3

Processes and Modelling

U. CUBASCH, R.D. CESS

Contributors:
F. Bretherton; H. Cattle; J.T. Houghton; J.F.B. Mitchell; D. Randall; E. Roeckner; J.D. Woods; T. Yamanouchi.

CONTENTS

Executive Summary 73

3.1 Introduction 75

3.2 Climate System 75
- 3.2.1 The Atmosphere 75
- 3.2.2 The Ocean 76
- 3.2.3 The Cryosphere 77
- 3.2.4 The Biosphere 77
- 3.2.5 The Geosphere 77
- 3.2.6 Timescales 77

3.3 Radiative Feedback Mechanisms 77
- 3.3.1 Discussion of Radiative Feedback Mechanisms 77
- 3.3.2 Water Vapour Feedback 78
- 3.3.3 Snow-ice Albedo Feedback 78
- 3.3.4 Cloud Feedback 79

3.4 Predictability of the Climate System 80

3.5 Methods of Predicting Future Climate 80
- 3.5.1 The Palaeo-Analogue Method 83
  - 3.5.1.1 Estimate of temperature sensitivity to CO_2 changes 83
  - 3.5.1.2 Construction of the analogue patterns 83
- 3.5.2 Atmospheric General Circulation Models 84
  - 3.5.2.1 Physical parameterizations 84
  - 3.5.2.2 Radiation and the effect of clouds 84
  - 3.5.2.3 Sub grid-scale transports 85
  - 3.5.2.4 Land surface processes 85
  - 3.5.2.5 Boundary conditions 86
- 3.5.3 Ocean Models 86
- 3.5.4 Carbon Cycle Models 86
- 3.5.5 Chemical Models 87
- 3.5.6 Coupled Models of the Atmosphere and the Ocean 87
- 3.5.7 Use of Models 87
  - 3.5.7.1 Equilibrium response experiments 88
  - 3.5.7.2 Time-dependent response experiments 88

3.6 Illustrative Equilibrium Experiments 88

3.7 Summary 89

References 90

EXECUTIVE SUMMARY

The climate system consists of the five components

- atmosphere
- ocean
- cryosphere (ice)
- biosphere
- geosphere

The fundamental process driving the global climate system is heating by incoming short-wave solar radiation and cooling by long-wave infrared radiation into space The heating is strongest at tropical latitudes, while cooling predominates at the polar latitudes of each winter hemisphere The latitudinal gradient of heating drives the large scale circulations in the atmosphere and in the ocean, thus providing the heat transfer necessary to balance the system

Many facets of the climate system are not well understood, and a significant number of the uncertainties in modelling atmospheric, cryospheric and oceanic interactions are directly due to the representation or knowledge of interactive climate feedback mechanisms Such feedback mechanisms can either amplify or reduce the climate response resulting from a given change of climate forcing

In order to predict changes in the climate system, numerical models have been developed which try to simulate the different feedback mechanisms and the interaction between the different components of the climate system

So far most climate simulations have been carried out with numerical Atmospheric General Circulation Models (AGCMs) which have been developed or derived from weather forecast models For investigations of climate change due to increased greenhouse gas concentrations, they have generally been run coupled with simple representations of the upper ocean and, in some cases, with more detailed, but low resolution, dynamical models of the ocean to its full depth Relatively simple schemes for interactive land surface temperature and soil moisture are also usually included Representations of the other elements of the climate system (land-ice, biosphere) are usually included as non-interactive components The resolution of these models is as yet too coarse to allow more than a limited regional interpretation of the results

Unfortunately, even though this is crucial for climate change prediction, only a few models linking all the main components of the climate system in a comprehensive way have been developed This is mainly due to a lack of computer resources, since a coupled system has to take the different timescales of the sub-systems into account An atmospheric general circulation model on its own can be integrated on currently available computers for several model decades to give estimates of the variability about its equilibrium response when coupled to a global ocean model (which needs millennia to reach an equilibrium) the demands on computer time are increased by several orders of magnitude The inclusion of additional sub-systems and the refinement of resolution needed to make regional predictions demands computer speeds several orders of magnitude faster than is available on current machines

It should be noted that current simulations of climate change obtained by incomplete models may be expected to be superseded as soon as more complete models of the climate system become available

An alternative to numerical model simulations is the palaeo-analogue method (the reconstruction of past climates) Although its usefulness for climate prediction is questioned because of problems involving data coverage and the validity of past climate forcing compared with future scenarios, the method gives valuable information about the possible spectrum of climate change and it provides information for the broader calibration of atmospheric circulation models in different climate regimes

3.1 Introduction

The aim of this section is to provide background understanding of the climate system, to explain some of the technical terms used in climate research (i e , what is a transient and what is an equilibrium response), and to describe how climate change can be predicted In the limited space available to this Section it is impossible to give more than a brief description of the climate system and its prediction The discussion will therefore be limited to the most relevant aspects More detailed description are found in the references and in, for example, the books of Gates (1975) and Houghton (ed) (1984)

A section has been devoted to feedback processes which introduce the non-linearities into the climate system, and which account for many of the difficulties in predicting climate change Climate models and their technical details are discussed where relevant to subsequent Sections of the Report For more detailed information the reader is referred to the book by Washington and Parkinson (1986)

To illustrate some of the difficulties and uncertainties which arise in climate change predictions from numerical models we compare results from two independent numerical simulations at the end of the Section

3.2 Climate System

The climate system (see Figure 3 1) consists of the five components

atmosphere
ocean
cryosphere
biosphere
geosphere

The fundamental processes driving the global climate system are heating by incoming short wave solar radiation and the cooling by long-wave radiation into space The heating is strongest at tropical latitudes, while cooling predominates in the polar regions during the winter of each hemisphere The latitudinal gradient of heating drives the atmosphere and ocean circulations, these provide the heat transfer necessary to balance the system (see Simmons and Bengtsson, 1984)

3.2.1 The Atmosphere

The bulk of the incoming solar radiation is absorbed not by the atmosphere but by the Earth's surface (soil, ocean, ice) Evaporation of moisture and direct heating of the surface generate a heat transfer between the surface and the atmosphere in the form of latent and sensible heat The atmosphere transports this heat meridionally, mainly via transient weather systems with a timescale of the order of days

The following processes are important in determining the behaviour of the atmospheric component of the climate system

Turbulent transfer of heat , momentum and moisture at the surface of the Earth,
The surface type (i e , its albedo), which determines the proportion of incoming to reflected solar radiation
Latent heat release when water vapour condenses, clouds, which play an important role in reflecting

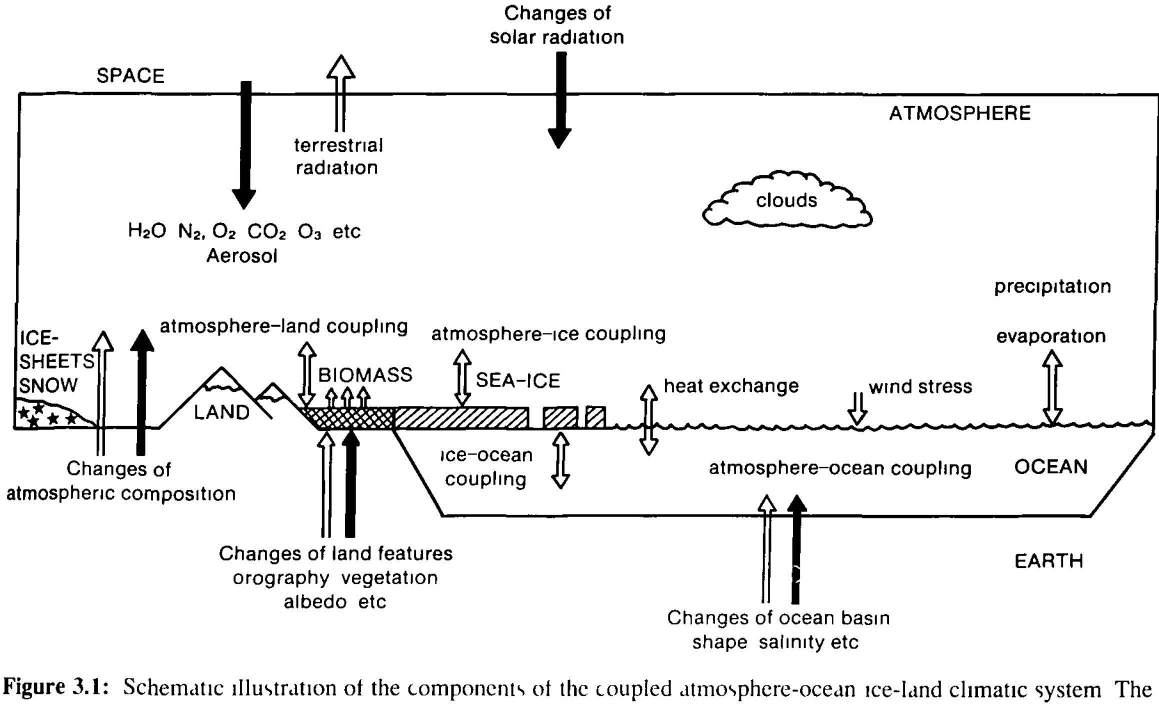


Figure 3.1: Schematic illustration of the components of the coupled atmosphere-ocean ice-land climatic system The full arrows are examples of external processes, and the open arrows are examples of internal processes in climatic change (from Houghton, 1984)

incoming solar short-wave radiation and in absorbing and emitting long-wave radiation,

The radiative cooling and heating of the atmosphere by CO_2, water vapour, ozone and other trace gases,

Aerosols (such as volcanic dust), the orbital parameters, mountain ranges and the land-sea distribution.

Atmospheric processes are also influenced by a number of feedback mechanisms which involve interactions between the atmospheric processes themselves (radiation and clouds, for example) and between these processes and the underlying surface. Such feedback mechanisms are discussed in more detail in 3.3 1

The problems concerning the impact of human activities on the greenhouse effect has broadened in scope from a CO_2 climate problem to a trace gas climate problem (Ramanathan et al., 1987). The climatic effects of trace gases are strongly governed by interactions between chemistry, radiation and dynamics. The nature of the trace gas radiative heating and the importance of chemical-radiative interactions has been already discussed in Section 2

3.2.2 The Ocean

The ocean also plays an essential role in the global climate system. Over half of the solar radiation reaching the Earth's surface is first absorbed by the ocean, where it is stored and redistributed by ocean currents before escaping to the atmosphere, largely as latent heat of evaporation, but also as long-wave radiation. The currents are driven by the exchange of momentum, heat and water between the ocean and atmosphere. They have a complicated horizontal and vertical structure determined by the pattern of winds blowing over the sea and the distribution of continents and submerged mountain ranges. The vertical structure of the ocean comprises three layers:

The Seasonal Boundary Layer, mixed annually from the surface, is less than 100 metres deep in the tropics and reaches hundreds of metres in the sub-polar seas (other than the North Pacific) and several kilometres in very small regions of the polar seas in most years;

The Warm Water Sphere (permanent thermocline), ventilated (i. e., exchanging heat and gases) from the seasonal boundary layer, is pushed down to depths of many hundreds of metres in gyres by the convergence of surface (Ekman) currents driven directly by the wind; and

The Cold Water Sphere (deep ocean), which fills the bottom 80% of the ocean's volume, ventilated from the seasonal boundary layer in polar seas

The ocean contains chemical and the biological mechanisms which are important in controlling carbon dioxide in the climate system. Carbon dioxide is transferred from the atmosphere into the interior of the ocean by the physical pump mechanism (described in the previous Section) caused by differences in the partial pressure of carbon dioxide in the ocean and the lowest layers of the atmosphere. Furthermore the annual ventilation of the seasonal boundary layer from the surface mixed-layer controls the efficiency of the biological pump by which ocean plankton convert dissolved carbon dioxide into particulate carbon, which sinks into deep water. These two pumps are responsible for extracting carbon dioxide from the global carbon cycle for periods in excess of a hundred years. The ocean branch of the carbon cycle involves a flux of carbon dioxide from the air into the sea at locations where the surface mixed layer has a partial pressure of CO_2 lower than the atmosphere and vice versa. Mixed-layer partial pressure of CO_2 is depressed by enhanced solubility in cold water and enhanced plankton production during the spring bloom. The rate of gas exchange depends on the air-sea difference in partial pressure of CO_2 and a coefficient which increases with wind speed.

The following processes control the the climate response of the ocean.

The small-scale (of order 50 km) transient eddies inside the ocean influence the structure of permanent gyres and streams and their interaction with submerged mountain ranges. The eddies also control the horizontal dispersion of chemicals (such as CO_2) dissolved in seawater.

The small-scale (tens of kilometres) patches of deep winter convection in the polar seas and the northernmost part of the North Atlantic, which transport heat and dissolved carbon dioxide below one kilometre into the deep reservoir of the cold water sphere, and the slow currents which circulate the newly implanted water around the world ocean.

The more extensive mechanism of thermocline ventilation by which some of the water in the surface mixed-layer flows from the seasonal boundary layer into the warm water sphere reservoir of the ocean, which extends for several hundreds of metres below most of the ocean's surface area.

The global transport of heat, freshwater and dissolved chemicals carried by ocean currents which dictate the global distributions of temperature, salinity, sea-ice and chemicals at the sea surface. Fluctuations in the large-scale circulation have modulated these patterns over years and decades. They also control the regional variations in sea surface properties which affect climate at this scale.

The biological pump in the seasonal boundary layer by which microscopic plants and animals (the plankton) consume some of the carbon dioxide dissolved in the seawater and sequester the carbon in the deep ocean away from the short term (up to a hundred years) interactions between ocean and atmosphere

3.2.3 *The Cryosphere*

The terrestrial cryosphere can be classified as follows (Untersteiner, 1984)

Seasonal snow cover, which responds rapidly to atmospheric dynamics on timescales of days and longer In a global context the seasonal heat storage in snow is small The primary influence of the cryosphere comes from the high albedo of snow covered surfaces

Sea ice, which affects climate on time scales of seasons and longer This has a similar effect on the surface heat balance as snow on land It also tends to decouple the ocean and atmosphere since it inhibits the exchange of moisture and momentum In some regions it influences the formation of deep water masses by salt extrusion during the freezing period and by the generation of fresh water layers in the melting period

Ice sheets of Greenland and the Antarctic, which can be considered as quasi-permanent topographic features They contain 80% of the existing fresh water on the globe, thereby acting as a long term reservoir in the hydrological cycle Any change in size will therefore influence the global sea level

Mountain glaciers are a small part of the cryosphere They also represent a freshwater reservoir and can therefore influence the sea level They are used as an important diagnostic tool for climate change since they respond rapidly to changing environmental conditions

Permafrost affects surface ecosystems and river discharges It influences the thermohaline circulation of the ocean

3.2.4 *The Biosphere*

The biosphere on land and in the oceans (discussed above) controls the magnitude of the fluxes of several greenhouse gases including CO_2 and methane, between the atmosphere, the oceans and the land The processes involved are sensitive to climatic and environmental conditions, so any change in the climate or the environment (e g , increases in the atmospheric abundance of CO_2) will influence the atmospheric abundance of these gases A detailed description of the feedbacks and their respective magnitudes can be found in Section 10

3.2.5 *The Geosphere*

The land processes play an important part in the hydrological cycle These concern the amount of fresh water stored in the ground as soil moisture (thereby interacting with the biosphere) and in underground reservoirs, or transported as run-off to different locations where it might influence the ocean circulation, particularly in high latitudes The soil interacts with the atmosphere by exchanges of gases, aerosols and moisture, and these are influenced by the soil type and the vegetation, which again are strongly dependent on the soil wetness Our present knowledge about these strongly interactive processes is limited and will be the target of future research (see Section 11)

3.2.6 *Timescales*

While the atmosphere reacts very rapidly to changes in its forcing (on a timescale of hours or days), the ocean reacts more slowly on timescales ranging from days (at the surface layer) to millennia in the greatest depths The ice cover reacts on timescales of days for sea ice regions to millennia for ice sheets The land processes react on timescales of days up to months, while the biosphere reacts on time scales from hours (plankton growth) to centuries (tree-growth)

3.3 Radiative Feedback Mechanisms

3.3.1 *Discussion of Radiative Feedback Mechanisms*

Many facets of the climate system are not well understood, and a significant number of the uncertainties in modelling atmospheric, cryospheric and oceanic interactions are directly due to interactive climate feedback mechanisms They can either amplify or damp the climate response resulting from a given climate forcing (Cess and Potter, 1988) For simplicity, emphasis will here be directed towards global-mean quantities, and the interpretation of climate change as a two-stage process forcing and response This has proved useful in interpreting climate feedback mechanisms in general circulation models It should, in fact, be emphasized that the conventional concept of climate feedback applies only to global mean quantities and to changes from one equilibrium climate to another

As discussed in Section 2, the radiative forcing of the surface-atmosphere system ΔQ is evaluated by holding all other climate parameters fixed, with $G = 4\ Wm^{-2}$ for an instantaneous doubling of atmospheric CO_2 It readily follows (Cess et al , 1989) that the change in surface climate, expressed as the change in global-mean surface temperature ΔT_s, is related to the radiative forcing by $\Delta T_s = \lambda \times \Delta Q$, where λ is the climate sensitivity parameter

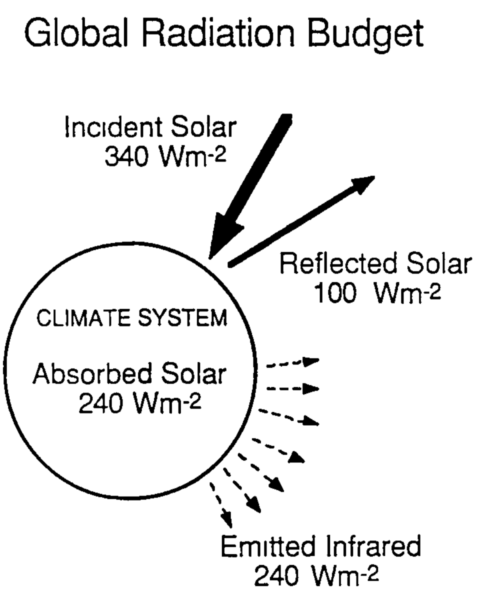
$$\lambda = \frac{1}{\Delta F/\Delta T_s \quad \Delta S/\Delta T_s}$$

where F and S denote respectively the global-mean emitted infrared and net downward solar fluxes at the Top Of the

Atmosphere (TOA). Thus ΔF and ΔS are the climate-change TOA responses to the radiative forcing ΔQ. An increase in λ thus represents an increased climate change due to a given radiative forcing ΔQ (= ΔF - ΔQ).

The definition of radiative forcing requires some clarification. Strictly speaking, it is defined as the change in net downward radiative flux at the tropopause, so that for an instantaneous doubling of CO_2 this is approximately 4 Wm^{-2} and constitutes the radiative heating of the surface-troposphere system. If the stratosphere is allowed to respond to this forcing, while the climatic parameters of the surface-troposphere system are held fixed, then this 4 Wm^{-2} flux change also applies at the top of the atmosphere. It is in this context that radiative forcing is used in this section.

A doubling of atmospheric CO_2 serves to illustrate the use of λ for evaluating feedback mechanisms. Figure 3.2 schematically depicts the global radiation balance. Averaged over the year and over the globe there is 340 Wm^{-2} of incident solar radiation at the TOA. Of this, roughly 30% or 100 Wm^{-2} is reflected by the surface - atmosphere system. Thus the climate system absorbs 240 Wm^{-2} of solar radiation, so that under equilibrium conditions it must emit 240 Wm^{-2} of infrared radiation. The CO_2 radiative forcing constitutes a reduction in the emitted infrared radiation, since this 4 Wm^{-2} forcing represents a heating of the climate system. Thus the CO_2 doubling results in the climate system absorbing 4 Wm^{-2} more energy than it emits, and global warming then occurs so as to increase the emitted radiation in order to re-establish the Earth's radiation balance. If this warming produced no change in the climate system other than temperature, then the system would return to its original radiation balance, with 240 Wm^{-2} both absorbed and emitted. In this absence of climate feedback mechanisms, $\Delta F/\Delta Ts$ = 3.3 $Wm^{-2} K^{-1}$ (Cess et al., 1989) while $\Delta S/\Delta Ts$ = 0, so that λ = 0.3 $Km^2 W^{-1}$. It in turn follows that ΔTs = $\lambda \times \Delta Q$ = 1.2°C. If it were not for the fact that this warming introduces numerous interactive feedback mechanisms, then ΔTs = 1.2°C would be quite a robust global-mean quantity. Unfortunately, such feedbacks introduce considerable uncertainties into ΔTs estimates. Three of the commonly discussed feedback mechanisms are described in the following sub-sections.



	Absorbed	Emitted
Instantaneous CO2 doubling	240 Wm^{-2}	236 Wm^{-2}
New Equilibrium with no other change	240 Wm^{-2}	240 Wm^{-2}

Figure 3.2: Schematic illustration of the global radiation budget at the top of the atmosphere.

3.3.2 Water Vapour Feedback

The best understood feedback mechanism is water vapour feedback, and this is intuitively easy to comprehend. For illustrative purposes a doubling of atmospheric CO_2 will again be considered. The ensuing global warming is, of course, the result of CO_2 being a greenhouse gas. This warming, however, produces an interactive effect, the warmer atmosphere contains more water vapour, itself a greenhouse gas. Thus an increase in one greenhouse gas (CO_2) induces an increase in yet another greenhouse gas (water vapour), resulting in a positive (amplifying) feedback mechanism.

To be more specific on this point, Raval and Ramanathan (1989) have recently employed satellite data to quantify the temperature dependence of the water vapour greenhouse effect. From their results it readily follows (Cess, 1989) that water vapour feedback reduces $\Delta F/\Delta Ts$ from the prior value of 3.3 $Wm^{-2} K^{-1}$ to 2.3 $Wm^{-2} K^{-1}$. This in turn increases λ from 0.3 $Km^2 W^{-1}$ to 0.43 $Km^2 W^{-1}$ and thus increases the global warming from ΔTs = 1.2°C to ΔTs = 1.7°C. There is yet a further amplification caused by the increased water vapour. Since water vapour also absorbs solar radiation, water vapour feedback leads to an additional heating of the climate system through enhanced absorption of solar radiation. In terms of $\Delta S/\Delta Ts$ as appears within the expression for λ, this results in $\Delta S/\Delta Ts$ = 0.2 $Wm^{-2} K^{-1}$ (Cess et al., 1989), so that λ is now 0.48 $Km^2 W^{-1}$ while ΔTs = 1.9°C. The point is that water vapour feedback has amplified the initial global warming of 1.2°C to 1.9°C, i.e., an amplification factor of 1.6.

3.3.3 Snow-Ice Albedo Feedback

An additional well-known positive feedback mechanism is snow-ice albedo feedback, by which a warmer Earth has less snow and ice cover resulting in a less reflective planet which in turn absorbs more solar radiation For simulations in which the carbon dioxide concentration of the atmosphere is increased, general circulation models produce polar amplification of the warming in winter, and this is at least partially ascribed to snow-ice albedo feedback The real situation, however, is probably more complex as, for example, the stability of the polar atmosphere in winter also plays a part Illustrations of snow-ice albedo feedback, as produced by general circulation models, will be given in Section 3 5 It should be borne in mind, however, that there is a need to diagnose the interactive nature of this feedback mechanism more fully

3.3.4 Cloud Feedback

Feedback mechanisms related to clouds are extremely complex To demonstrate this, it will be useful to first consider the impact of clouds upon the present climate Summarized in Table 3 1 are the radiative impacts of clouds upon the global climate system for annual mean conditions These radiative impacts refer to the effect of clouds relative to a 'clear-sky Earth, as will shortly be described this is termed cloud-radiative forcing

The presence of clouds heats the climate system by 31 Wm^{-2} through reducing the TOA infrared emission Note the similarity to trace-gas radiative forcing, which is why this impact is referred to as cloud radiative forcing Although clouds contribute to the greenhouse warming of the climate system, they also produce a cooling through the reflection and reduction in absorption of solar radiation As demonstrated in Table 3 1, the latter process dominates over the former, so that the net effect of clouds on the annual global climate system is a 13 Wm^{-2} radiative cooling As discussed below with respect to cloud feedback components, cloud-radiative forcing is an integrated effect governed by cloud amount, cloud vertical distribution, cloud optical depth and possibly the cloud droplet distribution (Wigley, 1989, Charlson et al, 1987)

Although clouds produce net cooling of the climate system, this must not be construed as a possible means of offsetting global warming due to increasing greenhouse gases As discussed in detail by Cess et al (1989), cloud feedback constitutes the ***change*** in net CRF associated with a change in climate Choosing a hypothetical example, if climate warming caused by a doubling of CO_2 were to result in a change in net CRF from 13 Wm^{-2} to -11 Wm^{-2}, then this increase in net CRF of 2 Wm^{-2} would amplify the 4 Wm^{-2} initial CO_2 radiative forcing and would so act as a positive feedback mechanism It is emphasized that this is a hypothetical example, and there is no *a priori* means of determining the sign of cloud feedback To emphasize the complexity of this feedback mechanism, three contributory processes are summarized as follows

Cloud Amount: If cloud amount decreases because of global warming, as occurs in typical general circulation model simulations, then this decrease reduces the infrared greenhouse effect attributed to clouds Thus as the Earth warms it is able to emit infrared radiation more efficiently, moderating the warming and so acting as a negative climate feedback mechanism But there is a related positive feedback, the solar radiation absorbed by the climate system increases because the diminished cloud amount causes a reduction of reflected solar radiation by the atmosphere There is no simple way of appraising the sign of this feedback component

Cloud Altitude: A vertical redistribution of clouds will also induce feedbacks For example, if global warming displaces a given cloud layer to a higher and colder region of the atmosphere, this will produce a positive feedback because the colder cloud will emit less radiation and thus have an enhanced greenhouse effect

Cloud Water Content There has been considerable recent speculation that global warming could increase cloud water content thereby resulting in brighter clouds and hence a negative component of cloud feedback Cess et al (1989) have recently suggested that this explanation is probably an oversimplification In one case, they demonstrated that this negative solar feedback induces a compensating positive infrared feedback In a more recent study they further indicate that in some models the net effect might thereby be that of positive feedback (see also Schlesinger and Roeckner, 1988, Roeckner et al 1987)

The above discussion clearly illustrates the multitude of complexities associated with cloud feedback and the uncertainties due to this feedback will further be emphasized in Section 3 5 In that both cloud and snow ice albedo feedbacks are geographical in nature then these feedback mechanisms can only be addressed through the use of three-dimensional numerical circulation models

Table 3.1: *Infrared, solar and net cloud-radiative forcing (CRF) These are annual-mean values*

Infrared CRF	31 Wm^{-2}
Solar CRF	- 44 Wm^{-2}
Net CRF	- 13 Wm^{-2}

3.4 Predictability Of The Climate System

The prediction of change in the climate system due to changes in the forcing is called climate forecasting In the climate system the slow components (for example the oceanic circulation) are altered by the fast components (for example the atmosphere) (Hasselmann, 1976, Mikolajewicz and Maier Reimer 1990) which again are influenced by the slow components, so that the complete system shows a considerable variance just by an interaction of all components involved This effect is an illustration of "natural variability

Taking the climate system as a whole we note that some elements of the system are chaotic viewed on a century to millennium time scale, while other parts are remarkably stable on those time scales The existence of these (in the time frame considered) stable components allows prediction of global change despite the existence of the chaotic elements The chaotic elements of the climate system are the weather systems in the atmosphere and in the ocean

The weather systems in the atmosphere have such a large horizontal scale that it is necessary to treat the whole of the atmospheric circulation as chaotic, nevertheless there are stable elements in the atmosphere as witnessed by the smooth seasonal cycle in such phenomena as the temperature distributions over the continents, the monsoon storm tracks, inter-tropical convergence zone etc That stability gives us hope that the response of the atmospheric climate (including the statistics of the chaotic weather systems) to greenhouse forcing will itself be stable and that the interactions between the atmosphere and the other elements of the climate system will also be stable even though the mechanisms of interaction depend on the weather systems

This leads to the common assumption used in climate prediction that the climate system is in equilibrium with its forcing That means, as long as its forcing is constant and the slowly varying components alter only slightly in the time scale considered, the mean state of the climate system will be stable and that if there is a change in the forcing, the mean state will change until it is again in balance with the forcing This state is described as an equilibrium state the transition between one mean and another mean state is called a transient state

The time-scale of the transition period is determined by the adjustment time of the slowest climate system component, i e the ocean The stable ('quasi stationary") behaviour of the climate system gives us the opportunity to detect changes by taking time averages Because the internal variability of the system is so high the averaging interval has to be long compared to the chaotic fluctuations to detect a statistically significant signal which can be attributed to the external forcing

A number of statistical test have been devised to optimize the detection of climate change signals (v Storch & Zwiers, 1988, Zwiers, 1988, Hasselmann, 1988, Santer and Wigley, 1990) (see Section 8)

Studies of the completed change from one mean state to another are called equilibrium response' studies Studies of the time evolution of the climate change due to an altered forcing, which might also be time dependent, are called transient response experiments

The weather systems in the ocean have much smaller horizontal scales (less than one hundred kilometres) than in the atmosphere leaving the large-scale features of the world ocean circulation to be non-chaotic The success of classical dynamical oceanography depends on that fact Observations of the penetration of transient tracers into the ocean show that the large-scale ocean currents are stable over periods of several decades Palaeo-oceanographic evidence shows that the currents and gyres adjusted smoothly to the ice age cycle That evidence and theoretical understanding of the large-scale ocean circulation suggests that we are indeed dealing, in the ocean, with a predictable system at least on timescales of decades The question is whether the existence of predictability in the ocean component of the Earth s climate system makes the system predictable as a whole However, this seems to be a reasonable working hypothesis, which receives some support from the smooth transient response simulated by coupled ocean-atmosphere models (see Section 6)

3.5 Methods Of Predicting Future Climate

Two approaches have been taken to predict the future climate

a) the analogue method , which tries to estimate future climate change from reconstructions of past climates using palaeo-climatic data,

b) climate simulations with numerical models (GCM s) of the atmospheric general circulation, which have been derived from weather forecast models They include representations of the other elements of the climate system (using ocean models, land surface models, etc) which have varying degrees of sophistication A comprehensive list of the models employed and the research groups involved can be found in Table 3 2(a) and (b)

Table 3.2(a): Summary of results from global mixed layer ocean atmosphere models used in equilibrium 2 x CO_2 experiments

ENTRY	Group	Investigators	Year	RESOLUTION No of waves or °lat x °long	No of Vertical Layers	Diurnal Cycle	Convection	Ocean Heat Transport	Cloud	Cloud Properties	ΔT (°C)	ΔP (%)	COMMENTS
A.	**Fixed, zonally averaged cloud; no ocean heat transport**												
1	GFDL	Manabe & Stouffer	1980	R15	9	N	MCA	N	FC	F	2.0	3.5	Based on 4 x CO_2 simulation
2		Wetherald & Manabe	1986 8	R15	9	N	MCA	N	FC	F	3.2	n/a	
B.	**Variable cloud; no ocean heat transport**												
3	OSU	Schlesinger & Zhao	1989	4° x 5°	2	N	PC	N	RH	F	2.8	8	
4			1989	4° x 5	2	N	PC	N	RH	F	4.4	11	As (3) but with revised clouds
5	MRI	Noda & Tokioka	1989	4° x 5°	5	Y	PC	N	RH	F	4.3 *	7 *	* Equilibrium not reached
6	NCAR	Washington & Meehl	1984	R15	9	N	MCA	N	RH	F	3.5 *	7 *	* Excessive ice Estimate ΔT = 4°C at equilibrium
7			1989	R15	9	N	MCA	N	RH	F	4.0	8	As (6) but with revised albedos for sea-ice, snow
8	GFDL	Wetherald & Manabe	1986 8	R15	9	N	MCA	N	RH	F	4.0	9	As (2) but with variable cloud
C	**Variable cloud; prescribed oceanic heat transport**												
9	AUS	Gordon & Hunt	1989	R21	4	Y	MCA	Y	RH	F	4.0	7	
10	GISS	Hansen et al	1981	8° x 10°	7	Y	PC	Y	RH	F	3.9	n/a	
11		Hansen et al	1984	8° x 10°	9	Y	PC	Y	RH	F	4.2	11	
12		Hansen et al	1984	8° x 10°	9	Y	PC	Y	RH	F	4.8	13	As (11) but with more sea-ice control
13	GFDL	Wetherald & Manabe	1989 †	R15	9	N	MCA	Y	RH	F	4.0	8	
14	MGO	Meleshko et al	1990	T21	9	N	PC	Y	RH	F	n/a	n/a	Simulation in progress
15	UKMO	Wilson & Mitchell	1987	5° x 7.5°	11	Y	PC	Y	RH	F	5.2	15	
16		Mitchell & Warrilow	1987	5° x 7.5°	11	Y	PC	Y	RH	F	5.2	15	As (15) but with four revised surface schemes
17		Mitchell et al	1989	5° x 7.5°	11	Y	PC	Y	CW	F	2.7	6	As (16) but with cloud water scheme
18			1989	5° x 7.5°	11	Y	PC	Y	CW	F	3.2	8	As (17) but with alternative ice formulation
19			1989	5° x 7.5°	11	Y	PC	Y	CW	V	1.9	3	As (17) but with variable cloud radiative properties
D.	**High Resolution**												
20	CCC	Boer et al	1989	T32	10	Y	MCA	Y	RH	V	3.5	4	* Soft convective adjustment
21	GFDL	Wetherald & Manabe	1989 †	R30	9	N	MCA	*	RH	F	4.0	8	* SSTs prescribed, changes prescribed from (13)
22	UKMO	Mitchell et al	1989	2.5°x3.75°	11	Y	PC	Y	CW	F	3.5	9	As (18) but with gravity wave drag

All models are global with realistic geography, a mixed-layer ocean, and a seasonal cycle of insolation Except where stated, results are the equilibrium response to doubling CO_2

R T = Rhomboidal/Triangular truncation in spectral space
ΔT = Equilibrium surface temperature change on doubling CO_2
ΔP = Percentage change in precipitation
N = Not included
Y = Included
MCA = Moist convective adjustment
PC = Penetrative convection
CA = Convective adjustment
CW = Cloud water
FC = Fixed cloud
RH = Condensation or relative humidity based cloud
V = Variable cloud radiative properties
F = Fixed cloud radiative properties
† = Personal communication
n/a = Not available
GFDL = Geophysical Fluid Dynamics Laboratory Princeton USA
NCAR = National Center for Atmospheric Research Boulder CO USA
MRI = Meteorological Research Institute Japan
MGO = Main Geophysical Observatory Leningrad USSR
CCC = Canadian Climate Center
UKMO = Meteorological Office United Kingdom
AUS = CSIRO Australia

Table 3.2(b): Summary of experiments carried out with global coupled ocean-atmosphere models

ENTRY	Group	Investigators	Year	RESOLUTION No. of Spectral Waves	Atmos. Levels	Diurnal Cycle	Conv-ection	Ocean Levels	Cloud	COMMENTS
1.	GFDL	Stouffer et al.	1989	R15	9	N	RH	12	MCA	100 Years, 1% CO_2 increase compounded.
2.	NCAR	Washington & Meehl	1989	R15	9	N	RH	4	FC	30 Years, 1% CO_2 increased linear.
3.	MPI	Cubasch et al.	1990	T21	19	Y	PC	11	CW	25 Years, instantaneous CO_2 doubling.
4.	UHH	Oberhuber et al.	1990	T21	19	Y	PC	9	CW	25 Years, instantaneous CO_2 doubling.

All models are global, with realistic geography and a seasonal cycle of insolation.

R, T = Number of waves in spectral space;
N = Not included;
MCA = Moist convective adjustment;
CA = Convective adjustment ;
FC = Fixed cloud;
Y = Included;
PC = Penetrative convection;
CW = Cloud water;
V = Variable cloud radiative properties.

GFDL = Geophysical Fluid Dynamics Laboratory,Princeton, USA;
MPI = Max Planck Institut für Meteorologie, Hamburg, FRG;
UHH = Met Institute, University of Hamburg, FRG;
NCAR = National Center for Atmospheric Research, Boulder, Co, USA.

3.5.1 The Palaeo-Analogue Method

This method has two distinct and rather independent parts. The first derives an estimate of global temperature sensitivity to atmospheric CO_2 concentrations based on estimates of CO_2 concentrations at various times in the past and the corresponding global average temperatures, adjusted to allow for past changes in albedo and solar constant. In the second part regional patterns of climate are reconstructed for selected past epochs, and they are regarded as analogues of future climates under enhanced greenhouse conditions. For a further discussion of the method, see for example Budyko and Izrael (1987).

3.5.1.1 Estimate of temperature sensitivity to CO_2 changes

There are three stages (Budyko et al., 1987):

i) determining the global mean changes for past palaeoclimates. This is done for four periods (Early Pliocene, early and middle Miocene, Palaeocene-Eocene and the Cretaceous). The temperature changes are based on isotopic temperatures obtained by Emiliani (1966) and maps derived by Sinitsyn (1965,1967) (see Budyko, 1982).

ii) subtracting the temperature change attributed to changes in the solar constant which is assumed to have increased by 5% every billion years, and to changes in surface albedo. A 1% increase in solar constant is assumed to raise the global mean surface temperature by 1.4°C. The changes in albedo are derived from the ratio of land to ocean, and each 0.01 reduction in albedo is assumed to have raised global mean temperature by 2°C. These corrections contribute to between 25% and 50% of the total change.

iii) relating the residual warming to the estimated change in atmospheric CO_2 concentrations. The CO_2 concentrations are derived from a carbon cycle model. The concentrations during the Eocene are estimated to be more than five times greater than present, and for the Cretaceous nine times greater (Budyko et al., 1987). On the other hand Shackleton (1985) argued that it is possible to constrain the total CO_2 in the ocean, and suggests that atmospheric CO_2 concentrations were unlikely to have been more than double today's value.

The result is a sensitivity of 3.0°C for a doubling of CO_2, with a possible range of ± 1°C, which is very similar to that obtained on the basis of numerical simulations (Section 5).

3.5.1.2 Construction of the analogue patterns

In their study, Budyko et al. (1987) used the mid-Holocene (5-6 kbp), the Last Interglacial (Eemian or Mikulino,125 kbp) and the Pliocene (3-4 mbp) as analogues for future climates. January, July and mean annual temperatures and mean annual precipitation were reconstructed for each of the above three epochs (see Figures 7.3, 7.4 and 7.5). Estimates of the mean temperatures over the Northern Hemisphere exceeded the temperature at the end of the pre-industrial period (the 19th century) by approximately 1°, 2° and 3-4°C during the mid-Holocene, Eemian and Pliocene respectively. These periods were chosen as analogues of future climate for 2000, 2025 and 2050 respectively.

Although the nature of the forcing during these periods was probably different, the relative values of the mean latitudinal temperature change in the Northern Hemisphere for each epoch were similar in each case (Figure 3.3). Note however that the observational coverage was rather limited, especially for the Eemian when the land-based data came essentially from the Eastern Hemisphere (see Section 7.2.2). Correlations were also calculated between estimated temperature anomalies for 12 regions of the Northern Hemisphere in each of the three epochs. These were found to be statistically significant in most cases, despite the limited quality and quantity of data in the earlier epochs.

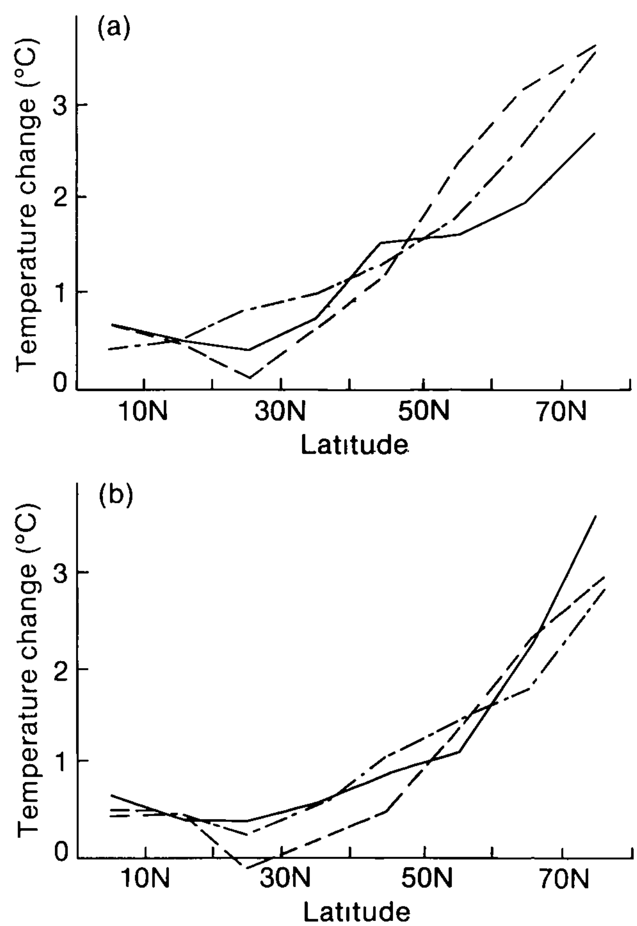


Figure 3.3: Relative surface air temperature changes in different latitudes of the Northern Hemisphere during the palaeo-climatic warm epochs. **(a)** winter **(b)** summer. Full line = Holocene. Dashed line = last interglacial. Dash-dotted line = Pliocene.

The considerable similarity between the temperature anomaly maps for the three different epochs suggests that the regional temperature anomaly changes are, to a first approximation, directly proportional to increasing mean global temperature If this is true, then the regional distributions of surface air temperature anomalies are analogous to each other and the similarities between these maps also suggest that the empirical methods for estimating the spatial temperature distribution with global warming may be relatively robust

Similarly, annual mean precipitation changes have been reconstructed, though the patterns in the mid Holocene differ from those found for the other two periods (see Section 5 4, Section 7 2 2)

When reconstructions of past climate conditions are accurate and thorough, they can provide relatively reliable estimates of self-consistent spatial patterns of climatic changes Weaknesses in developing these relationships can arise because of uncertainties

i) in reconstructing past climates
ii) in extending limited areal coverage to global scales
iii) in interpreting the effects of changing orography and equilibrium versus non-equilibrium conditions
iv) in determining the relative influences of the various factors that have caused the past climatic changes

3.5.2 Atmospheric General Circulation Models

General circulation models are based on the physical conservation laws which describe the redistribution of momentum, heat and water vapour by atmospheric motions All of these processes are formulated in the primitive equations which describe the behaviour of a fluid (air or water) on a rotating body (the Earth) under the influence of a differential heating (the temperature contrast between equator and pole) caused by an external heat source (the Sun) These governing equations are non-linear partial differential equations whose solution cannot be obtained except by numerical methods These numerical methods subdivide the atmosphere vertically into discrete layers wherein the variables are carried and computed For each layer the horizontal variations of the predicted quantities are determined either at discrete grid points over the Earth as in grid point (finite difference) models or by a finite number of prescribed mathematical functions as in spectral models The horizontal resolution of a typical atmospheric model used for climate studies is illustrated by its representation of land and sea shown in Figure 3 4

The values of the predicted variables (wind temperature humidity surface pressure rainfall etc) for each layer (including the surface) and grid point (or mathematical function) are determined from the governing equation by marching (integrating) forward in time in discrete time steps starting from some given initial conditions To

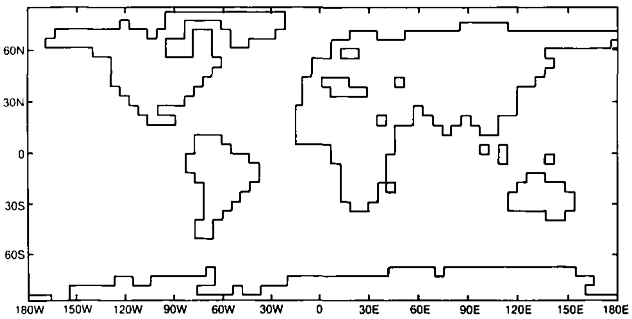


Figure 3.4: The model land sea mask for a typical climate model (T21, ECHAM, after Cubasch et al, 1989)

prevent the solution from becoming numerically unstable, the time step must be made smaller than a value that depends on the speed of the fastest moving disturbance (wave), the the grid size (or smallest resolved wavelength), and the integration method

The spatial resolution of GCM's is constrained for practical reasons by the speed and memory capacity of the computer used to perform the numerical integrations Increasing the resolution not only increases the memory required (linearly for vertical resolution, quadratically for horizontal resolution), but also generally requires a reduction in the integration time step Consequentially, the computer time required increases rapidly with increasing resolution Typical models have a horizontal resolution of 300 to 1000 km and between 2 and 19 vertical levels These resolutions are sufficient to represent large-scale features of the climate, but allow only a limited interpretation of results on the regional scale

3 5 2 1 Physical parameterizations

Due to their limited spatial resolution, GCM's do not (and will not with any foreseeable increase of resolution) resolve several physical processes of importance to climate However the statistical effects of these sub grid-scale processes on the scales resolved by the GCM have to be incorporated into the model by relating them to the resolved scale variables (wind, temperature, humidity and surface pressure) themselves Such a process is called parametrization, and is based on both observational and theoretical studies Figure 3 5 shows the physical processes parameterized in a typical GCM, and their interactions

3 5 2 2 Radiation and the effect of clouds

The parametrization of radiation is possibly the most important issue for climate change experiments, since it is through radiation that the effects of the greenhouse gases are transferred into the general circulation A radiation parametrization scheme calculates the radiative balance of

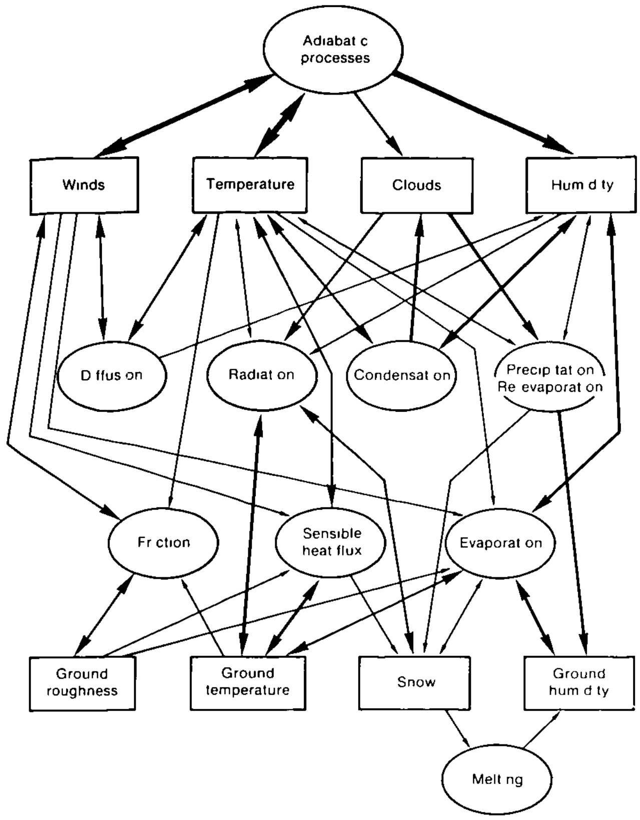


Figure 3.5: The processes parametrized in a numerical atmosphere model (ECMWF) and their interaction. The thickness of the arrows indicates the strength of the interaction (from Houghton 1984).

the incoming solar radiation and the outgoing terrestrial long-wave radiation and, as appropriate, the reflection, emission and absorption of these fluxes in the atmosphere. Absorption and emission are calculated in several broad spectral bands (for reasons of economy) taking into account the concentration of different absorbers and emitters like CO_2, water vapour, ozone and aerosols.

One sensitive part in any radiation scheme is the calculation of the radiative effect of clouds. In early GCM experiments clouds were prescribed using observed cloud climatologies (fixed cloud (FC) experiments), and were not allowed to alter during the experiments with (for example) changed CO_2 concentration. Later schemes contained interactive cloud parametrizations of various sophistication, but mostly based on an estimate of the cloud amount from the relative humidity (RH experiments). Only the most advanced schemes calculate the variation of cloud optical properties by the cloud water content (CW experiments). Capital letters in brackets indicate abbreviations used in Table 3.2 (a) and 3.2 (b).

The seasonal variation of the solar insolation is included in almost all experiments, but a diurnal cycle is omitted in many simulations. Climate experiments run without a seasonal cycle are limited in scope and their reliability for climate change experiments is therefore doubtful. The inclusion of the diurnal cycle improves the realism of some feedback mechanisms and therefore the quality of the climate simulations.

3.5.2.3 Sub-grid-scale transports

Most of the solar radiation absorbed by the climate system is absorbed at the surface. This energy becomes available for driving the atmospheric general circulation only after it has been transferred to the atmosphere through the planetary boundary layer (PBL), primarily by small-scale turbulent and convective fluxes of sensible and latent heat, but also by net long-wave radiative exchange. On the other hand, the general circulation is slowed down by frictional dissipation which basically takes place in the PBL through vertical transport of momentum by turbulent eddies.

In most GCMs the turbulent fluxes of heat, water vapour and momentum at the surface are calculated from empirical "bulk formulae" with stability dependent transfer coefficients. The fluxes at the PBL top (at a fixed height generally) are either neglected or parametrized from simple mixed-layer theory. In GCMs that resolve the PBL, the eddy diffusion approach is generally employed. Considerable efforts are made to incorporate into the PBL parametrizations the effects of cloud, vegetation and sub grid-scale terrain height.

Cumulus convection in a vertically unstable atmosphere is one of the main heat producing mechanisms at scales which are unresolvable in GCMs. A common procedure is to adjust the temperature and water vapour profile to a conditionally stable state (Moist Convective Adjustment MCA). The second class of cumulus parameterizations often employed in GCMs is based on a "moisture convergence" closure (KUO). Other GCMs use Penetrative Convection (PC) schemes to mix moist conditionally unstable air from lower model layers with dry air aloft. The question of how sophisticated convective parameterizations in GCMs need be, and how much the sensitivity of climate change experiments depends on their formulation, is still open.

3.5.2.4 Land surface processes

Another important parametrization is the transfer of heat and water within the soil, for instance the balance between evaporation and precipitation, snow melt, storage of water in the ground and river runoff. This parametrization is of extreme relevance for climate change predictions, since it shows how local climates may change from humid to arid and vice versa depending on global circulation changes. It furthermore reflects, in some of the more sophisticated schemes, the changes that could occur through alterations in surface vegetation and land-use.

Most soil moisture schemes used to date are based either on the so-called "bucket" method or the force-restore method In the former case, soil moisture is available from a single reservoir, or thick soil layer When all the moisture is used up, evaporation ceases In the latter method, two layers of soil provide moisture for evaporation, a thin, near-surface layer which responds rapidly to precipitation and evaporation, and a thick, deep soil layer acting as a reservoir If the surface layer dries out, deep soil moisture is mostly unavailable for evaporation and evaporation rates fall to small values However, in the presence of vegetation, realistic models use the deep soil layer as a source of moisture for evapotranspiration

At any given grid-point over land, a balance between precipitation, evaporation, runoff and local accumulation of soil moisture is evaluated If precipitation exceeds evaporation, then local accumulation will occur until saturation is achieved After this, runoff is assumed and the excess water is removed The availability of this runoff as fresh water input to the ocean has been allowed for in ocean models only recently (Cubasch et al , 1990) Most models differ in the amount of freshwater required for saturation, and few treat more than one soil type The force-restore method has recently been extended to include a range of soil types by Noilham and Planton (1989)

3 5 2 5 Boundary conditions

To determine a unique solution of the model equations, it is necessary to specify a set of upper and lower boundary conditions These are

- input of solar radiation (including temporal variation) at the top of the atmosphere,
- orography and land sea distribution,
- albedo of bare land,
- surface roughness,
- vegetation characteristics

The lower boundary over the sea is either prescribed from climatological data or, as this is not very appropriate for climate change experiments it has to be calculated by an ocean model As comprehensive ocean models are expensive to run (see Section 3 5 3) the most commonly used ocean model coupled to atmospheric models is the mixed-layer model This model describes the uppermost layer of the ocean where the oceanic temperature is relatively uniform with depth It is frequently modelled as a simple slab for which a fixed depth of the mixed layer is prescribed and the oceanic heat storage is calculated the oceanic heat transport is either neglected or is carried only within the mixed layer or is prescribed from climatology Sea ice extents are determined interactively usually with a variant of the thermodynamic sea ice model due to Semtner (1976) Such an ocean model evidently has strong limitations for studies of climate change particularly as it does not allow for the observed lags in heat storage of the upper ocean to be represented Variations of mixed-layer depth, oceanic heat flux convergence, and exchanges with the deep ocean, which would entail an additional storage and redistribution of heat, are all neglected as well Attempts have been made to couple atmospheric models to ocean models of intermediate complexity Thus, for example, Hansen et al (1988) have used a low resolution atmospheric model run with a mixed layer model coupled diffusively to a deep ocean to simulate the time dependent response to a gradual increase in trace gases

3.5.3 Ocean Models

To simulate the role of the ocean more adequately, a number of dynamical ocean models have been developed (Bryan, 1969, Semtner, 1974, Hasselmann, 1982, Cox, 1984, Oberhuber, 1989) The typical ocean model used for climate simulations follows basically the same set of equations as the atmosphere if the equation defining the water vapour balance is replaced by one describing salinity As with atmospheric GCMs, numerical solutions can be obtained by applying finite difference techniques and specifying appropriate surface boundary conditions (i e , fluxes of heat, momentum and fresh water) either from observations (uncoupled mode) or from an atmospheric GCM (coupled mode see Section 3 5 6) The vertical and horizontal exchange of temperature, momentum and salinity by diffusion or turbulent transfers is parametrized

The formation of sea-ice is generally treated as a purely thermodynamic process However, some models already include dynamical effects such as sea ice drift and deformation caused by winds and ocean currents (Oberhuber et al 1989)

One of the problems of simulating the ocean is the wide range of time and length scales involved The models for climate sensitivity studies resolve only the largest time and length scales (horizontal resolution 200 to 1000 km, time scale hours to 10 000 years, vertical resolution 2 to 20 levels) High resolution models, which can resolve eddies, are now being tested (Semtner and Chervin, 1988) but with the currently available computer power cannot be run sufficiently long enough to simulate climate changes

3.5.4 Carbon Cycle Models

The exchange of carbon dioxide between the ocean and atmosphere can be simulated by adding equations to the ocean component for the air-sea gas flux, the physics of gas solubility the chemistry of carbon dioxide buffering in sea water and the biological pump (Maier Reimer and Hasselmann 1989) This extension of the coupled ocean-atmosphere model will permit diagnosis of the fractionation of carbon dioxide between the atmosphere and ocean in the last hundred years and changes to that fractionation in the future as the ocean begins to respond to global warming, in particular through changes in the ocean mixed layer depth, which affects both the physical uptake of carbon dioxide and the efficiency of the biological pump The physical and chemical equations are well established, but more work is needed to establish equations for the biological pump The latter must parametrize the biological diversity, which varies regionally and seasonally and is likely to vary as the climate changes, ideally the equations themselves must cope with such changes without introducing too many variables Candidate sets of such robust biological equations have been tested in one-dimensional models and are now being used in ocean circulation models with encouraging results It seems likely that they will have to be incorporated into eddy-resolving ocean circulation models in order to avoid biases due to the patchy growth of plankton They will also have to pay special attention to the seasonal boundary layer (the biologist s euphotic zone) and its interaction with the permanent thermocline in order to deal with nutrient and carbon dioxide recirculation Such models are computationally expensive and complete global models based on these equations will have to await the arrival of more powerful supercomputers later in the 1990s Besides the biological organic carbon pump the biological calcium carbonate counter pump and interactions between the seawater and carbon sediment pools must be considered First results with models which include the organic carbon pump with a sediment reservoir indicate the importance of these processes (Heinze and Maier Reimer 1989)

3.5.5 Chemical Models

Due to the increasing awareness of the importance of trace gases other than CO_2 a number of research groups have now started to develop models considering the chemical interactions between a variety of trace gases and the general circulation (Prather et al 1987) At the time of writing, these models have not yet been used in the models discussed so far to estimate the global climate change It will be interesting to see their impact on future climate change modelling

3 5.6 Coupled Models of the Atmosphere and the Ocean

Due to the dominating influence of the ocean atmosphere link in the climate system, realistic climate change experiments require OGCM s and AGCM s to be coupled together by exchanging information about the sea surface temperature, the ice cover, the total (latent ,sensible and net longwave radiative) heat flux, the solar radiation and the wind stress

One basic problem in the construction of coupled models arises from the wide range of time scales from about one day for the atmosphere to 1000 years for the deep ocean Synchronously coupled atmosphere ocean models are extremely time consuming and limited computer resources prohibit equilibrium being reached except with mixed-layer models Various asynchronous coupling techniques have been suggested to accelerate the convergence of a coupled model However, the problem is far from being solved and can only really be tackled by using faster computers

A second basic problem that arises through such coupling is model drift The coupled model normally drifts to a state that reflects the systematic errors of each respective model component because each sub model is no longer constrained by prescribed observed fluxes at the ocean-atmosphere interface Therefore flux correction terms are sometimes introduced to neutralize the climate drift and to obtain a realistic reference climate for climate change experiments (Sausen et al, 1988 Cubasch, 1989) (c f Section 4 9) However, these terms are additive and do not guarantee stability from further drift, They are also prescribed from present-day conditions and are not allowed to change with altered forcing from increased CO_2

Carbon cycle models have already been coupled to ocean models, but coupling to an AGCM-OGCM has not yet been carried out

3.5.7 Use of Models

Despite their shortcomings, models provide a powerful facility for studies of climate and climate change A review of such studies is contained in Schlesinger (1983) They are normally used for investigations of the sensitivity of climate to internal and external factors and for prediction of climate change by firstly carrying out a control integration with parameters set for present day climate in order to establish a reference mean model climatology and the necessary statistics on modelled climatic variability These can both be verified against the observed climate and used for examination and assessment of the subsequently modelled climate change The climate change (pert urbation) run is then carried out by repeating the model run with appropriately changed parameters (a doubling of CO_2 for example) and the differences between this and the parallel control run examined The difference between the control and the perturbed experiments is called the response The significance of the response must be assessed against the model s natural variability (determined in the control run) using appropriate statistical tests These enable an assessment to be made (usually expressed in terms of probabilities) of the confidence that the changes obtained represent an implied climatic change, rather than simply a result of the natural variability of the model

Typical integration times range from 5 to 100 years depending on the nature of the investigation Until now most effort to study the response to increased levels of greenhouse gas concentrations has gone into determining the equilibrium response of climate to a doubling of CO_2 using atmospheric models coupled to slab ocean models A comparatively small number of attempts have been made to determine the transient (i e time-dependent) climate response to anthropogenic forcing using coupled atmosphere and ocean circulation models

3 5 7 1 Equilibrium response experiments

In an equilibrium response experiment both simulations, i e , the control experiment with the present amount of atmospheric CO_2 and the perturbation experiment with doubled CO_2, are run sufficiently long to achieve the respective equilibrium climates A review of such experiments is given in Schlesinger and Mitchell (1987) For a mixed-layer ocean the response time to reach equilibrium amounts to several decades, which is feasible with present day computers For a fully coupled GCM the equilibrium response time would be several thousand years and cannot be achieved with present day computers A comprehensive list of equilibrium response experiments can be found in Table 3 2(a)

3 5 7 2 Time dependent response experiments

Equilibrium response studies for given CO_2 increases are required as standard benchmark calculations for model intercomparison The results may be misleading, however, if applied to actual climate change caused by man s activities, because the atmospheric CO_2 concentrations do not change abruptly but have been growing by about 0 4% per year Moreover, the timing of the atmospheric response depends crucially on the ocean heat uptake which might delay the CO_2 induced warming by several decades Thus, for realistic climate scenario computations, not only have the atmospheric processes to be simulated with some fidelity but also the oceanic heat transport which is largely governed by ocean dynamics First experiments with coupled dynamical atmosphere ocean models have been performed (Table 3 2(b)) and will be discussed later in Section 6

3.6 Illustrative Equilibrium Experiments

In this section climate sensitivity results, as produced by a large number of general circulation models, are summarized for two quite different climate change simulations The first refers to a simulation that was designed to suppress snow-ice albedo feedback so as to concentrate on the water vapour and cloud feedbacks The second consists of a summary of global warming due to a CO_2 doubling

The first case, addressing only water vapour and cloud feedbacks consists of a perpetual July simulation in which the climate was changed by imposing a 4°C perturbation on the global sea surface temperature while holding sea ice fixed Since a perpetual July simulation with a general circulation model results in very little snow cover in the Northern Hemisphere this effectively eliminates snow-ice albedo feedback The details of this simulation are given by

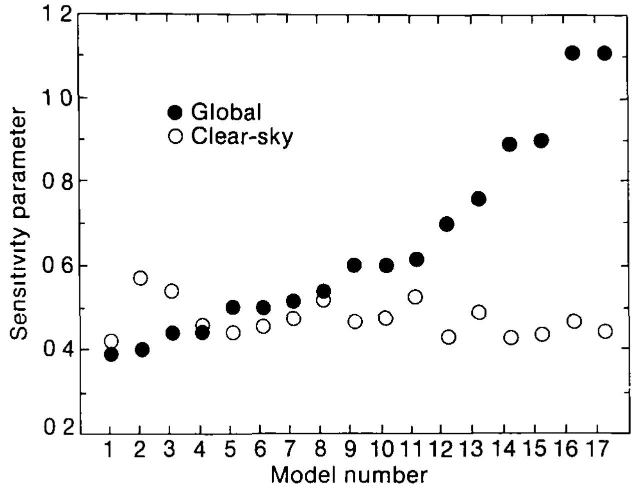


Figure 3.6: Summary of the clear sky and global sensitivity parameters for 17 general circulation models

Cess et al (1989), the main point is that it was chosen so as to minimize computer time and thus allow a large number of modelling groups to participate in the intercomparison This procedure is in essence an inverse climate change simulation Rather than introducing a radiative forcing into the models and then letting the model climates respond to this forcing, the climate change was instead prescribed and the models in turn produced their respective radiative forcings

Cess et al (1989) have summarized climate sensitivity parameters as produced by 14 atmospheric general circulation models (most of them are referenced in Table 3 2(a) and 3 2(b) This number has since risen to 17 models and their sensitivity parameters (λ as defined in Section 3 3 1) are summarized in Figure 3 6 The important point here is that cloud effects were isolated by separately averaging the models clear sky TOA fluxes, so that in addition to evaluating the climate sensitivity parameter for the globe as a whole (filled circles in Figure 3 6), it was also possible to evaluate the sensitivity parameter for an equivalent clear-sky Earth (open circles)

Note that the models are in remarkable agreement with respect to the clear sky sensitivity parameter, and the model average $\lambda = 0\ 47\ Km^2W^{-1}$ is consistent with the discussion of water vapour feedback (Section 3 3 2), for which it was suggested that $\lambda = 0\ 48\ Km^2W^{-1}$ There is, however a nearly threefold variation in the global sensitivity parameter and since the clear sky sensitivity parameters are in good agreement, then this implies that most of the disagreements can be attributed to differences in cloud feedback A more detailed demonstration of this is given by Cess et al (1989) The important conclusions from this intercomparison are that the 17 models agree well with an observational determination of water vapour feedback, whereas improvements in the treatment of cloud feedback are needed if general circulation models are ultimately to be used as reliable climate predictions

The second type of simulation refers to a doubling of atmospheric CO_2, so that in proceeding from one equilibrium climate to another, snow-ice albedo feedback is additionally activated in the general circulation models It must be cautioned however, that cloud feedback in this type of simulation should not be expected to be similar to that for the perpetual July simulation Furthermore, one should anticipate interactive effects between cloud feedback and snow-ice albedo feedback

Summarized in Table 3 2(a) are ΔTs results, as well as the related changes in global precipitation, for CO_2 doubling simulations using a number of general circulation models All models show a significant increase in global-mean temperature which ranges from 1 9°C to 5 2°C As in the perpetual July simulations, cloud feedback probably introduces a large uncertainty, although here it is difficult to quantify this point

Most results lie between 3 5°C and 4 5°C, although this does not necessarily imply that the correct value lies in this range Nor does it mean that two models with comparable ΔTs values likewise produce comparable individual feedback mechanisms For example consider the Wetherald and Manabe (1988) and Hansen et al (1984) simulations for which the respective ΔTs values are 4 0°C and 4 2°C Summarized in Table 3 3 are their diagnoses of individual feedback mechanisms These two models (labelled GFDL and GISS respectively) produce rather similar warmings in the absence of both cloud feedback and snow-ice albedo feedback The incorporation of cloud feedback, however, demonstrates that this is a stronger feedback in the GISS model, as is consistent with the perpetual July simulations But curiously the additional incorporation of snow-ice albedo feedback compensates for their differences in cloud feedback Thus, while the two models produce comparable global warming, they do so for quite different reasons

Table 3.3: *Comparison of ΔTs (°C) for the GFDL and GISS models with the progressive addition of cloud and snow ice feedbacks*

FEEDBACKS	GFDL	GISS
No cloud or snow-ice	1 7	2 0
Plus cloud	2 0	3 2
Plus snow ice	4 0	4 2

It should be emphasized that Table 3 3 should not be used to appraise the amplification factor due to cloud feedback since feedback mechanisms are interactive For example, from Table 3 3 the cloud feedback amplifications for the GFDL and GISS models might be inferred to be 1 2 and 1 6 respectively But, these are in the ***absence*** of snow-ice albedo feedback Conversely, if snow-ice albedo feedback is incorporated before cloud feedback, then the respective amplification factors are 1 3 and 1 8 These larger values are due to an amplification of cloud feedback by snow-ice albedo feedback

3.7 Summary

Many aspects of the global climate system can now be simulated by numerical models The feedback processes associated with these aspects are usually well represented, but there appear to be considerable differences in the strength of the interaction of these processes in simulations using different models Section 4 examines results from various models in more detail

Unfortunately, even though this is crucial for climate change prediction, only a few models linking all the main components of the climate system in a comprehensive way have been developed This is mainly due to a lack of computer resources, since a coupled system has to take the different timescales of the sub-systems into account, but also the task requires interdisciplinary cooperation

An atmospheric general circulation model on its own can be integrated on currently available computers for several model decades to give estimates of the variability about its equilibrium response, when coupled to a global ocean model (which needs millennia to reach an equilibrium) the demands on computer time are increased by several orders of magnitude The inclusion of additional sub-systems and a refinement of resolution needed to make regional predictions demands computer speeds several orders of magnitude faster than is available on current machines

We can only expect current simulations of climate change to be broadly accurate and the results obtained by existing models may become obsolete as more complete models of the climate system are used Results from fully coupled atmosphere ocean models are now beginning to emerge, these are given in Section 6

The palaeo-analogue method, although of limited use for detailed climate prediction (see Section 5), nevertheless gives valuable information about the spectrum of past and future climate changes and provides data for the calibration of circulation models in climate regimes differing from the present Results from these calibrations are shown in Section 4

References

Boer G et al , 1989 Equilibrium response studies with the CCC climate model Pers Com

Bryan, K 1969 A numerical method for the study of the circulation of the world ocean *J Comp Phys* , **4**, 347-376

Budyko, M I , 1982 The Earth s climate ,past and future (English translation) Academic Press

Budyko, M I , A B Ronov and A L Yanshin, 1987 History of the Earth's atmosphere (English translation) Springer-Verlag

Budyko, M I and Y Izrael, 1987 Anthropogenic climate change , Gidrometeorizdat, 406pp

Cess, R D and G L Potter, 1988 A methodology for understanding and intercomparing atmospheric climate feedback processes in general circulation models *J Geophys Res* , **93**, 8305 8314

Cess R D , G L Potter, J P Blanchet G J Boer, S J Ghan, J T Kiehl, H Le Treut Z X Li, X Z Liang J F B Mitchell, J J Morcrette, D A Randall, M R Riches, E Roeckner, U Schlese A Slingo, K E Taylor, W M Washington R T Wetherald and I Yagai, 1989 Interpretation of cloud climate feedback as produced by 14 atmospheric general circulation models *Science*, **245**, 513 516

Cess, R D , 1989 Gauging water vapour feedback *Nature* **342** 736 737

Cox, M D 1984 A primitive equation, three dimensional model of the ocean GFDL ocean group, *Techn Rep* No **1** 250pp

Charlson, R J , J E Lovelock, M O Andrae and S G Warren, 1987 Oceanic plankton, atmospheric sulphur, cloud albedo and climate *Nature*, **326**, 655-661

Cubasch U , 1989 Coupling a global atmosphere model with a global ocean model using the flux correction method, In *Aspects of coupling atmosphere and ocean models* Rep No 6 , Met Inst Univ Hamburg, FRG

Cubasch U , E Maier Reimer, B Santer, U Mikolajewicz, E Roeckner and M Boettinger, 1990 The response of a global coupled O-AGCM to CO_2 doubling MPI report, MPI fuer Meteorologie Hamburg, FRG

Emiliani C 1966 Isotopic palaeotemperatures, *Science* **154**, 851-857

Gates I 1975 The physical basis of climate and climate modelling GARP Publication Series No 16 WMO, Geneva

Gordon and Hunt, 1989 Australian equilibrium response experiments Pers Com

Hansen J D Johnston, A Lacis, S Lebedeff, P Lee D Rind and G Russel, 1981 Climate impact of increasing atmospheric carbon dioxide *Science*, **213**, 957 966

Hansen J , A Lacis, D Rind, L Russel P Stone I Fung, R Ruedy J Lerner, 1984 Climate sensitivity analysis of feedback mechanisms In *Climate processes and climate sensitivity*, (ed J Hansen and T Takahashi), Geophys Monogr Ser **29**, 130-163 AGU, Washington

Hansen J I Fung, A Lacis, D Rind S Lebedeff, R Ruedy and G Russel, 1988 Global climate changes as forecast by GISS s three dimensional model *J Geophys Res* **93**, 9341 9364

Hasselmann, K 1976 Stochastic climate models Part I Theory *Tellus*, **28**, 6 473 485

Hasselmann, K 1982 An ocean model for climate variability studies *Prog Oceanogr* **11** 69-92

Hasselmann, K , 1988 PIPs and POPs - A general formalism for the reduction of dynamical systems in terms of Principal Interaction and Patterns and Principal Oscillation Patterns *J Geophys Res* , **93**, 11,015 - 11,022

Heinze, C and E Maier Reimer, 1989 Glacial pCO_2 reduction by the world ocean Experiments with the Hamburg Carbon Cycle model, In *3 World conference on Analysis and Evaluation of atmospheric CO_2 data present and past* WMO, Environmental Pollution Monitoring and Research Program, Rep No **59**, 9 14

Houghton, J (ed), 1984 The Global Climate Cambridge University Press, Cambridge, UK 233pp

Manabe, S and R J Stouffer, 1980 Sensitivity of a global climate model to an increase in the CO_2 concentration in the atmosphere *J Geophys Res* **85**, 5529-5554

Manabe, S , K Bryan and M Spelman, 1989 Transient response of a global ocean atmosphere model to a doubling of atmospheric carbon dioxide submitted to *J Geophys Res*

Maier Reimer, E and K Hasselmann, 1987 Transport and storage of CO_2 in the ocean - an inorganic ocean circulation carbon cycle model, Climate Dynamics, **2**, 63-90

Maier Reimer E and U Mikolajewicz, 1989 Experiments with an OGCM on the cause of the Younger Dryas", MPI Report No 39 MPI fur Meteorologie Hamburg

Meehl, G A and W M Washington, 1988 A comparison of soil moisture sensitivity in two global climate models *JAS*, **45**, 1476 1492

Meleshko, 1990 CO_2 equilibrium response studies with the CCCP climate model Pers Com

Mikolajewicz, U and E Maier Reimer, 1990 Internal secular variability in an ocean general circulation model, MPI Report No 50, MPI fur Meteorologie, Hamburg

Mitchell, J F B and D A Warrilow, 1987 Summer dryness in northern mid latitude due to increased CO_2 *Nature*, **330**, 238-240

Mitchell, J F B , C A Senior and W J Ingram, 1989 CO_2 and climate A missing feedback? *Nature*, **341**, 132-134

Noda, A and T Tokioka, 1989 The effect of doubling CO_2 concentration on convective and non-convective precipitation in a general circulation model coupled with a simple mixed layer ocean *J Met Soc Japan*, **67**, 95 110

Noilham, J and S Planton 1989 A simple parametrization of land surface processes for meteorological models *MWR*, **117**, 536 549

Oberhuber, J , 1989 An isopycnic global ocean model MPI report, MPI fuer Meteorologie, Hamburg

Oberhuber, J , F Lunkeit and R Sausen, 1990 CO_2 doubling experiments with a coupled global isopycnic ocean-atmosphere circulation model Rep d Met Inst Univ Hamburg, Hamburg, FRG

Prather, M , M McElroy, S Wolfsy, G Russel and D Rind, 1987 Chemistry of the global troposphere Fluorocarbons as tracers of air motion *JGR*, **92**, 6579 6611

Ramanathan V L Callis, R Cess J Hansen, F Isaksen, W Kuhn, A Lacis, F Luther J Mahlman, R Reck and M Schlesinger 1987 Climate-chemical Interactions and effects of changing atmospheric trace gases *Rev of Geophysics*, **25**, 1441 1482

Raval, A and V Ramanathan, 1989 Observational determination of the greenhouse effect *Nature*, **342**, 758 761

Roeckner E, U Schlese, J Biercamp and P Loewe, 1987 Cloud optical depth feedbacks and climate modelling *Nature*, **329**, 138-140

Santer, B and T M L Wigley, 1990 Regional validation of means, variances and spatial patterns in general circulation model control runs *JGR*, **95**, 829 850

Sausen, R, K Barthel and K Hasselmann, 1988 Coupled ocean-atmosphere models with flux corrections *Climate Dynamics*, **2**, 154-163

Schlesinger, M E, 1983 A review of climate model simulations of CO_2 induced climatic change OSU Report 41, Climatic Research Institute and Department of Atmospheric Sciences

Schlesinger, M E and J F B Mitchell, 1987 Climate Model Simulations of the Equilibrium Climatic Response to Increased Carbon Dioxide *Reviews of Geophysics*, **25**, 760-798

Schlesinger, M E and E Roeckner 1988 Negative or positive cloud optical depth feedback *Nature*, **335** 303-304

Schlesinger, M E and Z C Zhao, 1989 Seasonal climatic change introduced by doubled CO_2 as simulated by the OSU atmospheric GCM/mixed-layer ocean model *J Climate*, **2**, 429-495

Semtner A J, Jr, 1974 An oceanic general circulation model with bottom topography Numerical Simulation of Weather and Climate, Technical Report No 9, Department of Meteorology, University of California, Los Angeles

Semtner A J, Jr, 1976 A model for the thermodynamic growth of sea ice in numerical investigations of climate *J Phys Oceanogr*, **6**, 379 389

Semtner A J, Jr and R M Chervin, 1988 A simulation of the global ocean circulation with resolved eddies *J Geophys Res*, **93**, 15,502 15,522

Shackleton, N J, 1985 Oceanic carbon isotope constraints on oxygen and carbon dioxide in the Cenozoic atmosphere "The carbon cycle and atmospheric CO_2, natural variations Archean to present *Geophysical monograph*, **32**, 412 417

Simmons A J and L Bengtsson 1984 Atmospheric general circulation models, their design and use for climate studies In *The Global Climate* (ed J Houghton) Cambridge University Press 233 pp

Sinitsyn V M, 1965 Ancient climates in Eurasia, Part 1 Palaeogene and Neogene Izd LGU, Leningrad(R)

Sinitsyn, V M 1967 Introduction to Palaeoclimatology Nedra,Leningrad (R)

Stouffer R J, S Manabe and K Bryan 1989 On the climate change induced by a gradual increase of atmospheric carbon dioxide *Nature*, **342**, 660-662

Untersteiner, N, 1984 The Cryosphere In *The Global Climate* (ed J Houghton), Cambridge University Press, 233 pp

v. Storch H and F Zwiers, 1988 Recurrence analysis of climate sensitivity experiments *Journal of Climate*, **1**, 157-171

Washington, W M and G A Meehl, 1984 Seasonal Cycle experiments on the climate sensitivity due to a doubling of CO_2 with an atmospheric GCM coupled to a simple mixed layer ocean model *J Geophys Res* **89** 9475 9503

Washington W M and G A Meehl, 1989 Climate sensitivity due to increased CO_2 experiments with a coupled atmosphere and ocean general circulation model *Climate Dynamics*, **4**, 1 38

Washington, W M and C L Parkinson, 1986 An introduction to three dimensional climate modeling University Science Books, 20 Edgehill Road, Mill Valley, CA 94941, 422pp

Wetherald, R T and S Manabe, 1986 An investigation of cloud cover change in response to thermal forcing *Climatic Change*, **10**, 11 42

Wetherald, R T and S Manabe, 1988 Cloud feedback processes in a general circulation model *J Atmos Sci*, **45**, 1397-1415

Wigley, T M L, 1989 Possible climatic change due to SO_2-derived cloud condensation nuclei *Nature*, **339**, 365 367

Wilson, C A and J F B Mitchell, 1987 Simulated climate and CO_2 induced climate change over western Europe *Climatic Change*, **10**, 11 42

Zwiers, F W, 1988 Aspects of the statistical analysis of climate experiments with multiple integrations MPI Report No 18, MPI fur Meteorologie, Hamburg.