow/ice also temp dependent land geographically
	$S15_a$	9	wave no 15 ($4.8^{o} \times 8^{o}$)	Yes	No	prescribed as zonally uniform	slab ocean, 68m fixed mixed layer, -2^{o} C below sea ice	simple soil bucket, runoff when saturated	sea with latitude, snow and sea ice fixed but different above and below -10^{o} C, land geographically
	$S15_b$	9	wave no 15 ($4.8^{o} \times 8^{o}$)	No	No	prescribed as zonally uniform	slab ocean, 68m fixed mixed layer, -2^{o} C below sea ice	simple soil bucket, runoff when saturated	sea with latitude, snow and sea ice fixed but different above and below -10^{o} C, land geographically
	G21	9	wave no 21 ($3.4^{o} \times 5.7^{o}$)	Yes	No	prescribed as zonally uniform	slab ocean, 68m fixed mixed layer, -2^{o} C below	simple soil bucket, runoff when saturated	sea, snow & sea ice vary with latitude,

						sea ice		snow/ice also temp dependent land varies geographically
FC	9	wave no 15 (4.8^{o} x 8^{o})	Yes	No	prescribed using more realistic scheme	slab ocean, 50m fixed mixed layer, -2^{o} C below sea ice	simple soil bucket, runoff when saturated	sea with latitude, snow and ice fixed but with different values above and below -10^{o} C, land geographically
VC	9	wave no 15 (4.8^{o} x 8^{o})	Yes	No	predicted, (RH-type scheme)	slab ocean, 50m fixed mixed layer, -2^{o} C below sea ice	simple soil bucket, runoff when saturated	sea with latitude, snow and ice fixed but with different values above and below -10^{o}C, land geographically
COAM	9	wave no 15 (4.8^{o} x 8^{o})	No	No	prescribed as zonally uniform	OGCM, predicted SSTs and sea ice	simple soil bucket, runoff when saturated	sea, land snow & sea ice vary with latitude, snow/ice also temp

									dependent
	$COAM_b$	9	wave no 15 ($4.8^o \times 8^o$)	No	No	prescribed as zonally uniform	OGCM, predicted SSTs and sea ice	simple soil bucket, runoff when saturated	sea, land snow and sea ice vary with latitude, snow/ice also temp dependent
OSU gridpoint	2LM	2	$4^o \times 5^o$	Yes	?	predicted, RH scheme 1	swamp ocean, prescribed SSTs and sea ice	simple soil bucket	vary with veg type and snow cover
	$2LM_b$	2	$4^o \times 5^o$	Yes	?	predicted, RH scheme 2	slab ocean, 60m mixed layer	simple soil bucket	vary with veg type and snow cover
	CGCM	2	$4^o \times 5^o$	Yes	Yes	predicted	OGCM, predicted SSTs and sea ice	simple soil bucket	vary with veg type and snow cover
	$CGCM_{ext}$	2	$4^o \times 5^o$	Yes	Yes	predicted	OGCM, predicted SSTs and sea ice	simple soil bucket	vary with veg type and snow cover

Table 2: Comparison of boundary conditions and other characteristics of model runs

Model	Version	Run	CO_2	SST/sea ice	Run length	Results averaged over last	References
UKMO	5LM	control	modern	modern	3.27 yr[1]	3 yr	Mitchell,1983
		$2xCO_2$	2x modern	modern	1.3 yr	1 yr	Mitchell,1983
		2C2S	2x modern	modern + 2K	2.34 yr	2 yr	Mitchell,1983
		$10xCO_2$	10x modern	modern	1.3 yr	1 yr	Mitchell,1983
		C4SL	4x modern	SSTs vary latitudinally	3.11 yr	3 yr	Mitchell & Lupton,1984
		$control^{ext}$	modern	modern	4.01 yr	3 yr	Wilson & Mitchell,1987a
		$C4SL^{ext}$	4x modern	SSTs vary latitudinally	3.61 yr	3 yr	Wilson & Mitchell,1987a
	$11LM_a$	control	modern	modern	8 yr	?	Mitchell et al.,1987
		2C2S	2x modern	modern + 2K	3 yr	2 yr	Mitchell et al.,1987
	$11LM_b$	control	323 ppmv	interactive	20 yr	15 yr	Wilson & Mitchell,1987b
		$2xCO_2$	646 ppmv	interactive	38 yr	15 yr	Wilson & Mitchell,1987b
	$11LM_c$	$control_{cly}$	323 ppmv	interactive	6 yr	5 yr	Mitchell & Warrilow,1987
		$control_{med}$	323 ppmv	interactive	6 yr	5 yr	Mitchell & Warrilow,1987
		$control_{snd}$	323 ppmv	interactive	6 yr	5 yr	Mitchell & Warrilow,1987
		$control_{frz}$	323 ppmv	interactive	6 yr	5 yr	Mitchell & Warrilow,1987
		$2xCO_{2cly}$	646 ppmv	interactive	6 yr	5 yr	Mitchell & Warrilow,1987
		$2xCO_{2med}$	646 ppmv	interactive	6 yr	5 yr	Mitchell & Warrilow,1987
		$2xCO_{2snd}$	646 ppmv	interactive	6 yr	5 yr	Mitchell & Warrilow,1987
		$2xCO_{2frz}$	646 ppmv	interactive	6 yr	5 yr	Mitchell & Warrilow,1987
	$11LM_d$	control	320 ppmv	interactive	?	5 yr	Mitchell et al.,1989
		$2xCO_2$	640 ppmv	interactive	?	5 yr	Mitchell et al.,1989
	$11LM_e$	control	320 ppmv	interactive	?	5 yr	Mitchell et al.,1989
		$2xCO_2$	640 ppmv	interactive	?	5 yr	Mitchell et al.,1989
	$11LM_f$	control	320 ppmv	interactive	?	5 yr	Mitchell et al.,1989
		$2xCO_2$	640 ppmv	interactive	?	5 yr	Mitchell et al.,1989
NCAR CCM	WM83a	control	modern	interactive	1.67 yr[2]	1 yr	Washington & Meehl,1983
		$2xCO_2$	2x modern	interactive	1.67 yr	1 yr	Washington & Meehl,1983
		$4xCO_2$	4x modern	interactive	1.67 yr	1 yr	Washington & Meehl,1983
	WM83b	control	modern	interactive	1.86 yr	1 yr	Washington & Meehl,1983
		$2xCO_2$	2x modern	interactive	1.86 yr	1 yr	Washington & Meehl,1983
		$4xCO_2$	4x modern	interactive	670 days	1 yr	Washington & Meehl,1983

	$WM83_{SSIA}$	control	modern	interactive	2.89 yr	1 yr	Washington & Meehl,1986
		$2xCO_2$	2x modern	interactive	2.89 yr	1 yr	Washington & Meehl,1986
	$WM83_{DSC}$	control	modern	interactive	1.89 yr	1 yr	Washington & Meehl,1986
		$2xCO_2$	2x modern	interactive	1.89 yr	1 yr	Washington & Meehl,1986
	WM84	control	modern	interactive	27 yr	3 yr	Washington & Meehl,1984
		$2xCO_2$	2x modern	interactive	15 yr	3 yr	Washington & Meehl,1984
	M88a	$SPEC^2$ SST	modern	modern	5 yr	3 yr	Meehl,1988
	M88b	MIX1	modern	interactive	12 yr[3]	3 yr	Meehl,1988
		MIX2	2x modern	interactive	12 yr[3]	3 yr	Meehl,1988
	WM89	control	330 ppmv	predicted	30 yr	5 yr	Washington & Meehl,1989
		$2xCO_2$	660 ppmv	predicted	30 yr	5 yr	Washington & Meehl,1989
		transient	increases by 1% per year from 330 ppmv to 429 ppmv	predicted	30 yr	5 yr	Washington & Meehl,1989
GISS	IIa	control	315 ppmv	interactive	35 yr	10 yr	Hansen et al.,1984
		$2xCO_2$	630 ppmv	interactive	35 yr	10 yr	Hansen et al.,1984
	IIb	control	315 ppmv	interactive	?	5 yr	Hansen et al.,1984
		$2xCO_2$	630 ppmv	interactive	?	5 yr	Hansen et al.,1984
	IIc	control	315 ppmv	averages from coarse grid control	3.5 yr	3 yr	Rind,1987,1988
		2CO2	630 ppmv	averages from coarse grid $2xCO_2$ run	3.5 yr	3 yr	Rind,1987,1988
		ALT2CO2	630 ppmv	prescribed, with increased high latitude, decreased low latitude signal	3.5 yr	3 yr	Rind,1987,1988
	IId	control	315 ppmv[4]	interactive	100 yr	100 yr	Hansen et al.,1988
		Case A	exponent'l increase in forcing	interactive	44 yr	44 yr	Hansen et al.,1987
		Case B	exponent'l increase in forcing with large	interactive	44 yr	44 yr	Hansen et al.,1987

			increases in CFCs				
		Scenario A	exponent'l increase in forcing	interactive	102 yr	102 yr	Hansen et al.,1988
		Scenario B	linear increase in forcing	interactive	102 yr	102 yr	Hansen et al.,1988
		Scenario C	drastic reduction in forcing	interactive	102 yr	102 yr	Hansen et al.,1988
GFDL	MW75	control	"normal"	interactive	800 days	100 days	Manabe & Wetherald,1975
		$2xCO_2$	2x"normal"	interactive	800 days	100 days	Manabe & Wetherald,1975
	MW80	control	modern	interactive	1200 days	500 days	Manabe & Wetherald,1980
		$2xCO_2$	2 x modern	interactive	1200 days	500 days	Manabe & Wetherald,1980
		$4xCO_2$	4 x modern	interactive	1200 days	500 days	Manabe & Wetherald,1980
	G15	control	300 ppmv	interactive	12 yr	3 yr	Manabe & Stouffer,1979, Manabe & Stoufer,1980
		$4xCO_2$	1200 ppmv	interactive	14 yr	3 yr	Manabe & Stouffer,1979, Manabe & Stoufer,1980
	$S15_a$	control	300 ppmv	interactive	19 yr	4 yr	Wetherald & Manabe,1981
		$4xCO_2$	1200 ppmv	interactive	20 yr	4 yr	Wetherald & Manabe,1981
	$S15_b$	control	300 ppmv	interactive	400 days/ 26 yr	200 days	Wetherald & Manabe,1981
		$4xCO_2$	1200 ppmv	interactive	400 days/ 26 yr	200 days	Wetherald & Manabe,1981
	G21	control	300 ppmv	interactive	13+5 yr	3 yr	Manabe et al.,1981
		$4xCO_2$	1200 ppmv	interactive	13+5 yr	3 yr	Manabe et al.,1981
	FC	1X-FC	300 ppmv	interactive	ca 40 yr	10 yr	Manabe & Wetherald,1987
		2X-FC	600 ppmv	interactive	ca 40 yr	10 yr	Manabe & Wetherald,1987
		3X-FC	1200 ppmv	interactive	ca 40 yr	10 yr	Manabe & Wetherald,1987
	VC	1X-VC	300 ppmv	interactive	ca 40 yr	10 yr	Manabe & Wetherald,1987
		2x-VC	600 ppmv	interactive	ca 40 yr	10 yr	Manabe & Wetherald,1987
	COAM	control	300 ppmv	interactive	25 yr[5]	?	Bryan et al.,1982
		$4xCO_2$	1200 ppmv	interactive	25 yr[5]	?	Bryan et al.,1982
		X/2	150 ppmv	interactive	?	600 days/ 300 yr	Manabe & Bryan,1985
		X/sqrt(2)	212 ppmv	interactive	?	600 days/ 300 yr	Manabe & Bryan,1985

		1X	300 ppmv	interactive	?	600 days/ 300 yr	Manabe & Bryan,1985
		2X	600 ppmv	interactive	?	600 days/ 300 yr	Manabe & Bryan,1985
		4X	1200 ppmv	interactive	?	600 days/ 300 yr	Manabe & Bryan,1985
		8X	2400 ppmv	interactive	?	600 days/ 300 yr	Manabe & Bryan,1985
	$COAM_b$	control	300 ppmv	interactive	8.2 yr[6]	?	Bryan et al.,1988
		$2xCO_2$	600 ppmv	interactive	8.2 yr[6]	?	Bryan et al.,1988
		control	300 ppmv	interactive	110 yr	110 yr	Bryan et al.,1988
		$2xCO_2$	600 ppmv	interactive	50 yr	50 yr	Bryan et al.,1988
OSU	2LM	control	322 ppmv	modern	39 months	?	Schlesinger & Gates,1979
		$2xCO_2$	644 ppmv	modern	9 months	?	Gates et al.,1981
		$4xCO_2$	1288 ppmv	modern	9 months	?	Gates et al.,1981
	$2LM_b$	$1xCO_2$	326 ppmv	interactive	45+24 yr	10 yr	Schlesinger & Zhao,1989
		$2xCO_2$	652 ppmv	interactive	45+16 yr	10 yr	Schlesinger & Zhao,1989
	CGCM	$1xCO_2$	326 ppmv	interactive	16 yr	16 yr	Schlesinger et al.,1985
		$2xCO_2$	652 ppmv	interactive	16 yr	16 yr	Schlesinger et al.,1985
	$CGCM_{ext}$	$1xCO_2$	326 ppmv	interactive	20 yr	20 yr	Schlesinger & Jiang,1988
		$2xCO_2$	652 ppmv	interactive	20 yr	20 yr	Schlesinger & Jiang,1988

Notes:
1 UKMO year length equals 365 days
2 NCAR year length equals 360 days
3 preceded by 2 phases of accelerated annual cycles
4 simulation includes other radiative forcings (CH_4, N_2O, CCl_3F, CCl_2F_2)
5 length of synchronously coupled run, with non-synchronously coupled run up
6 length of run for atmospheric component, oceans coupled asynchronously

Table 3: Radiative forcing scenario used in Case A of Hansen et al. (1987)

CO_2	1958-1984	observed values
	1980's	delta CO_2 set to 15 ppm
	1990's	delta CO_2 set to 19 ppm
CH_4	1960's	increases by 0.5% a year
	1970's	increases by 1% a year
	1980s	increases by 1.5% a year
	1990's	increases by 1.5% a year
N_2O	1970's	increases by 0.2% a year
	1980's	increases by 0.3% a year
	1990's	increases by 0.3% a year
CCl_2F_2	emission rates are taken as constant at the mean rate for the 1970's with an atmospheric lifetime of 75 years.	
CCl_3F	emission rates are taken as constant at the mean rate for the 1970's with an atmospheric lifetime of 150 years	

stratospheric aerosol opacities, measured changes between 1958 and 1984 (the two substantial events being Mt Agung, 1963-1965, and El Chichon, 1982-1984)

Notes:
There is a discrepancy between the quoted atmospheric lives of CCl_3F and CCl_2F_2 given in this paper and that quoted in Hansen et al. (1988).

Table 4: Radiative forcing scenarios used in Scenarios A, B and C of Hansen et al. (1988)

Scenario A

CO_2	1958-1981, measured values 1981-2060, 1.5% growth of annual increment per year
CH_4	1.4 ppbv in 1958 1959-1970, increases at 0.6% per year 1970's, increases at 1% per year 1980-2060, increases at 1.5% per year
N_2O	increases by 0.1% year in 1958 increases by 0.2% year in 1980 increases by 0.4% year in 2000 increases by 0.9% year in 2030
CCl_3F	reported rates, with 3% increased emission per year in the future, and an atmospheric lifetime of 75 years
CCl_2F_2	reported rates, with 3% increased emission per year in the future, and an atmospheric lifetime of 150 years

Potential effects of other CFCs, O_3, stratospheric H_2O etc. approximated by doubling amounts of CCl_3F and CCl_2F_2.

Stratospheric aerosol opacities: no additional volcanic aerosols are included after those from El Chichon have decayed to background level.

Scenario B

CO_2	1958-1981, measured values 1.5% until 1990, reduced to 1% year in 1990 reduced to 0.5% year in 2000 reduced to 0 in 2010; thus after 2010 the annual increment of CO_2 is 1.9 ppmv per year.
CH_4	1.4 ppbv in 1958 1959-1970, increases at 0.6% per year 1970's, increases at 1% per year 1980's, annual growth rate 1.5% year reduced to 1% year in 1990 reduced to 0.5% in 2000
N_2O	annual growth 3.5% per year today reduced to 2.5% per year in 1990 reduced to 1.5% per year in 2000 reduced to 0.5% per year in 2010.
CCl_3F	3% year increase today reduced to 2% in 1990 reduced to 1% in 2000 reduced to 0 in 2010.
CCl_2F_2	3% year increase today reduced to 2% in 1990 reduced to 1% in 2000 reduced to 0 in 2010.

Other CFCs, O_3, stratospheric H_2O etc. not included.

Stratospheric aerosol opacities: affected by a volcanic event with properties identical to El Chichon in 1995 and a volcanic event with properties identical to Mt. Agung in 2015.

Scenario C

CO_2	1958-1981, measured values growth rate 1.5% per year up to 1985 1985-2000, growth increment 1.5 ppmv per year 2000-2060, constant at 368 ppmv
CH_4	1.4 ppbv in 1958 1959-1970, increases at 0.6% per year 1970's, increases at 1% per year 1980-1990, growth rate 1% per year 1990-2000, growth rate 0.5% per year 2000-2060, abundance constant at 1916 ppbv
N_2O	details not specified in paper
CCl3F	abundances are the same as A and B until 1990, thereafter emissions decrease linearly to zero in 2000
CCl_2F_2	abundances are the same as A and B until 1990, thereafter emissions decrease linearly to zero in 2000

Other CFCs, O_3, stratospheric H_2O etc. not included.

Stratospheric aerosol opacities: affected by a volcanic event with properties identical to El Chichon in 1995 and a volcanic event with properties identical to Mt. Agung in 2015.