CLIMATE CHANGE: AN INTEGRATED PERSPECTIVE

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Chapter 3

MODELLING OF THE CLIMATE SYSTEM

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3.1 Introduction

To better understand the earth's climate, climate models are constructed by expressing the physical laws, which govern climate mathematically, solving the resulting equations, and comparing the solutions with nature. Given the complexity of the climate, the mathematical model can only be solved under simplifying assumptions, which are a priori decisions about which physical processes are important. The objective is to obtain a mathematical model, which both reproduces the observed climate and can be used to project how the earth's climate will respond to changes in external conditions.

There are several factors, which must be taken into account. The earth and its enveloping atmosphere have a spherical geometry, the atmosphere and oceans are gravitationally attracted to the centre of the earth, the earth rotates on its axis once per day and revolves about the sun once per year, and the composition of the earth's atmosphere includes several radiatively active gases, which absorb and emit energy. All of these factors introduce effects on the climate, which may vary with longitude, latitude, altitude, and time of the day or season of the year. Additionally, some of these factors may feedback on other processes (see Chapter 2 and 4), making the climate system non-linear in the sense that feedbacks among diverse physical processes make it difficult to predict the collective response to the processes from their individual influences.

3.2 Simple climate modelling

For simplicity, existing climate models can be subdivided into two main categories: (1) General Circulation Models (GCMs), which incorporate three dimensional dynamics and all other processes (such as radiative transfer, sea-ice processes, etc.) as explicitly as possible, and (2) simple models in which a high degree of parameterisation of processes is used. Both types exist side by side and have been improved during the past several years. Both types of models have been able to benefit from each other: GCM-results provide insight into climate change processes, which allow useful parameterisations to be made for the simple models; simple models allow quick insight into large-scale processes on long time scales since they are computationally fast in comparison with GCMs. Moreover, due to computational limits, only simple climate models can currently be used to study interaction with processes on long time scales (e.g., the slow adjustment of the biosphere, and the glacial cycles).

There are several choices of simplifying assumptions, which may be applied. For example, one can integrate the mathematical equations either horizontally or vertically or both in order to simplify the system. In the simplest possible climate model, a single number is obtained to describe the entire climate system. In the two sections that follow, two simple climate modelling schemes; energy balance and radiative-convective balance, are described.

3.2.1 Energy balance climate models

In the case of an energy balance climate model, the fundamental laws, which are invoked, are conservation of total energy and total mass. No appreciable mass is assumed to escape from the top of the earth's atmosphere, and the earth and its atmosphere are assumed to be in thermal equilibrium with the space environment. These are robust assumptions, which can be validated by observations. It is also possible to assume that the energy flux is in equilibrium at the earth's surface, although this is not strictly true since there may be considerable heat storage in the ocean on millennial and shorter time scales. Such models have been used to determine the sensitivity of the earth's climate to variations in the solar radiation at the top of the atmosphere. Analytic solutions to the energy balance equations have been obtained in some classes of models.

The simplest possible model, a zero dimensional model in which the global average, time average fluxes at the top of the atmosphere are in balance, may be solved for the equilibrium temperature (Chapter 2). By assuming that the energy flux from the sun is a constant, and that the earth

conforms to the Stefan-Boltzman "black body" law for radiative emission, the energy balance may be written as:

$$\frac{S_0}{4}(1-\alpha)=\sigma T^4 \qquad (3.1)$$

where S_0 is the energy from the sun (solar constant), α is the planetary albedo which is the ratio of energy flux which is scattered to that which is absorbed, σ is the Stefan-Boltzman constant, and T is the effective temperature of the earth-atmosphere system. The factor of 4 on the left hand side represents the ratio of the surface area of the spherical earth (emitting surface) to the surface area of the circular disk of solar radiation intercepted by the earth (absorbing surface). Given a measurement for the solar constant (1,372 Wm^{-2}), the model may be solved for the effective temperature at the top of the atmosphere up to the parameter α. Measurements from space indicate that the earth's radiant temperature is 255 K and the albedo is 0.3 (Chapter 2).

This simple model can be used as a means to test the sensitivity of the earth's climate to changes in either the solar energy flux reaching the top of the atmosphere or the planetary albedo, which is a function of the cloud cover and the snow and ice cover at the surface. For example, a one percent change in the solar energy reaching the top of the atmosphere results in a 0.65 K change in the earth's effective temperature. In order to establish a quantitative relationship between the radiative energy flux at the top of the atmosphere and the climate near the surface, it is necessary to take into account the effects of the atmosphere, particularly its vertical structure, and the effects of surface conditions, particularly feedbacks associated with snow, ice and clouds (chapter 2).

Figure 3.1 shows the earth's radiation energy balance with the incoming solar energy flux normalised to 100 units (100 units = 343 W m^{-2} = 1,372/4 W m^{-2}). As may be seen in the figure, the solar energy is scattered to space by clouds or by the surface (28%), absorbed by the atmosphere (25%) or by the earth's surface (47%). In order to preserve the thermal equilibrium, the energy absorbed at the surface must be transported to the atmosphere, where it can be re-emitted to space. This is accomplished by surface radiative emission and sensible and latent heat transfers. The surface emits 391 W m^{-2}, primarily in the infrared portion of the electromagnetic spectrum, to the atmosphere, which absorbs 374 W m^{-2} and allows 17 W m^{-2} to pass into space. The atmosphere, in turn, emits 229 W m^{-2} to space and 329 W m^{-2} back to the earth's surface. This downward emission by clouds and radiatively active atmospheric gases is termed the "greenhouse effect" by analogy to a greenhouse, which glass walls permit solar radiation to pass through, but inhibit the transmission of infrared radiation from inside.

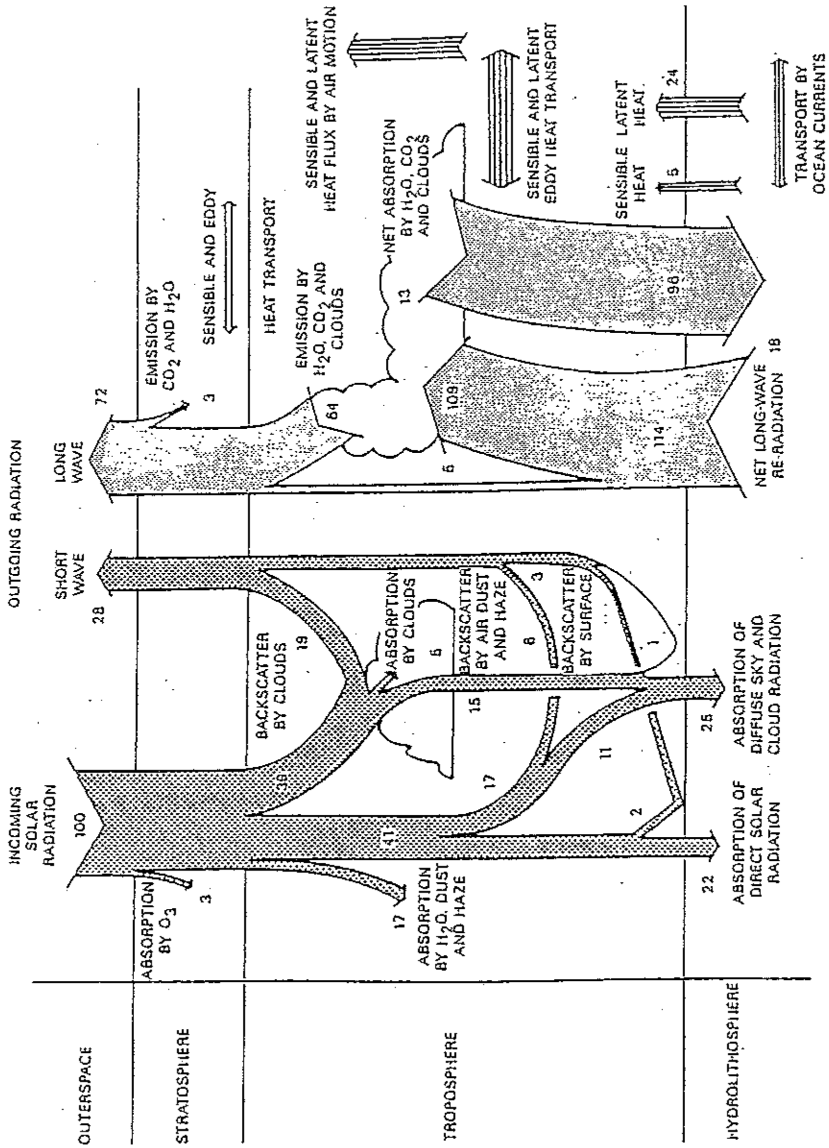

Figure 3.1 Schematic diagram of earth radiation budget components. Incoming solar energy normalised to 100 units. Adapted from "Understanding Climate Change", U.S. National Academy of Sciences, Washington, D.C., p. 14, 1975.

Thus, the surface energy balance is strongly influenced by the composition of the atmosphere, the amount of cloudiness, and the transport of water vapour (latent heat).

The effects of surface conditions can also profoundly influence the surface energy balance, primarily by the variations of the snow and ice cover at the surface and their feedback on the climate. Since snow and ice are bright, they contribute to the planetary albedo by scattering solar radiation back to space before it is absorbed. If snow or ice cover were to increase for some reason, then the scattering of solar radiation would increase, the planetary albedo would increase and it may be seen that the effective temperature of the earth would decrease. If that lower temperature at the top of the atmosphere were related to a similarly reduced surface temperature, then there would be a resultant increase in snow and ice, creating a positive feedback with the albedo effect.

3.2.2 Radiative-convective models

The second simplest climate model is one in which the effect of the vertical structure of the atmosphere is considered. Since the atmosphere is a fluid, the physical mechanism, which is absent in energy balance climate models, but present in a model with vertical structure, is the vertical motion of the air. The relevant forces in such a motion are the gravitational attraction of the atmosphere toward the centre of the earth and convection.

As was shown in the previous section, the atmosphere absorbs 86 Wm^{-2} of the solar energy and 374 Wm^{-2} of the terrestrial energy it receives, and it emits 229 Wm^{-2} to space and 329 Wm^{-2} back to the surface of the earth. The latter is referred to as the "greenhouse effect" and is primarily due to water vapour and clouds, with smaller contributions by other radiatively active gases, such as carbon dioxide, ozone and methane. The atmosphere is a net exporter of radiant energy at a rate of 98 Wm^{-2}. Therefore, there is a radiative cooling of the atmosphere with a corresponding radiative heating of the earth's surface.

When a fluid is heated from below and cooled internally, the result is convection (Chapter 2). Convection is the destabilisation of fluid stratification by heating and the resultant overturning circulation of the fluid to restore stable stratification. The overturning of the fluid may take place by large-scale circulation or by small-scale turbulent transfers of heat and water vapour. Given the radiative heating of the atmosphere from below by emission from the earth's surface and the radiative cooling of the atmosphere by emission to space and back to the earth, the earth's atmosphere is prone to convective overturning. The temperature of the atmosphere tends to have its maximum near the earth's surface and to decrease with altitude. The declining temperature of the atmosphere with height above the surface is called the lapse rate (Chapter 2). It is possible to determine from the lapse rate whether the atmosphere is stably, neutrally, or unstably stratified. It is also possible to construct a mathematical climate model on the basis of balancing the two atmospheric processes of radiative cooling and convection.

By assumption, the convective overturning of the atmosphere is assumed to be efficient, so that the equilibrium state of the atmosphere is a neutrally stable lapse rate. Convection dominates the lower portion of the atmosphere, called the troposphere, and radiation dominates the balance in the upper portion of the atmosphere, called the stratosphere. A radiative-convective climate model, then, is one in which a radiative balance is assumed in the stratosphere, a convectively neutral lapse rate is assumed in the troposphere and the surface temperature may then be determined. The radiative equilibrium may be quite complicated, due to the diversity of absorbing and emitting radiative gases.

The most important advantage that radiative-convective models have over energy balance climate models is that they can be used to quantify the cloud albedo feedback mechanism under various assumptions about cloud formation. The climate sensitivity to variations in cloudiness may then be examined critically using such models.

3.3 General Circulation Models (GCMs)

3.3.1 Introduction

Climate models may be organised into a hierarchy, based on the complexity of the models, which also bears upon the simplifying assumptions, which must be made. The simplest model is the zero dimensional energy balance model described in Chapter 2 and Section 3.2.1. Next in the hierarchy are one and two dimensional energy balance models (Section 3.2.1) in which the atmosphere is treated as a single layer, and the one and two dimensional radiative-convective models (Section 3.2.2) in which deviations from the global or zonal area mean are neglected, but vertical structure within the atmosphere is considered. At the top of the hierarchy are three dimensional general circulation models (GCM). A GCM is a model in which all horizontal and vertical motions on scales larger than a chosen "resolved" scale are included (see Section 3.3.2). Motions, which take place on scales smaller than the resolved scale, are represented parametrically in terms of the large-scale climate variables. Parametric representation (or parameterisation) involves devising a set of mathematical

rules, which relate phenomena occurring on unresolved scales to the large-scale variables that are computed directly. In general, such parameterisations are based on a combination of empirical (i.e., drawn from observations) and theoretical studies. Also included are the effects of radiative heating and cooling, convective overturning (both in the resolved large scales and in the unresolved or parameterised scales), thermodynamic conversions of water vapour to liquid and back, and surface effects associated with surface ice, snow, vegetation, and soil.

General circulation models are used in place of energy balance models or radiative-convective models, when the horizontal and vertical structures or transient nature of the atmosphere are important considerations. Energy balance models can yield valuable insights into climate sensitivity and different feedback processes (Chapter 2) can be investigated very easily. However, the effects of clouds, aerosols, vertical heat transport, meridional heat transport and momentum transports can not be modelled adequately using energy balance models.

The starting point for a GCM is the set of governing laws. The laws of conservation of energy and mass are postulated, as is Newton's law (changes in momentum are related to the sum of external forces acting on a body), which applies with the slightly more restrictive assumption that all motions are hydrostatic (defined below). Newton's law for fluids is expressed mathematically in what are called the Navier-Stokes equations. With the hydrostatic assumption, changes in density are related to changes in pressure, and the downward gravitational force is balanced by the upward pressure gradient force, regardless of the motion of the fluid. The hydrostatic approximation was developed to filter sound waves, which have no importance on climate time and space scales. Mathematical equations may be written, which describe the conservation of atmospheric mass (also called the continuity equation), the conservation of energy (expressed by the first law of thermodynamics), and the changes in momentum, due to external forces, which include gravity, the pressure gradient force caused by differences in pressure from place to place, and the Coriolis force (Section 2.3.1). This set of equations, called the primitive equations of motion, is a set of non-linear, partial differential equations that have been known for centuries.

The spatial and temporal derivatives in the resulting equations, which are continuous in nature, are then approximated by discrete forms, which are suitable for a numerical treatment. The discrete equations are algebraic and may be solved by computer to determine the three dimensional distribution of temperature and winds. While various discrete forms of the primitive equations have been known for some time, only since the 1960s have computational resources become available to make their solution feasible. In addition, since the temporal dimension is also treated discretely and numerically, it is possible to solve the equations for their time dependent part, so that such processes as the annual cycle associated with the revolution of the earth about the sun, the inter-annual variation of climate, and the slow response of the climate to changes in external forcing, such as the Milankovitch orbital changes (Chapter 2) or the composition of the atmosphere, may be examined. The techniques of discretisation and numerical solution were originally developed for the problem of weather prediction. The first such successful application was attempted in the 1950s with a one layer atmospheric model.

3.3.2 Basic characteristics

Space and time are represented as continuous in the Navier-Stokes and the primitive equations. In order to allow solutions to be computed, space and time in the model world are each represented by discrete sets of points. The distance between these points defines the *resolution* of the model; high resolution represents the fields in finer detail, while low resolution can capture only the largest scale spatial or temporal structures. Those structures that can be seen at the given resolution are the *resolved scales*, while those structures that are too small to be seen are the *unresolved scales*.

The stable stratification in the atmosphere and oceans allows one to consider each as series of fluid layers, among which there is very little interaction. The representation of vertical derivatives selected, depends upon the problem being considered, but is typically effected by means of finite differences between layers or levels which are pre-selected. The choice of a co-ordinate to represent the vertical structure of the atmosphere or ocean can be complicated, because of the substantial irregularity of the earth's surface and ocean bathymetry. The hydrostatic approximation suggests that the most natural atmospheric vertical co-ordinate would be pressure, but the very steep topography at many places on the earth's surface make this a poor choice, since co-ordinate surfaces of some constant pressure are pierced by mountains. A more successful choice for the vertical co-ordinate is the s co-ordinate, which is pressure normalised by its value at the earth's surface. Ocean models make use of either distance from the sea surface (Z co-ordinate) or density (isopycnal or S co-ordinate) as the vertical co-ordinate.

The horizontal discretisation may be effected in a number of ways. The simplest formulation is an application of finite difference approximations to the continuous derivatives in both the longitudinal and latitudinal directions. Finite element approximations, which are useful in the vicinity of irregular boundaries, have been applied in limited domain models, as well as in sub-domains of global GCMs in order to more accurately simulate the

atmospheric flow over and around mountains. Another class of discretisation techniques is the set of spectral methods in which the basic variables (temperature, moisture, wind speed, etc.) are expressed as series expansions in ortho-normal basis functions.

The temporal discretisation is typically effected by means of finite differences, but a complication arises because the atmosphere and oceans, being fluids, are capable of supporting waves. The complication is due to the fact that the speed, at which the phase of atmospheric and oceanic waves propagates, must be accurately resolved. This means that the time resolution (time step) and spatial resolutions are related.

Typically, the time step is chosen to be as large as possible, without causing the computational solution to become unstable, due to inaccuracies in representing the propagation speed of waves. A finer spatial resolution requires a proportionally smaller time step. For example, doubling the spatial resolution reduces the time step by a factor of two, so that doubled resolution in each direction increases the number of computations by a factor of 16 and the storage requirement by a factor of eight. An order of magnitude increase in computational resources supports only a modest increase in model resolution.

Once the continuous differential equations are transformed to a discrete set of algebraic equations, they may be solved computationally if boundary conditions and initial conditions are specified. Boundary conditions establish the values of model variables at the edges of the model domain. For example, since fluid cannot flow through a solid wall, the velocity component, normal to the earth's surface or the coasts and bottom of the oceans, is specified to be zero. Given the values of the variables at a specific moment in time and all points in the model domain, the values of the variables can be advanced one step. This process, known as time marching, is repeated time step by time step until the desired length of climate simulation is obtained. To start time marching, a set of initial conditions must be specified.

The specification of boundary conditions at the ocean-atmosphere interface is of particular importance. The appropriate air-sea boundary conditions to be specified for the oceanic GCM, as part of a coupled climate model, are the wind stress in the zonal and meridional directions, the net heat flux and the net fresh water flux. The ocean circulation is in large measure a response to these fluxes. Additionally, the flow and properties of water entering the ocean from land areas (i.e. river flow) need to be specified, as well as the heat flow across the solid boundaries (i.e. geothermal heating), is usually taken to be zero.

The result of the computational solution of the discrete primitive equations is a simulation of the three dimensional structure of the earth's atmosphere or oceans. Such a simulation should be capable of reproducing the observed characteristics of climate, including the global mean vertical structure of temperature and humidity, the zonal mean structure of the pressure and wind fields with subtropical jets near the top of the tropopause, and the longitudinally varying distribution of pressure, temperature and humidity, which is observed. In addition to the mean fields, the simulation should realistically represent the temporal variability of the main features, such as the degree to which the Polar front meanders, the seasonal shifts of major pressure belts, such as the subtropical highs, the annual cycle of tropical features, such as the Inter-tropical Convergence Zone and the Asian monsoon, and the progression of weather systems (Chapter 2). A reasonable ocean simulation reproduces the observed mean distributions of temperature and salinity, as well as the major currents. Important inter-annual variations such as El Niño (Sections 3.3.5 and 3.3.7) and the decadal oscillation of the North Atlantic should also be simulated.

3.3.3 Climate sensitivity

General circulation models are extremely useful tools for studying climate sensitivity. The procedure to study climate sensitivity is quite straightforward. First, the climate model is integrated to simulate the current climate. This simulation is referred to as the *control* run. Once a satisfactory simulation of the current climate has been obtained, an input parameter to the model is changed for the desired sensitivity experiment and the model is integrated again. This integration is referred to as the *experiment* run. The difference between the two model simulations (*experiment* minus *control*) is referred to as the model response (or sensitivity) to the particular parameter that was changed.

During the past 20 years, about 30 climate modelling groups in the world have conducted hundreds of climate sensitivity experiments using atmospheric GCMs. These numerical experiments are carried out to test a certain hypothesis about climate sensitivity, something that can not be done by analysing the past data alone. Although a detailed description of these experiments is beyond the scope of this chapter, they can be classified into the following broad categories:

a) Sensitivity to boundary conditions at the earth's surface:
A control integration is made with one set of values of sea surface temperature (SST), soil moisture, vegetation, albedo, snow cover, sea ice, height of mountains, etc. The integration is then repeated by changing one or more of the boundary conditions and the difference between the two simulations is interpreted as the effect of the change in that particular boundary condition. The similarity between the observed and model simulated anomaly patterns validates not only the hypothesis but also the model.

Sensitivity experiments in a manner described above have been carried out by removing all the mountains (mountain - no mountain experiment); removing all the land masses (aqua-planet experiment); changing the configuration of the land masses (paleo-climatic experiments); replacing the forests by grass (deforestation experiment); expanding the deserts (desertification experiment); changing the extent and depth of snow and sea ice (snow, sea ice experiment) and changing the wetness of the ground (soil moisture experiment).

b) Sensitivity to changes in the chemical composition of the atmosphere:
Of particular interest are the experiments to study the sensitivity of the earth's climate to changes in the concentration of greenhouse gases (see Sections 3.3.8 and 3.3.9). Sensitivity experiments have also been carried out to study the impact of a large number of nuclear explosions (nuclear winter experiment) and effects of volcanic eruptions.

c) Sensitivity to changes in physical parameterisations and numerical techniques:
The examples include sensitivity to parameterisations of convection, liquid water and ice crystals, cloudiness, radiation formulation, boundary layer schemes, surface roughness, land-surface processes, vertical mixing of momentum, heat, salt and water in the oceans and atmosphere, and numerical formulations for solving the mathematical equations that describe the climate model.

One of the major limitations of climate sensitivity studies is that the model simulated control climate, in many instances, has large errors, compared to the observed current climate, and therefore, the simulated anomalies (*experiment - control*) might be erroneous. It is generally assumed that deficiencies of the model and the control climate cancel in the subtraction of the *experiment* and *control*. However, this may not always be the case. Generally, it is up to the researcher to decide if a particular model, being used for a particular climate sensitivity study, is good enough to test the particular hypothesis.

3.3.4 Atmospheric modelling

In addition to the issues raised in Section 3.3.2, regarding transforming the continuous primitive equations to a discrete set of algebraic equations, there are several other problems, which must be addressed in modelling the

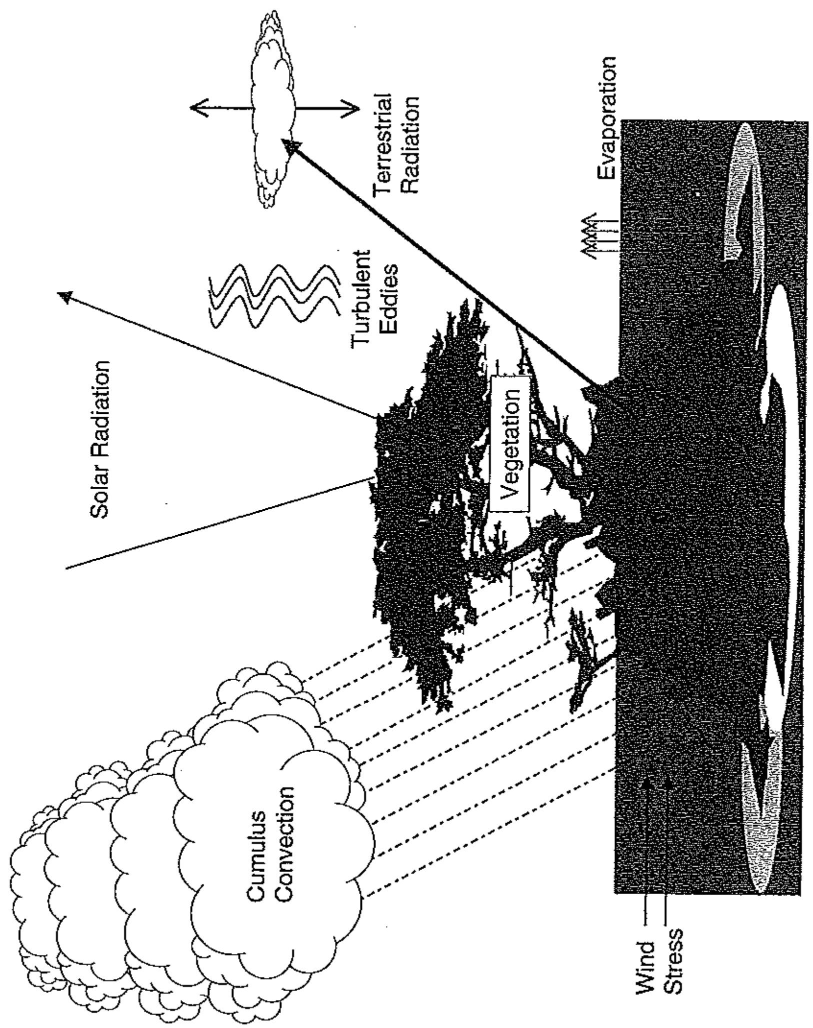

Figure 3.2 Schematic diagram of sub-grid scale processes, which must be parameterised in General Circulation Models

earth's atmosphere. These form a class of problems of parametrically representing the physical processes that act on scales smaller than the resolved scale. Such representations are referred to as sub-grid scale parameterisations. A schematic showing the various processes is given in Figure 3.2. In most cases, these phenomena involve small-scale structures or processes, which collective effect on the large-scale variables may not be simply related to those variables. For example, transports of heat or momentum are accomplished by turbulent wave motions, also known as eddies, at very different rates, depending on whether the large-scale flow is stably or unstably stratified. As a result, some assumptions, referred to as closure assumptions, must be made to establish the relationship between the sub-grid scale processes and the large-scale variables. The relationships so constructed may then be validated using observational data and incorporated into atmospheric GCMs (AGCMs). The most important sub-grid scale parameterisations - radiation, convection and the planetary boundary layer - are described below.

Radiation

In order to represent the energy balance at the earth's surface and to accurately simulate the heating by solar energy and the cooling by infrared emission to space, it is necessary to parameterise the absorption and emission of radiation in the atmosphere. The wide difference in temperature between the photosphere of the sun and the surface of the earth means that the radiation from the two bodies, if assumed to conform to the black body law, is in completely distinct frequency bands of the electromagnetic spectrum. The atmosphere is nearly transparent at the ultraviolet and visible frequencies, at which solar energy is emitted, and nearly opaque to the infrared radiation band in which the earth's surface emits. The typical AGCM radiative transfer model embodies three characteristics of the atmosphere: (1) the atmosphere is nearly transparent to solar radiation, (2) the atmosphere is nearly opaque to terrestrial radiation, and (3) the radiative flux may be characterised by a two stream approximation, in which all radiation propagates in either an upward or downward direction. The degree to which a column of air is cloudy, considerably complicates this model of radiative transfer, and, therefore, a GCM must include some model of sub-grid scale cloud formation in order to accurately take account of the effects of clouds. Cloud models based on relative humidity, upward air velocity, and other large-scale variables have been introduced into existing GCMs. More explicit cloud formation models, in which cloud condensation nuclei and droplet aggregation are included to simulate cloud liquid water content, have been formulated and are being incorporated into climate models at the present time.

Planetary Boundary Layer

Transfers of heat, moisture and momentum within the atmosphere, and exchanges of these quantities between the atmosphere and the underlying surface (land, ocean or ice), can be carried out by two types of motion. The "resolved" scales of motion are those, which are explicitly treated by the atmospheric model, while the "sub-grid" scale motions are those, which have a spatial scale too small to be resolved, but because of their importance must be "parameterised" in terms of the variables describing the resolved flow. In the context of a GCM the sub-grid scale motions are considered to be turbulence and are especially important in the layer just adjacent to the earth's surface, termed the "planetary boundary layer" (PBL). In this layer, in which friction plays a major role in the balance of forces, vertical turbulent transfers are highly significant.

There are currently two predominant classes of parameterisations of the PBL turbulent mixing. "Turbulence Closure" modelling is a general mathematical approach, in which the system of equations, describing the very small scale flow, is formulated in a non-convergent series of statistical moments. The system may then be closed by a set of consistent approximations. This approach is actually applied throughout the model atmosphere and allows the GCM to internally generate its own PBL, with a depth that varies in response to both the forcing from the surface and the influence of the resolved flow. In contrast, "Mixed Layer Theory" treats most of the PBL as a single layer. The basic idea here is that since heat and moisture are so well mixed in this layer, the transition between the top of the PBL and the atmosphere above is quite discrete, often accompanied by a discontinuity in temperature and moisture. The profiles of heat, moisture and momentum within the PBL, their fluxes into the free atmosphere and even the depth of the PBL itself can be approximated as functions of the resolved variables.

Convection

One of the very significant sources of heating in the interior of the atmosphere is the convection that takes place in many areas of the tropics. This "cumulus" convection is quite extensive in the vertical domain, often reaching to the top of the troposphere, and yet it occurs on horizontal scales of only 1 to 10 kilometres, far too small to be resolved (explicitly treated) by the atmospheric GCM (The word "cumulus" refers to clouds with a clumped structure). The transfer of heat, moisture and momentum by turbulence within the cumulus clouds is a very complex problem, analogous to the problem of parameterising the PBL discussed in the above section.

The approaches to this problem vary widely. The "Moist Convective Adjustment" scheme eliminates the gravitational instability of moist air by

suitably adjusting the temperature and moisture in the vertical, whenever the instability occurs. The "CISK" class of parameterisations uses the horizontal convergence of moisture within the lowest layers of the atmosphere to determine when convection will occur. A parameterisation for the partition of sensible vs. latent heating (that is, heating, which increases the temperature vs. heating, which evaporates liquid water) is generally incorporated into this type of scheme, as is a treatment of the heat exchange between clouds and the environment. Yet a third class of parameterisation, the "Cloud-Buoyancy" or "Mass-Flux" schemes, attempt to explicitly model an ensemble of cumulus clouds. The basic kinematics of the clouds, including the "entrainment" (bringing in) of air near the cloud base, and the "detrainment" (letting air out) near the cloud top are treated. The implementations of this type of scheme vary widely in the degree to which phenomenology is used to simplify the complex physics. Atmospheric models appear to be sensitive to the details of these schemes.

Current state of atmospheric modelling

Among the components of climate models, AGCMs are probably the best verified subsystem models. The AGCM', which produce daily weather forecasts, are subjected to a prediction-analysis-verification cycle with well defined analysis techniques. The weather prediction models have much higher resolution than AGCMs used for climate modelling, because of the requirement for climate models to make much longer simulations. It is known that higher resolution improves the skill of weather predictions. The realism of the precipitation distribution appears to improve markedly with increasing resolution at climate time scales.

The prediction of clouds and the representation of their effect on the short and long wave radiation fields is both empirical and very crude in AGCMs. The cloudiness problem is important for global warming, because the change in the amount of solar radiation reflected to space, due to a few percent change in cloudiness, could compete with the CO_2 induced greenhouse effect. The spatial and temporal inhomogeneity of cloudiness makes the parameterisation of cloud effects for the large scale very difficult.

Dust particles in the atmosphere and their influence on the hydrological cycle, as condensation nuclei and on the radiation budget as reflectors and absorbers of solar radiation, have not been satisfactorily included in atmospheric models.

The representation of processes that maintain the water vapour distribution in the upper troposphere in certain geographical regions is inadequate. Since upper tropospheric water vapour provides the major positive feedback to the greenhouse effect, it is important that the dynamics and physics of the processes that maintain this field be correct. Numerical problems with water vapour are a main reason for the development of a new class of AGCM, just beginning to be used for climate simulations, the "semi-Lagrangian model".

3.3.5 Ocean modelling

Ocean models have varying degrees of complexity, analogous to the climate model hierarchy discussed in Section 3.3.1. The simplest ocean models treat the ocean as a motionless slab of water of fixed depth that stores heat uniformly throughout its depth. This simplest ocean model is known as a *slab mixed layer ocean*. This kind of ocean model, used as the oceanic component of a climate model, can produce a reasonable representation of the amplitude and phase of the annual cycle of SST in much of the extra-tropical ocean, when the depth of the slab is taken to be about 50m. A somewhat more realistic ocean model is the *mixed layer ocean*. This model still takes the ocean to be a motionless slab, but the depth of the slab is calculated internally, rather than specified, and the temperature can be a function of depth in the slab. These additional characteristics of the mixed layer are based on formulae, developed by extrapolation from measurements or deduced from intuitively plausible assumptions. The mixed layer ocean can respond more realistically to atmospheric forcing by storms, for example, than the slab mixed layer ocean.

The mixed layer models address the thermal interchanges between the upper ocean and atmosphere. However, in reality ocean currents transport heat both horizontally and vertically. When these transports are neglected, the climate model, using a mixed layer ocean model, will produce large errors. One enhancement to the mixed layer model that has been used in climate sensitivity experiments (Section 3.3.8) and transient climate experiments (Section 3.3.9) is to include specified heat fluxes by the currents in the mixed layer or slab mixed layer ocean. Inclusion of specified heat fluxes improves simulation of the observed annual mean and annual cycle of SST, but the resulting climate model will not be able to correctly simulate or predict many types of coupled climate variability, such as the large inter-annual variability of SST in the Eastern tropical Pacific (part of the El Niño/Southern Oscillation or ENSO phenomenon) or variability associated with the ocean's thermohaline circulation. When the mixed layer ocean model is used for a transient climate experiment, it should be kept in mind that the experiment is simulating the future climate, assuming that the heat transports by ocean currents do not change in response to the changing forcing.

Ocean models that simulate the time evolution of the ocean currents and the physical properties of the water contained in those currents can be

termed *ocean general circulation models (OGCM)*. The OGCM begins with the discretised Newton's laws for fluids (the Navier-Stokes equations), the equation of state, which gives the density as a function of thermodynamic variables, and the mathematical statement of conservation of energy, just as the atmospheric GCM (AGCM) discussed in Section 3.3.4. The OGCM differs from the AGCM in two important respects. First, the density of sea water depends on salinity (the concentration of dissolved salts), as well as temperature, whereas the density of air depends primarily on temperature and water vapour concentration (humidity). Therefore the OGCM contains budget equations that determine salinity changes, while the AGCM calculates humidity changes. The other major difference is that the continents divide the ocean into basins, which communicate through narrow passages, whereas the atmosphere has no such boundaries. Then representation of topography is a more serious issue in the OGCM and has a great influence on the technical aspects of OGCM design.

As noted above, the physical laws have been known for hundreds of years. However, computational resources that allow models of the global ocean circulation to be developed, which can be considered potentially realistic, have only recently become available Current computers allow climate simulations using OGCMs with horizontal resolution of about 4° in the longitudinal and latitudinal directions and less than 20 levels in the vertical direction. Features with horizontal scales less than 1000 km are unresolved with this grid structure and must be parameterised if they contribute significantly to the budget equations on the resolved scales.

The parameterisations commonly used in OGCMs are vertical and horizontal diffusion of momentum, heat, and salt. These parameterisations are based on theories of turbulence that have not been verified over most of the parameter range at which they are applied. In the simplest formulations, vertical diffusion coefficients are specified and sometimes enhanced near the upper surface of the ocean, to represent the turbulent mixed layer forced by air-sea interactions. Horizontal diffusion is used primarily to damp small-scale motions, so that the resolved scale motions look smooth. Horizontal diffusion schemes are employed for numerical reasons and have little physical basis. Another parameterisation is used to obtain the vertical structure of absorption of solar radiation below the sea surface.

The typical horizontal scale of oceanic storms (meso-scale eddies, discussed in Chapter 2) is small compared to the OGCM resolution. Parameterisation of the effect of the oceanic meso-scale eddies on the resolved scales is a serious research question for the ocean, until computers are sufficiently powerful to resolve the eddies in climate simulations, whereas atmospheric storms are resolved by the coarse resolution AGCMs, used for climate simulation and their effects need not be parameterised.

Whereas the AGCM can be viewed as driven from below by SST, the OGCM is driven from its upper boundary by wind stress, which transfers momentum from the atmosphere to the ocean, heat flux, which adds heat, and fresh water flux, which affects the salinity. Conceptually, the response of the OGCM to the wind stress forcing is referred to as *the wind driven circulation* and the collective response to the heat and fresh water flux is referred to as the *thermohaline circulation*. The existence of the thermohaline circulation presents significant technical obstacles to climate simulation, since the time scales for climate variability, associated with the thermohaline circulation, can be thousands of years. The OGCM may require thousands of years of simulation to achieve equilibrium to a change in the surface forcing, whereas the time scale for the AGCM to approach equilibrium is on the order of months.

Developmental research for OGCMs is active in the areas of improvement of numerical methods for the resolved scales and improvement of the parameterisation schemes. Advances in computing have allowed the resolution of OGCMs to be increased to the point of resolving the oceanic meso-scale eddies *(eddy-resolving OGCM)*. However, the eddy resolving OGCM is not practical for climate modelling, because of the limitations of current computing technology.

3.3.6 Modelling other subsystems.

While the models of the atmosphere and ocean are the basic building blocks of the climate model, and the climate GCM in particular, there are several other subsystems with potentially important roles in determining climate. Models of various levels of sophistication have been developed for these subsystems. These subsystem models are described below.

- **Land:**

  A model of the land surface is required to calculate land surface temperature, evaporation, snow cover, and rainfall runoff, quantities of obvious importance for human society. Until recently, land surface models used in GCM climate modelling have been simple but crude "bucket models." As the name implies, the bucket model has a certain capacity for storing water. When precipitation exceeds evaporation sufficiently, so that water accumulates beyond that capacity, the bucket overflows into runoff. Evaporation is taken to be some fraction of potential evaporation, where the fraction is proportional to the percentage of capacity that the bucket is filled.

  Recently, more sophisticated land surface biosphere models have been developed that include the influence of local soil and plant properties on the evaporation and water holding capacity. While constructed more realistically, these biosphere models introduce a very large number of new parameters into the model that either have not or cannot be measured. An additional issue with the biosphere models is the manner in which parameterisation of land surface heterogeneity on the climate model scale is handled. If climate changes, the vegetation distribution will of course also change, which may in turn feed back on the climate. Empirical models that predict vegetation distribution as a function of climate parameters are being developed to study this issue.

- **Hydrology:**

  More rain falls on land than re-evaporates to the atmosphere. The excess is either stored in lakes or underground, or flows to the ocean in rivers. The fresh water river flow that reaches the ocean affects the ocean salinity, and hence the oceanic thermohaline circulation. Little is known about the influence of this process on climate variability. A hydrology model is needed to partition the runoff from the land surface into the various components and can be extremely complex. The hydrology models used in global coupled GCMs are very crude, basically assigning the runoff at each point over land to a river outlet into the ocean, instantaneously transporting the runoff to that point, and freshening the sea water at that point.

- **Sea ice:**

  Energy balance climate models demonstrate the potential importance of sea ice feedbacks to climate change. Formation and melting of sea ice is in large part a thermodynamic process. However, simulation of the motion of the ice under the joint influence of the ocean currents and the wind has been found to be a necessary ingredient for realistic simulation of the seasonal variations of the sea ice extent. Some first generation thermodynamic/dynamic models of sea ice exist and are in the process of being verified and included in climate GCMs.

- **Atmospheric chemistry:**

  Realistic simulation of the coupled interactions of atmospheric dynamics, photochemistry, and transport of trace species is important for understanding phenomena such as the "ozone hole" and "acid rain." The chemistry model locally calculates sources and sinks of various species from a system of chemical and photochemical reactions.

The number of reactions considered can be very large. The atmospheric winds carry these species from place to place, so that an additional budget equation must be added to the AGCM for each species for which the sources and sinks are not in local equilibrium. The coupled AGCM/chemistry model can be many times more expensive than the AGCM by itself. Coupled AGCM/chemistry models exist for study of stratospheric ozone, but are not yet used in coupled climate GCMs.

- **Aerosols:**

  Both naturally and anthropogenically produced aerosols have potentially important effects on climate and climate change. Depending on particle size, these aerosols can scatter or absorb both solar and long wave radiation, altering the radiation balance of the climate system. Volcanically produced particles appear to have a substantial effect on global climate for several seasons after they are produced, by reducing solar radiation reaching the ground. Dust raised by Saharan dust storms can be carried for thousands of miles. There is indirect evidence that anthropogenically produced sulphate-aerosols are reducing global warming from increasing CO_2-concentrations. Simulation of these effects requires determination of the sources and radiative properties of the particles, and calculation of the transport by winds, fallout, and removal by precipitation processes. Some early experiments have been conducted with GCMs, simulating the climatic consequences of volcanic eruptions and "nuclear winter."

- **Glaciers:**

  Formation and melting of glaciers can have catastrophic climatic consequences. Ice sheets, several kilometres thick, covered much of North America and Eurasia during the last Ice Age. Melting of the Antarctic ice sheet could inundate many of the world's major cities. Some coupled climate GCMs incorporate simple models for glacial accumulation, flow and melting.

3.3.7 Choices in the philosophy and design of GCMs.

An enormous range of phenomena are encompassed under the umbrella of the "general circulation", which can be defined as the set of circulations involving time scales of a few days and longer, and spatial scales of order about a thousand kilometres and longer. If we think of a hierarchy of time scales, three broad categories emerge. On the seasonal time scale, atmospheric "blocking" (the persistence for many days of anomalous mid-latitude high pressure systems), and the timing and intensity of the

Indian/Asian monsoons are examples of phenomena which affect the climate. On the seasonal-to-inter-annual time scale we consider the tropical oscillation, known as El Niño and the Southern Oscillation (ENSO), in which the entire tropical Pacific atmosphere-ocean system undergoes dramatic shifts, as the intense convection (normally situated in the Western Pacific) moves towards the centre of the tropical Pacific basin (see Section 3.3.5). These "warm episodes" of ENSO occur irregularly with a period of about 2 - 4 years. In addition, the year-to-year changes in the Indian/Asian monsoon are both significant and are thought to be related to ENSO. On decadal to century time scales, the atmospheric response to increases in greenhouse gases and the problem of the formation and circulation of the deep water of the oceans (the thermohaline problem of Section 2.3.2) are of great interest. In addition, the slow increase in the extent of the African desert (in the Sahel region) is a major problem with societal implications, which takes place on these time scales. On even longer time scales (centuries), the glacial cycles manifest themselves, as discussed in Sections 2.4 and 2.5.

Each of these categories has very different modelling requirements. The atmospheric behaviour on the seasonal time scale can be understood in terms of fixed oceanic conditions (mainly sea surface temperature), while it is necessary to include a fairly complete spectrum of atmospheric motions. It is also necessary to include the interactions of the land surface with the atmosphere, particularly for the summer season. These interactions include the storage of precipitated water in the soil and the subsequent evaporation of this water back into the atmosphere, and play a key role in the hydrological cycle in summer over land. Since they are mediated to a large extent by vegetation (biosphere), the GCM used to study these problems should have a biosphere component. An atmosphere-biosphere GCM, utilising a moderate horizontal resolution with fixed oceanic boundary conditions, is appropriate here (a moderate horizontal resolution in this context consist of 42 or more global wave-numbers retained, corresponding roughly to a minimum grid resolution of 3 degrees in both latitude and longitude).

The simulation of the seasonal to inter-annual time scale (ENSO) must involve the upper layers of the ocean, in which the temperature and salinity are fairly well mixed. This is because the changes in the atmospheric circulation in the tropical Pacific that characterise ENSO are coupled to changes in the mixed layer of the ocean. The movement of the intense convection from Western to Central Pacific is in the short term caused by the extension of the very warm tropical sea surface temperatures normally in the Western Pacific to the east (where the sea surface temperature is normally much colder). However, considering the entirety of the ENSO oscillation, neither the atmosphere nor the ocean are the causative factors - they are coupled together. Thus what is needed is an atmosphere - ocean coupled GCM (CGCM), which includes at least the mixed layer of the oceans. If only the tropical atmospheric component of ENSO is to be simulated, the atmospheric component of the model can be quite simple and does not need to include many of the refinements of a full GCM. On the other hand, the simulation of the relationships between the Indian monsoon and ENSO, or the mid-latitude response to the warm episodes require a full atmosphere-biosphere GCM coupled to a mixed layer ocean model.

As the time scale of interest gets longer, increasingly deep layers of the ocean come into play. This is because the thermohaline (deep ocean) circulation becomes important on time scales of decades or longer, and affects the sea surface temperature on these time scales (Section 2.3.2). Since the problem of desertification (increase in desert extent) is thought to be critically linked to sea surface temperature and local land interactions on decadal time scales, the full ocean circulation should be taken into account. The problem of the response of the atmosphere to increases in greenhouse gases also cannot be studied without reference to the deep ocean, since it can store vast amounts of these gases as well as heat. Thus, the simulation of these time scales requires full, coupled oceanic and atmospheric GCMs.

Finally, on the time scales of centuries, in which the glacial cycles dominate, the dynamics of ice in its various forms (that is, the "cryosphere") become as important as the changes in state of the ocean and atmosphere, and coupled atmosphere-ocean-cryosphere models must be used (Section 2.5).

3.3.8 Equilibrium experiments

When the atmospheric composition is fixed and the incoming solar radiation is constant, climate model simulations eventually approach an *equilibrium*, in the sense that the annual mean surface air temperature does not systematically warm or cool, and similarly the precipitation does not show a systematic change of the same sign from year to year. The researcher can perform experiments changing the solar forcing, atmospheric composition, or some other aspect of the model such as resolution or a parameterisation scheme, and calculate the model equilibrium for each case. Then the change in the equilibrium climate gives the model *sensitivity* with respect to the change in the model.

There are some complications to the concept of equilibrium climate. Many simple climate models possess *multiple equilibria* for the same solar forcing, atmospheric composition, and model parameters. Typically, one equilibrium is close to the current climate, and the earth is ice covered in

another. The ice covered earth reflects much more of the incident solar radiation to space, leading to the cold surface temperatures and allowing the ice to persist. The climate predicted by the model that possesses multiple equilibria can then be very different for different choices of initial conditions. If the initial state is chosen to be ice covered, it will remain ice covered in equilibrium, while an ice free initial state may lead to the warm climate state. The existence of multiple equilibria is intuitively plausible. One coupled GCM is known to possess two equilibrium states (Manabe and Stouffer, 1988). These two states are both close to the current climate. One has a cold North Atlantic SST and weak thermohaline circulation, while the other has a more realistic warm North Atlantic and stronger thermohaline circulation.

An important class of experiments, the *greenhouse sensitivity* experiment, investigates the sensitivity to changing the concentration of CO_2 in the atmosphere. The typical greenhouse sensitivity experiment compares the equilibrium climate with current CO_2-concentration to the equilibrium climate with double the current CO_2-concentration. There are many possible variations on this theme. The greenhouse sensitivity experiments have been performed with all classes of climate models, from the simplest to the GCM. Most of the GCM greenhouse sensitivity experiments have used a slab mixed layer ocean (Section 3.3.5). In this case, the equilibrium climate is achieved in a few decades of model simulated time. When the oceanic component of the climate model is an OGCM, achieving equilibrium can take thousands of years of model simulated time. The slab mixed layer ocean consequently allows a much more comprehensive investigation of the model climate sensitivity to various changes, than does the OGCM. Of course, the results using the slab mixed layer ocean may differ from those obtained with the OGCM, but this is the sort of trade-off between realism and expediency that must be made because of the cumbersomeness of GCMs.

A detailed discussion of the results of greenhouse sensitivity experiments, as carried out with twenty different models, is given by Mitchell *et al.*, 1990. In all of the experiments the doubled CO_2 climate was warmer than the *control* (current CO_2-concentration) climate, as measured by the global mean surface temperature. The sensitivity of global mean surface temperature to a doubling of CO_2 ranges from 1.9°C to 5.2°C. All of the experiments also found an increase in the global mean precipitation, with warmer climates receiving more precipitation. These results are summarised in Figure 3.3 (after Mitchell *et al.*, 1990).

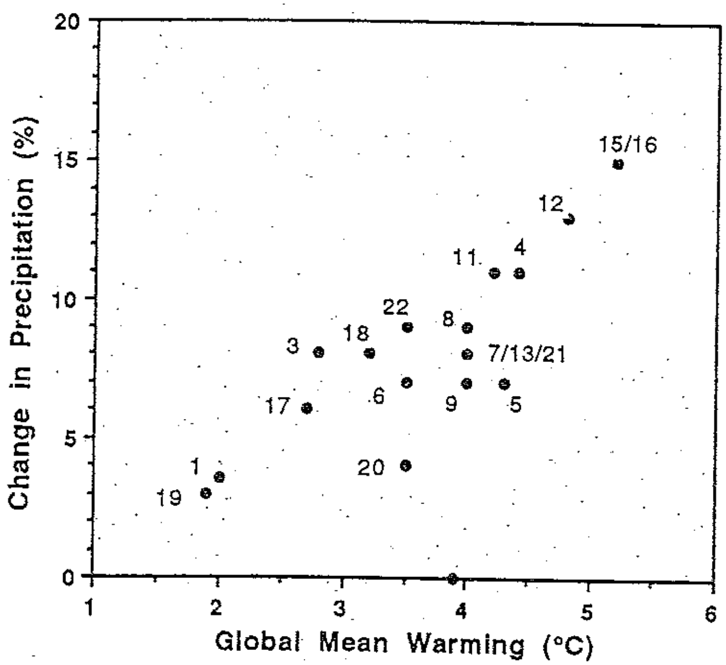

Figure 3.3. Percentage change in globally and annually averaged precipitation as a function of global mean warming from seventeen models (After Figure 5.1 in Mitchell *et al.*, 1990)

The warming directly attributable to the increased CO_2-concentration has been estimated to be about 1°C. The warming above that value is due to *positive feedback* in the climate models. The most important positive feedback involves water vapour. An increased CO_2-concentration increases downward thermal radiation at the surface. The surface and the atmosphere warm in response to this increase. The capacity of the atmosphere to store water vapour (the *saturation humidity*) increases strongly as temperature

increases and the amount of water vapour in the atmosphere also increases. Water vapour is an even stronger greenhouse gas than CO_2 , so the downward thermal radiation at the surface is increased, and the surface warms much more than it would if the water vapour amount did not increase. The water vapour feedback is so powerful that a simple climate model suggests that if the Earth were moved to the orbit of Venus, where solar radiation is about 30% higher, then a *runaway greenhouse* might occur, in which the oceans would completely evaporate (Ingersoll, 1969).

An increase in cloudiness with a doubled CO_2-concentration could either enhance or reduce the climate sensitivity. Increased cloudiness can act as a positive feedback by increasing the downward thermal radiation at the ground, but could also produce a *negative feedback* by reducing the solar radiation reaching the surface. Sea ice can also produce a positive feedback. In this scenario, the sea ice amount decreases as the climate warms, leading to a decrease in the amount of solar radiation reflected to space.

It is important to understand the causes of the wide range of results obtained by the different models. All of the models cannot be correct. Some of the models may contain coding errors, since GCM computer codes are extremely complex. Coding errors are almost unavoidable, but the hope is that they do not affect the results. Assuming that there are no coding errors, the differences must be explained by differences in the strength of feedbacks related to different model parameterisations. The strength of the water vapour feedback is similar in most GCMs, although this agreement does prove that the models are correct. Some important differences in climate sensitivity have been traced to differences between cloudiness parameterisations.

Equilibrium experiments allow the potential magnitude of climate change to be estimated, and provide a convenient tool for increasing understanding of the feedbacks influencing these changes. The mechanisms that are likely to produce climate change, and equally importantly those, which are likely to be unimportant, can be identified. Equilibrium experiments also are useful for model inter-comparison and eventually will help in understanding and resolving the causes of the differences between the models. Based on the results from equilibrium experiments, potential climate change due anthropogenic emissions of greenhouse gases must be taken seriously.

3.3.9 Transient experiments

The sensitivity experiments described in Section 3.3.8 evaluate the equilibrium response of the climate model to some specified change. For the equilibrium state to be achieved, the solar forcing, and atmospheric composition are held fixed. Following the behaviour of the model, as it adjusts to the equilibrium state, is not the object of the equilibrium experiment. After the model reaches equilibrium, the climate does not change systematically, and further simulation with the model should give the same climate. The initial condition for the simulation is then irrelevant for an equilibrium experiment, with the caveat that if the model has multiple equilibria, the initial condition should be chosen, so that the simulation produces the equilibrium state relevant to the Earth's current climate. This presents little difficulty in practice.

Predicting the time evolution of the climate is important for practical reasons in developing strategies for dealing with climate change. In the case of simulation of the greenhouse effect, due to an increasing CO_2-concentration, the atmospheric composition may be changing so rapidly that the response of the model surface climate will lag behind the equilibrium climate change, found by equilibrium experiments with the relevant CO_2-concentrations, by decades. The time lag could be due to heat storage in the ocean, for example. If the CO_2-concentration is continually changing, the equilibrium state will never be reached, and the equilibrium experiment may not be useful for quantitative climate prediction. In this case a *transient experiment* will be more relevant. The transient experiment follows the evolution of the climate response to the variations in the CO_2-concentration or other time dependent specified quantity in the model. The transient experiment simulates the time scales for the climate change and the sequence of events in the climate change process. The choice of the initial condition will influence the results from the transient experiment for some time after the beginning of the integration. In contrast to the equilibrium experiment, the choice of the initial condition for the transient experiment, particularly for the initial state of the ocean, may be important.

A common transient experiment is to specify a time dependent scenario for the increase of atmospheric CO_2-concentration, such as a 1% increase per year. An initial condition, representative of the current climate, is used to initiate the experiment. The results for the 1% scenario at the time of CO_2-doubling (about 70 simulated years) will differ from the results of an experiment with a 2% increase of CO_2 per year after the same length of simulation, since the change of the CO_2-concentration will be twice as large as in the latter case. Some of this difference may be explained by appealing to results from equilibrium experiments. However, the difference between the two transient experiments will not necessarily be the same as that obtained from doubled and quadrupled CO_2 equilibrium experiments, due to the time lag effect. Similarly, due to the time lag effect, the results for the two scenario's could differ significantly if compared at the time of CO_2-doubling, which will occur after 35 simulated years in the 2% case, and 70 years in the 1% case.

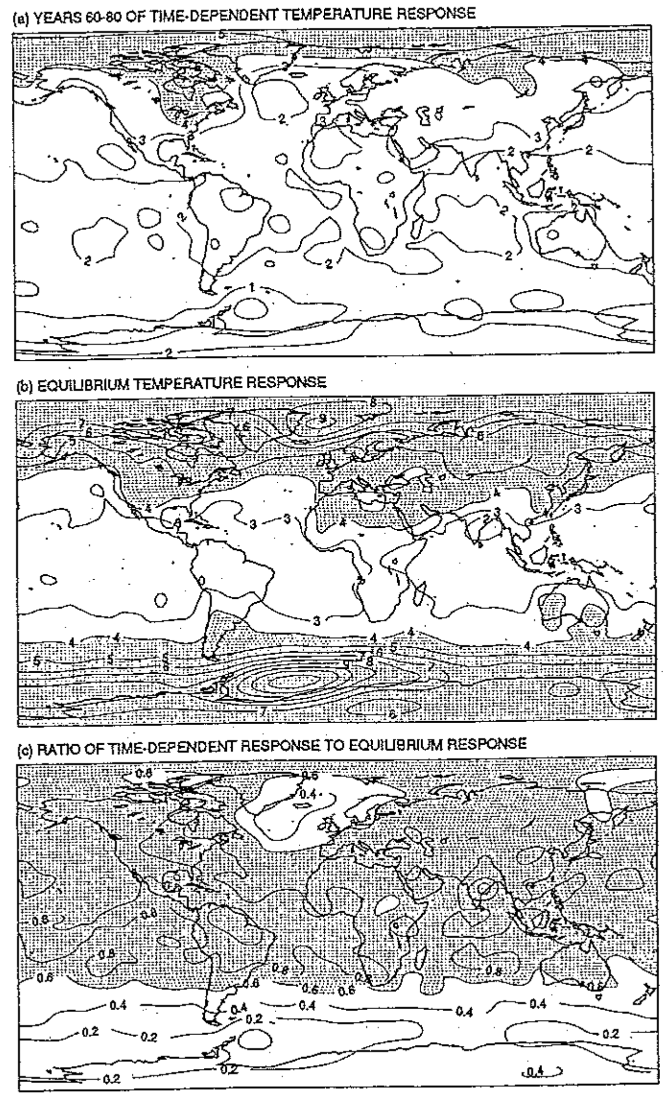

Figure 3.4. (a) The time-dependent response of surface air temperature (°C) in a coupled ocean-atmosphere model to a 1%/yr increase of atmospheric CO_2. The difference between 1%/yr perturbation run and years 60-80 of the control run, when the atmospheric CO_2-concentration approximately doubles, is shown. (b) The equilibrium response of surface air temperature (°C) in the atmosphere-mixed-layer ocean model to a doubling of atmospheric CO_2. (c) The ratio of the time-dependent to equilibrium responses shown above. (After Figure 6.5 in Bretherton *et al.*, 1990)

In a transient experiment the slowly varying elements of the climate system (e.g. the deep ocean) are not in equilibrium and this disequilibrium leads to the time lag effect. In the case of the response to increasing atmospheric CO_2-concentrations, the global warming at any time will be less than the equilibrium response evaluated at the relevant CO_2 value, with the magnitude of the difference being a measure of the imbalance in the climate system. This characteristic relationship between transient and equilibrium experiments is shown in Figure 3.4 (after Bretherton *et al.*, 1990). This figure also shows that transients experiments can give very different results for the spatial distribution of the climate change than equilibrium experiments, due to the interactions between components of the climate system with different intrinsic time scales.

The transient experiment can be viewed as a step towards prediction of climate evolution in the near and long term. A model, which realistically represents the evolution of the components of the climate system that vary on the long time scales of interest, in particular the oceans and sea ice, as well at the behaviour of the atmosphere with its much shorter intrinsic time scale, is necessary for predictive purposes. Therefore, a full dynamical GCM and some type of interactive sea ice are desirable features for the transient experiment model.

The coupled AGCM/OGCM, at its current state of development, generally does not reproduce the current climate when run to equilibrium with current atmospheric CO_2-concentrations in *control simulations*. This phenomenon is known as *climate drift*. The climate drift can be large and is thought to indicate deficiencies in the GCM parameterisations. Since excessive climate drift may lead to unrealistic sea ice distribution and distort the model sensitivity, specified corrections have been added at the atmosphere-ocean interface to force the equilibrium climate to remain close to the observed climate. This procedure is known as *flux correction* or *flux adjustment*. Then, the same flux correction is applied to the model in the transient experiment as in the control. Flux correction is controversial (see Section 3.7). An alternative procedure is to remove the effect of the climate drift *a posteriori* by taking the difference between the transient experiment and the drifting uncorrected control.

Figure 3.5 shows the results from transient experiments carried out with increasing CO_2 scenarios in a number of global coupled AGCM/OGCMs. The curve labelled "IPCC A" shows the results from the IPCC's "business-as-usual" CO_2 scenario (close to the 1% per year case discussed above) using a simplified upwelling-diffusion ocean model coupled to a one dimensional energy balance atmospheric model.

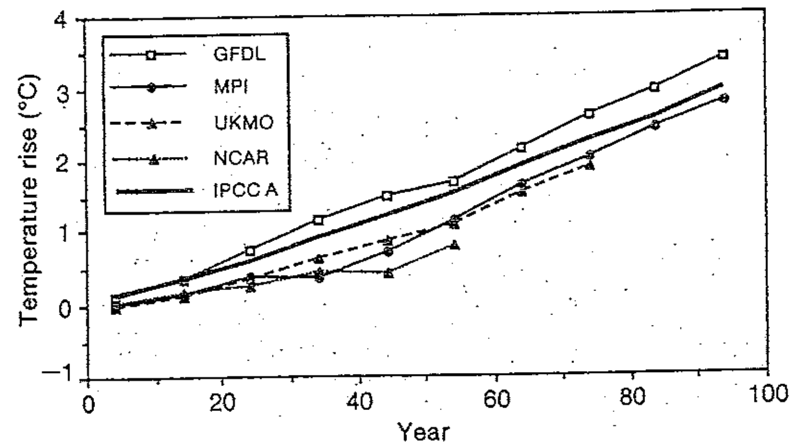

Figure 3.5 Decadal mean changes in globally averaged surface temperature (°C) in various transient coupled ocean-atmosphere experiments. Note that the scenarios employed differ from model to model, and the effect of temperature drift in the control simulation has been removed. The different models are denoted GFDL = Geophysical Fluid Dynamics Laboratory, MPI = Max Plank Institut für Meteorologie, UKMO = United Kingdom Met. Office, NCAR = National Center for Atmospheric Research, IPCC = IPCC 1990 Scenario A "best estimate". (After Gates *et al.*, 1992)

The term "transient experiment" will eventually be replaced by "climate prediction" when sufficient progress is made in the areas of initialisation, model error, and verification for coupled climate models.

3.4 Model calibration

In an effort to ensure that GCMs are as faithful as possible to the observations of the ocean and atmosphere, a process of calibration is required. As the models evolve, they are continually scrutinised in the light of observations to determine if significant physical processes are missing, and to refine those that are present. The latter may involve a reworking of the existing physical parameterisations or simply a new choice for the adjustable parameters that inevitably are part of these parameterisations. Changing the latter is also called "tuning".

The calibration of GCMs to the present climate is a complex and subtle process, for the optimal configuration of one parameterisation depends both on the overall characteristics of the model's resolved flow, and on the behaviour of all the other parameterisations. One important example of this arises when the model's resolution is changed. This will lead to a systematic alteration in the resolved flow, which serves as the input to all the parameterisations, necessitating a re-calibration. Since all the parameterisations simultaneously affect the predicted fields of momentum, temperature and moisture in the atmosphere (momentum, temperature and salinity in the ocean), and hence effectively communicate to each other through these fields, a change in one parameterisation often requires changes in others.

In this section we present three specific examples of model calibration. The first example concerns the relationship of the cumulus convection in the atmosphere to the large scale tropical 30-60 day oscillation, also referred to as the Madden-Julian Oscillation. It consists of a very large scale (zonal wave-number one[1]) tropical circulation which moves eastward around the globe with a variable amplitude and a period of roughly 30 to 60 days. This circulation is most strikingly seen in the divergent component of the circulation. This is the component involving convergence (coming together) and divergence (drawing apart) of air. Low level convergence leads to rising motion, while low level divergence is associated with sinking motion. In the phase of the Madden-Julian oscillation involving low level convergence, there is a corresponding enhancement of the precipitation, while there is a suppression of precipitation in the divergent phase. The oscillation is geographically variable in the sense that when the convergent portion moves over areas of relatively cold tropical sea surface temperatures (such as the Eastern Pacific) it is attenuated, but it redevelops over those areas more favourable for tropical cumulus convection.

Since this oscillation dominates tropical intra-seasonal variability in the atmosphere, much effort has gone into studying the ability of GCMs to simulate it. Models employing a mass flux scheme for cumulus convection originally failed to simulate the correct period for the oscillation, simulating instead a much faster (15 day) eastward travelling circulation (mass flux

[1]A zonal wave-number-one-circulation has a sinusoidal variation around a latitude circle with just one complete sine wave.

schemes of cumulus convection are discussed in Section 3.3.4). Theoretical studies of the oscillation indicated that the lower the level at which atmospheric diabatic heat release occurred in conjunction with cumulus convection, the slower the oscillation ("diabatic heating" is heating, which occurs not simply due to compression of dry air, but in this case due to the condensation of water and release of latent heat). Since the mass flux schemes were predicting almost uniform latent heat release from lower levels to upper levels, a re-calibration of the scheme was suggested. An adjustment in the value of a single parameter, which controls the entrainment of environmental air into the cumulus clouds, led to a profile of diabatic heating more peaked at mid-levels, and consequently a more realistic 30-60 day oscillation.

A good example of a problem in parameterisation, occurring in ocean models, is the effect of the vertical diffusion scheme on the depth of the oceanic mixed layer. Correctly simulating changes in the mixed layer depth is important for modelling ocean climate. The sub-grid scale motions (turbulence), which carry out the vertical diffusion of temperature and salinity necessary to keep the mixed layer homogenous must be parameterised, and there are several methods of doing this. The simplest is to employ classical diffusion with a constant diffusion coefficient determined empirically. A more sophisticated version of this allows the diffusion coefficient to be dependent upon a measure of the local vertical stability of the ocean, with less stable areas implying larger mixing. Finally, a full turbulence closure scheme (similar to that discussed in the context of the atmospheric boundary layer in Section 3.3.4) represents the most complex treatment.

Unfortunately, the choice of mixing scheme has a large influence on the predicted depth of the mixed layer in the ocean, and the nature of this influence is highly dependent upon other factors, such as simulated flux of heat between the ocean and atmosphere and the number of vertical layers used in the ocean model. Since the physical nature of this influence is not completely understood, each ocean model must be tuned individually with respect to the treatment of vertical mixing.

The final example in this section illustrates how a missing physical process can be identified. Higher resolution GCMs tended to generate mid- to high latitude upper level zonal (i.e. west to east) winds, which were too large (too much eastward motion) (see Section 2.3.1 for background on the upper level zonal winds). This phenomenon was far more noticeable in the Northern Hemisphere than in the Southern Hemisphere. Simultaneously the wave motions in the higher resolution GCMs were diagnosed as systematically transporting more eastward momentum towards the poles than the lower resolution GCMs. The poleward transport into this region was consistent in the sense of maintaining the east-west momentum balance. Yet the increase in the model error in the upper level zonal winds, as the resolution was increased, suggested that a physical process had been missed, one that would operate more strongly at high resolution and in the Northern Hemisphere.

The hypothesis of "Gravity Wave Drag" was successful as an explanation on both counts. The basic notion is that the impinging of the wind on rugged terrain (mountains) generates vertically propagating gravity waves, which break at the level of maximum vertical derivative of the zonal wind (known as wind shear). This occurs near the jet level at 200 hPa in the upper troposphere ("Gravity Waves" are waves involving periodic displacements of parcels due only to density differences with their surroundings). Thus, their relatively small eastward momentum is deposited in the jet, providing an effective drag on the very strong winds (with large eastward momentum) at this level. When implemented in GCMs, this mechanism was seen to alleviate the excessive upper level winds.

The problem noted above (excessive upper level winds) was never noted in low resolution models (even without gravity wave drag), because the wave motions in these GCMs fail to transport sufficient momentum towards the pole. Thus, the momentum budget in mid and high latitudes could be balanced without the need for additional drag, and the lack of a gravity wave drag mechanism was never noticed. In effect, the two errors cancelled each other. In current practice a parameterisation of gravity wave drag is present in most models.

These examples suggest some general limitations to the class of models currently used to simulate the ocean and atmosphere. While there are cases, in which physical reasoning leads directly to improvement in the simulation, there are many counter-examples, in which alternative physically based parameterisations lead to very different results, with no clear physical explanation available. In addition, the sensitivity of these parameterisations to a few tuneable parameters suggests that some underlying physics have been missed.

3.5 Model validation

Simplification is the essence of modelling. Moreover, relatively simple parameterisations are usually both easier to understand and to formulate in terms that a computer can use, than are more complicated or physically more complete descriptions. On the other hand, it is clearly important to check, whether the simplified formulation in existing models of the ocean and atmosphere seriously undermine their simulation skills - a step referred to as "model validation."

In order to determine how well GCM simulations capture the behaviour of the Earth's atmosphere and ocean, we compare these simulations with data sets constructed from observations (this is also referred to as "verification"). We compare simulations of different GCMs in order to discover both the problems and successes common to various models, and to learn how the distinct modelling philosophies (Section 3.3.6) compare in their realism. In Section 3.5.1 we discuss some of the many types of observations against which GCMs are compared, and in Section 3.5.2 we introduce a few of the more recent inter-model comparison studies.

3.5.1 Comparison with observational datasets

The process of comparing GCM simulations with datasets constructed from observations of the atmosphere and ocean, a process often referred to as verification, is as important as the modelling itself. Broadly speaking, there are two classes of observational datasets. Primary observational datasets come more or less directly from the observing instruments themselves. Good atmospheric examples are the routine measurements made by radiosondes (instruments carried into the atmosphere by balloons) and aircraft, and the more intensive measurements made by instrument towers during special field experiments.

Another class of primary observations comes from satellite ("remotely sensed") data. The satellites measure only radiation in various wavelength bands, and this information can be used to verify the outgoing short wave radiation (solar radiation reflected from the earth-atmosphere system) and the outgoing long wave radiation (due to emission from clouds and the ground) that are given by the GCMs radiative calculations. Since these quantities play a pivotal role in defining the overall energy balance of the Earth-atmosphere system, they are of vital importance in modelling the climate. See the discussion in Section 2.2.

Satellite measurements of radiance can be used to estimate time mean atmospheric precipitation, although this requires an intermediate physical model. Tropical rainfall can be estimated from the outgoing long wave radiation (OLR), as follows: deep convection leads to very high cloud tops, which are quite cold and hence emit relatively less upward long wave radiation. Since the satellite measures the OLR from the top of the clouds, there will be a correlation between small tropical values of OLR and convective precipitation. Recent microwave measurements can be used to estimate precipitation globally, since these measurements are sensitive to the total amount of liquid water in the atmosphere. These can be combined with the primary measurements of rainfall and snowfall over land to give a global picture of the total precipitation on a season by season basis. This is of great relevance to atmospheric dynamics, for the precipitation gives a measure of the vertically integrated heating, due to latent heat release. While the above discussion has given examples from the atmosphere, similar types of primary data are available for the oceans.

Primary measurements alone are not adequate, however, to define the full four dimensional (three spatial dimensions plus time) structure of either the atmosphere or ocean. This can only be done by using derived observational datasets, in which the primary datasets are merged, using the GCM itself, to produce uniform, gridded four-dimensional datasets called "analyses". The primary data are combined in a process called "data assimilation", which produces values for all the basic variables (horizontal momentum, temperature and moisture for the atmosphere; horizontal momentum, temperature and salinity for the ocean) at regular time intervals. A short-range forecast made by the GCM from the previous analyses is used to provide information in regions where no data are available. Data assimilation is critical for obtaining realistic "initial states" from which to run forecasts with GCMs, whether for the atmosphere alone, the ocean alone, or the coupled system.

The most basic set of statistics of the general circulation that is verified from analyses consists of the (three dimensional) time mean fields, where the averaging period is usually a month or a season, or even a year. Seasonal time means form part of the annual cycle, the regular, smooth component of the atmospheric circulation which is due to the annual cycle of the solar heating. Because of the rotation of the earth there is a very approximate longitudinal independence of the large scale fields such as temperature and zonal wind, particularly in the nearly all ocean covered Southern Hemisphere. It has thus proven useful to separate out the "zonal mean" (average around a latitude circle), with the remaining part of the field

$$[A] = \frac{1}{2\pi}\int_0^{2\pi} A d\lambda \qquad (3.2)$$

referred to as the "eddy component". The zonal mean is denoted by square brackets, where A refers to a two-dimensional field and l is longitude. The eddy component of A is denoted by an asterisk and is defined by A* = A - [A].

The time mean of the field A over a period of time T is denoted by an overbar, as in:

$$\overline{A} = \frac{1}{T}\int_0^T A dt \qquad (3.3)$$

and the departure from the time mean (known as the "transient" field) is denoted by a prime, so that

$$A' = A - \overline{A} \qquad (3.4)$$

The time average of the zonal (west to east) wind u can be written as

$$\overline{u} = [\overline{u}] + \overline{u}^* \qquad (3.5)$$

The first term is the zonal and time average, which is given as a function of latitude and pressure in Figure 3.6. This represents a basic state of the atmosphere (or ocean) in terms of which many wave quantities can be computed, using linearised versions of the basic equations of motion. These wave quantities include the "stationary eddy field" denoted by the second term in the above equation. The stationary eddy field plays a large role in defining regional climate. In the atmosphere it is forced both by the presence of mountains and by latent and radiative heating, and it can be thought of as consisting of waves, which vertical and horizontal propagation is controlled by the mean zonal wind (first term above).

In the context of the overall balance of momentum, heat and moisture in the general circulation of the atmosphere (or ocean), both the stationary eddies and the transient field introduced above, play a large role. We have for example the following expression for the northward transport of sensible heat in the atmosphere:

$$[\overline{VT}] = [\overline{v}][\overline{T}] + [\overline{v}^*\overline{u}^*] + [\overline{v'T'}] \qquad (3.6)$$

in which T is the temperature, v is the meridional (south to north) wind and all the remaining notation has been explained above. In words, the total northward transport of sensible heat is due to the "time mean meridional averaged circulation" (first term), the stationary eddies (second term), and the transients (third term). Together they make up the atmospheric heat

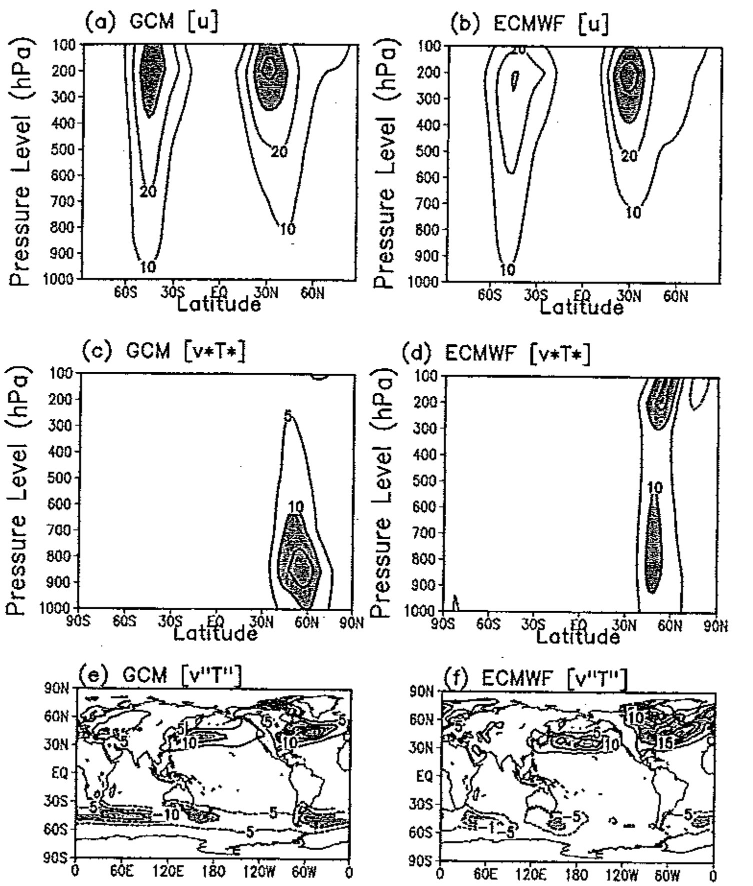

Figure 3.6 The *top panels* show the time mean and zonally averaged u-wind for the Center of Ocean-Land-Atmosphere Studies (COLA) atmospheric GCM (left) and the European Centre for Medium-Range Weather Forecasts (ECMWF) analyses (right panel). The time period is Jan. through Mar. 1983. Units are in meters/sec, the contour interval is 10 m/s and values greater than 30 m/s are shaded. The *middle panels* show the stationary eddy heat flux, time averaged (over the same period as above) and zonally averaged, with the GCM again on the left and the analyses on the right. The units are (°K)(m/s), the contour interval is 5 and values greater than 10 are shaded. The *bottom panels* show maps of the transient heat flux at the 850 hPA level, filtered to retain periods of less than 10 days. The time period is the same as above, and the units are (°K)(m/s). Values grater than 10 in the Northern Hemisphere (less than -10 in the Southern Hemisphere) are shaded, and the contour interval is 10. The GCM is on the left, the ECMWF analyses on the right. In all figures dotted lines denote negative values.

transport discussed in Section 2.3.1. In the tropics the first term is associated with the Hadley cell, which refers to the near-equatorial zonally averaged cell of upward and poleward motion (see Section 2.3.1). These are shown for both the COLA atmospheric GCM and analyses of analyses of the ECMWF in Figure 3.6, where the time mean used is the winter season.

The transient component in general measures the intensity of both the day-to-day weather and the less rapid varying components. The day-to-day weather circulations are in large part due to instabilities, in which small perturbations grow rapidly (see Figure 2.6). The mechanisms involved in this rapid growth can be studied both by controlled experiments with GCMs and by detailed examination of analyses. The instabilities tend to be associated with the upper level jets (strong maxima in the u-wind), which are centred over the east coasts of Asia and North America in the Northern Hemisphere, and over the Indian and Pacific Oceans in the Southern Hemisphere (the signature of the jets in the zonal mean is seen in panels (a) and (b) of Figure 3.6). The regions of instability associated with these jets are known as "storm tracks", and can be diagnosed from GCM output or analyses by time filtering the transient data to retain only periods shorter than about 10 days. The geographical distribution of transient heat transport associated with these short time scales can be seen in panels (e) and (f) of Figure 3.6, which indicates that the GCM simulates the systematic heat transport associated with these transients in a reasonable manner.

3.5.2 Inter-model comparison

The compelling scientific rationales for comparing the results of different GCMs are firstly to isolate the effects that specific physical processes or interactions have on the general circulation, and secondly to document whether the current state of the art in general circulation modelling is adequate to simulate particular aspects of the general circulation. Are important physical processes being entirely ignored or misrepresented? (see, for example, Section 3.4). Are common assumptions made in the modelling process faulty? The first goal is addressed by controlled comparisons, in which the models entering the comparison differ from each other only in a few aspects, such as a parameterisation, or in having different resolution. Unrestricted model comparisons, on the other hand, compare the performance of a set of widely varying GCMs.

The clearest and most extensive controlled comparisons among atmospheric GCMs are those that have examined the effects of varying the horizontal resolution. Integrations of the same GCM at differing resolutions are run from the same initial conditions, under the influence of the same boundary conditions (most notably sea surface temperature). We must recognise that changing the resolution implies more than increasing the number of spectral components retained (or decreasing the grid size). The complex topography (mountains) of the Earth's surface and the gradients of sea surface temperature are much more sharply defined at a higher resolution. The improved forcing of the stationary waves by the mountains, and the added atmospheric sensitivity to sharp anomalies of sea surface temperature are strong arguments for using a higher resolution. Results from such comparisons indicate a consistent improvement in the realism of the stationary wave simulation with increasing resolution only up to a point, beyond which little improvement is seen in the stationary waves. This saturation point occurs at about T42 spectral resolution, corresponding to a 2.8° x 2.8° grid spacing (T stands for triangular truncation in a spectral discretisation technique; 42 for the maximum global wave-number retained). The level of transient activity (measured for instance by the temporal variance of the basic fields) also tends to increase with resolution, but this effect does not have such a well defined saturation point.

A good example of an unrestricted model comparison is that carried out by the Monsoon Numerical Experimentation Group (MONEG), who compared the simulation of the summer Indian Monsoon in a large number of GCMs and in two different sets of analyses. The period is the summer (June through August) of 1988, a year in which the Indian Monsoon was very good in the sense of having significantly higher than normal rainfall. Very important for the Indian Monsoon is the wind at lower levels (850 hPa) in the Indian Ocean.. Here the flow goes almost due eastward in the Arabian Sea, transporting the moisture, needed for the Monsoon rainfall to India. Thus, the amount of rainfall simulated is sensitive to the precise configuration of this current, as it approaches and crosses the west coast of India. Two simple parameters, which describe this flow, are the maximum wind speed over India itself and the latitude at which this maximum is attained. Each model simulation is represented by its own acronym in Figure 3.7, and the two analyses (indicated by the letters "ECMWF" and "NMC") by dark squares. Not only is there a great deal of variation between the GCMs in representing the Monsoon flow, but the two sets of analyses (which each represent a valid set of observations) disagree. Research, seeking to relate the Monsoon circulation to the parameterisations of the boundary layer and cumulus convection as well as the treatment of orography, is ongoing.

The most extensive unrestricted comparison between models that has occurred to date is the Atmospheric Modeling Intercomparison Project (AMIP), being sponsored by the World Climate Research Program (WCRP). A large number (30) of modelling groups throughout the world integrated their GCMs for 10 years (starting from 1 January, 1979), using a common

set of sea surface temperature and sea-ice fields as boundary conditions. The range of models used was very wide, both in terms of horizontal and vertical resolution and in terms of the philosophy behind the parameterisations of physical processes (see Section 3.3.6). The basic outcome of this comparison is that although variations between models of course exist, there are a number of common errors made by many of the models.

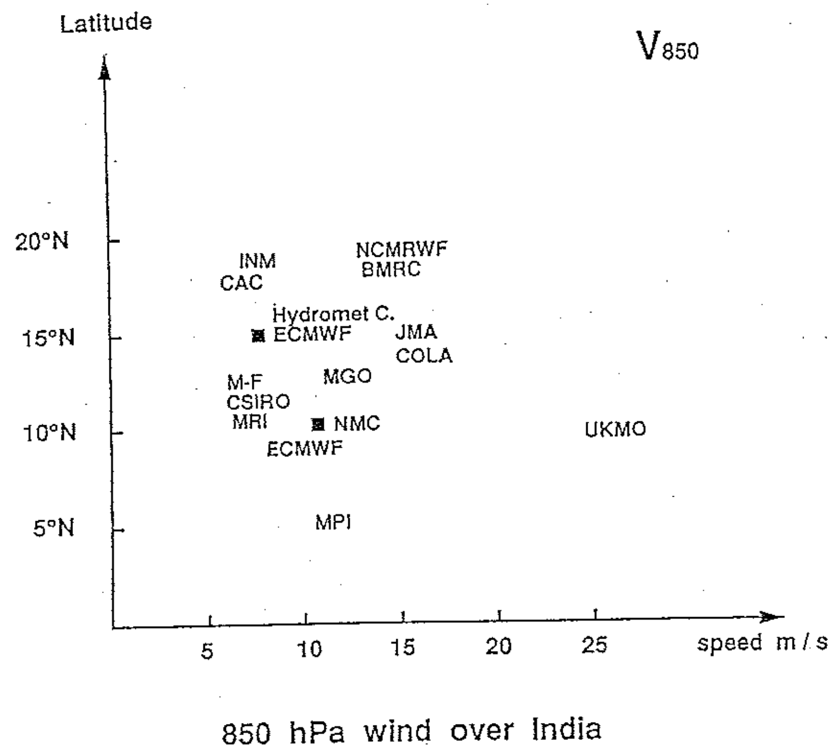

Figure 3.7 A comparison between two different analyses (ECMWF and NMC, labelled by squares) and many GCM simulations of the low-level (850 hPa) winds over India during the period of June through August of 1988. The abscissa shows the maximum wind speed over India and the ordinate the latitude at which this wind speed is reached.

One of the more interesting AMIP comparisons involves the meridional energy transports by the atmosphere-ocean system, by the atmosphere alone and the implied oceanic transports. The latter is seen to vary dramatically from model to model and in some cases lies outside the range estimated from the all available observations. It turns out that the treatment of the cloud-radiation interaction is the key player here and that it is apparently not being well handled by many models.

The annual mean meridional transport of energy by the ocean can be calculated from net ocean surface energy flux, which is the sum of the net downward flux of radiation at the ocean surface and the fluxes of sensible heat (internal energy) and latent heat (energy tied up in water vapour) from the atmosphere to the ocean. All are quantities predicted by the GCMs, so that the implied oceanic transport can be computed. In the Northern Hemisphere all the GCMs give northward transport, although the magnitude varies by as much as a factor of 10. In the Southern Hemisphere even the sign is in doubt, with most models yielding northward (that is, equatorward) transport, which is in disagreement with the estimates from observations, which show southward (poleward) heat transport in this region.

Another estimate of this transport may be obtained by using satellite data to estimate the total (atmosphere plus ocean) annual mean heat transport, which can be calculated from the net downward radiation at the top of the atmosphere (see Figure 2.7). Subtracting from this the atmospheric energy transport calculated directly from the simulated atmospheric fields of temperature, pressure and specific humidity give a hybrid implied ocean heat transport. This quantity shows much greater agreement between GCMs, in particular giving southward (poleward) transport everywhere in the Southern Hemisphere. Since the only term involving the radiative heating is taken from observations in this method, it is free of the model errors in cloud-radiation interaction. Clouds influence atmospheric radiative heating primarily by trapping longwave energy within and beneath the cloud layer. See the discussion in Section 2.4.4.

The errors in the net ocean surface energy flux due to the errors in cloud radiative forcing do not affect the atmospheric simulations in these atmospheric GCMs, because the sea-surface temperature (SST) is prescribed. However, when atmospheric and oceanic GCMs are coupled and the SST is predicted, these errors become significant, causing the coupled models to "drift" away from conditions typical of the current climate.

3.6 Climate predictions

Although it is well known that the day to day changes of weather cannot be predicted for periods beyond 1-2 weeks, it has been suggested that the climate variations (climate being defined as the space-time average of weather) can be predicted if averaged over certain space and time scales.

How well we can predict climate variations or if we can predict them at all, depends upon our ability to understand and model the mechanisms that produce climate variations. For brevity, the mechanisms that produce climate variations can be described in two categories.

a) Internal:
These are climate variations that are produced by the internal dynamics of, and interactions among the atmosphere, biosphere, cryosphere and hydrosphere components of the coupled climate system. This category will include climate variations due to tropical ocean-atmosphere interactions (El Niño-Southern Oscillation or ENSO), land atmosphere interactions (droughts, desertification, deforestation), deep ocean-atmosphere interactions (thermohaline circulation), and ocean-land-atmosphere interactions (monsoon floods and droughts, heat waves and cold spells).

b) External:
These are climate variations that are caused by factors which are external to the climate system itself, and are not initially caused by the internal dynamical and physical processes. It should be recognised that this classification is only for convenience and even if the primary causes are external, the details of the regional climate variations are determined by interactions and feedbacks among the external and internal processes. As an example, this category will include climate change and climate variations due to changes in solar forcing (either due to changes in the solar constant or changes in Earth's orbit around the sun or Earth's axis of rotation), and changes in the chemical composition of the Earth's atmosphere (either due to human induced changes in the concentration of greenhouse gases or due to volcanoes).

It should also be noted that the predictability of climate variations, due to the *internal* mechanism, sensitively depends upon the initial state of the climate system, and therefore there is a finite limit of predictability of these variations. However, it may be possible to predict a new equilibrium climate for a different external forcing, which is independent of the initial state of the climate system. The above two types of predictions were earlier suggested by E. Lorenz, the originator of the ideas referred to as the "Butterfly Effect" and Chaos, as the climate prediction of the first kind and the second kind, respectively.

In the following two subsections (3.6.1 and 3.6.2) we describe the state of our current knowledge about our ability to predict seasonal to inter-annual variations and decadal variations respectively. It is generally agreed that seasonal to inter-annual variations are predominantly due to *internal* mechanisms and, in fact, successful predictions of certain aspects of seasonal to inter-annual variations have been made without any consideration of external forcing, due to greenhouse gases or volcanoes. The mechanisms for decadal to century variations, however, are not well understood and the relative roles of *internal* and *external* mechanisms in producing decadal to century scale variations and their predictability is a topic of current research.

3.6.1 Prediction of seasonal to inter-annual variations

As discussed before, seasonal to inter-annual variations are caused by the internal dynamical processes of the coupled climate system. However, it is convenient to further subdivide these processes into two categories: the fast atmospheric variations associated with the day to day weather; and slowly varying changes in sea surface temperature (SST), soil wetness, snow cover and sea ice at the Earth's surface. The latter act as slowly varying boundary conditions for the fast weather variations. It is now well known that seasonal mean atmospheric circulation and rainfall are strongly influenced by changes in the boundary conditions at the Earth's surface, and this has provided a scientific basis for dynamical prediction of seasonal and inter-annual variations.

Thus, if it were possible to predict the boundary conditions themselves, it would be possible to predict the atmospheric circulation and rainfall. The extent, to which slowly varying boundary conditions influence atmospheric circulation and rainfall, strongly depends upon the latitudinal position of the region under consideration. Therefore, we will describe the predictability of seasonal to inter-annual variations separately for the tropical and the extra-tropical regions. The predictability of the seasonal mean circulation and rainfall for the current season and for the same season one year in advance is also very different, depending upon the region and season under consideration. Therefore, we will describe this subsection under four separate items:

a) Prediction of seasonal variations (tropics)

The tropical circulation is dominated by large-scale east-west (Walker circulation) and north-south (Hadley circulation) overturnings. These large-scale features have a well defined annual cycle, associated with the annual cycle of SST and the solar heating of land masses. Weak tropical disturbances (for example, easterly waves, lows and depressions) are superimposed on these large-scale features. Changes in the location and intensity of these large-scale overturnings are caused by changes in the boundary conditions at the Earth's surface. For example, when the Central Pacific ocean is warmer than normal, the climatological mean rainfall

maximum shifts eastward and produces floods over the Central Pacific Islands and drought over Australia and India. When the Northern tropical Atlantic, in the months of March, April and May, is warmer than normal, the inter-tropical convergence zone does not move as far south as normal and gives rise to severe droughts over Northeast Brazil. Likewise, seasonal rainfall over sub-Sahara Africa is strongly influenced by the location of the inter-tropical convergence zone, which in turn is influenced by the boundary conditions over the global tropical oceans and local land conditions. A simple conceptual model to understand the causes of tropical floods and droughts is to consider space and time shifts of the climatological annual cycle of rainfall. These shifts are caused by anomalous boundary conditions at the Earth's surface.

A large number of climate model simulations have well established the validity of the conceptual model described above. It is now well accepted that the potential for dynamical prediction of the seasonal mean circulation and rainfall in the tropics is quite high. The most important dynamical reason for high seasonal predictability in the tropics is the absence of strong dynamical instabilities, which produce large amplitude weather fluctuations. Day to day weather fluctuations are relatively weak in the tropics, and therefore, it is possible for changes in boundary conditions to exert a large influence on the seasonal mean circulation and rainfall.

b) Prediction of seasonal variations (extra-tropics)

Extra-tropical weather fluctuations, especially during the winter season, are caused by strong dynamical instabilities, which produce very large day to day changes. The seasonal mean circulation in the extra-tropics, therefore, is not influenced by slowly varying boundary conditions to the same extent as it does in the tropics. The inherent chaotic nature of the extra-tropical circulation makes it less likely that useful seasonal predictions can be made. In some special cases, especially during the winter season, when tropical SST anomalies are large in amplitude and in spatial scale, and there is a significant change in the dominant tropical heat sources, it has been found that these tropical changes also affect the extra-tropical circulation. For example, it has been shown that during the years of strong El Niño events in the tropical Pacific, there is a well defined predictable pattern of winter season climate anomalies over the Pacific North America region.

The potential for prediction of the seasonal mean circulation over the extra-tropics is higher during spring and summer season, because, not unlike the tropics, the day to day weather changes are not strong and therefore it is possible that seasonal variations are controlled by changes in the boundary conditions. It are the local land boundary conditions, which are more important during the spring and summer seasons, because the solar forcing is large, and the large scale dynamical environment is not favourable for propagation of remote influences from tropical SST and heating changes.

c) Prediction of inter-annual variations (tropics)

The El Niño-Southern oscillation (ENSO) phenomenon is the most outstanding example of tropical **inter-annual** variability, for which there are sufficient oceanic and atmospheric observations to describe, and which has been successfully predicted by several dynamical models of the coupled tropical ocean-atmosphere system. ENSO is an a-periodic (with quasi-periodicity of 3-5 years) phenomenon characterised by alternating episodes of warmer than average and colder than average SST in the Central and Eastern Pacific. When the SST is warmer than average, the surface pressure is higher than average in the Central Pacific ocean and lower than average in the Eastern Indian ocean, droughts occur over Australia and India, and floods occur over the west coast of South America and Central Pacific Islands. ENSO is produced by an interaction between the upper ocean and the overlying atmosphere. Several dynamical models of the coupled ocean-atmosphere system have successfully simulated the ENSO related **inter-annual** variations and it has been demonstrated that the range of predictability of the coupled tropical ocean-atmosphere system is about 1-2 years. It has been further recognised that, just as the memory for predictability of weather resides in the (initial) structure of atmosphere, and boundary conditions are crucial for predictability of seasonal averages, the memory for predictability of ENSO primarily resides in the (initial) structure of the upper layers of the tropical Pacific ocean.

Variations in the Indian monsoon rainfall are another example of **inter-annual** variability, which has been successfully predicted, using empirical techniques. It has been found that winter seasons with excessive snowfall over Eurasia are followed by below average monsoon rainfall over India and vice versa. This relationship, along with a strong association between warmer than normal equatorial Pacific SST and deficient monsoon rainfall over India, is routinely used to predict summer monsoon rainfall over India.

d) Prediction of inter-annual variations (extra-tropics)

With the exception of the influence of tropical Pacific SST anomalies on winter season circulation over North America, there are no well recognised and generally accepted mechanisms that can be invoked to predict inter-annual variations over the extra-tropics. There are strong correlations between the extra-tropical SST/sea ice and extra-tropical circulation. However, these correlations occur either for simultaneous variations or for atmospheric anomalies, forcing (i.e. ahead of) SST and sea ice anomalies. It is likely, although it can not be proven, due to lack of appropriate land-

surface datasets, that anomalous soil wetness, albedo and vegetation cover during spring and summer season can produce significant anomalies of circulation and rainfall over land.

3.6.2 Prediction of decadal variations

As described before, decadal variations can occur either due to internal dynamical mechanisms of the coupled climate system, or due to external forcing. For either case, it is reasonable to state that our understanding of the physical mechanisms is insufficient, and our ability to model the decadal variations is inadequate, and therefore, at present, a scientific basis to make decadal predictions does not exist. However, it is instructive to review our current knowledge of the evidence for and understanding of the nature of decadal variations and the potential for their predictability. We present this discussion separately for *internal* and *external* decadal variations.

Internally forced decadal variations

Analysis of past observations in the atmosphere and oceans has revealed several examples of decadal variability, which can be attributed to the internal dynamics and interactions among the atmosphere, ocean and land processes. The examples include fluctuations of the thermohaline circulation (especially in the Atlantic); persistent droughts in sub-Sahara Africa; multi-decadal variability of the Indian monsoon rainfall; changes in the level of the Great Salt Lake, Utah and SST anomalies in the North Atlantic and the North Pacific. In addition, there are decadal changes in the intensity and frequency of El Niño events, which can produce, in turn, decadal changes in global circulation. As yet, no systematic study of predictability of decadal variations has been carried out. There have been only a few, if any, model simulations of decadal variability using realistic models of the atmosphere and ocean. Currently, there are several national and international efforts underway to observe, model and, if possible, predict decadal variability.

Externally forced decadal variations

The most extensively discussed example of decadal (and longer) variation of climate is that due to the increase in the concentration of greenhouse gases. Figure 3.8 shows a time series of the observed global mean surface temperature anomaly. It can be seen that there is a well defined long-term trend, as well as decadal variations in the global mean surface temperature. GCM sensitivity experiments (see Section 3.3.8) have been carried out to predict the long-term trend. However, it is unclear at this stage if decadal variations due to greenhouse gases are predictable or not. For one thing, the magnitude of these changes is quite small, especially compared to the error in the model's ability to simulate present climate. While the magnitude of the observed and model simulated regional anomalies is quite large compared to the fluctuations of the global mean, it is far more difficult to predict regional averages, because these fluctuations are largely determined by internal dynamics mechanisms, which are inherently unpredictable.

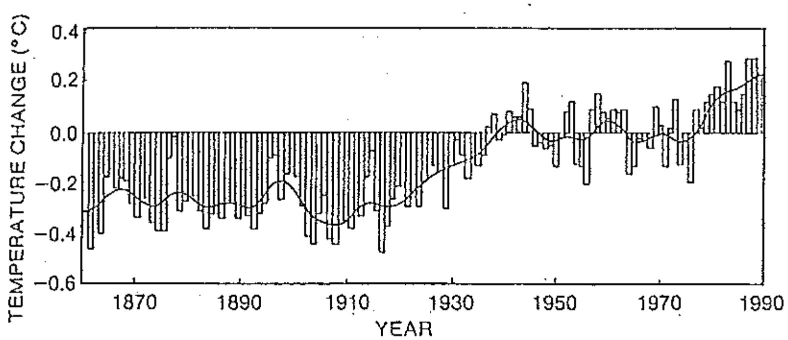

Figure 3.8 Global-mean combined land-air and sea-surface temperatures, 1861-1989, relative to the average for 1951-80 (Source: IPCC, 1990)

3.6.3 Prediction of changes in variability due to climate change

Till now we have discussed the predictability of seasonal, inter-annual and decadal averages. It is also of interest, scientifically as well as from a societal point of view, what, if any, changes in the frequency of extreme events (viz frequency and intensity of hurricanes, severe floods and droughts), changes in the amplitude of diurnal cycle, and changes in the intensity and frequency of El Niño events can be predicted. This question has been addressed mainly in the context of climate change, due to increase in greenhouse gases. There are no conclusive results yet, as some model calculations show some change in hurricane frequency (and amplitude), whereas some other calculations do not show any significant change. Actual observations show a clear tendency for reduced hurricane frequency in the Atlantic during El Niño years.

3.7 Limitations in present climate-modelling

The two major limitations in present-day climate modelling are lack of important ocean data and inadequate models. The subsurface ocean has been observed only sporadically in space and time. This hinders verification of OGCMs and the development of initial conditions for climate prediction.

All types of climate models suffer from both conceptual and practical problems. The simple climate models discussed in Section 3.2 rely for their solution on a small number of assumptions that are known to be violated by the observed climate system. For example, terms in the equations of motion are neglected or approximated. Despite this, simple climate models are useful in helping to understand the results obtained from GCMs. However, GCMs are also based to an uncomfortable extent on assumptions about the behaviour on the unresolved scales, the parameterisations. While there is general agreement on which processes need to be parameterised, there usually exist multiple parameterisation schemes for each process and few compelling reasons for choosing between them. Associated with the parameterisations, the GCM has many more adjustable parameters than the simple climate model. The appropriate range of values for these parameters is in many cases poorly known. Even the number of adjustable parameters in climate GCMs is not well known. Ideally, the values for the parameters would come from measurements. In practice, many parameter values are chosen to reduce errors in the GCM simulation of the current climate. This procedure is known as model tuning. The parameters that are considered tuneable are a matter of taste, sometimes including the solar constant. The parameters of simple climate models are also tuned, but the range of options is much more limited, and the cause and effect relationships better understood than in GCMs. Despite extensive tuning, simulations of the current climate with GCMs have large errors.

A climate GCM contains many complex interactions between the component models. The coupled model may exhibit large errors that are not apparent from inspection of the results from the component models, and which are resistant to tuning. For example, the uncoupled AGCM is integrated by specifying observed SST. The results may have important errors in the heat flux or wind stress at the atmospheric ocean interface. However, the consequences of these errors may not be obvious, until the AGCM is coupled to an OGCM. The coupled model simulation of SST will then have large errors that are immediately apparent. The SST errors can adversely affect the simulation of the sea ice, and lead to potentially incorrect estimates of climate sensitivity. This has in fact been the experience with many coupled GCMs. The process of coupling highlights those processes, which are important for the interactions between the subsystems.

When tuning fails, flux adjustment (Section 3.3.9) is sometimes used to correct errors in the mean quantities of coupled GCMs, such as annual mean SST and annual cycle of SST. With flux correction, specified fluxes are added at the atmosphere-ocean interface to those calculated internally by the model. These fluxes are chosen to minimise the error in the coupled simulation, usually of SST, and the requirement that physical laws be locally or globally satisfied is suspended. Flux correction is obviously a questionable procedure, and the results from flux corrected models should be viewed with suspicion. Many objections have been raised to flux correction, based on results from simpler climate models. On the other hand, there are strong arguments that climate projections with GCMs will be wrong if the model simulation of current climate is too far from the actual climate. These conflicting points of view will be reconciled when climate GCMs are developed that produce good simulations of the current climate without flux correction.

The ultimate test of the climate GCM will be the verification of a large enough number of predictions on the time scale of interest. This may be possible in a few decades for prediction of inter-annual variability of SST in the tropical Pacific associated with ENSO. However, action taken to deal with inter-decadal climate change, produced by anthropogenic greenhouse emissions, will have to rely on projections from models that have not been extensively verified on that time scale.

3.7.1 The different subsystems

The uncoupled component models have been developed independently from each other. Each component model is generally the product of a single scientific discipline. The effect of the limitations of the different subsystems of climate simulation will not be known, until these limitations have been superseded - the limitations introduce uncertainty, but not necessarily error into climate model projections.

a) Ocean

The major limitation to ocean models is currently poor resolution. Ocean models, with the resolution currently used for climate studies, have difficulty in simulating the steep temperature gradients in the tropical upper oceans, called the thermocline, and do not resolve motions on the scale of oceanic "storms", which is known as the Rossby radius. In the ocean the Rossby radius is on the order of 50 km, an order of magnitude smaller than the corresponding scale for the atmosphere. Ocean models that can resolve the Rossby radius are known as "eddy-resolving". The ocean models used in

climate modelling are not eddy resolving, and require much stronger parameterised horizontal mixing to obtain realistic-looking large scale results than the eddy-resolving models. The magnitude and structure of the errors introduced by the low resolution is not known. Computational limitations have prevented the use of eddy-resolving OGCMs in climate modelling. Ocean modelling also suffers from a shortage of data, with which to construct realistic oceanic initial conditions and for use in model verification.

b) Land

Land surface models that have been used in climate modelling have been very crude. Models that include the important effects of plants on heat transfer between the land and atmosphere, essentially by representing the plant and soil properties in each atmospheric grid cell by a single huge plant, are now being implemented for study of climate change. Obviously it would be more realistic to model a collection of many different plant species. The plants in the current land surface models are specified in their distribution and physical properties as a function of time of year. If the climate is affected by the vegetation type, it is important to allow the model climate to influence the distribution of the different types of vegetation. Realistic models, in which the vegetation type is determined by the local climate, do not yet exist.

The surface hydrological models, currently used in climate models, are extremely crude. River flow models, important in both the response of the land surface to climate change and in the fresh water balance of the oceans, are just beginning to be developed for coupling into climate models.

c) Sea ice

Energy balance models demonstrate the potential importance of sea ice feedbacks to climate change. Formation and melting of sea ice is in large part a thermodynamic process. However, simulation of the motion of the ice under the joint influence of the ocean currents and the wind has been found to be a necessary ingredient for realistic simulation of the seasonal variations of the sea ice extent. Some probably oversimplified thermodynamic/dynamic models of sea ice exist and are in the process of being verified and included in climate models.

d) Atmospheric chemistry

Current climate models do not include representations of the transport and chemistry of the important naturally occurring and anthropogenically produced trace species.

3.7.2 The complex interaction

The separate component models are constructed and verified with specified realistic external conditions, those conditions that are the outputs from the other component models. The verification quantities are usually variables internal to the component model, such as the mid-tropospheric circulation field in the case of the AGCM, and not those quantities that are important when the component models are coupled together, such as surface wind stress and heat flux over the oceans, produced by the AGCM. When the component models are coupled together and the coupling quantities are determined internally, the coupled models can and do develop unrealistic climates. The process of coupling highlights those processes, which are important for the interactions between the subsystems. The identification of the important coupling processes and the improvement of their representation can only be accomplished within the framework of the coupled model.

The development of the current coupled atmosphere-ocean-land-sea ice climate models has created a number of problems that have to be overcome. Climate adjustment due to poor initialisation and climate drift due to errors in the model physics have been discussed in Section 3.3. Serious errors occur in all AGCMs in areas like the stratospheric circulation and mean precipitation distribution. When mixed layer oceans are included at the lower boundary, the surface temperature begins to differ from the observed present climate, especially with warmer and more zonally homogenous tropical oceans. These differences are removed by specifying a heat flux "below" the mixed layer ocean. When a fully interactive OGCM is used, the surface temperature simulation can become radically different from the present climate. This could be indicative of serious errors in the model formulations, but the most severe effects can be apparently removed by applying specified fluxes at the air-sea interface. These flux adjustments can correct errors in the mean quantities, such as monthly mean surface temperature and its annual cycle. However, the effect of flux adjustment on the climate anomalies, the prediction of which is the object of climate modelling, are not known.

An active area of research, employing coupled atmosphere-ocean GCMs, is the simulation and prediction of the El Niño Southern Oscillation (ENSO) phenomenon, which dominates the inter-annual variability of tropical climate in the Pacific (see Section 3.6.1). A good simulation of both the annual cycle of SST and its inter-annual variability in the near-equatorial Pacific has proven elusive. Models, which simulate the annual cycle well, do poorly at simulating inter-annual variability and vice versa. The errors are so severe that models, which have mediocre simulations of both phenomena, can perhaps claim to be the best overall. Models, which use flux adjustment to assure the verisimilitude of the annual cycle, have been as successful or more successful than models, which do not use flux adjustment in ENSO prediction.

Coupled model sea ice simulations without flux adjustment are not yet satisfactory. The AGCMs and OGCMs have particular mathematical problems near the poles, and the simulations by the component models near the North Pole are usually poor by themselves, which leads to poor simulations of the sea ice extent and duration.

3.8 Discussion

The current climate models do not take into account several feedbacks. This is partly because many of the feedbacks are not well understood and therefore difficult to model, and partly because the computational needs of such models can be prohibitive. The biogeochemical cycles in particular have a very long time scale and it is not clear if a complex model of weather and climate needs to be integrated for hundreds of thousands of years to investigate the possible interaction between biogeochemical cycles and cycles of water and energy (see Chapter 4). The current models also do not have an adequate treatment of solar cycles and glacial cycles. In order to investigate the role of long-period global cycles, it may be necessary to develop simpler models of the fast components of climate i.e. atmosphere, land and upper oceans, as discussed in the next chapters.

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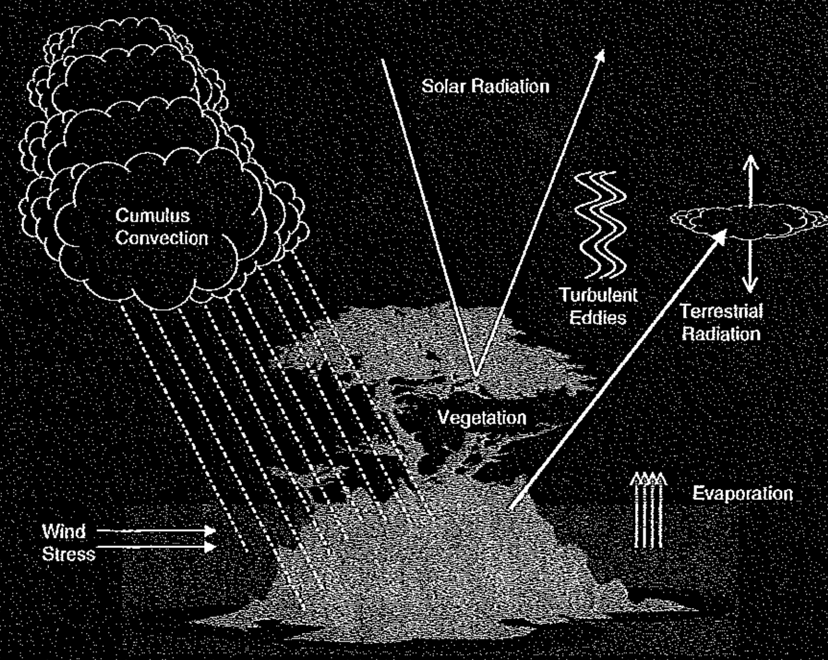

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