CHAPTER 2

A History of and Introduction to Climate Models

> The mathematical problem is not yet defined: there are more unknowns than equations.
>
> C. G. Rossby (1946)

2.1 INTRODUCING CLIMATE MODELLING

Any climate model is an attempt to represent the many processes that produce climate. The objective is to understand these processes and to predict the effects of changes and interactions. This characterization is accomplished by describing the climate system in terms of basic physical, chemical and biological principles. Hence, a numerical model can be considered as being comprised of a series of equations expressing these laws. Climate models can be slow and costly to use, even on the fastest computer, and the results can only be approximations.

The need for simplification

For several reasons, a model must be a simplification of the real world. The processes of the climate system are not fully understood, although they are known to be complex. Rossby was alluding to this problem in the quotation at the start of this chapter. Furthermore, the components of the climate system interact with each other, producing feedbacks (Section 1.4), so that any solution of the governing equations must involve a great deal of computation. The solutions that are produced start from some initialized state and investigate the effects of changes in a particular component of the climate system. The boundary conditions, for example the solar radiation, sea-surface temperatures or vegetation distribution in the case of the atmosphere, or the bathymetry and atmospheric wind field in the case of the ocean, are set from observational data or other simulations. These data are rarely complete or of adequate accuracy to specify completely the environmental conditions, so that there is inherent uncertainty in the results.

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Today's large-scale coupled climate system models, designed to simulate the climate of the planet, take into account the whole climate system (see Figure 1.2). All of the interactions between the components must be integrated in order to develop such a model. This presents great problems because the various interactions operate on different time-scales. For example, the effects of changes in deep water formation in the ocean may be very important when considering climate averaged over decades to centuries, while local changes in wind direction may be unimportant on this time-scale. If, on the other hand, monthly time-scales are of concern, the relative importance would be reversed.

Early global models were of the atmosphere alone and were initially used to generate average conditions for January and July. This was usually done by maintaining forcing appropriate to one particular month and running the model for hundreds of days. These models typically did not include the diurnal cycle and were termed 'perpetual January' or 'perpetual July' (depending on forcing). This is not to imply that a particular January in the period for which a climate model prediction is made would have these conditions, only that the conditions apply to an average January. The latest climate models now include many components, most importantly the ocean and atmosphere, and are now routinely run for hundreds of years with diurnally varying radiation and for multi-year seasonal cycles, and these are used to produce 'climate' averages. The availability of faster computers has introduced the idea of 'ensemble runs'. In such experiments, the modellers carefully perturb initial conditions for each of a collection of model runs, producing an ensemble set. It is always implied that any 'new' climate predicted will have variation about the mean, just as with the present climate. Such experiments help place limits on the variation in climate. This is important when the results of global-scale models are used to estimate the possible impact of climatic change in a local or regional area, or in detection of a climatic change.

The simplifications that must be made in the laws governing climatic processes can be approached in several ways. Consequently, numerous different global-scale climate models are available. In general, two sets of simplifications need to be made. The first involves the processes themselves. It is usually possible to treat in detail some of the processes, specifying their governing equations fairly fully. However, other processes must be treated in an approximate way, either because of our lack of exact information, lack of understanding or because there are still inadequate computer resources to deal with them. For example, it might be decided to treat the radiation processes in great detail, but only approximate the horizontal energy flows associated with regional-scale winds. The approximation may be approached either by using available observational data, the empirical approach, or through specification of the physical laws involved, the theoretical (or conceptual) approach.

Resolution in time and space

The second set of simplifications involves the resolution of the model in both time and space (see Figure 1.1). While it is generally assumed that finer spatial resolutions produce more reliable results, constraints of both data availability and computational time may dictate that a model may have to have, for example, latitudinally averaged values as the basic input. In addition, too fine a resolution may be inappropriate because processes acting on a smaller scale than the model is designed to resolve may be inadvertently incorporated. Similar considerations are involved in the choice of temporal resolution. Most computational procedures require a 'timestep' approach to calculations. The processes are allowed to act for a certain length of time and the new conditions are calculated. The process is then repeated using these new values. This continues until the conditions at the required time have been established. Timestepping is a natural consequence of there not being a steady-state solution to the model equations. Although accuracy potentially increases as the timestep size decreases, there are constraints imposed by data, computational capacity and the design of the model. The time and space resolutions of the model are also linked, as will be explained in Chapter 5.

Although models are designed to aid in predicting future climates, performance can only be tested against the past or present climate. Usually when a model is developed, an initial objective is to test the sensitivity of the model and to ascertain how well its results compare with the present climate. Thereafter it may be used to simulate past climates, not only to see how well it performs but also to gain insight into the causes and features of these climates. Although such past climates are by no means well known, this comparison provides a very useful step in establishing the validity of the modelling approach. After such tests, the model may be used to gain insight into possible future climates.

2.2 TYPES OF CLIMATE MODELS

The important components to be considered in constructing or understanding a model of the climate system are:

1. *Radiation* – the way in which the input and absorption of solar radiation by the atmosphere or ocean and the emission of infrared radiation are handled;
2. *Dynamics* – the movement of energy around the globe by winds and ocean currents (specifically from low to high latitudes) and vertical movements (e.g. small-scale turbulence, convection and deep-water formation);
3. *Surface processes* – inclusion of the effects of sea and land ice, snow, vegetation and the resultant change in albedo, emissivity and surface–atmosphere energy and moisture interchanges;
4. *Chemistry* – the chemical composition of the atmosphere and the interactions with other components (e.g. carbon exchanges between ocean, land and atmosphere);
5. *Resolution in both time and space* – the timestep of the model and the horizontal and vertical scales resolved.

The relative importance of these processes and the theoretical (as opposed to empirical) basis for parameterizations employed in their incorporation can be discussed using the climate modelling pyramid (Figure 2.1). The edges represent the basic elements of the models, with complexity shown increasing upwards. Around the base of the pyramid are the simpler climate models which incorporate only one primary process. There are four basic types of model.

1. Energy balance models (EBMs) are zero- or one-dimensional models predicting the surface (strictly the sea-level) temperature as a function of the energy balance of the Earth. Simplified relationships are used to calculate the terms contributing to the energy balance in each latitude zone in the one-dimensional case.
2. One-dimensional models such as radiative–convective (RC) models and single column models (SCMs) focus on processes in the vertical. RC models compute the (usually global average) temperature profile by explicit modelling of radiative processes and a 'convective adjustment' which re-establishes a predetermined lapse rate. SCMs are single columns 'extracted' from a three-dimensional model and include all the processes that would be modelled in the three-dimensional version but without any of the horizontal energy transfers.
3. Dimensionally constrained models now take a wide variety of forms. The oldest are the statistical dynamical (SD) models, which deal explicitly with surface processes and dynamics in a zonally averaged framework and have a vertically resolved atmosphere. These models have been the starting point for the incorporation of reaction chemistry in global models and are still used in some Earth Models of Intermediate Complexity (EMICs).
4. Global circulation models (GCMs). The three-dimensional nature of the atmosphere and ocean is incorporated. These models can exist as fully coupled ocean–atmosphere models or 'coupled climate system models' or, for testing and evaluation, as independent ocean or atmospheric circulation models. These models attempt to simulate as many processes as possible and produce a three-dimensional picture of the time evolution of the state of the ocean and atmosphere. Vertical resolution is typically much finer than horizontal resolution but, even so, the number of layers is usually much less than the number of columns.

The vertical axis in Figure 2.1 shows increasing complexity (i.e. more processes included and linked together) and also indicates increasing resolution: models appearing higher up the pyramid tend to have higher spatial and temporal resolutions.

There is ambiguity concerning the expansion of GCM. Two possible terms are the more recent 'global climate model' and the older 'general circulation model'. The latter also refers to a weather forecast model so that in climate studies GCM is understood to mean 'general circulation climate model'. A further distinction has historically been drawn between oceanic general circulation models and atmospheric general circulation models by terming them OGCMs and AGCMs. As the pyramid is ascended, more processes are integrated to develop a coupled ocean–atmosphere global model (OAGCM or CGCM). It has been suggested that, as processes that are currently fixed come to be incorporated into GCMs, the coupling will be more complete, say including changing biomes (an AOBGCM) or changes in atmospheric,

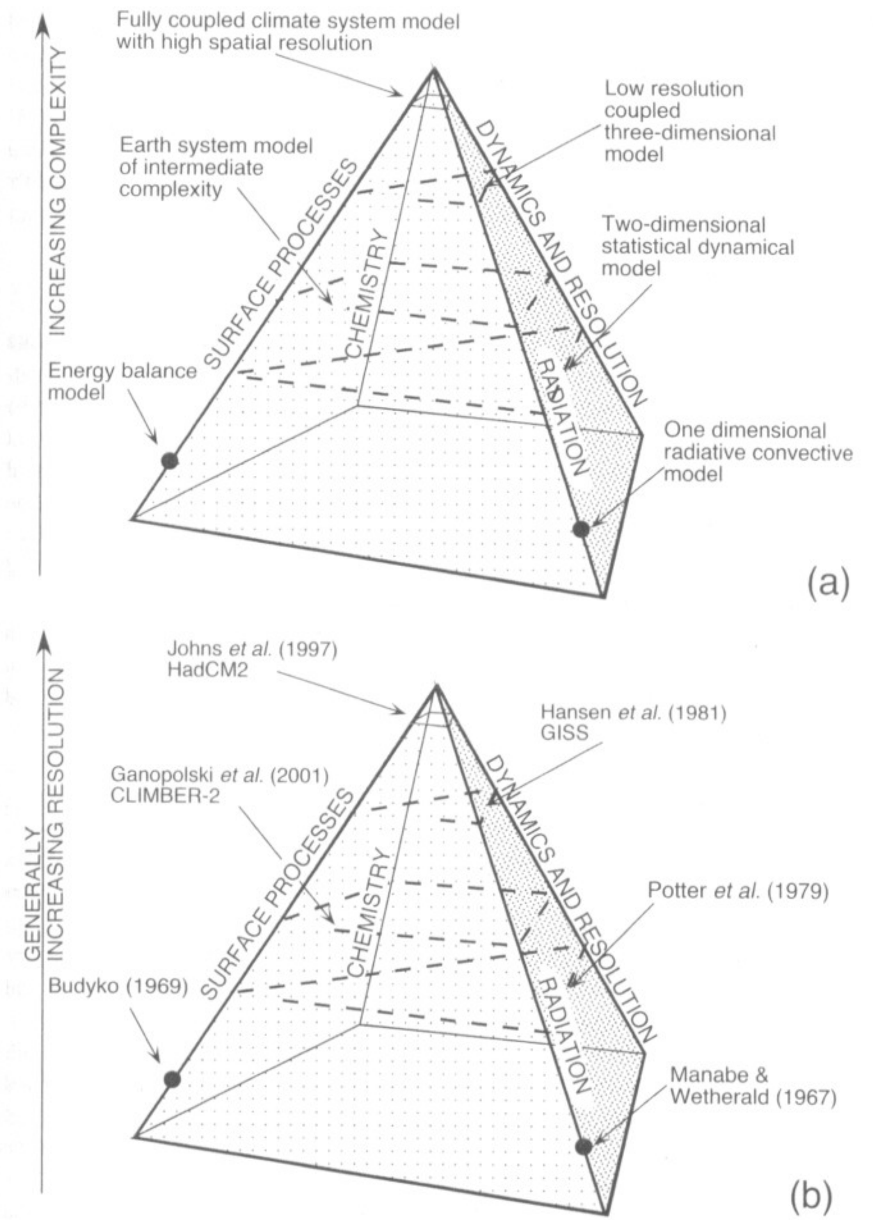

Figure 2.1 The climate modelling pyramid. The position of a model on the pyramid indicates the complexity with which the four primary processes (dynamics, radiation, surface and oceans and chemistry) interact. Progression up the pyramid leads to greater interaction between each primary process. The vertical axis is not intended to be quantitative. (a) The position of various model types; (b) Examples from the literature and their positions on the pyramid

ocean and even soil chemistry. Such models are becoming known as 'coupled climate system models' or 'Earth system models'. From being the only components in GCMs, the atmosphere and ocean are now parts of modular software packages designed to tackle a wide variety of problems. In this book, the generic term 'GCM' is used to mean a complex three-dimensional model of the atmosphere and ocean incorporating other components and used for climate simulation. As in the broader literature, the particular meaning will be clear from the context.

2.2.1 Energy balance climate models

These models have been instrumental in increasing our understanding of the climate system and in the development of new parameterizations and methods of evaluating sensitivity for more complex and realistic models. This type of model can be readily programmed and implemented on most small computers and the inherent simplicity of EBMs combined with the ease of interpreting results make them ideal instructional tools. They are widely used to investigate the sensitivity of the climate system to external changes and to interpret the results of more complex models. Energy balance models are discussed more fully in Chapter 3 and codes are included on the Primer CD (see Appendix C).

Energy balance models are generally one-dimensional, the dimension in which they vary being latitude. Vertical variations are ignored and the models are used with surface temperature as the predicted variable. Since the energy balance is allowed to vary from latitude to latitude, a horizontal energy transfer term must be introduced, so that the basic equation for the energy balance at each latitude, ϕ, is

$$C_m[\Delta T(\phi)/\Delta t] = R\downarrow(\phi) - R\uparrow(\phi) + \text{net transport into zone } \phi \qquad (2.1)$$

where C_m is the heat capacity of the system and can be thought of as the system's 'thermal inertia' and $R\downarrow$ and $R\uparrow$ are the incoming and outgoing radiation fluxes respectively.

The radiation fluxes at the Earth's surface must be parameterized with care since conditions in the vertical are not considered in this type of model. To a large extent the effects of vertical temperature changes are treated implicitly. In a clear atmosphere, convective effects tend to ensure that the lapse rate remains fairly constant. However, cloud amount depends only weakly on surface temperature, so that cloud albedo is only partially incorporated in the model. In particular, clouds in regions of high temperatures, such as the intertropical convergence zone, are ignored in the parameterization of albedos in EBMs.

Atmospheric dynamics are not modelled in an EBM; rather it is assumed that a 'diffusion' approximation is adequate for including heat transport. This approximation relates energy flow directly to the latitudinal temperature gradient. This flow is usually expressed as being proportional to the deviation of the zonal temperature, T, from the global mean, $\bar{T}$. When using the model for annual average calculations, the surface albedo can be regarded as constant for a given latitude. This type of model, however, can also be used for seasonal calculations. In this case, it is usual to allow the albedo to vary with temperature to simulate the effects of changes in sea ice and snow extent.

Early EBMs were originally found to be stable only for small perturbations away from present-day conditions. For instance, they predicted the existence of an ice-covered state for the Earth for only slight reductions in the present solar constant. This result prompted studies of the sensitivity of various climate model types to perturbations (see Section 2.4).

2.2.2 One-dimensional radiative–convective climate models

One-dimensional RC models represent an alternative approach to relatively simple modelling of the climate and they also occur at the bottom of the modelling pyramid (Figure 2.1). In this case the 'one dimension' in the name refers to altitude. One-dimensional RC models are designed with an emphasis on the global average surface temperature, although temperatures at various levels in the atmosphere can be obtained.

The main emphasis in these models is on the explicit calculation of the fluxes of solar and terrestrial radiation (the radiation streams). Given an initially isothermal atmosphere, the heating rates for a number of layers in the atmosphere are calculated, although the cloud amount, optical properties and the albedo of the surface generally need to be specified. The temperature change in each layer which results from an imbalance between the net radiation at the top and bottom of the layer is calculated. At the end of each timestep a revised radiative temperature profile is produced. If the calculated lapse rate exceeds some predetermined 'critical' lapse rate, the atmosphere is presumed to be convectively unstable. An amount of vertical mixing, sufficient to re-establish the prescribed lapse rate, is carried out and the model proceeds to calculate the next radiative timestep. This procedure continues until convective readjustment is no longer required and the net fluxes for each layer approach zero. One-dimensional RC models operate under the constraints that at the top of the atmosphere there must be a balance of shortwave and longwave fluxes and that surface energy gained by radiation equals that lost by convection. However, they vary in the way they incorporate the critical lapse rate. Some use the dry adiabatic lapse rate, some the saturated one, while many use a value of 6.5 K km^{-1}, which is the value in an observed standard atmospheric profile. Similarly, different humidity and cloud formulations are possible.

Radiative–convective models (discussed more fully in Chapter 4) can be constructed either as equilibrium models or in a time-dependent form. FORTRAN code for the latter type is included on the Primer CD – see Appendix C. These models can also be given an additional dimension and applied to zonally averaged conditions, by including a description of the horizontal energy transport. The main use of radiative–convective models is to study the effects of changing atmospheric composition and to investigate the likely relative influences of different external and internal forcings. They are the basis for the 'column' models that have recently begun to be used to evaluate aspects of the parameterizations of the atmospheric (and surface) 'columns' in more complex GCMs. Column models are, in effect, single columns from a GCM and include the sophisticated physics usually found in these models.

2.2.3 Dimensionally-constrained climate models

Dimensionally-constrained climate models typically represent either two horizontal dimensions or the vertical plus one horizontal dimension. The latter were originally more common, combining the latitudinal dimension of the energy balance models with the vertical one of the radiative–convective models. These models also tended to include a more realistic parameterization of the latitudinal energy transports. In such models, the general circulation is assumed to be composed mainly of a cellular flow between latitudes, which is characterized using a combination of empirical and theoretical formulations. A set of statistics summarizes the wind speeds and directions while an eddy diffusion coefficient of the type used in EBMs governs energy transport. As a consequence of this approach, these models are called 'statistical dynamical' (SD) models. These 2D SDs can be considered as the first attempts at Earth modelling with intermediate complexity – the EMICs.

EMICs are about one-third of the way up the modelling pyramid (Figure 2.1), being more complicated than the vertically or latitudinally resolved one-dimensional models. Indeed, as we shall see in Chapter 4, many EMICs now claim to represent fractionally more than two dimensions and some even represent all three but with very coarse spatial or temporal resolution. Their use has provided insight into the operation of the present climate system, for example showing that the relatively simple diffusion coefficient approach for poleward energy transports is appropriate, provided that the coefficient, as well as the transport, is allowed to vary with the latitudinal temperature gradient. Advances in the understanding of baroclinic waves were achieved from studies of the results of 2D SD models. Dimensionally-constrained models have been employed to make simulations of the chemistry of the stratosphere and mesosphere. These models typically involve the modelling of tens to hundreds of chemical species and many hundreds of different reactions, and are much more demanding of computer time than atmosphere-only 2D models. Although traditional two-dimensional models are insensitive to changes within a latitude band, a compromise (and fractionally increased dimensionality) may be obtained by considering each zone as being divided into a land and ocean part. This type of 'two-channel' approach is discussed with reference to a more complex EBM in Section 4.9.

As a result of the lack of full three-dimensional resolution and the increased availability of computer resources enabling many more people to run GCMs, two-dimensional SD models have been largely superseded for consideration of the effect of perturbations on the present climate and for purposes such as IPCC. However, use of this type of model has blossomed recently in applications involving socio-economic change and climate assessments. These modern dimensionally-

constrained models, the EMICs, proudly abandon physical dimensions specifically to incorporate human systems, their impacts and susceptibilities.

2.2.4 General circulation models

The aim of GCMs is the calculation of the full three-dimensional character of the atmosphere or ocean (Figure 2.2). The solution of a series of equations (Table 2.1)

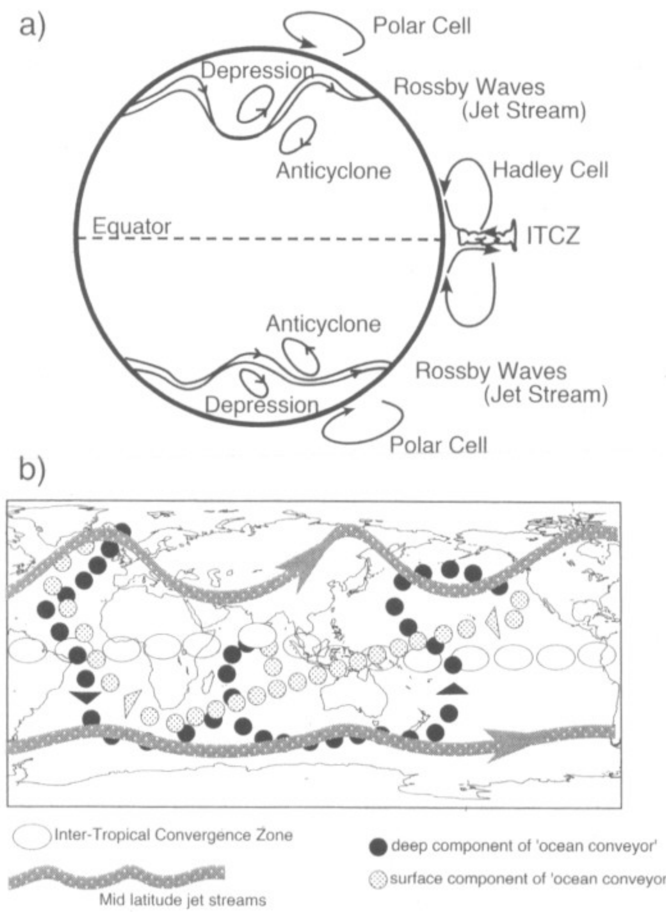

Figure 2.2 Illustration of the main features of the atmospheric (a) atmospheric and oceanic (b) circulation. The atmospheric circulation is determined primarily by the net radiation budgets (excess in the tropics and deficit near the poles) and the rotation of the Earth (especially the Rossby waves). The thermohaline circulation of the ocean (lines of shaded circles), often referred to as the 'ocean conveyor belt', results in the movement of water throughout the major ocean basins of the world over periods of hundreds to thousands of years. The black circles show the deep ocean conveyor, and the grey the surface component (see also Figure 1.17)

Table 2.1 Fundamental equations solved in GCMs

1. *Conservation of energy* (the first law of thermodynamics)
   i.e. Input energy = increase in internal energy plus work done
2. *Conservation of momentum* (Newton's second law of motion)
   i.e. Force = mass × acceleration
3. *Conservation of mass* (the continuity equation)
   i.e. The sum of the gradients of the product of density and flow-speed in the three orthogonal directions is zero. This must be applied to air and moisture for the atmosphere and to water and salt for the oceans, but can also be applied to other atmospheric and oceanic 'tracers' such as cloud liquid water.
4. *Ideal gas law* (an approximation to the equation of state – atmosphere only)
   i.e. Pressure × volume is proportional to absolute temperature × density

that describe the movement of energy, momentum and various tracers (e.g. water vapour in the atmosphere and salt in the oceans) and the conservation of mass is therefore required. Generally the equations are solved to give the mass movement (i.e. wind field or ocean currents) at the next timestep, but models must also include processes such as cloud and sea ice formation and heat, moisture and salt transport. The first step in obtaining a solution is to specify the atmospheric and oceanic conditions at a number of 'grid points', obtained by dividing the Earth's surface into a series of rectangles, so that a traditionally regular grid results (Figure 2.3). Conditions are specified at each grid point for the surface and several layers in the atmosphere and ocean. The resulting set of coupled non-linear equations is then solved at each grid point using numerical techniques. Various techniques are available, but all use a timestep approach.

Although GCMs formulated in this way have the potential to closely approach the real oceanic and atmospheric situation, at present there are a number of practical and theoretical limitations. The prime practical consideration is of the time needed for the calculations. For example, one particular low-resolution AGCM requires around 48 Mbytes of memory, whereas a more recent, higher resolution, version of the model requires over 160 Mbytes. Much of this stored information must be accessed and updated at each model timestep and this places a strain on the resources of even the largest and fastest computers (cf. Figure 1.5). Since the accuracy of the model partly depends on the spatial resolution of the grid points and the length of the timestep, a compromise must be made between the resolution desired, the length of integration and the computational facilities available. At present, atmospheric grid points are typically spaced between 2° and 5° of latitude and longitude apart and timesteps of approximately 20–30 minutes are used. Vertical resolution is obtained by dividing the atmosphere into between six and fifty levels, with about twenty levels being typical.

The ocean is a three-dimensional fluid that must be modelled using the same principles as for the atmosphere. As well as acting as a thermal 'fly-wheel' for the climate

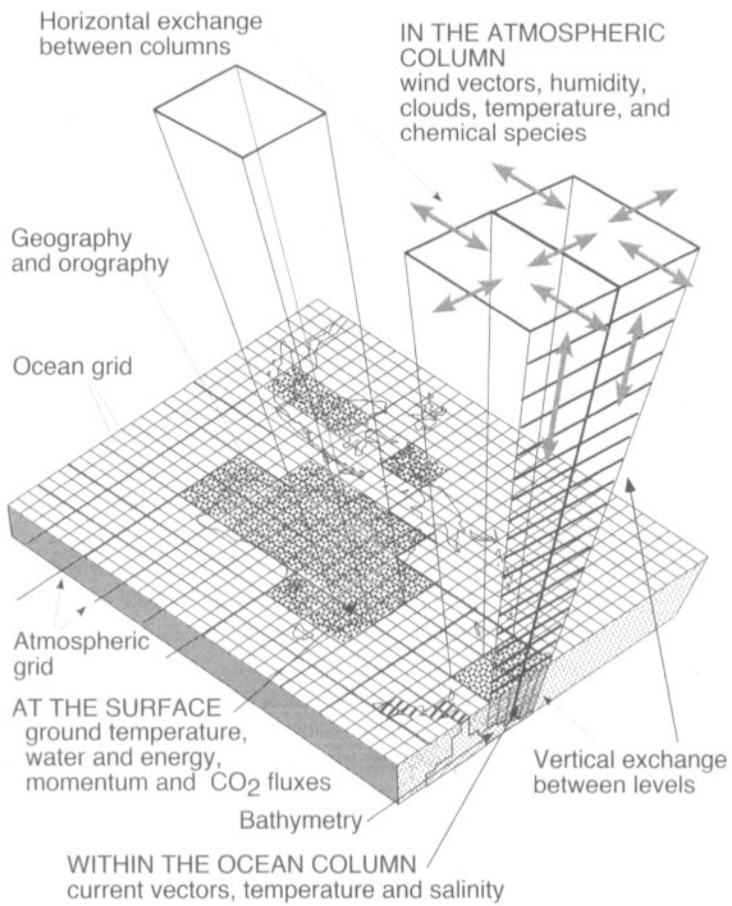

Figure 2.3 Illustration of the basic characteristics of a three-dimensional climate model, showing the manner in which the atmosphere and ocean are split into columns. Both atmosphere and ocean are modelled as a set of interacting columns distributed across the Earth's surface. The resolutions of the atmosphere and ocean models are usually different

system, the ocean also plays a central role in the carbon cycle, accounting for approximately half of the carbon absorbed from the atmosphere every year. The dynamics of the ocean are governed by the amount of radiation available at the surface and by the wind stresses imposed by the atmosphere. Ocean modellers must also track the salt in the ocean. Evaporation, precipitation, sea ice formation and river discharge affect the salinity of the ocean, which in turn affects the density of the water. The flow of ocean currents is also constrained by the positions and shapes of the continents (Figure 2.2). Ocean GCMs calculate the temporal evolution of oceanic variables (velocity, temperature and salinity) on a three-dimensional grid of points spanning the global ocean domain. Although early climate model simulations incorporated only very simple models of the ocean, which do not explicitly include ocean dynamics, the incorporation of a dynamic ocean is now an essential part of any state-of-the-art climate model.

Modelling a full three-dimensional ocean is made difficult by the fact that the scale of motions in the oceans is much smaller than in the atmosphere (ocean eddies are around 10–50 km compared to around 1000 km for atmospheric eddies) and that the ocean also takes very much longer to respond to external changes (cf. Table 1.2). The deep water circulation of the ocean (Figure 2.2) can take hundreds or even thousands of years to complete. Ocean models that include these dynamic processes are now routinely coupled with atmospheric GCMs to provide our most detailed models of the climate system. The formation of oceanic deep water is closely coupled to the formation and growth of sea ice, so that representative ocean dynamics demands effective modelling of the dynamics and thermodynamics of sea ice. Modelling groups are continuously faced with the problem of dealing with a complex, interacting and diverse collection of models, demanding new skills and approaches.

Originally, computational constraints dictated that global circulation models could only run for very short periods. For the atmosphere this meant only simulating a particular month or season, rather than a full seasonal cycle, although now all models include a seasonal cycle and most include a diurnal cycle. For the oceans, restrictions of computer power meant that the models were used before they had fully equilibrated. This could result in the 'drift' of the ocean climate away from present-day conditions, which was often corrected by applying adjusting fluxes at the ocean surface to compensate for systematic errors which persist at equilibrium. This was a particular problem for early coupled OAGCMs, but most modern coupled models have overcome this problem. The importance of removing such arbitrary adjustments and of including realistic time-dependent phenomena is now well established, and modellers have striven to include increasing numbers of these phenomena as well as using the increased computer power to provide higher resolution and better physics (cf. Figure 2.1).

It is important to identify the very different aims of those developing and using GCMs as compared to the designers of numerical weather forecast models. The latter are prediction tools, while GCMs can represent only probable conditions. For this reason, many GCM integrations must be performed and their results averaged to generate an ensemble before a climate prediction can be made.

Computational constraints lead to problems of a more theoretical nature. With a coarse grid spacing, small-scale atmospheric motions (termed sub-gridscale), such as thundercloud formation, cannot be modelled, however important they may be for real atmospheric dynamics. Fine grid models can be used for weather prediction because the integration time is short. In contrast, climate models must mostly rely on some form of parameterization of sub-gridscale processes (see Section 5.2.4). Some progress has been made in incorporating cloud-resolving models into GCMs and this is discussed in Chapter 5.

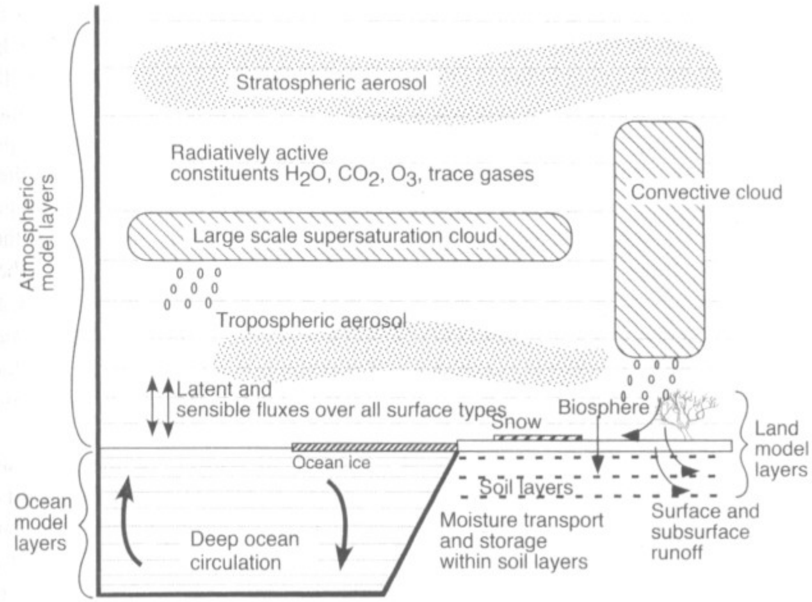

Figure 2.4 Schematic illustration of the processes in a single column of a global circulation climate model. In most models, two types of cloud are treated. In this example, soil moisture is modelled in a number of layers, and tropospheric and stratospheric aerosols are included. (Reproduced with permission from Hansen *et al.* (1983), *Mon. Wea. Rev.*, **111**, 609–662

Some of the processes usually incorporated into global circulation climate models are shown in Figure 2.4. Within the atmosphere, modellers adopt an approach similar to that used for the RC models in calculating heating rates (although they are often computationally simpler), but also often include cloud formation processes as part of the convection and consider in detail the effects of horizontal transport. Ocean models must take into account how the radiation from the atmosphere is absorbed in the upper layers of the ocean in an analogous manner along with the factors that affect the ocean salinity. The interaction between the land or ocean surface and the near-surface layer of the atmosphere, however, must be parameterized. Detailed consideration of these transfer processes is computationally too demanding for explicit inclusion. Commonly, the surface fluxes of momentum, sensible heat and moisture are taken to be proportional to the product of the surface wind speed and the gradient of the property away from the surface. More detailed aspects of ocean and atmospheric circulation models will be considered in Chapter 5.

2.2.5 Stable isotopes and interactive biogeochemistry

The many roles of the biosphere of importance to the climate include the exchange of carbon and other elements; the transfer of moisture from the soil into the atmosphere; modification of the albedo, which changes the amount of radiation absorbed by the climate system; and modification of the surface roughness, which alters the exchange of momentum. The interactive nature of the plant life of the planet has only fairly recently been included in climate models. The first approach was to delineate geographic boundaries of biomes (species characterized by similar climate demands) using simple predictors available from GCMs such as temperature, precipitation and possibly sunshine or cloudiness. Attempts made to evaluate these methods included using palaeo-reconstructions of vegetation cover during past epochs. Recently, modellers have included ecological succession models into their GCMs and have been able to make sub-gridscale features of the terrestrial biosphere interactive. These interactive biosphere models are still in their infancy but are beginning to provide useful predictions of responses of the biosphere to climate including the issue of possible future CO_2 fertilization of the biosphere. Tracking various isotopes in the water cycle has illuminated diverse aspects of bio-climate modelling and model validation.

Isotopic measurements have been used to illuminate aspects of the water and chemical budgets of the Amazon Basin. The Amazon drains around one-third of the continental area of South America generating a massive discharge totalling about 20 per cent of the freshwater influx to the world's oceans. Understanding such an important source of non-saline water is critical for the ocean's climate, but the Amazon puzzled mid-twentieth century climate scientists. It was known that the basin-average Amazon precipitation is about $2200\,\mathrm{mm\,yr^{-1}}$ (which, multiplying by the basin's area of $6.5 \times 10^6\,\mathrm{km}^2$, implies a total water influx to the basin of $\sim 14 \times 10^{12}\,\mathrm{m^3\,yr^{-1}}$) but the Amazon's ultimate water discharge to the sea is 'only' $6 \times 10^{12}\,\mathrm{m^3\,yr^{-1}}$ – still a massive flow. So, something happens to $8 \times 10^{12}\,\mathrm{m}^3$ of water every year in the Amazon system. This mystery of the almost 60 per cent of rainfall that does not run to the sea was solved in the 1970s using measurements of the stable isotopes of water.

The dominant atmospheric flow over the Amazon is along the equator from east to west. Water evaporates from the equatorial Atlantic and this moist air is carried by the trade winds up-river to the Andes. Precipitation falls as the air passes over the land and is lifted towards the mountains (Figure 2.5a). If this were as simple as depicted, all the precipitation would appear as river discharge instead of 60 per cent being 'lost'. Also the rainfall would display a straightforward decrease in heavy water isotopes, $^1HD^{16}O$ and $^1H_2{}^{18}O$, because these form precipitation more readily than the common and lighter water molecule $^1H_2{}^{16}O$. Measurements of D and ^{18}O enrichments do show fairly steady decreases inland over all continents but, in the Amazon, the slopes are much shallower than anywhere else. It seems that some of the 'heavy' rain falling in the Amazon re-enters the atmosphere. Efficient recycling of moisture re-inserts heavier isotopes (as well as normal water) back into the atmosphere, and this is the reason that the depletions of D and ^{18}O measured in Amazon rainfall reduce more slowly inland than in other continents (Figure 2.5b). This means that most evaporation is not from water bodies such as lakes and the river itself, because these would preferentially evaporate light isotopes. The majority of the Amazon's water recycling must be transpiration through plants or re-evaporation of

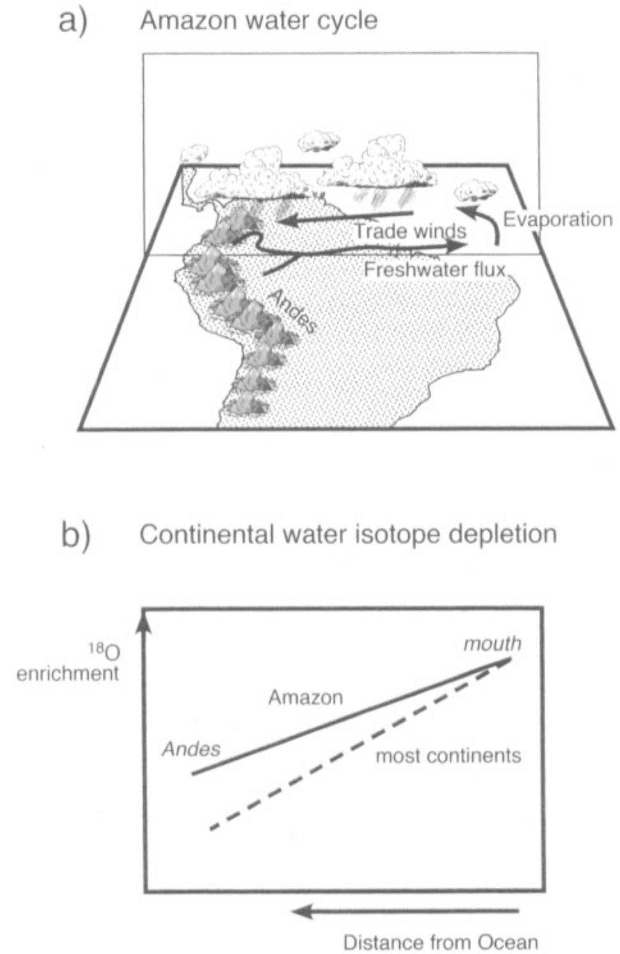

Figure 2.5 (a) Schematic illustration of the water cycle in the Amazon Basin. The Andes Mountains provide an effective barrier to moisture from the Pacific Ocean, meaning that moisture in the upper basin is transported from the Atlantic Ocean and is returned to the ocean by the river. (b) The progressive recycling of moisture by non-fractionating processes (transpiration and canopy evaporation) as it travels from the mouth to the Andes means that the gradient of heavy isotope enrichment is less than for other, less heavily vegetated continents

water caught on foliage: both are non-fractionating processes i.e. they do not distinguish between light and heavy isotopes.

The isotopic measurements showed that the Amazon Basin recycles about half its water. Specifically, the central Amazon has a water recycling time of about 5.5 days and, during this period, about half of all rainfall is re-evaporated or transpired and, of this, around 50 per cent falls again as precipitation. This moisture recycling within the Amazon Basin leads to a seasonally averaged downward gradient of only 1.5‰ per 1000 km in ^{18}O going inland on an east to west transect as compared with 2.0‰ decrease observed for other continents. So, the puzzle of the missing Amazonian water was really an illusion. The river outflow really equals the available water but it is counted as precipitation many times.

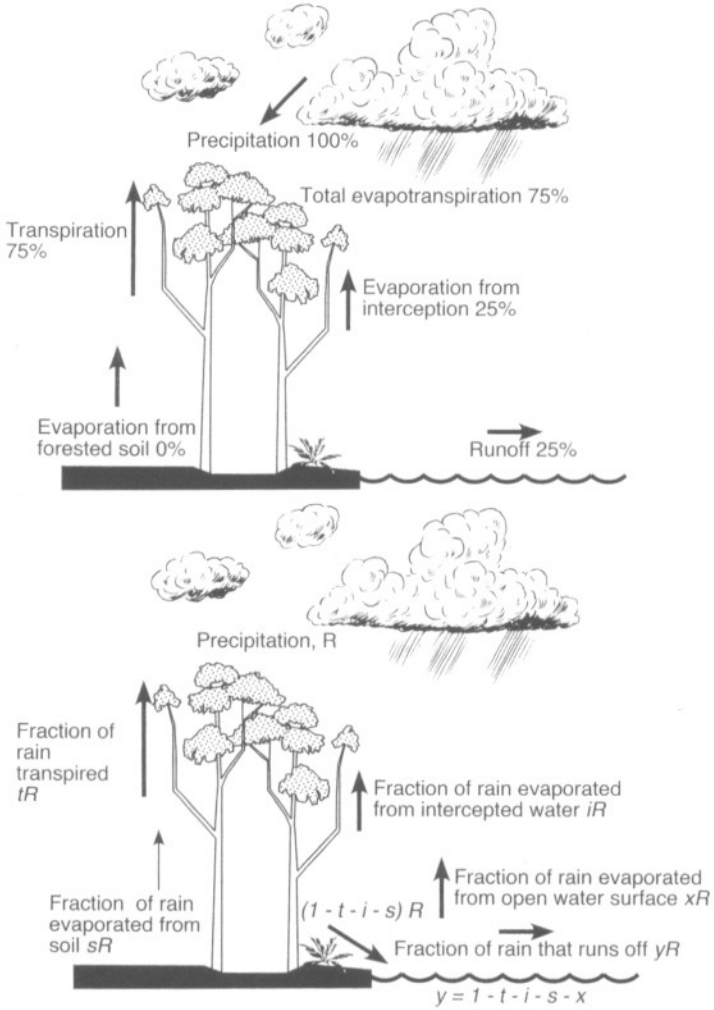

Figure 2.6 The hydrological cycle of the Amazon forest from a traditional viewpoint (top) and from an isotopic viewpoint (lower). In an isotopic view, the moisture fluxes must be differentiated into fractionating (separates heavier and lighter isotopes) and non-fractionating (no preferential separation) processes. This fractionation can be seen in the values of the fractions (y, t, i, s and x) of total rainfall (R). Typical values for y are between 0.25 and 0.35 and in the Amazon $t + i >> s + x$ because non-fractionating processes dominate (resulting in the gradient shown in Figure 2.5b)

This isotopic dimension focused attention on the importance of the biosphere in this major basin's hydrology around the time that GCMs acquired the ability to simulate some aspects of land–atmosphere interactions. The challenge for GCMs, to simulate the partition of Amazonian rainfall into appropriate proportions of evaporation, transpiration and runoff so that the gross basin hydrology is correct (i.e. only one-third of the rain going into runoff) and so that the isotopic recycling occurs through non-fractionating processes, remains today. The different representations of the relative proportions of runoff, re-evaporation from the canopy, transpiration and other evaporative components (Figure 2.6) may account for the range of temperature sensitivities among the large number of GCM simulations of Amazonian deforestation (see Figure 1.15b). GCMs have only recently begun to include open water elements such as lakes and rivers and, as yet, very few track isotopic ratios.

The stable isotopes of carbon (^{13}C and ^{12}C) have also begun to be incorporated into some biospheric components of GCMs. This inclusion is to try to improve understanding of the substantial year-to-year variation in the annual increase in atmospheric carbon dioxide despite the relatively constant input due to fossil fuel emissions. Interannual variations in the uptake of carbon by the biosphere are, very likely, responsible for this observed variation. The biosphere, particularly in heavily forested regions such as the Amazon, responds strongly to seasonal and interannual variations in the environment. The isotopic fractionation of stable carbon isotopes in various processes in the biosphere provides a means of studying the seasonal and interannual variations in biospheric activity. The $^{13}C/^{12}C$ ratio in plant material provides information about the physiological characteristics of the plant over the time the carbon was fixed and, together with atmospheric measurements of isotopes in CO_2, biospheric activity can thus be quantified.

2.3 HISTORY OF CLIMATE MODELLING

As climate models are readily described in terms of an hierarchy (e.g. Figure 2.1), it is often assumed that the simpler models were the first to be developed, with the more complex GCMs being developed most recently. This is not the case. Norman Phillips performed the classic global circulation computations in the mid-1950s. His model was limited as he had only 5 kilobytes of computer memory available (barely enough to store the textual information on this page) but it was successful. His model atmosphere was a cylindrical sheet to avoid complex geometry, with heating at the bottom and cooling at the top. His results demonstrated that it was possible to simulate the motion of the atmosphere on monthly and longer time-scales. This experiment led directly to the first atmospheric general circulation climate models (as we know them) being developed in the early 1960s, concurrently with the first RC models. Energy balance climate models, as they are currently known, were not described in the literature until 1969, and the first discussion of two-dimensional SD models was in 1970. The latter metamorphosed into EMICs in the 1990s and now represent the fastest evolving model group.

The first atmospheric general circulation climate models were derived directly from numerical models of the atmosphere designed for short-term weather forecasting. These had been developed during the 1950s and, around 1960, ideas were being formulated for longer period integrations of these numerical weather prediction schemes. It is in fact rather difficult to identify the transition point in many modelling groups. For example, Syukuro Manabe joined the National Oceanic and Atmospheric Administration's Geophysical Fluid Dynamics Laboratory (GFDL) in the USA in 1959 to collaborate in the numerical weather prediction efforts, and was to go on to become one of the world leaders in the climate modelling community. Scientists concerned with extending numerical prediction schemes to encompass hemispheric or global domains were also studying the radiative and thermal equilibrium of the Earth–atmosphere system. It was these studies that prompted the design of the RC models, which were once again spearheaded by Manabe, the first of these being published in 1961.

Other workers, such as Julián Adem, also expanded the domain of numerical weather prediction schemes in order to derive global climate models. The low-resolution thermodynamic model first described by Adem in 1965 is an interesting type of climate model, since it lies part-way towards the apex of the climate modelling pyramid (Figure 2.1) although the methodology is simpler in nature than that of an atmospheric GCM. Similar in basic composition to an EBM, Adem's model includes, in a highly parameterized way, many dynamic, radiative and surface features and feedback effects, giving it a higher position on the modelling pyramid.

Mikhail Budyko and William Sellers published descriptions of two very similar EBMs within a couple of months of each other in 1969. These models did not depend upon the concepts already established in numerical weather prediction schemes, but attempted to simulate the essentials of the climate system in a simpler way. The EBMs drew upon observational data derived from descriptive climatology, suggesting that major climatic zones are roughly latitudinal. As a consequence of the intrinsically simpler parameterization schemes employed in EBMs, they could be applied to longer time-scale changes than the atmospheric GCMs of the time. It was the work by Budyko and Sellers, in which the possibility of alternative stable climatic states for the Earth was identified, that prompted much of the interest in simulation of geological time-scale climatic change. Concurrently with these developments, RC models, usually globally averaged, were being applied to questions of atmospheric disturbance including the impact of volcanic eruptions and the possible effects of increasing atmospheric CO_2.

The desire to improve numerical weather forecasting abilities also prompted the fourth type of climate model, the SD model. A primary goal for dynamical climatologists was seen to be the need to account for the observed state of averaged atmospheric motion, temperature and moisture on timescales shorter than seasonal but longer than those characteristic of mid-latitude cyclones. One group of climate modellers preferred to design relatively simple low-resolution SD models to be used to illuminate the nature of the interaction between forced stationary longwaves and travelling weather systems. Much of this work was spearheaded in the early 1970s

by John Green. Theoretical study of large-scale atmospheric eddies and their transfer properties combined with observational work led to the parameterizations employed in two-dimensional climate models.

By 1980, this diverse range of climate models seemed to be in danger of being overshadowed by one type: the atmospheric GCM. Although single-minded individuals persevered with the development of simpler models, considerable funding and almost all the computational power used by climate modellers was being consumed by atmospheric GCMs. However, by the mid- to late 1980s, a series of occurrences of apparently correct results being generated for the wrong reason by these highly non-linear and highly complex models prompted many modelling groups to move backward, in an hierarchical sense, in order to try to isolate essential processes responsible for the results that are observed from more comprehensive models. When only the most topical (e.g. doubled CO_2) model experiments are considered, the trend has been for GCM experiments to replace simpler modelling efforts. For example, in 1980–81, from a total of 27 estimates of the global temperature change due to CO_2 doubling, only seven were made by GCMs. By 1993–4, GCMs produced 10 out of 14 estimates published. The IPCC science working group has underlined the value of results from simple models such as the 'box' models (described in Chapter 3) while its impacts and responses groups have spawned many EMICs (see Chapter 4). The strategy of intentionally utilizing an hierarchy of models was originally proposed in the 1980s by scientists such as Stephen Schneider at the US National Center for Atmospheric Research. More recently, the soundness of an hierarchy of climate modelling tools has been championed by Tom Wigley.

In 1969, Kirk Bryan at GFDL developed the ocean model that has become the basis for most current ocean GCMs. The model has been modified and has become widely known as the Bryan–Cox–Semtner model. Albert Semtner and Robert Chervin constructed a model version which is 'eddy resolving' and as a consequence pushed the simulations to higher and higher resolution (currently 1/6 degree). Others have chosen to implement the model in non-eddy resolving form and have been able to run the model at 2° resolution for direct coupling with an atmospheric model.

Even though this three-dimensional ocean model dates back to the late 1960s, most global climate models treated the oceans in much simpler ways until the early 1990s. The original GCMs used fixed ocean temperatures based on observed averaged monthly or seasonal values. This 'swamp' model allows the ocean to act only as an unlimited source of moisture. Naturally, it is very difficult in such a model to disturb the climate away from present-day conditions when such large areas of the globe remain unchanged. Following this, in the late 1980s, computation of the heat storage of the mixed layer of the ocean (approximately 70–100 m) was the most common approach. In this model the lower deep ocean layer acts only as an infinite source and sink for water. The mixed layer approach is appropriate for time-scales ≤30 years, beyond which the transfer of heat to lower levels becomes significant. The mixed layer model does not include the transport of heat by ocean currents. GCMs with mixed layer models either needed to specify ocean heat transports to each grid square as a function of season, or make do with poor simulation of the ocean surface temperature in many areas.

The nature of climate model experiments has changed considerably as climate model complexity has increased. Early modellers were restricted to short 'experiment' and 'control' integrations, where the effects of a perturbation could be viewed in isolation. The inclusion of interactive oceans, biosphere, aerosols and clouds together with historical volcanic and solar forcings has led to the development of more complex experimental strategies. For example, early GCM experiments studying the effect of increased CO_2 were based on equilibrium experiments, where a model was allowed to equilibrate with the enhanced forcing. Modellers then subtracted the mean 'experiment' climate from the mean 'control' climate to determine the effect of the imposed change in CO_2. However, in the real world, climate forcings such as volcanic aerosols, solar variability, CO_2 and land-surface changes are transient, and different components of the model will react with different time-scales. Modellers must now focus on this aspect of the climate system and develop transient forcing datasets to be applied to their model.

The desire to make climate models more realistic has led to the involvement of many disciplines in the framework of climate modelling and hence to the realization that no one discipline can assume constancy in the variables prescribed by the others. Joseph Smagorinsky, who pioneered much of the early development in numerical weather prediction and steered the course of one of the flagships of climate modelling, NOAA's Geophysical Fluid Dynamics Laboratory, when commenting on the exponential growth in climate modelling research, noted that at the international conference on numerical weather prediction held in Stockholm in June 1957, which might be considered the first international gathering of climate modellers, the world's expertise comprised about 40 people, all loosely describable as physicists. In 2001, the IPCC Third Assessment Report (Working Group I alone) comprised hundreds of contributors and authors. A complete list of all who might term themselves climate modellers would today number tens of thousands and encompass a wide variety of disciplines. Interdisciplinary ventures have led to both rapid growth in insight and near-catastrophic blunders. Also, increasing complexity in narrowly defined areas such as land-surface climatology has forced upon modellers the recognition that other characteristics of their models, such as the diurnal cycle of precipitation, are being poorly predicted. The inclusion of more complex parameterizations of various subsystems, for example sea ice, is of little value if the atmospheric forcing in polar regions is inadequate. The tuning process that accompanies the addition of new model components might, in this situation, soak up these errors. Modellers must maintain an holistic view of their model.

2.4 SENSITIVITY OF CLIMATE MODELS

An important stage in the development of climate models is a series of sensitivity tests. Modellers examine the behaviour of their modelled climate system by altering one component and studying the effect of this change on the model's climate.

Equilibrium climatic states

As an example of a change in an internal variable we can consider the variation in the albedo, α, as a function of the mean global temperature in an EBM. Above a certain temperature, T_g, the planet is ice-free and the value of the albedo is independent of temperature. As it becomes colder we expect the albedo to increase as a direct result of increases in ice and snow cover. Eventually the Earth becomes completely ice-covered, at temperature T_i, and further cooling will produce no further albedo change. This could be expressed in the form

$$\begin{aligned} \alpha(T) &= \alpha_i && \text{for } T \le T_i \\ \alpha(T) &= \alpha_g && \text{for } T \ge T_g \\ \alpha(T) &= \alpha_g + b(T_g - T) && \text{for } T_i < T < T_g \end{aligned} \tag{2.2}$$

where b is the rate of change of α as the temperature decreases. T_i is usually assumed to be 273 K but may range between 263 and 283 K. If we are concerned with equilibrium conditions (i.e. when the left-hand side of Equation (2.1) is zero) we can calculate $R\uparrow$ for a series of temperatures and $R\downarrow$ for a series of albedos and show the results graphically. The points of intersection of the curves occur when emitted and absorbed radiation fluxes balance (i.e. $R\downarrow = R\uparrow$) which represent the equilibrium situations (Figure 2.7). Any slight imbalances between the fraction of the incident solar radiation, S, absorbed, $S(1 - \alpha(T))$, and the emitted longwave flux at the top of the atmosphere, approximated by $\varepsilon\sigma T^4$ where ε is the emissivity, lead to a

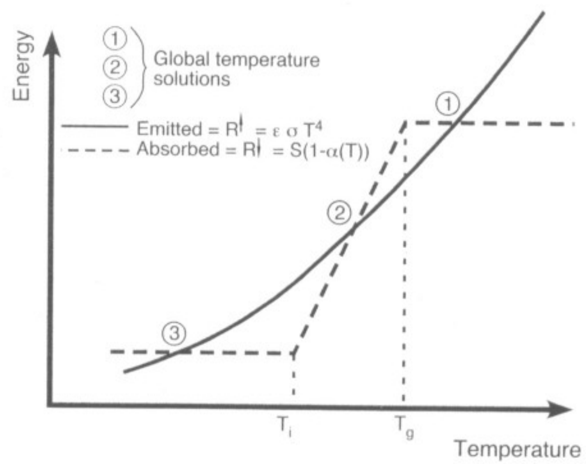

Figure 2.7 The three equilibrium temperature solutions for a zero-dimensional global climate model are shown at the intersection between the curves of emitted infrared radiation $R\uparrow$ and absorbed solar radiation $R\downarrow$. They are: (1) an ice-free Earth; (2) an Earth with some ice; (3) a completely ice-covered Earth (reproduced with permission from Crafoord and Källén (1978), *J. Atmos. Sci.*, **35**, 1123–1125

change in the temperature of the system at the rate $\Delta T/\Delta t$, the changes serving to return the temperature to an equilibrium state. However, there are three equilibrium solutions, as shown in Figure 2.7: an ice-free Earth (1), a completely glaciated (or 'Snowball') Earth (3) and an Earth with some ice (2) (e.g. the present situation of the planet). All are possible.

Stability of model results

Great care must be taken in choosing the constants for any parameterization scheme in any model. If values have been determined solely from empirical evidence, it may be that they are appropriate only for the present day, with the result that the model is likely to be constrained to predict the present-day situation and thus the less likely it is to be able to respond realistically to perturbations.

For 'external stability', we can test the response of the model to perturbations in the solar constant, since this is a convenient method of exploring climate model structure. Figure 2.8 shows the way in which $\overline{T}$ changes as the total incident radiation, μS, changes. Reduction of the solar constant to some critical value ($\mu_c S$) means that the number of solutions is reduced from two to one. Below $\mu_c S$, no solution is possible. This point is termed the bifurcation point. For values of incoming radiation, μS, less than $\mu_c S$, temperatures are so low that the albedo, $\alpha(T, \phi)$, becomes very close to or equal to 1 and thus it is impossible to regain energy balance. However, if some limit is put on how high the albedo may become, as is usually the case, e.g. $\alpha \leq 0.75$, the solution becomes what might be described as an ice-covered Earth.

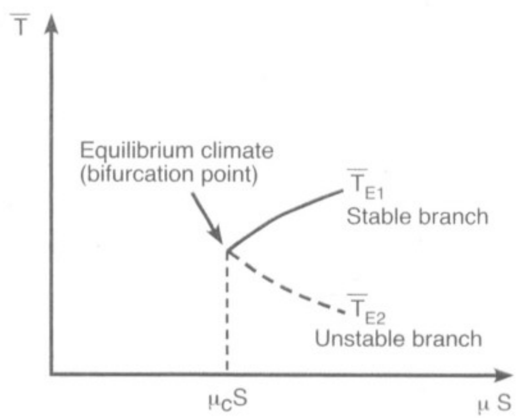

Figure 2.8 The equilibrium climate bifurcation point. For values of the solar luminosity given by μS where μ is a fractional premultiplier of the solar constant S, such that $\mu S > \mu_c S$, there are two solutions, whereas below this critical value no solutions exist. Changes in solar radiation lead to either a stable or an unstable equilibrium climate, illustrated here by the two equilibrium branches

'Internal stability' concerns the response of each branch in Figure 2.8 to perturbations from equilibrium which are created by internal factors. To determine if temperatures will return to equilibrium after the perturbation, we can use a time-dependent formulation and postulate a new value for $\overline{T}$ that is close to the equilibrium climate already calculated at that level of μS. This change can be computed iteratively until it is determined whether the values do regain the original $\overline{T}$ solution. If it is regained, then the solution is said to be internally stable. In Figure 2.8, only the top branch is stable because the model preserves $\overline{T}$ as proportional to μS. Using this method, it is possible to determine whether the model is transitive or intransitive, these terms being defined in Figure 2.9. The identification of almost intransitivity, also defined in Figure 2.9, is not possible in this manner.

Equilibrium conditions and transitivity of climate systems

Such a simple model has some very obvious limitations. However, it not only shows one means of analysing the results of climate models, it also indicates some of the more general problems associated with the solutions; in particular, the question of whether or not all three equilibrium states identified are 'stable' and capable of persisting for long periods of time. Many non-linear systems, even ones that are far simpler than the climate system, have a characteristic behaviour termed almost intransitivity. This behaviour is illustrated in Figure 2.9. If two different initial states of a system evolve to a single resultant state as time passes, the system is termed a transitive system. State A for this transitive system would then be considered the solution or normal state and all perturbed situations would be expected to evolve to

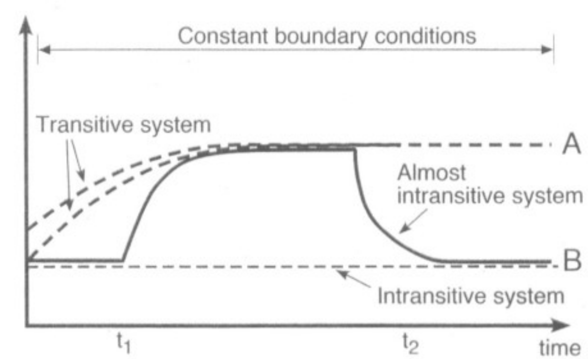

Figure 2.9 The behaviour of three types of climate system: transitive, intransitive and almost intransitive with respect to the initial state. In a transitive system, two different initial states evolve into the same resultant state, A. An intransitive system exhibits the 'opposite' behaviour, with more than one alternative resultant state. The characteristic of an almost intransitive state is that it mimics transitive behaviour for an indeterminate length of time and then 'flips' to an alternative resultant state (reproduced by permission of National Academies Press, 1975)

it. At the other extreme, an intransitive system has at least two equally acceptable solution states (A and B), depending on the initial state.

Difficulty arises when a system exhibits behaviour which mimics transitivity for some time, then flips to the alternative state for another (variable) length of time and then flips back again to the initial state and so on. In such an almost intransitive system it is impossible to determine which is the normal state, since either of two states can continue for a long period of time, to be followed by a quite rapid and perhaps unpredictable change to the other. At present, geological and historical data are not detailed enough to determine for certain which of these system types is typical of the Earth's climate. In the case of the Earth, the alternative climate need not be so catastrophic as complete glaciation or the cessation of all deep ocean circulation. It is easy to see that, should the climate turn out to be almost intransitive, successful climate modelling will be extremely difficult. Current studies of the climate as a chaotic system have focused on determining the characteristics of a climate attractor. The behaviour of the simple model of Edward Lorenz (Figure 2.10) has been used as an example of such an attractor, but no definitive conclusions have been reached on the nature of this attractor (if it exists) and no clear statements can be made regarding the transitivity of the climate system.

Measures of climate model sensitivity

The magnitude and direction of the sensitivity of any climate model to a known forcing are important characteristics. Although the term 'sensitivity' has recently

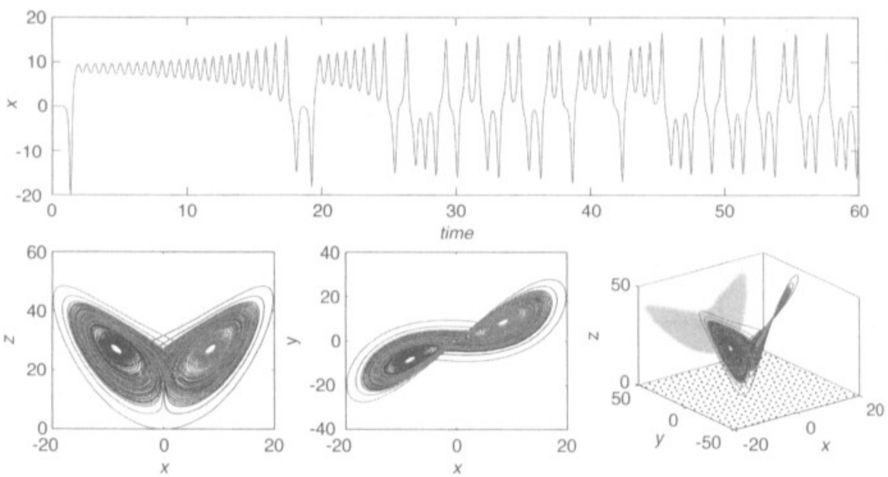

Figure 2.10 The 'Lorenz Butterfly'. A Poincaré section, showing the 'climate attractor' for the simple climate model constructed by Edward Lorenz in the 1960s. The system is characterized by three variables (x, y and z), which pinpoint the state of the system in a three-dimensional space. The apparently disordered behaviour of the system indicated in the graph in the top left conceals the structure which is apparent when the system is examined in three dimensions. Since the system never repeats itself exactly, the track never crosses itself

acquired mystique, the concept is straightforward. Most people, if pricked by a pin, exhibit a sensitivity and demonstrate this by a recognizable and quantifiable response. This response, although not identical in all subjects (a child might cry, while an adult would not), is readily differentiable from the generalized response to being hit by a flying cricket ball or baseball. The direction of both responses is generally negative and the magnitudes differ. The same is true for climate models.

Ideally, a climate model to be used for prediction should exhibit sensitivities that are commensurate with equivalent observable responses. However, this is not easy to check. Thus, for us to have confidence in model predictions of temperature increases in response to doubling or quadrupling of CO_2, we would like to know whether models of Venus, which has a massive greenhouse, are correct, or whether models of the Earth can correctly hindcast past periods when CO_2 and other greenhouse gas concentrations were much higher than today. Even for the single situation of doubled CO_2, there is a range of different measures of climate (and climate model) sensitivity including:

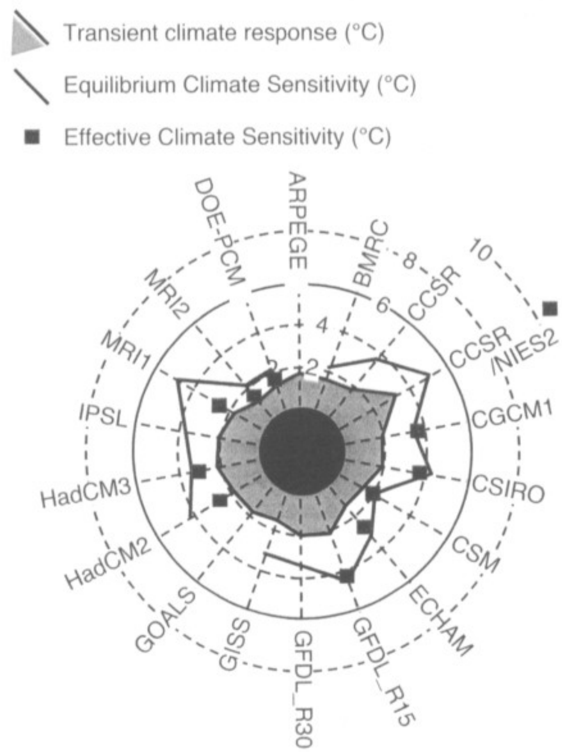

Figure 2.11 The three segments of the circle contain three different measures of modelled climate sensitivity derived for the IPCC Third Assessment Report. These sensitivities (shown by the length of the radial lines) are: (i) transient climate response; (ii) equilibrium climate sensitivity and (iii) effective climate sensitivity. Within segment measures are comparable but between segment comparisons are not valid (created from Table 9.1 from IPCC TAR WGI)

- transient climate response;
- equilibrium climate sensitivity (mixed layer ocean);
- effective climate sensitivity (deep ocean); and
- equivalent climate sensitivity.

These are illustrated in Figure 2.11 to describe the climate change predictions included in the IPCC Third Assessment Report. Different modellers choose different sensitivity measures and the result is a scatter of estimates that must be fed to policymakers.

Climate sensitivity measures can take many other forms, some of which were discussed in Chapter 1. Many modellers now prefer to evaluate models by reviewing their simulation of the twentieth-century climate. Sensitivity measures can be contrived that evaluate regional responsiveness to known forcings, such as the extent of the monsoonal activity or variations in seasonal snow cover.

Usually, the greatest confidence tends to be placed in climate models that exhibit sensitivities most like those observed. However, even this, apparently reasonable, view may produce excessive confidence because of the rather narrow climatic experiences during the observable record.

2.5 PARAMETERIZATION OF CLIMATIC PROCESSES

The climate system is a physical/chemical/biological system possessing infinite degrees of freedom. Any attempt to model such a highly complex system is fraught with dangers. It is (unfortunately) necessary to represent a distinct part, or more usually many distinct parts, of the complete system by imprecise or semi-empirical mathematical expressions. Worse still is the need to neglect completely many parts of the complete and highly complex system. This process of neglect/semi-empirical or imprecise representation is termed parameterization. Parameterization can take many forms. The simplest form is the null parameterization where a process, or a group of processes, is ignored. The decision to neglect these can only be made after a detailed consideration of their importance relative to other processes being modelled. Unnecessary computing time should not be spent on processes that can be adequately represented in some simpler way, or on processes that have relatively little effect on the climate at the scale of the model. Processes treated in this way are always candidates for improvement in later versions of the model.

Climatological specification, usually by prescribing observed averages, is a form of parameterization widely used in most types of model. In the 1970s, it was not uncommon to specify oceanic temperatures (with a seasonal variation) and in some of these models the clouds were also specified. When considering climate sensitivity experiments, it is important to recognize all such prescriptions because feedback features of the climate system will have been suppressed. Even today, most models specify the land-surface characteristics and few models permit the soil or vegetation to change in response to climate forcing. Only slightly less hazardous than this is the procedure by which processes are parameterized by relating them to present-day

observations: the constants or functions describing the relationship between variables are 'tuned' to obtain agreement. It is important that physically unrelated processes are not tuned together by this method. For example, the association of gradients in two different variables need not mean that the two are physically related. At best, this procedure presumes that constants and relationships appropriate to today's climate will still be applicable should some aspect of the climate alter.

The most advanced parameterizations have a theoretical justification. For instance, in some two-dimensional zonally averaged dynamical models, the fluxes of heat and momentum are parameterized via baroclinic theory (in which the eddy fluxes are related to the latitudinal temperature gradient). The parameterization of radiative transfer in clear skies is another example. To a good approximation, the atmosphere is like a set of parallel sheets of air with different properties. All that needs to be known is the vertical variation of temperature and humidity. Unfortunately, these parameterizations can lead to problems of uneven weighting because another process of equal importance cannot be adequately treated. In the case of heat and momentum transport by eddies, the contribution to these fluxes from stationary waves forced primarily by the orography and the land/ocean thermal contrast cannot be so easily considered. In radiation schemes, since clouds are three-dimensional and horizontal interactions are important, the parameterization of cloudy sky processes is not as advanced as for clear skies.

Interactions in the climate system

The interactions between processes in any model of the climate are crucially important. Wiring diagrams which show all these interactions are often used to illustrate the complexity of incorporating them all adequately. A most important concept in climate modelling is that the relative importance of processes and the way that different processes interlink is a strong function of the time-scale being modelled. The whole concept of parameterization is subsumed by this assertion. Establishing whether a system is likely to be sensitive to the parameterization used for a particular process often depends upon the response time of that feature as compared with other 'interactive' features. It is pointless to invoke a highly complex, or exceedingly simplistic, parameterization if it has been constructed for a time-scale different from that of the other processes and linkages in the model. The adage 'choosing horses for courses' is fundamental to the art of climate modelling.

As the climate system depends upon scales of motion and interactions ranging from molecular to planetary, and from time-scales of nanoseconds to geological eras, parameterizations are a necessary part of the modelling process. A decision is generally made very early in model construction about the range of space- and time-scales which will be modelled explicitly. Figure 2.12 illustrates the difficulty faced by all climate modellers. The constraints of computer time and costs and data availability restrict the prognostic (or predictive) mode. Outside this range there are 'frozen' boundary conditions and 'random variability'. Thus the two examples shown in Figure 2.12 illustrate the range of prognostic computations for (i) an Earth

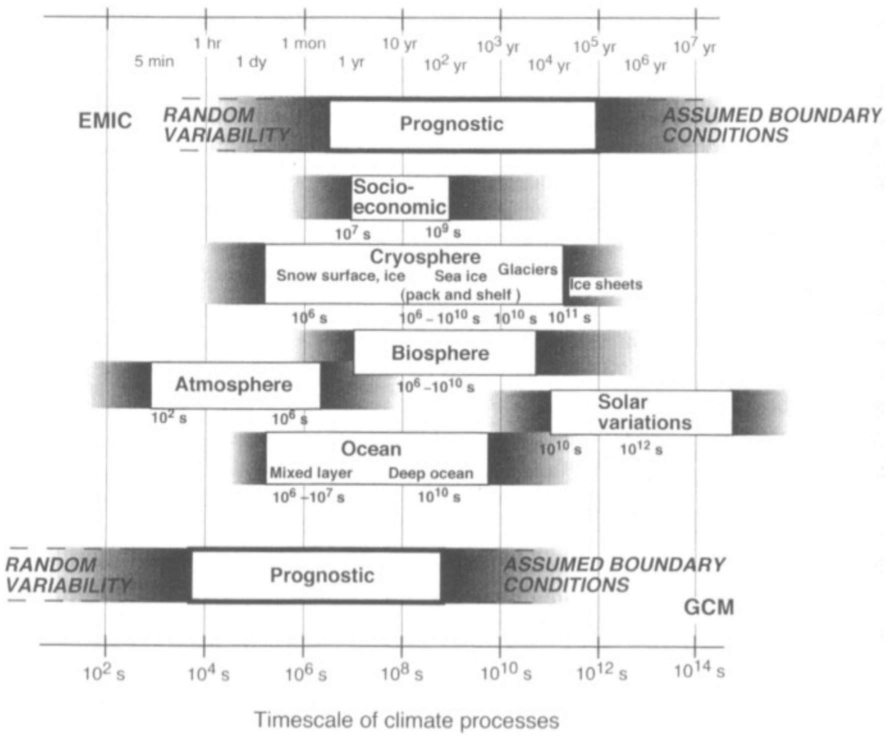

Figure 2.12 The importance of different temporal scales changes as a function of the type of model. The domain in which the model simulates the behaviour of the system is called 'prognosis'. It is expected that processes which fluctuate very rapidly compared with the prognostic time-scales will contribute only small random variability to the model predictions, while processes which fluctuate very slowly compared with the prognostic time-scale can be assumed to be constant. Two types of model are illustrated: an EMIC and a coupled ocean atmosphere general circulation model

System Model of Intermediate Complexity (EMIC, see Chapter 4), and (ii) a GCM focused on examining the effects of greenhouse gases on climate. In both cases, longer time-scales than those of concern to the modeller are considered as invariant and shorter time-scales are neglected as being random fluctuations, the details of which are of too short a period to be of interest.

Parameterizations must be mutually consistent. For instance, if two processes produce feedback effects of opposite sign, it is important that one process is not considered in the other's absence. An example is the effect that clouds have on the radiative heating of the atmosphere. Longwave radiation causes a comparatively rapid cooling at the cloud top, whereas the absorption of solar radiation results in heating. To consider the effect of clouds on only one of the two radiation fields may be worse than neglecting the effect of clouds entirely.

Figure 2.13 portrays an hierarchical averaging scheme for the climate system. The averaging processes are described in terms of a single variable, which could be as simple a component of the climate system as temperature, but could alternatively be, for example, representative of the carbon budget. There are two averaging subsystems in the lower part of the diagram, the one on the right-hand side being based on an initial averaging of the mean state in the vertical, followed by zonal and/or meridional averaging, while the one on the left-hand side is averaged first around latitude zones.

A traditional view of the averaging diagram in Figure 2.13 would be that the simplest approximations to the climate system (models) lie at the bottom of the diagram (*cf.* the base of the climate modelling pyramid: Figure 2.1) with increasing resolution being synonymous with increasing (and perhaps more desirable) complexity on ascent through the diagram. The apex of this diagram would be presumably that radiative and diffusive processes would be described at the molecular level in GCMs. Clearly such an ultimate goal is absurd, although it sometimes seems to be consistent with the desire for increasing complexity in a few GCM modelling groups. An alternative view might be that some of the more sophisticated lower-resolution SD models might contain the maximum information currently available/verifiable for very long-term integration periods. These would, therefore, be adequate and appropriate models since the climate system over long time-scales would be deemed to be insensitive to higher-resolution features. Thus, the key element in any model is the method of parameterization, whereby processes that cannot be treated explicitly

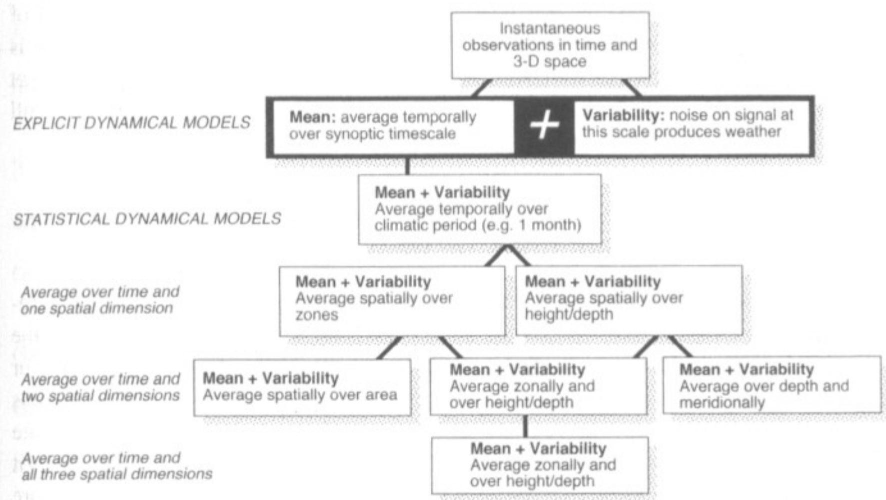

Figure 2.13 An hierarchical scheme for the averaging of climatological variables. In the lower half of the figure the representations of the climate system on the right-hand side involve averaging first over the atmospheric column, whereas the representations on the left-hand side involve zonal averaging first (adapted from Saltzman, 1978)

are instead related to variables that are considered directly in the model. An example is in EBMs where only the surface temperature is calculated explicitly. Since poleward transport of heat by atmospheric motions is important, this transport has to be parameterized in some way relating to the surface temperature, such as the latitudinal temperature gradient. In GCMs, those processes that operate on scales too small to be resolved by the model (sub-gridscale processes), like convective clouds, can, and do, exert influence on the atmosphere and must be parameterized in terms of available model variables.

The need for observations

All climate models need observed values for part of their input, especially in order to specify the boundary conditions, and all require observational data with which to compare their results. Some variables, such as surface pressure, are available worldwide and pose only the problem of evaluating the accuracy of the observed dataset. Others, however, are sparse in either time or space. Knowledge of sea ice extent is largely dependent on satellite observations, so that there is only a short observational record and, although satellites offer information on extent and concentration of sea ice, there is little they can say about ice thickness. Thus it is difficult to compare such observations with any long-term average values obtained from models. As modellers include ever more sophisticated components of the climate system in their experiments, there is a growing need for information on other parameters for validation of models. One particular example is 'soil moisture'. The term could mean all the water in a soil column (which might, technically, include large reserves of groundwater not accessed by the biosphere) or might be limited to the amount of water accessible to the biosphere (possibly termed 'available soil water'). There is no consistent definition between different modelling groups and no validation set comparable to traditional observations of pressure and temperature. There is still much to be done in the field of model validation.

2.6 SIMULATION OF THE FULL, INTERACTING CLIMATE SYSTEM: ONE GOAL OF MODELLING

Despite their limitations, coupled climate system models (*cf.* Section 2.2.4) represent the most complete type of climate model currently available. They illustrate the tremendous advances in our understanding of the atmosphere and ocean and our ability to model them over the 40–50 years since the first numerical climate models were produced. They do not yet, however, incorporate all aspects of the climate system and are therefore not at the apex of the pyramid in Figure 2.1. Indeed, it seems reasonable to suppose that the apex is unattainable. There will always be more features to include in the model. These models can, however, provide a great deal of information about the present climate and the possible effects of future perturbations. That these predictions are often contradictory is inevitable, given our incomplete knowledge of present conditions and developing understanding of the

controlling processes and interactions. If a model is built on sound theoretical principles, incorporates rational, and balanced, parameterization schemes, accounts for the major processes acting in the climate system and has been adequately tested against the available data, its results should be treated with respect. The results provide at least an indication of the possible future climate conditions created by a perturbation in the forces controlling our present climate.

The rest of the book is structured so that the concepts upon which full three-dimensional models are based are introduced sequentially. Chapter 3 underlines the fundamental basis of climate modelling: the energy balance. Chapter 4 describes models which operate with intermediate complexity, often by reducing the problem to one or two dimensions, and which help to provide insight into the operation of the full climate system over protracted periods or pay particular attention to specific aspects.

The overt goal of the text is therefore clear: we are aiming towards Chapters 5 and 6 in which the big players, the coupled atmosphere–ocean models, are explained and the process of evaluating and using climate model results is described. The other equally valid and important goal is less obvious. Throughout the book we have tried to choose examples to illustrate and enhance understanding of the mechanisms controlling the climate, their complexities, time- and space-scales and interactions. Both goals are worthy of considerable effort.

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Web resources

http://www.met-office.gov.uk/research/hadleycentre/	UK Meteorological Office: Hadley Centre
http://www.giss.nasa.gov/	NASA Goddard Institute for Space Studies
http://www.gfdl.gov/	The Geophysical Fluid Dynamics Laboratory
http://www.cgd.ucar.edu	US National Center for Atmospheric Research
http://www.cccma.bc.ec.gc.ca/eng_index.shtml	Canadian Centre for Climate Modelling and Analysis
http://www.dkrz.de/	Deutsches Klimarechenzentrum GmbH
http://www.bom.gov.au/bmrc/	Australian Bureau of Meteorology Research Centre

http://ugamp.nerc.ac.uk/	UK Universities Global Atmospheric Modelling Programme
http://www.mri-jma.go.jp/	Japanese Meteorological Agency: Meteorological Research Institute
http://www.climateprediction.net/	Climate Prediction.net
http://www.lmd.jussieu.fr/	Laboratoire de Météorologie Dynamique du CNRS
http://www.aip.org/history/climate/GCM.htm	The American Institute of Physics History of GCMs
http://www.cdc.noaa.gov/cdc/reanalysis/	NCEP/NCAR Reanalysis
http://www.cru.uea.ac.uk/cru/data	Climatic Research Unit, University of East Anglia, UK
http://nsidc.org/data/index.html	US National Snow and Ice Data Center