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FORTY YEARS OF NUMERICAL CLIMATE MODELLING

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ABSTRACT

Climate modelling is now a mature discipline approaching its fortieth birthday. The need for valid climate forecasts has been underlined by the recognition that human activities are now modifying the climate. The complex nature of the climate system has resulted in the development of a surprisingly large array of modelling tools. Some are relatively simple, such as the earth systems and energy balance models (EBMs), while others are highly sophisticated models which challenge the fastest speeds of the most powerful supercomputers. Indeed, this discipline of the latter half of the twentieth century is so critically dependent on the availability of a means of undertaking powerful calculations that its evolution has matched that of the digital computer. The multi-faceted nature of the climate system demands high quality, and global observations and innovative parameterizations through which processes which cannot be described or calculated explicitly are captured to the extent deemed necessary. Interestingly, results from extremely simple, as well as highly complex and many intermediate model types are drawn upon today for effective formulation and evaluation of climate policies. This paper discusses some of the important developments during the first 40 years of climate modelling from the first models of the global atmosphere to today's models, which typically consist of integrated multi-component representations of the full climate system. The pressures of policy-relevant questions more clearly underline the tension between the need for evaluation against quality data and the unending pressure to improve spatial and temporal resolutions of climate models than at any time since the inception of climate modelling.

KEY WORDS: Archaean; climate models; climate system; deforestation; earth models of intermediate complexity (EMICs); feedbacks; global climate models (GCMs); greenhouse warming; land-use change; last glacial maximum; mid-Holocene; Milankovitch; models; ozone hole; palaeoclimate

1. THE CLIMATE SYSTEM

Today, the atmosphere of planet Earth is undergoing changes unprecedented in human history and, although changes as large as those we are witnessing now have occurred in the geological past, relatively few have happened with the speed which characterizes today's climate changes (e.g. Pearman, 1992). Concentrations of greenhouse gases are increasing, stratospheric ozone is being depleted, and the changing chemical composition of the atmosphere is reducing its ability to cleanse itself through oxidation (e.g. Keeling *et al*., 1976, 1995; WMO, 1994; Houghton *et al*., 1996; MacKay *et al*., 1997). These global changes are threatening the balance of climatic conditions under which life evolved and is sustained. Temperatures are rising, ultraviolet radiation is increasing at the surface and air pollutant levels are increasing. Many of these changes can be traced to industrialization, deforestation and other activities of a human population that is itself increasing at a very rapid rate (e.g. Bruce *et al*., 1996; Giambelluca and Henderson-Sellers, 1996; Watson *et al*., 1996).

Now, for the first time in the history of our planet, emissions of some trace gases from human activities equal, and for some even exceed, emissions from natural sources. This is important for the climate system

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because the atmosphere's primary constituents, molecular nitrogen and molecular oxygen, are transparent to infrared radiation, and so the greenhouse gases (mainly water vapour, carbon dioxide (CO_2), ozone (O_3), methane (CH_4), nitrous oxide (N_2O) and the chlorofluorocarbons, or CFCs, which have no natural sources) present in much smaller amounts, play a major role in the Earth's energy budget and climate. Trace gases also govern the production and destruction of ozone, affect many biospheric processes, and play other important roles in the climate system. It is of great importance to determine where and how these constituents enter the atmosphere, how they are distributed and transformed by the complex interactions of sunlight, air, land, sea and living organisms; and how they behave in the climate system (cf. Litfin, 1994). Trace gases are carried from their surface sources to the upper atmosphere and around the world by numerous, interdependent processes, such as mixing in the atmospheric boundary layer, vertical exchanges associated with weather systems, moist convection in the mid latitudes, deep convection in tropical storms, and the mean circulation of the atmosphere (e.g. Houghton, 1984).

Possibly the most important species in the atmosphere and hence in the climate system is the hydroxyl radical (OH). Generated by interactions among ozone, water vapour and ultraviolet radiation, OH is the atmosphere's primary agent of oxidation and the means by which many compounds are transformed into others more readily removed from the atmosphere. The role of the OH radical in climate change is difficult to quantify because it is short-lived in the atmosphere (less than a second), and concentrations can only be inferred by examining the concentrations of other participants in its reactions. The concentration of OH has probably changed since pre-industrial times, as concentrations of CH_4, NO_x and O_3 have increased, but because of the complex chemistry involved and competing pathways of destruction and generation (e.g. Wang and Jacob, 1998) it is difficult to quantify these changes. Climate changes cannot be fully understood without improved determination of the net effect of these complex interactions on the abundance of this dominant oxidant (e.g. Brasseur and Solomon, 1986; McKeen *et al*., 1997; Carslaw *et al*., 1999; Kohlmann *et al*., 2000).

Global climate system changes resulting from human impact on the atmosphere and surface have been described as 'creeping climate crises'. Their characteristics seem to be that they are slow to develop and, therefore, may not become apparent until their effects have become dangerously far advanced. The iconic demonstration of this dates back to 1985 when British scientists (Farman *et al*., 1985) discovered that the mean ozone column abundance for October over the Antarctic station of Halley Bay had been decreasing very rapidly since the late 1970s, forming the so-called Antarctic ozone hole. The phenomenon is now a well established feature of the Antarctic atmosphere (e.g. Solomon, 1999). Stratospheric ozone concentrations at the South Pole in spring are now very much less than half of the values of only 30 years ago. Worldwide, stratospheric ozone has declined noticeably (now by a few percent) and, in the Northern Hemisphere, where the stratospheric circulation is more complicated, springtime depletions similar to those over Antarctica have developed over the last decade (McKenzie *et al*., 1992; Goutail *et al*., 1999; Hansen and Chipperfield, 1999).

Climate was once defined, rather simply, as 'average weather', but 'the climate system' has come to be defined more completely over the last few decades. In 1975, the Global Atmospheric Research Programme (GARP) of the World Meteorological Organization (WMO) stated that the climate system is composed of the atmosphere, hydrosphere, cryosphere, land surface and biosphere (WMO, 1975). The United Nations Framework Convention on Climate Change (FCCC), signed in March 1992, coming into force in March 1994, provided an updated definition of the climate system: the totality of the atmosphere, hydrosphere, biosphere and geosphere and their interactions. While these definitions are similar, the emphasis on interactions can be seen to have grown in the 19 years which separate them. The atmosphere, the land surface, the oceans and surface water (the hydrosphere), those parts of the Earth covered with ice and snow (the cryosphere) and the biosphere (the vegetation and other living systems on the land and in the ocean) are all very strongly coupled. This coupled climate system presents a special challenge for modellers, and this has led to a number of very significant volumes which detail the construction of models of the global climate (e.g. Schlesinger, 1988; Trenberth, 1992; Jacobson, 1998; Mote and O'Neill, 2000).

Concepts about the climate system are also concerned with personal and societal issues of habitability and sustainability. Most people evaluate climate in fairly simple terms such as temperature: is it too hot or too cold?; chemistry: is the air breathable?; sustenance: is there enough water for drinking and for growing crops; and ambient environment: does it feel comfortable, i.e. not too humid nor too dry? Trying to predict answers to these questions and to the larger question, can this planet continue to sustain life, is the goal of numerical climate modelling. The science of climate modelling, now just 40 years old, is currently being tested in attempts to understand past climates, relate the present climate to human activities, as well as predicting future climates (e.g. Budyko, 1969; Manabe and Bryan, 1969; McGuffie and Henderson-Sellers, 1997).

To first order, the Earth's climate is controlled by the amount of incident solar radiation that is absorbed by the planet and by the thermal absorptivity of the gases in the atmosphere which controls the balancing emitted infra-red radiation (e.g. Paltridge and Platt, 1976; Goody and Yung, 1996). Solar radiation is absorbed principally at the surface of the Earth, and over the mean annual cycle, this absorption is balanced by radiation emitted from the Earth (Figure 1). This global radiative balance, which is controlled by the surface and atmospheric characteristics, by the Earth's orbital geometry (e.g. Berger, 1981, 1988) and by the variability of solar radiation itself over time (Shindell *et al.*, 1999), controls the habitability of the Earth, mean temperatures and the existence of water in its three phase states (e.g. Robinson and Henderson-Sellers, 1998). These characteristics, together with the effects of the rotation of the Earth on its axis, determine the dynamics of the atmosphere and ocean, and the development of snow and ice masses. This combination of a distributed radiative budget and the forces resulting from the planet's axial rotation characterize any 'snapshot' of the Earth's climate (Figure 1).

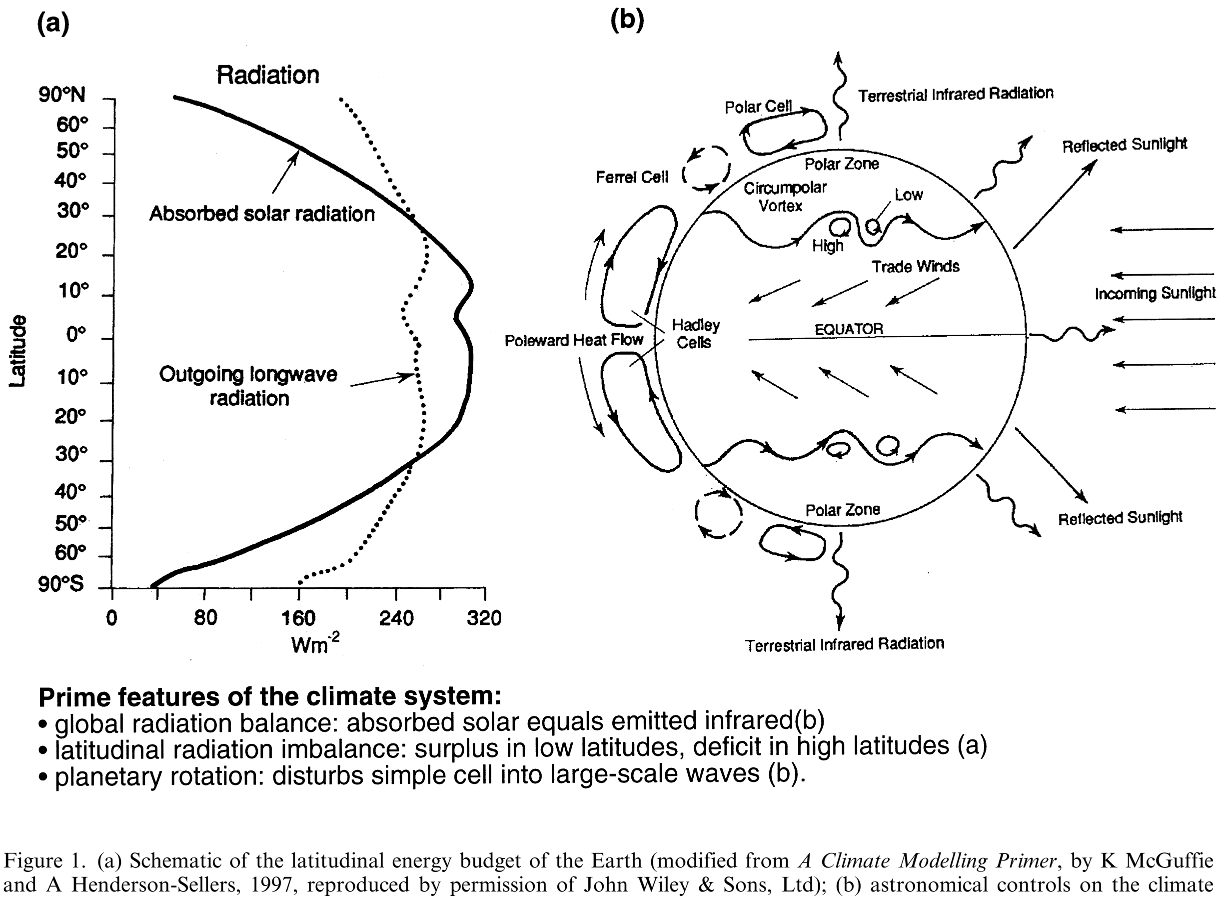

Figure 1. (a) Schematic of the latitudinal energy budget of the Earth (modified from *A Climate Modelling Primer*, by K McGuffie and A Henderson-Sellers, 1997, reproduced by permission of John Wiley & Sons, Ltd); (b) astronomical controls on the climate system include the planetary radiation balance (absorbed solar equals emitted infrared) and the effect on atmospheric and oceanic circulation of the planet's spin around its axis (after Henderson-Sellers, 1995). The prime characteristics of the global climate are listed

The second and complementary timeframe that needs to be borne in mind in characterizing the climate system is the evolutionary time-scale which controls the very long-term aspects of the climate components and those factors which force it, such as the physics and chemistry of the planet itself and the luminosity of the Sun. Viewed in this timeframe, the Earth's climate is prey to the forces of astronomical, geological and biological processes which control the persistence of ice caps and glaciers; the biota; rock structures and global geochemical cycling (e.g. Crowley, 1983; Schneider and Boston, 1991). For example, the

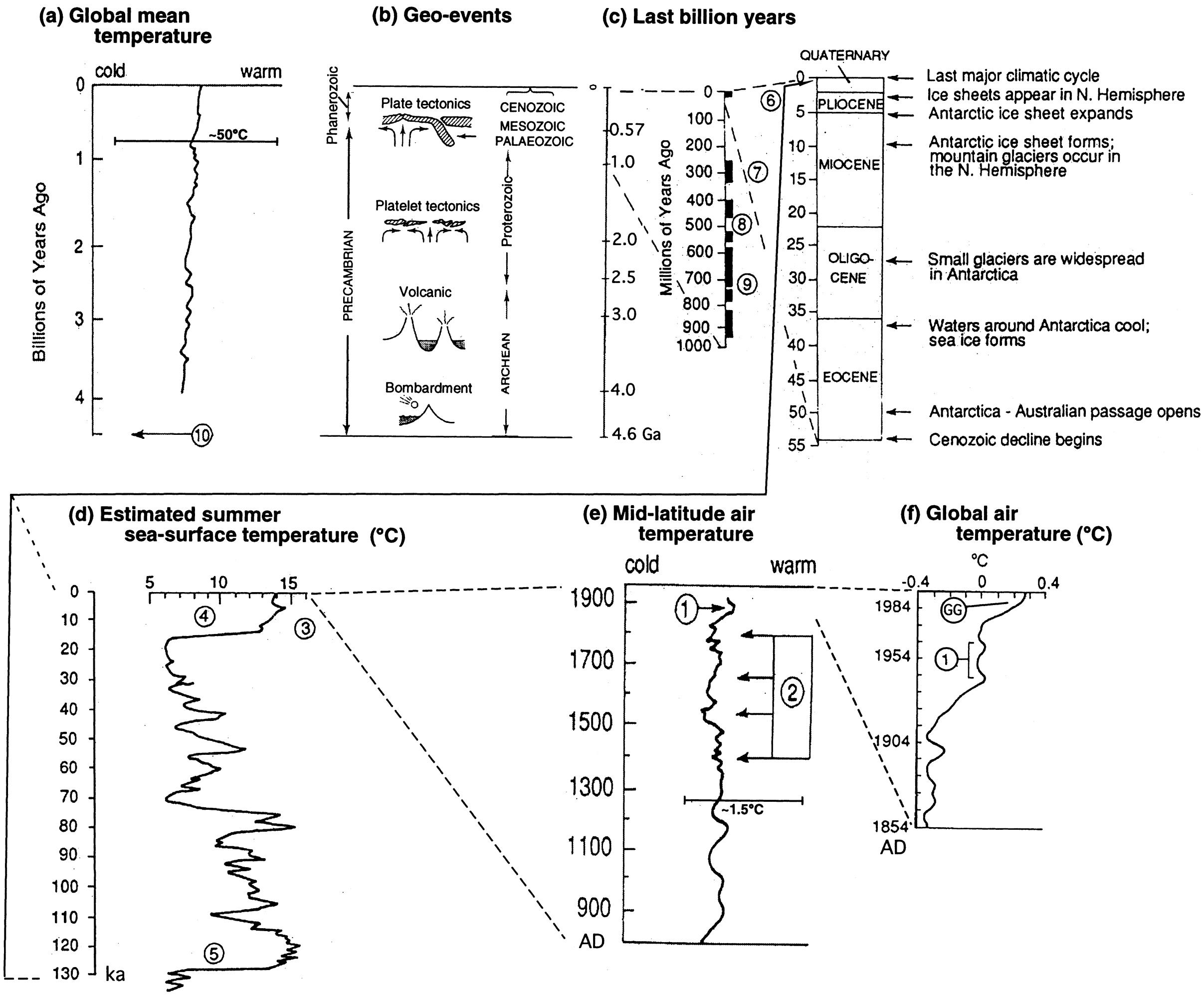

Figure 2. The Earth's climatic history is shown (a) over the planet's lifetime of 4.6 billion years (reprinted from after *Future Climates of the World*: *A Modelling Perspective*, World Survey of Climatology Series, Vol.16, by A Henderson-Sellers (ed.), Chapter 1 – Climates of the Future, Figure 1f, copyright (1995), with permission from Elsevier Science); (b) in terms of the geo and astrophysical events during its 4.5 billion year life (from Schneider and Boston, 1991, reproduced by permission of The MIT Press; (c) geochronology of the last billion years (figure reproduced from A Berger, *Milankovitch theory and climate*, copyright © 1988, by the American Geophysical Union, reproduced with permission), data from Baron, 1995; (d) estimated summer sea surface temperature (°C) over the last 130 000 years (reprinted from *Future Climates of the World: A Modelling Perspective*, World Survey of Climatology Series, Vol. 16, by A Henderson-Sellers (ed.), Chaper 1 – Climates of the Future, Figure 1d, Copyright (1995), with permission from Elsevier Science); (e) mid-latitude air temperatures (°C) since 800 AD (figure reproduced from *Climate Change*, **26,** 1994, pages 109–142, Was there a 'warm period', and, if so, where and when, by K Hughes and F Diaz, Figure 3, with kind permission of Kluwer Academic Publishers); and (f) global air temperatures (°C) since 1854 AD (reprinted from *Future Climates of the World: A Modelling Perspective*, World Survey of Climatology Series, Vol. 16, by A Henderson-Sellers (ed.), Chapter 1 – Climates of the Future, Figure 1a, Copyright (1995), with permission from Elsevier Science

The keyed events are: 1. Thermal maximum of the 1940s. 2. Little ice age. 3. Cold interval (Younger Dryas). 4. Present interglacial. 5. Previous interglacial. 6. Present glacial age. 7. Permo-Carboniferous glacial age. 8. Ordovician glacial age. 9. Late Precambrian glacial ages. 10. Earth's origin. GG, global greenhouse

Archaean, which represents almost half of the Earth's history from the formation of the Earth around 4.6 billion years ago to the Proterozoic transition 2.5 billion years ago, enjoyed a variety of geospheres before plate tectonics became established (Figure 2(b)). During this period, the planetary surface was first bombarded by cometary and meteoritic material, as the debris from the planetesimals was 'swept up'. This was followed by a period of surface vulcanism and then by a platelet tectonic regime. Life must have become established around or before 3.5 billion years ago because rocks from that era contain fossil evidence of viable and diverse microbial communities (e.g. Schopf *et al*., 1983). The oldest known rocks, the Isua supercrustal, are dated to 3.83 billion years, and show evidence of a global climate system not grossly different from that of the present. In particular, they contain sedimentary material that was waterborne. This suggests that the Earth's climate has been very stable when viewed in this evolutionary timeframe (e.g. Lovelock, 1991).

Figure 2 illustrates two interesting observations about the Earth's climate system. The first is that, over the lifetime of the planet and despite massive upheavals, the climate has remained remarkably stable (Figure 2(a)). The second is that excursions in temperature, and presumably any other climatic variables, have been large and aperiodic over the whole history of the Earth. The two characteristics of very long stability upon which short and medium term excursions are superposed are themselves a function of another fundamental quality of the climate system: the time required for equilibration, i.e. the time needed to adjust to a new forcing. The equilibration times for different subsystems of the Earth's climate system differ very markedly (Table I). The longest equilibration times are those for the deep ocean, the glaciers and ice sheets (10^{10}–10^{12} s), while the remaining elements of the climate system have equilibration times nearer 10^5–10^7 s.

The result of these two time-scales (evolution and equilibration) and of the complex interactions between these components of the climate system is a rich spectrum of climatic variability (Figure 2). The largest peaks in this spectrum relate to astronomical forcings: the Earth's rotation, its revolution around the Sun, variations in this orbit and the formation of the solar system (Berger, 1981, 1995). Coupling and feedbacks amongst processes within the climate system components, the atmosphere with the oceans, surface water with ice masses and the biosphere are responsible for the myriad of variations in this climate system spectrum. To try to understand, analyse and predict such variations, climate scientists have developed numerical models.

The Intergovernmental Panel on Climate Change (IPCC) reports on climate change have underlined the extent of our dependence on numerical climate models. For example, Gates *et al*. (1996, p. 233) state:

> The most powerful tools available with which to assess future climate are coupled climate models, which include three-dimensional representations of the atmosphere, ocean, cryosphere and land surface . . . [and] . . . More detailed and accurate simulations are expected as models are further developed and improved.

Table I. Representative equilibration times for components of the Earth's climate system

Climatic domain	s	Equivalent
Atmosphere		
Free	10^6	11 days
Boundary layer	10^5	24 h
Hydrosphere		
Ocean mixed layer	10^6–10^7	Months–years
Deep ocean	10^{10}–10^{11}	300–3000 years
Lakes and rivers	10^6	11 days
Cryosphere		
Snow and surface ice layer	105	24 h
Sea ice	10^6–10^{10}	Days–100s of years
Mountain glaciers	10^{10}	300 years
Ice sheets	10^{12}	3000 years
Biosphere		
Soil/vegetation	10^6–10^{10}	11 days–100s of years
Lithosphere	10^{15}	30 million years

This recognized dependence on the results of today's numerical climate models must be matched by an appreciation of their history and weaknesses, as well as of their benefits and strengths. In this paper, we offer a review of some aspects of these characteristics of climate models.

2. CLIMATE MODELS

The date of the true origin of 'climate modelling' depends, of course, on both the definition of 'climate' (e.g. local, regional or global over weeks, months or millennia) and of modelling (e.g. physical construction, correlation with, say, latitude or numerically based). In this paper, we review only numerical and global climate modelling, i.e. representations of the global climate constructed by calculations arising from an equation-based characterization. The father of today's climate models was Richardson. He published the first description of a method for constructing a weather forecast by means of numerical calculations (Richardson, 1922). It is well known that this method was at least thirty years ahead of even the very modest capability of the earliest computers. Richardson's farsighted parenthood of climate modelling is further underlined in his recognition of climate components other than the atmosphere. In common with many climate modellers since his time, Richardson knew and acknowledged the importance of currently neglected aspects, in his case the ocean, via sea surface temperatures. He also shared the aspiration of many of today's climate modellers when he wrote of the concept of developing a numerical model of the ocean similar to that which he had developed for the atmosphere:

> It may come to that, but let us hope that something simpler will suffice (Charnock, 1993, p. 32).

Climate models are tools employed to enhance understanding of the climate system and to aid prediction of future climates. Although there have been great advances made in the discipline of climate modelling over its forty year history, even the most sophisticated models remain very much simpler than the full climate system. Indeed, such simplicity is an unavoidable and, for some, also an intended, attribute of climate models (e.g. Washington and Parkinson, 1986; McGuffie and Henderson-Sellers, 1997). Modelling of a system which encompasses such a wide variety of components as the climate system is a formidable task, and it requires co-operation between many disciplines if reliable conclusions are to be drawn. Even the most elementary characteristics of the atmosphere vary considerably between climate models as illustrated in Figure 3, although whenever such a figure is shown, there are immediate explanations of the differences among the illustrated results. Intercomparisons such as these are now an integral part of climate science, and an important means for advancement of understanding of the climate system. Indeed, it is a genuine measure of the maturity of the climate modelling community that such intercomparisons occur.

An essential ingredient for all climate modelling is the speed with which calculations can be made (cf. Richardson, 1922). The rapid increase in computing power over the last 40+ years has meant that climate models have expanded both in terms of complexity, as measured by the total time they can simulate, and in the spatial and temporal resolution they can achieve (e.g. Trenberth, 1992).

Multi-decadal to millennial simulations are now common; full diurnal and seasonal cycles are now standard in climate experiments; and transient changes in, for example, the atmospheric trace gases such as CO_2 are becoming commonplace (e.g. Manabe and Stouffer, 1996; Boer *et al*., 2000). As knowledge increases, more aspects of the climate system are being and will be incorporated into climate models, and the resolution and length of integrations will further increase. At the same time, the goal of developing and evaluating the modelling tool most appropriate to each task will remain (cf. Shackley *et al*., 1998).

The simplest possible way of constructing a model of the Earth's climate is to consider the radiative balance of the globe as a whole (cf. Figure 1). This is a zero dimensional model often written in the form of a pair of equations

$$S(1-\alpha)=\sigma T_e^4 \tag{1}$$

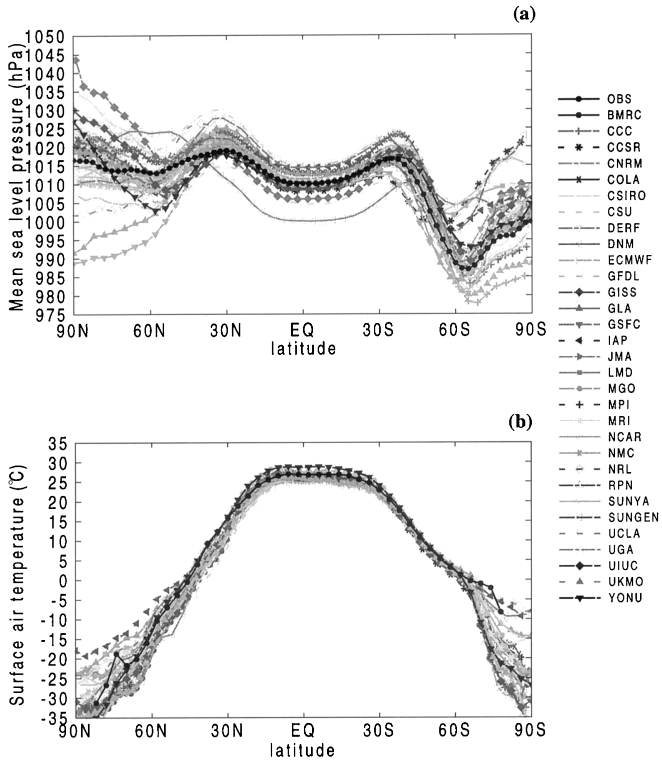

Figure 3. The zonally-averaged distribution of selected variables simulated by the AMIP models for December–January–February of 1979–1988, and that given by observations (solid black line) from Gates *et al.* (1999): (a) is the sea-level pressure, with observed data from the European Centre for Medium-range Weather forecasts (ECMWF) reanalysis; (b) is the surface air temperature, with observed data from Jones (1988) and COADS (Comprehensive Ocean-Atmosphere Data Set; da Silva *et al.*, 1994). Many of the differences among the model simulations are due to differences in spatial amd temporal resolution and in parameterization sophistication. See Gates *et al.* (1999) for model identification. (This figure is reproduced from PCMDI Report No. 45, Gates *et al.*, 1998. The U.S. Government's right to retain non-exclusive, royalty-free licence in and to any copyright covering this figure is acknowledged. Credit is given the University of California, Lawrence Livermore National Laboratory, and the Department of Energy under whose auspices the work was performed.)

plus

$$T_s = T_e + T_{\text{greenhouse}} \tag{2}$$

Here, S (the amount of solar radiation instantaneously incident at the planet per unit area of its (spherical) surface) has a value of about 342 W m^{-2} and the Earth's albedo, α, is 0.3. Thus, the effective blackbody radiating temperature of the Earth, T_e, is found to be around 255 K. This is lower than the current global mean surface temperature, T_s, of 288 K, the difference, about 33 K, being a result of the greenhouse effect. In Figure 4, the components of the Earth's globally and annually averaged radiation budget are presented as percentages of the average solar constant (342 W m^{-2}) at the top of the atmosphere. Nearly half the incoming solar radiation penetrates the clouds and greenhouse gases to the Earth's surface. These gases and clouds re-radiate most (i.e. 88 units) of the absorbed energy back down toward the surface. This is the basis of the mechanism of the greenhouse effect. The magnitude of the greenhouse effect is commonly measured as the difference between the blackbody emission at the surface temperature (a global average of 288 K gives 390 W m^{-2}), and the outgoing infrared radiation at the top of the atmosphere (here 70 units or 239 W m^{-2}), i.e. 151 W m^{-2}.

Within the very long time-scale of the Earth's history, it is possible to take a 'snapshot' view of the climate system (e.g. Figure 1(b)). In this 'instantaneous' view, the shortest time-scale processes are most evident. Of these, the most important are the latitudinal distribution of absorbed solar radiation (large at low latitudes and much less near the poles) as compared with the emitted thermal infrared radiation which varies much less with latitude (Figure 1(a)). This latitudinal imbalance of net radiation for the surface-plus-atmosphere

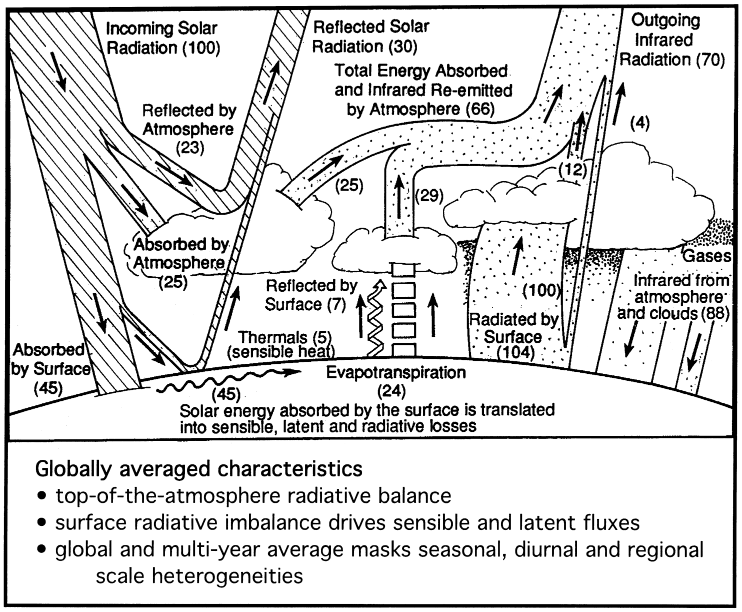

Figure 4. Schematic of the Earth's energy budget (modified from SH Schneider, 1992, *Coupled Climate System Modelling*, by KE Trenberth (ed.), reproduced by permission of Cambridge University Press). Units are percentage of the incident solar radiation, 342 W m^{-2}. Major characteristics of the climate system are listed

system as a whole (positive in low latitudes and negative in higher latitudes) is partitioned into energy fluxes at every location (e.g. Figure 4), and combines with the effect of the Earth's rotation on its axis to produce the dynamical circulation system of the atmosphere (Figure 1(b)) and the oceans.

The latitudinal radiative imbalance tends to warm air which rises in equatorial regions, and would sink in polar regions were it not for the rotation of the Earth. The westerly waves in the upper troposphere in mid-latitudes and the associated high- and low-pressure systems are the product of planetary rotation affecting the thermally-driven atmospheric circulation (Figure 1(b)). The overall circulation pattern comprises thermally direct cells in low latitudes, strong waves in the mid-latitudes and weak direct cells in polar regions (Peixoto and Oort, 1991). This circulation, combined with the vertical distribution of temperature, represents the major aspects of the atmospheric climate system (e.g. Schneider, 1992).

There is, today, a wide range of climate models available for the variety of simulation tasks associated with improving understanding of the climate system and predicting future (and past) climate changes. Currently the most highly developed tools available for climate assessment are the global climate models (GCMs) and the earth models of intermediate complexity (EMICs). These models, based on knowledge of physics, chemistry, biology, as well as economics and social science, portray this understanding in simplified representations, called parameterizations, of the processes they are designed to characterize. In a climate model, an atmospheric component is coupled to a model of the ocean, a representation of the biota and sometimes characterization of technological trends and food and water resources (e.g. Schlesinger, 1988; Wiebe and Weaver, 1999). The term GCM is nowadays taken to mean at least fully three-dimensional models of the atmosphere and oceans coupled together. If only the atmospheric (or oceanic) component is represented, the acronym AGCM (atmospheric GCM) or OGCM (oceanic GCM) is used. The difference in response (or equilibration) times of, for example, the ice masses and the carbon cycle compared to the atmosphere (Table I) means that different components are explicitly incorporated into different climate model types. For long time-scale simulations of future and past climates, the EMICs are used, while for periods of days and decades to a century or two, GCMs are employed. Although a few GCM integrations have extended over 10000 years or more (Broccoli, 2000), the main focus of GCM studies continues to be on the decadal to century scale.

This review is organized in a roughly historical narrative, which is summarized in Table II. This 40-year story of numerical climate modelling does not, however, fit tidily into either an evolutionary structure or

Table II. Historical evolution of climate models

Decade and landmark papers	Climate model status
≤1969	
Manabe and Möller (1961)	Numerical weather forecasts extended
Manabe and Strickler (1964)	RC models developed
Sellers (1969)	Dynamics and radiation virtually separate
Budyko (1969)	EBMs newly described
1969–1981	
Manabe and Bryan (1969)	Multi-layer oceans added to GCMs
Green (1970), Stone (1973)	SD models developed
Manabe and Wetherald (1975)	Greenhouse modelling with GCMs
CLIMAP (1981)	Palaeo datasets first employed for 'validation'
1981–1989	
Hansen *et al.* (1981)	GCMs becoming predominant model type
Sellers *et al.* (1986)	Surge in computational power and capacity
Oort and Peixoto (1983)	Satellites generate global observations
Luther *et al.* (1988)	Model intercomparisons suggested
1989–1999	
Houghton *et al.* (1990)	Simpler models required by IPCC
Semtner and Chervin (1992)	OAGCMs established but need flux correction
Flato and Hibler (1992)	Sea-ice and land-surface components evolving
Cubasch *et al.* (1994)	First ocean–atmosphere coupled ensemble
Santer *et al.* (1996)	Validation and attribution first described
2000s	
? ? ?	EMICs as important as GCMs
	Past climate simulations re-emerging for testing
	Observational need driven by evaluation demand
	Policy needs a major driver of numerical models

allow a neat sectionalization into systems' and components' descriptions. This is the result of two, sometimes competing, factors: increased computer power and sparcity of observations. First, the development of numerical climate modelling has always been dependent on the state of development of the numerical platforms, i.e. the computer. This interdependence, for climate modellers needs have also prompted computational developments, is a tangled affair, which has, at some points, seen computation undertaken without clear motives other than to use the power (e.g. Semtner and Chervin, 1988, 1992). At the same time, scientists challenging the 'received wisdom' have always disputed the predictions of numerical models and even their underpinning premises. These debates and disputes have, quite naturally, often been tied up with the issues of funding, influence and publicity (see Shackley *et al.*, 1998; Henderson-Sellers and McGuffie, 1999).

3. DEVELOPMENT OF CLIMATE MODELLING

Climate modelling has developed considerably since the first global atmospheric models were applied to climate simulation in the 1960s (cf. Table II). Models have been developed in response to scientific probing of existing model components, and have drawn on existing models. Throughout the 40 or so years of climate modelling, the growth in the complexity and physical realism of models has been facilitated by developments in computer technology (Henderson-Sellers and McGuffie, 1987). Here, we review the historical framework for some of the developments which have taken place over the last 40 years, and look at how the evolution of climate modelling has paced the emergence of each generation of high performance computers.

As climate models are sometimes described in terms of an hierarchy (e.g. McGuffie and Henderson-Sellers, 1997; Henderson-Sellers and McGuffie, 1999), 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 (Table II and Washington and Parkinson, 1986). The first atmospheric general circulation climate models were being developed in the early 1960s (e.g. Smagorinsky *et al*., 1965) concurrently with the first radiative convective (RC) models (Manabe and Möller, 1961; Manabe and Strickler, 1964). On the other hand, the simplest (energy balance) climate models, as they are currently recognized, were not described in the literature until 1969; the first discussion of two-dimensional statistical dynamical (SD) models was in 1970 (Saltzman, 1978) and the ESMIC are the youngest model type (e.g. Opsteegh *et al*., 1998; Rahmstorf and Ganopolski, 1999).

3.1. Complex climate models

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 (e.g. Charney *et al*., 1950; Smagorinsky, 1983) and, around 1960, as advances in computer technology allowed more extensive simulations, ideas were being formulated for long enough integrations of these numerical weather prediction schemes that they might be considered as climate models. Indeed, it is rather difficult to identify the timing of the transition from weather forecasting to climate prediction in these early modelling groups. The numerical requirements of weather prediction were extended to hemispheric domains (global calculations were not introduced until later) and the extension to longer integration periods sometimes became simply a matter of availability of computer resources. Indeed, to this day, climate modelling and numerical weather forecasting groups co-exist, especially in national meteorological bureaux. However, the needs and focus of the two disciplines differ: for example GCMs have to conserve mass, energy and moisture, while many forecast simulations are over too short a period for conservation to be an issue.

Many of the early pioneers of climate modelling came from numerical weather prediction. For example, Manabe joined the National Oceanic and Atmospheric Administration's (NOAA's) Geophysical Fluid Dynamics Laboratory in the USA in 1959 to collaborate in the development of numerical weather prediction models. He was to go on to become one of the pre-eminent leaders of the climate modelling community (e.g. Manabe and Bryan, 1969; Manabe and Wetherald, 1980; Manabe, 1985; Manabe and Bryan, 1985; Manabe and Stouffer, 1999). Scientists concerned with extending numerical prediction schemes to encompass hemispheric or, later, global domains were also studying the radiative and thermal equilibrium of the Earth–atmosphere system (Table II). It was these studies which prompted the design of the RC models, which were once again spearheaded by Manabe (Manabe and Möller, 1961). Other workers also expanded the domain of numerical weather prediction schemes in order to derive GCMs (Adem, 1965). The low-resolution thermodynamic model first described by Adem in 1965 is an interesting climate model type. Although the methodology is simpler in nature than that of an atmospheric GCM, it captures many aspects of a full GCM. Similar in basic composition to the energy balance models (EBM) developed later, Adem's model includes, in a highly parameterized way, many dynamic, radiative and surface features and feedback effects. It could be argued that Adem's model is the ancestor of today's EMICs.

3.2. Simpler climate models

Not all climate models originated from weather forecast models. In 1969, two very similar models were published within months of each other. Budyko and Sellers published descriptions of models which 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 simple way (Budyko, 1969; Sellers, 1969). These EBMs drew upon observational data derived from descriptive climatology; for example, the reasons why the major climatic zones are roughly latitudinal. In particular, EBMs are computationally very much faster than GCMs because instead of calculating the dynamical movement of the atmosphere, using the Navier–Stokes equations (as in GCMs), they employ much simpler parameterizations, and typically, much coarser grids. 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. Although the desire for longer simulation times is still an important driver for the development of simple climate model types, other demands, particularly associated with climate policy needs, are becoming increasingly important.

It was the work by Budyko and Sellers, in which the possibility of alternative stable climatic states for the Earth were identified, which prompted much of the interest in simulation of geological time-scale climatic change (cf. Ghil, 1984), and the recognition of chaos in the climate system (Lorenz, 1963). 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 (e.g. Hansen *et al*., 1981) and the very long-term evolution of the Earth (e.g. Rossow *et al*., 1982).

The desire to improve numerical weather forecasting abilities also prompted the fourth type of climate model: the SD model (e.g. Stone, 1973; Saltzman, 1983; MacKay and Khalil, 1994). 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 time-scales shorter than seasonal, but longer than those characteristic of mid-latitude depressions and anticyclones. To respond to this, one group of climate modellers opted to design relatively simple low-resolution SD models to be used to illuminate the nature of the interaction between forced stationary long waves and travelling weather systems. Much of this work was spearheaded in the early 1970s by Green (1970). 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 (e.g. Stone, 1973; Saltzman, 1978), and more recently, the seminal work of Hoskins (e.g. Hoskins *et al*., 1983; Held and Hoskins, 1985).

Although single-minded individuals persevered with the development of simpler models (Potter *et al*., 1981; Wigley and Schlesinger, 1985), by the early 1980s, this diverse range of climate models seemed to be in danger of being overshadowed by one type: the atmospheric GCM (Table II). Considerable funding, and almost all the computational power used by climate modellers, was being consumed by atmospheric GCMs, and an ethos of 'big is beautiful' was evident (e.g. Shackley *et al*., 1998). However, by the mid to late 1980s, a series of occurrences of apparently correct results being generated by these highly non-linear and highly complex models for obviously incorrect reasons prompted many modelling groups to move backwards, in an hierarchical sense, in order to try to isolate essential processes responsible for the results which are observed from more comprehensive models.

During the 1980s and 1990s, considering solely the then most topical (i.e. doubled CO_2) model experiments, there was a clear trend for GCM experiments to replace simpler modelling efforts. For example, in 1980–1981, from a total of 27 estimates of the global temperature change resulting from CO_2 doubling, only seven were made by GCMs. By 1993–1994, GCMs produced 10 of 14 published estimates. The IPCC process has employed box models in each of its reports since 1990, although it is arguable that the third report (preliminary findings published on the Internet at http://www.ipcc.ch on 20 January, 2001) will place greater emphasis on the value of results from simple models (e.g. Wigley and Schlesinger, 1985; Wigley, 1998). The strategy of intentionally utilizing an hierarchy of models was originally proposed in the 1970s by scientists such as Schneider at the US National Center for Atmospheric Research (Schneider and Dickinson, 1974). More recently, the soundness of an hierarchy of climate modelling tools has been championed by Wigley and the Third IPCC Assessment Report is rebalancing the relative emphasis by reporting developments of and results from EMICs compared with GCMs (e.g. Houghton *et al*., 2001).

3.3. Complexity of different types

The 1990s also saw the emergence of ensemble methods in climate studies. For example, Hansen *et al*. (1997) investigated the roles of climate forcings and chaos (unforced variability) in climate change via ensembles of climate simulations in which forcings were added one by one. This study is essentially an extension of an earlier study (Hansen *et al*., 1981). The ensemble technique involves letting a model run through a period of particular interest many times, so generating an 'ensemble' of climate realizations. The model can create more ensembles in which climate forcings, such as volcanoes and greenhouse gases, are added to the model one-by-one to study their effects. Hansen *et al.* (1997) studied the period 1979–1996. Ensemble techniques have also been applied to simulations of future climate (e.g. Vitart *et al.*, 1997; Tett *et al.*, 1999). One of the first ensemble predictions of climate changes using a coupled ocean-atmosphere GCM was made by Cubasch *et al.* (1994).

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. Smagorinsky, who pioneered much of the early development in numerical weather prediction (Smagorinsky *et al.*, 1965) and steered the course of one of the premier institutions of climate modelling NOAA's Geophysical Fluid Dynamics Laboratory (GFDL), 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 whole world's expertise comprised about 40 people, all loosely describable as physicists. In 2000, however, Working Group I, alone of the IPCC Third Scientific Assessment, has around 150 lead authors, and well over 600 contributing authors. The principal discipline of many of these is outside physics.

A complete list of those who consider themselves professional climate modellers would now number many, many thousands and encompass a wide variety of disciplines. Interdisciplinary ventures have led to both rapid growth in insight and near-catastrophic blunders (cf. Howe and Henderson-Sellers, 1997). Increasing complexity in narrowly defined areas such as land-surface parameterization has forced upon modellers the recognition that fundamental 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 (e.g. Flato and Hibler, 1992), is of little value if the atmospheric forcing in polar regions is inadequate.

In 1969, Bryan at GFDL developed the ocean model which has become the basis for most current ocean GCMs (Bryan, 1969, 1989a,b). This model has been modified, and has become widely known as the Bryan–Cox–Semtner model (e.g. Killworth *et al.*, 1991). Semtner and Chervin have constructed a model version which is 'eddy resolving', and as a consequence, have pushed the simulations to higher and higher resolution (currently 1/6 degree) (e.g. Semtner and Chervin, 1992; McCann *et al.*, 1994; Semtner, 1995). 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.

The 1990s has also seen the development of evaluation and intercomparison methodologies from relatively informal 'eyeball' comparisons of model and observations by individual groups to detailed intercomparison programs (e.g. Figure 3 and Gates *et al.*, 1999) and complex statistical techniques. Modern climate modellers must maintain a holistic view of their model and, as importantly, of the climate system itself. This view has to include at least (i) the available calculation power; (ii) the preferred inclusivity of the model (i.e. physics only; biogeochemical; economics); and (iii) the possibilities for demonstrating the verity or appropriateness of simulations either by evaluation against observations or by model intercomparison.

3.4. Computers and climate modelling

Throughout the development of computers, one of the major tasks to which they have been devoted is numerical weather and climate prediction. Operational weather forecasting is, perhaps, the largest single use to which modern computers are devoted. Thus, as the development of computational power has been exponential, so has the capability for forecasts. Desktop computers are developing at a comparable rate, so that the computational power of desktop machines is now as great as that of the supercomputers of 15 years ago. The combination of a massive increase in access to computers and the increased capacity of these machines has seen a corresponding rise in the number of climate modellers.

The earliest computer known to have been specifically used for weather forecasting is the IBM 701 of which only 19 were manufactured (Figure 5). One of the first installations of the IBM 701 was at the US Weather Bureau, where it was devoted to the problem of numerical weather prediction. Even today,

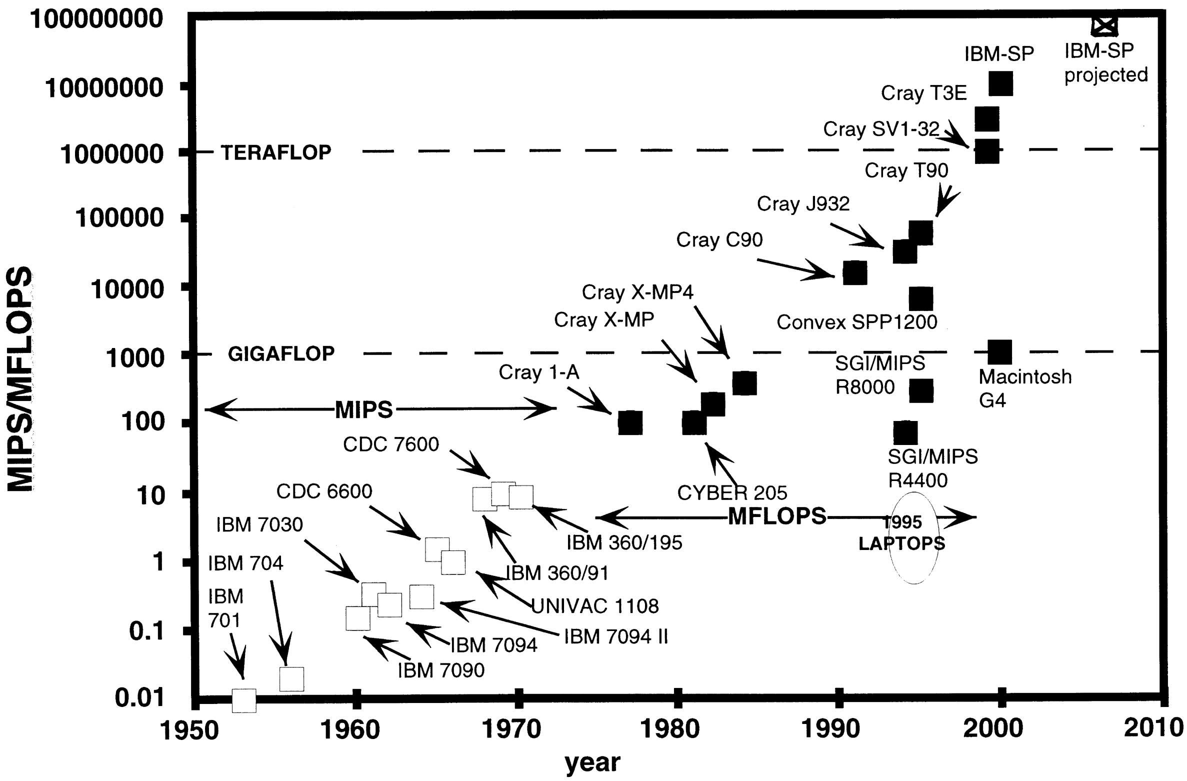

Figure 5. Development of computer power since 1950. Speeds are shown in millions of instructions per second (MIPS) up to 1974 and in millions of floating point operations per second (MFLOPS) from 1975 onwards. The rate of increase is exponential and shows no signs of tailing off (modified from *A Climate Modelling Primer*, by K McGuffie and A Henderson-Sellers, 1997, reproduced by permission of John Wiley & Sons, Ltd)

meteorological and climate research establishments have some of the fastest and most powerful computers available. In the 14th list of the top 500 computer sites, published at the 1999 supercomputer conference SC99 in Portland, Oregon, Deutscher Wetterdienst, Offenbach has the 9th fastest (SGI/Cray T3E1200), and the United Kingdom Meteorological Office, Bracknell has numbers 11 (SGI/Cray T3E900) and 13 (SGI Cray T3E1200). There is a constant cycle of upgrade and renewal as faster computers become available and the requirements of climate and meteorological research establishments increase.

Early computers, such as the IBM 701, were very basic by today's standards, but, as computer hardware developed in the 1960s, the introduction of solid state (transistor based) circuits led to an increase in speed and memory capacity. In turn, these solid state devices became more effective as techniques for placing more circuits on a single chip improved. As circuits became smaller, the components could be switched at a faster rate, and the resulting calculations became faster. The increased switching speed and decreased size of hardware resulted in increased difficulty in dissipating energy from the processor, and in significant engineering problems associated with differential heating of circuit boards. Just as modern desktop computer processors have an internal cooling fan attached, most supercomputers have large sophisticated cooling systems. Improvements in computer power in the late 1970s and early 1980s were dominated by improvements in processor design architecture, with the introduction of pipelining and vector processors to improve the efficiency of the computers at performing calculations. These developments placed increased emphasis on algorithmic design and climate modelling groups have devoted a great deal of effort to tuning their model to a particular computer architecture (e.g. Semtner and Chervin, 1992; Drake *et al.*, 1994).

As a computer processor becomes faster, there are physical (relativistic) limitations associated with synchronizing processor communications. From the mid 1970s to the early 1990s, the supercomputer market was dominated by the work of Seymour Cray at Cray Research Inc., and later Cray Computer Inc. (e.g. Murray, 1997) and SGI/Cray continues to be a major player in the supercomputer market. Seymour Cray made some of the most significant advances in the development of supercomputers during his time at Cray and prior to that at Control Data Corporation.

Because of the limitations in processor speed brought about by known barriers to speed, current advances in computer performance are coming from the development of scalable parallel devices, such as the Intel ASCI Red and the IBM SP series, which are currently delivering teraflop speeds. In practical terms, the limitations on effective computer speed now stem largely from the communication of results from one processor to another (the reason that each processor on the IBM SP has its own disk). The software available for compiling programs on supercomputers is now extremely sophisticated, and there is a significant amount of effort which must be devoted to formulating the problem (the climate model) in a manner which is transferable to the computer architecture. Thus, even today the dual evolution of computer power and climate model capacity continues to be tightly coupled (Figure 5).

The development of climate models mirrors the development of computers through the last four decades. This can be clearly seen by considering one component of the full climate system, for example, the oceans. Even though the earliest three-dimensional ocean model (at that time uncoupled from the atmosphere) dates back to the late 1960s, the immense amount of computer resources required to run such a model condemned most global climate modelling groups to treating the oceans in much simpler ways in early simulations (Figure 6).

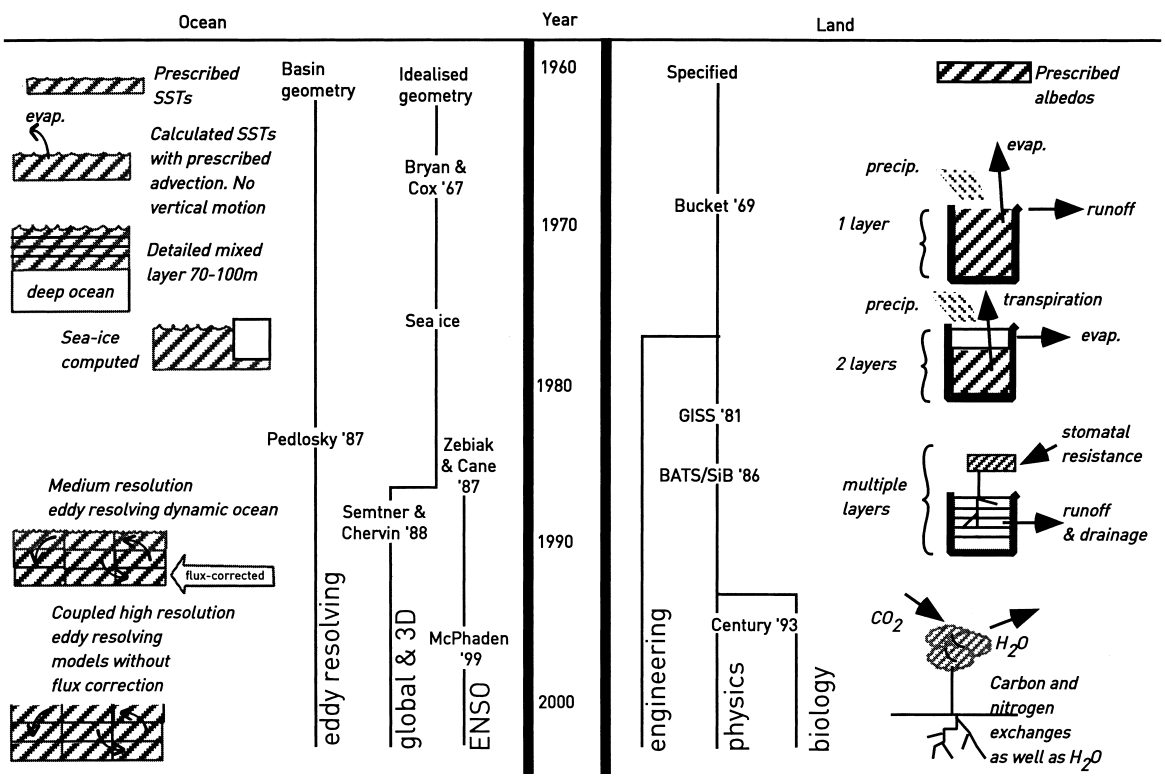

Figure 6. Schematic of the history of the development of parameterization in two sub components of today's coupled climate models. Both ocean and land surface components of climate models have developed from simple specified surfaces through highly parameterized models to today's sophisticated codes, which attempt to capture as much of the physical processes as understanding and parameterization and computer resources permit. Note that the timelines are generalized for the whole community, e.g. Bryan and Cox (1967) had a fully dynamic ocean cf. Manabe and Wetherald (1975), who continued to use a 'swamp' model

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.

The first coupled ocean–atmosphere model was the product of Manabe and Bryan in 1969. Although including a very low resolution ocean, it was multi-layered, and so provided a target for other modelling groups. Despite this, as late as the late 1980s, computation of the heat storage of the mixed layer of the ocean (approximately 70–100 m) was the most common approach (e.g. Hansen *et al*., 1983) (Figure 6). In this model, the lower deep ocean layer acts only as an infinite source and sink for water. The mixed layer approach is appropriate only for time-scales up to about 10–30 years, beyond which the transfer of heat to lower levels becomes significant.

Many of the important features of the climate system, such as the North Atlantic Drift and the seasonal growth and decay of Antarctic sea-ice, are dominated by dynamic effects and are absent from mixed layer models (e.g. Manabe and Stouffer, 1980). Thus, ocean models had to be global, fully three-dimensional, and of adequate resolution that eddies and mesoscale circulation features could be resolved. Their demands continue to challenge some of the world's fastest supercomputers even today (e.g. Washington *et al*., 2000).

A rather similar co-evolution of computing power and addition of necessary complexity to climate calculations can be traced in atmospheric chemistry (especially the inclusion of the OH radical) and carbon budgeting (cf. Harvey, 2000a,b).

3.5. Forcings and feedbacks

The state of the climate system at any time and its sensitivity to perturbations (both internal and external) is determined by the forcings acting upon it and the complex and interlocking internal feedbacks that these forcings prompt. In the broadest sense, a feedback occurs when a portion of the output from an action is added to the input so that the output is further modified. The basic theory for this can be found in any introductory text in electrical engineering. The result of such a looped system can either be an amplification (a positive feedback) of the process or a dampening (a negative feedback): positive feedbacks enhance a perturbation whereas negative feedbacks oppose the original disturbance (Figure 7).

3.5.1. Snow and ice: a surface feedback. If some external perturbation, say an increase in solar luminosity, or an internal perturbation, such as increased CO_2 concentration, acts to increase the global surface temperature, then snow and ice will melt and their overall areas reduce in extent. These cryospheric elements have high albedos, so reducing them means that a smaller amount of solar radiation will be reflected away from the planet. Increased absorption leads to higher temperatures. A further decrease in snow and ice results from this increased temperature and the process continues. This positive feedback mechanism is known as the ice-albedo feedback mechanism (Figure 7(a)). This feedback is the main source of the sensitivity of the simple climate models of Budyko (1969) and Sellers (1969), although it is also exhibited by more complex models. Cess *et al*. (1991) examined this feedback as exhibited by 17 GCMs, and found that additional amplification or moderation could occur.

3.5.2. Water vapour and clouds: atmospheric feedbacks. Water vapour contributes to greenhouse warming as a result of its absorption of infrared radiation emitted from the surface (Rind *et al*., 1991; Rind, 1998). Thus, as temperatures increase and the water vapour content of the atmosphere increases, it enhances the original temperature increase: a positive feedback effect.

Clouds have two radiative effects in the Earth's atmosphere that tend to act in opposite ways. Clouds act to cool the Earth by reflecting solar radiation, but they have a heating effect because they absorb infrared energy that is emitted from the surface, and which would otherwise escape to space. Knowledge of the height, coverage and thickness of cloud layers is essential for both modelling the complicated feedback processes between clouds, radiation and climate and understanding climate change. The net effect on (heating or cooling) Earth depends strongly on the vertical distribution of clouds: high clouds tend to warm; lower clouds to cool.

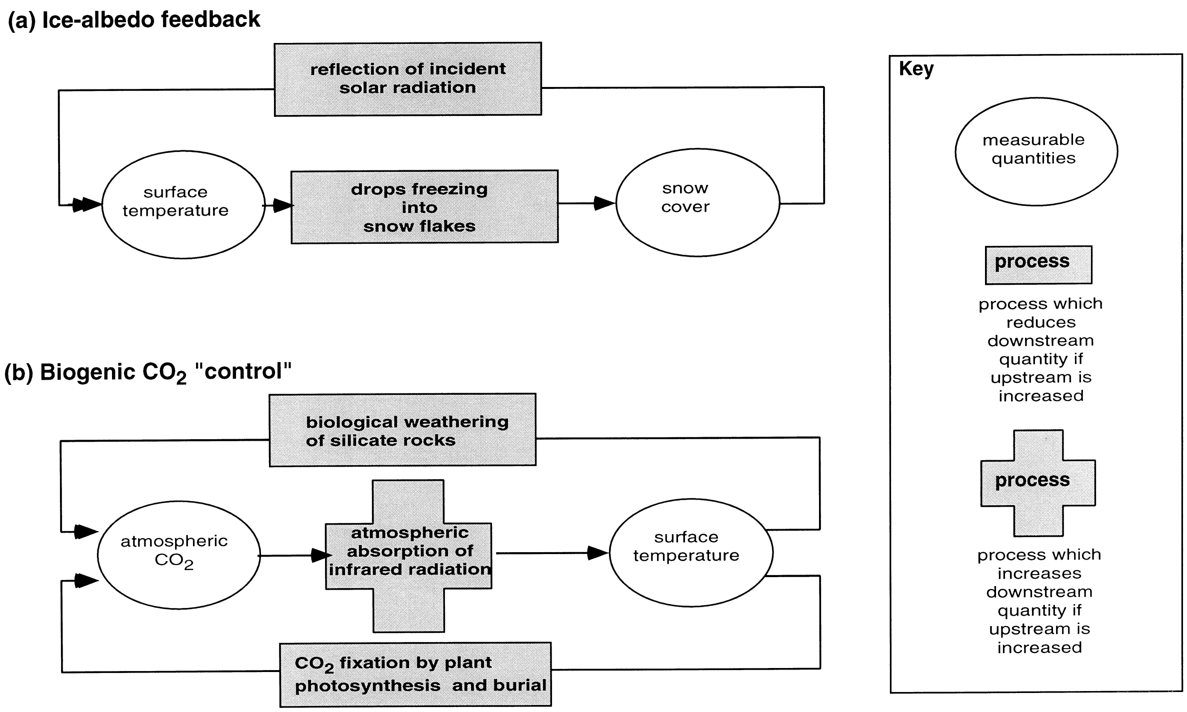

Figure 7. Schematic showing possible feedback loops with in the climate system. Measurable quantities are shown in ovals, and processes that modify them in shaped and shaded boxes displaying the sign of the change caused in the quantity immediately downstream of the shaped box in response to an increase in the upstream oval quantity (modified from Shearer, 1991). (a) Ice albedo feedback, and (b) biogenic feedback on the atmospheric CO_2 concentration. Note that the time-scale issue appears here also (cf. Table I). Both parts of this figure assume an intrinsic timeframe. Exploring other timeframes gives rise to different and perhaps additional feedbacks. For example, (b) has a long time-scale commensurate with rock breakup and plant carbon forming rocks. Looking over different time-scales, say the next 100 years, would show that plant respiration rates may change (a doubling for a 10 K warming, e.g. Raich and Schlesinger, 1992), which would be a negative feedback

The investigation of the effects of cloud cover has a long modelling history (Wetherald and Manabe, 1980, 1986, 1988; Rossow *et al*., 1982; Zhang *et al*., 1995; Randall *et al*., 1996). Indeed, the need for accurate information on clouds for climate modelling studies was the driving force behind the International Satellite Cloud Climatology Project (ISCCP) (Rossow and Garder, 1993a,b; Rossow *et al*., 1993; Rossow and Schiffer, 1999).

The role of cloud feedback in modelling the climate system is clearly important. If a climate change (say an increase in surface temperature caused by enhanced CO_2) increases the amount of high cloud, then this will increase the atmospheric 'greenhouse' by enhancing the amount of re-emitted radiation. Similarly, if the amount of low cloud is increased, then more solar radiation will be reflected from the Earth and the result will be a cooling. The combined effects of these cloud processes, coupled with the geographical variability of cloud cover types, means that, to date, a full understanding of the cloud radiation feedback has remained elusive (e.g. Cess *et al*., 1990; Senior and Mitchell, 1993). Specifically, the uncertainties surrounding cloud changes and cloud feedbacks are such that 'cloudiness' usually heads all lists of climate issues requiring further investigations, more resources and better observations. In the future, climate models may have to simulate the complex microphysics of clouds in order to capture and characterize responses to global warming and increased aerosol loading.

3.5.3. Biogenic feedbacks in the climate system. If CO_2 levels rise (or fall), the planet becomes warmer (or cooler) which increases (decreases) both the biological weathering of silicate rocks and the fixation of CO_2 by plants and its subsequent burial (Schneider and Boston, 1991). Both these effects reduce (increase) atmospheric CO_2, thus providing negative feedbacks (Figure 7(b)): i.e. the feedbacks tend to dampen the initial disturbance (e.g. Shearer, 1991). This hypothesis of 'bio-control' of geophysiology has prompted many climate model simulations relating to the much longer time-scales of climate history (and its future) (e.g. Henderson-Sellers *et al*., 1991).

The production of more (or less) dimethylsulphide (DMS) by marine phytoplankton has been proposed as a means by which the local sea surface temperatures and/or the incident solar irradiation at the ocean surface return to an earlier state following a disturbance. The mechanism involves DMS oxidation to water-soluble particles, which become cloud condensation nuclei (CCN). The production of more (fewer) CCN by the marine phytoplankton results in higher (lower) cloud albedos and hence increased (decreased) reflection of solar radiation reducing (increasing) the initial disturbance to ocean surface conditions (e.g. Charlson *et al*., 1987).

3.5.4. Combining climate system feedbacks. As many feedback effects operate within the climate system at any point in time in response to a variety of perturbations, simulating how feedbacks combine is an important attribute of climate models. In a system in which a change of surface temperature of magnitude ΔT is introduced, for no internal feedbacks, this temperature increment will represent the change in the surface temperature, but if feedbacks occur, there will be an additional surface temperature change and the new value of the surface temperature change will be

$$\Delta T_{\text{system}} = \Delta T + \Delta T_{\text{feedbacks}} \tag{3}$$

The value of ΔT_{system} can be related to the perturbation which caused it, providing a measure of the sensitivity of the climate system to disturbance. A convenient climate sensitivity parameter is given in terms of a perturbation in the global surface temperature ΔT which occurs in response to an externally generated change in planetary net radiative flux, ΔQ,

$$C[\delta(\Delta T)/\delta t] + \lambda \Delta T = \Delta Q \tag{4}$$

Here $\lambda \Delta T$ is the net radiation change applied to the climate system resulting from the internal dynamics, t is time and C represents the climate system's heat capacity. The equilibrium response ΔT and the forcing ΔQ are related by the feedback factor λ through $\Delta T = \Delta Q/\lambda$. Although Equation (4) represents a significant simplification of the system, it is useful for interpreting and summarizing the sensitivity of the overall climate system, and also provides a means of calibrating and intercomparing numerical climate models. Each process in the climate model will combine so that the overall feedback parameter will be given by

$$\lambda_{\text{system}} = \lambda_B + \lambda_{\text{icealbedo}} + \lambda_{\text{watervapour}} + \lambda_{\text{cloud}} + \lambda_{\text{biogenic}} \tag{5}$$

In Equation (5), λ_B is the base sensitivity of the climate system to changes in radiative input. Each of the other sensitivities (λ_i) represent the feedback factors associated with other aspects of the climate system (see McGuffie and Henderson-Sellers, 1997, for a more complete discussion of feedback factors, including definitions and ranges). Various estimates have been made of the feedback effects likely to be caused by biogenic feedbacks (e.g. Lashoff, 1991). The addition of such feedbacks to those considered above could raise the surface temperature increase due to increasing CO_2 or reduce it to a zero response. The importance of both forcings and feedback effects in the climate system depends upon the time-scale of behaviour of the subcomponents each affects (Table I).

4. CURRENT CLIMATE MODELLING DIRECTIONS

Development of climate models has been driven by a number of factors. As discussed in Section 3, the development of climate models has not proceeded through time from 'simplest' to 'most complex'. Many of the simpler models have been developed to isolate features of complex models or to conduct simulations on longer time-scales or including a larger number of climate components. Two model families which currently comprise a major component of climate research are the GCMs, which have a long pedigree, and the EMICs, which are designed to bridge the gap between GCMs and the simpler physical models such as the EBMs.

4.1. GCMs

The aim of GCMs is the calculation of the full three-dimensional character of the climate comprising at least the global atmosphere and the oceans. If a model were to be constructed which included the entirety of our knowledge on the atmosphere-ocean system, it would not be possible to run it on even the fastest computer. For this reason, even GCMs, currently the most complicated numerical models, can only be simplifications of our current knowledge of the climate system.

As discussed in Section 3, GCMs are the direct descendants of the numerical weather prediction models, the basis of which is the representation of the climate system as a series of differential equations representing the many processes in the atmosphere and oceans. The solution of these equations (Table III) 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, momentum and salt transport.

Treatments of individual components are generally complex (e.g. Briegleb, 1992; Flato and Hibler, 1992; Hack, 1993; Dickinson, 1995; Deardorff, 1978), although the general process is the same for all aspects of the climate system. The first step in obtaining a solution is to specify the atmospheric, oceanic and surface conditions at a number of 'grid points', obtained by dividing the Earth's surface into a series of patches, so that a global grid results (Figure 8). Conditions are specified at each patch for the surface and multiple layers in the atmosphere and ocean. The resulting set of coupled non-linear equations are then solved at each patch using numerical techniques. Various techniques are available, but all use a timestep approach (e.g. Haltiner and Williams, 1980; Hansen *et al*., 1983; Hack, 1992). Computational techniques divide atmospheric models into two main groups: spectral models and grid models. Spectral models (Bourke *et al*., 1977; Boer *et al*., 1984), make use of fast Fourier transforms (FFT) to conduct part of the calculation in a wave formulation; whereas grid point models (Manabe *et al*., 1979; Hansen *et al*., 1983; Mitchell *et al*., 1995) make use of a straightforward rectangular grid.

Modelling the full three-dimensional nature of the ocean is more difficult than capturing the atmosphere because the scales of motion which exist in the oceans are much smaller than those in the atmosphere (ocean eddies are around 10–50 km, cf. around 1000 km for atmospheric eddies), and the ocean also takes very much longer to respond than the atmosphere to changed forcing (Figure 6, cf. Table I). The smaller scales demand a smaller grid size. Hence, there are very many more points at which computations must be made. The dynamics of the ocean are governed by the amount of radiation which is available at the surface, and by the wind stresses imposed by the atmosphere, but the flow of ocean currents is also constrained by the positions and shapes of the continents (Gates, 1979). The formation of

Table III. Fundamental equations solved in GCMs (after *A Climate Modelling Primer*, by K McGuffie and A Henderson-Sellers, 1997, reproduced by permission of John Wiley & Sons, Ltd)

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 oceanic 'tracers' and to cloud liquid water
4. *Ideal gas law* (an approximation to the equation of state—atmosphere only), i.e. Pressure × volume = gas constant × absolute temperature

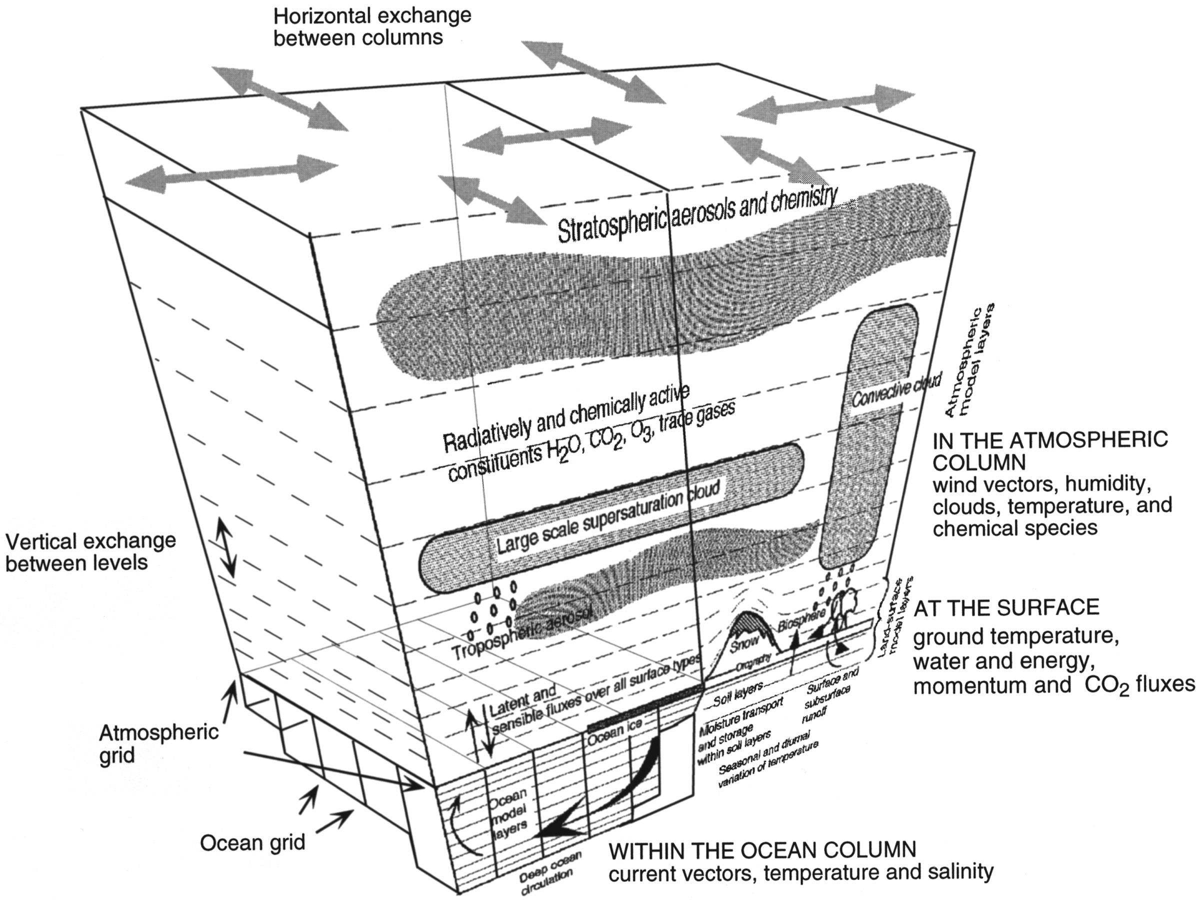

Figure 8. Illustration of the basic characteristics and processes within a GCM, showing the manner in which the atmosphere and ocean are split into columns. Both atmosphere and ocean are modelled as a set of interacting columns distributed across the Earth's surface. The resolutions of the atmosphere and ocean models are often different because the processes differ and have different time-scales and equilibration times. Typically, many types of cloud and land surface are treated. In this example, soil moisture is modelled in a number of layers and tropospheric and stratospheric aerosols are included (redrawn from *A Climate Modelling Primer*, by K McGuffie and A Henderson-Sellers, 1997, reproduced by permission of John Wiley & Sons, Ltd)

oceanic deep water is closely coupled through salinity to the formation and growth of sea-ice so that ocean dynamics demands effective inclusion of sea-ice dynamics and thermodynamics (Semtner, 1976; Hibler, 1979; Flato and Hibler, 1992). The Antarctic circumpolar current, for example, is largely controlled by topography, and errors in the path of this current can result in significant errors in sea surface temperature (Gent *et al*., 1998; Gordon *et al*., 2000). Furthermore, deep water circulation of the ocean can take hundreds or even thousands of years to complete, so that ocean models which include these dynamic processes often have to be asynchronously coupled with atmospheric components to provide the most detailed models of the physical climate system. However, Gordon *et al*. (2000) describes synchronously coupled simulation. Ocean models are generally constructed on a rectangular grid, as the sharp discontinuities at the edge of oceans make them unsuited to the spectral techniques which are used for some models of the atmosphere (Gates *et al*., 1996).

As well as acting as a thermal 'fly-wheel' for the physical climate system, the ocean also plays a central role in the carbon cycle, absorbing approximately half of the carbon which is released into the atmosphere every year (Schimel *et al*., 1996). Computational constraints and breakthroughs have dictated the evolution of carbon budgeting in numerical climate models. Originally, GCMs 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, while for the oceans, restrictions of computer power meant that the model output was employed before full equilibration could be achieved. This often resulted in the 'drift' of the ocean climate away from present-day conditions, which was initially corrected by applying adjusting fluxes at the ocean surface to compensate for persistent and systematic errors (Gates *et al*., 1996; Gregory and Mitchell, 1997). These simplifications and fixes have largely been overcome through additional computational capability (Gordon *et al*., 2000), which allows higher resolution in the ocean component. However, realistic surface fluxes from the atmosphere to the ocean were another important feature of achieving the improvements (e.g. Pope *et al*., 2000). This achievement of adequate present-day representation without *ad hoc* fix-ups means that the global system is able to be integrated for long periods and hence that global scale carbon budgeting can be attempted (e.g. Schimel *et al*., 1996)

Computational constraints lead to other problems for climate modellers (cf. Figure 5). With a coarse grid spacing, small-scale motions (termed sub-gridscale), such as thundercloud formation, soil moisture transfer or oceanic eddies, cannot be modelled, however important they may be for the climate system (Figure 8). Fine grid models (e.g. Bengtsson, 1996; Bengtsson *et al*., 1996) can be used for specific predictions because the integration time is short, but full climate models must rely on some form of parameterization of sub-gridscale processes. The definition of 'sub-grid' depends upon the type of climate model and the computational power available.

In a manner similar to time-scales, the parameterizations in numerical climate models must account for variations across space scales. In particular, the recognition of feedbacks between scales; edge effects; different dominant processes and non-linearities pose real challenges for those developing model parameterizations (Table IV(a)). The solutions by modellers range from acknowledging but choosing to ignore; through simple adjustments to thresholds and the use of lumped sub-models with tuned parameters; to calculating the effects explicitly, either by notching up the overall spatial resolution or by creating new high resolution subcomponent schemes (Table IV(b)).

Table IV. Sub-grid phenomena demanding parameterization solutions in numerical climate models (modified from *Climatic Change*, **44**, 2000(a), pages 225–263, Upscaling in global change research by DD Harvey, Table III, reproduced by kind permission of Kluwer Academic Publishers)

(a) Parameterization challenges	Examples
Different process dominate at different scales	Terrestrial and marine ecology
Edge effects	Sea-ice; terrestrial ecology
Spatial variability and process non-linearity	Surface hydrology; formation of clouds and precipitation; photosynthetic response to higher atmospheric CO_2; natural disturbances
Feedbacks between scales	Transpiration response to higher CO_2; economic costs of greenhouse gas emission abatement
Temporal lag dependent on pre-existing conditions	Surface runoff

(b) Parameterization solutions	Examples
Create a new model to integrate effects of next smallest scale	Cumulus cloud ensembles; ecophysiology
Greatly increase model resolution	Ocean GCMs
Refuse	Tropospheric chemistry (regional to global transition)
Ignore	CO_2 control on stomates
Adjust critical thresholds	Formation of clouds at relative humidities less than 100%
Use a lumped model with tuned parameter values	Surface hydrology; treatment of partial cloud cover with overlap; atmospheric chemistry (effective emissions)

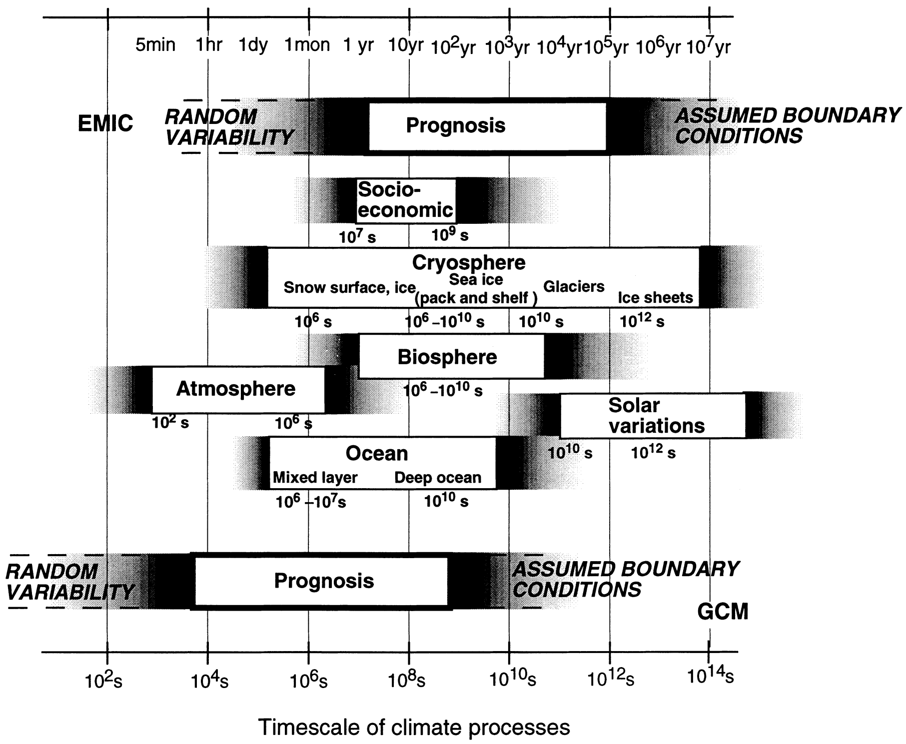

Figure 9. The different time-scales captured and encompassed by an EMIC and GCM, the two main types of climate models used for climate assessment. Both span a range of different domains (e.g. cryosphere and atmosphere) and temporal scales. The main emphasis tends to migrate as a function of the type of model. The domain in which the model simulates the behaviour of the system is labelled '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. This depiction is generalized for timeframes and domains across the community. There are GCMs which have been run 100000 years and EMICs that consider annual time-scales (modified from *A Climate Modelling Primer*, by K McGuffie and A Henderson-Sellers, 1997, reproduced by permission of John Wiley & Sons, Ltd)

4.2. EMICs

The sliding criteria from sub-grid variability to frozen (or specified) boundary conditions for global climate models and EMICs are shown in Figure 9. Within the EMICs, modellers intentionally adopt simple approaches to selected processes; for example, interactions between the surface and the near-surface layer of the atmosphere. Detailed consideration of the transfer processes at the surface of the Earth are computationally too demanding for explicit inclusion in EMICs, but are included in GCMs (Figure 9). Achieving the most beneficial trade-off between calculated, sub-grid variation and boundary specification defines the art of climate modelling (Table IV). Some modellers (e.g. Giorgi *et al.*, 1994) have used the approach of nesting a regional model within a global model to provide high resolution simulations of a particular region. Results from such studies are dependent on the quality of the global scale model used to supply the regional boundary conditions and upon the verity of the nesting procedure (e.g. Seth and Giorgi, 1998).

Aspects of the physical character of the climate system deemed to be critically important in GCMs, such as the radiation fluxes at the Earth's surface, are parameterized in EMICs. Cloud amount may be made to be dependent on surface temperature, and surface albedo regarded as constant for a given latitude: remarkably reminiscent of the early EBMs (e.g. Budyko, 1969). Atmospheric dynamics are often not modelled explicitly in EMICs. Instead, simple parameterizations such as a 'diffusion' approximation are employed to parameterize heat transport: the same approach as used in 2D SDs (e.g. Potter *et al.*, 1981; Potter and Cess, 1984).

Earth system models employ whatever resolutions and parametrization of the physical climate system are deemed necessary. Gallee *et al*. (1991) have used a two-dimensional, zonally averaged model, whereas Stocker *et al*. (1992) and Marchal *et al*. (1998) use 2.5-dimensional models with a simple atmospheric module. Statistical–dynamical atmospheric modules have been used by Petoukhov *et al*. (1999), whereas Opsteegh *et al*. (1998) choose a reduced-form option of a comprehensive (i.e. fully three-dimensional) numerical model. There are also EMICs which involve an energy-moisture balance model coupled to an OGCM and a number of both thermodynamic and dynamic/thermodynamic sea-ice models (e.g. Fanning and Weaver, 1997). Schnellnhuber (1999) provides an overview of the system-wide approach employed by EMICs.

The power of these EMICs is that they can be applied to very wide ranging time-scales (e.g. Loutre and Berger, 2000). Thus, while one or two GCMs and a number of EMICs (e.g. Broccoli, 2000) have been used to try to investigate the last glacial maximum (LGM), as well as the collapse of the ocean conveyor in greenhouse warming experiments (Rahmstorf and Ganopolski, 1999), both now being recognized as an important aspect of climate model evaluation/validation, only an EMIC has been used to evaluate the effect of historical land cover change over hundreds of years (Brovkin *et al*., 1999)

There are many ways in which the biosphere is of importance to the climate system. These include transfer of moisture from the soil into the atmosphere, modification of the albedo, which changes the amount of radiation absorbed by the climate system, responsibility for the exchange of carbon and other chemicals, and modification of the surface roughness which alters the exchange of momentum. Only recently has the interactive nature of the plant life of the planet been included in climate models and, although GCMs are attempting to compute some biospheric attributes, important developments have also been made in the lower dimensional EMICs (e.g. Joos *et al*., 1996; Kleidon *et al*., 2000). The first approach has been to delineate geographic boundaries of biomes (vegetation groups characterized by similar species) by using simple predictors available from the GCM such as temperature, precipitation and possibly sunshine or cloudiness. Currently, attempts are being made to evaluate these methods using palaeo-reconstructions of vegetation cover during past epochs (Ducoudre *et al*., 1993). Some modellers have included simplified 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 may provide useful predictions of future responses of the biosphere including the issue of possible future CO_2 fertilization of the biosphere (Henderson-Sellers and McGuffie, 1995; Zhang *et al*., 2001).

Many EMICs are two-dimensional, i.e. they represent either the two horizontal dimensions or the vertical and one horizontal dimension (Gallee *et al*., 1991; Petoukhov *et al*., 1999). It is the latter which are more common, combining the latitudinal dimension of the EBMs with the vertical one of the RC models (McGuffie and Henderson-Sellers, 1997). These models also include a more realistic parameterization of the latitudinal energy transports, so that the global circulation is assumed to be composed mainly of a cellular flow between latitudes (cf. Figure 1(b)).

In the 1980s, the lack of zonal resolution in two-dimensional SD models caused them to be replaced by GCMs when consideration of the effect of perturbations on the present climate was the main goal (Table II). Indeed, where the goal is to capture the impact on climate of the forcing by land-ocean contrasts and orography, a two-dimensional SD can never compete with a GCM. However, recent developments in EMICs rekindled interest in the value of 'stripped-down' or 'computationally efficient' climate modelling when there are additional aspects to include, such as the technology alternatives associated with global greenhouse gas emissions (e.g. Green, 2000). Similarly, advances in the understanding of baroclinic waves achieved from studies of physical SD models are now being incorporated into EMICs. Two-dimensional models have been employed to make simulations of the chemistry of the stratosphere and mesosphere; these models typically involving tens to hundreds of chemical species and many hundreds of different reactions.

4.3. Complexity versus conceptualization

Potential changes in the climate system can be expected to be of a wide variety and to operate and equilibrate over many different time-scales (cf. Figure 9 and Table I). Calculating many of the key processes operating in the climate system by means of computer programs requires simplifications which reduce complexity, uncouple or disregard some feedbacks and, hence, reduce computing, data and parameterization requirements (e.g. Table IV). The recognition that different aspects of the climate system demand different types of simplification has led to the development of a wide variety of climate model types (e.g. Shine and Henderson-Sellers, 1983; Henderson-Sellers and McGuffie, 1999). In selecting the type of climate model to be employed for a particular simulation task, it is necessary to balance socio-economic and scientific understanding and outcome demands against available computational capability, and review all three in the context of the data with which parameterizations and initializations will be established (cf. Shackley *et al*., 1998; Harvey, 2000b). It is also deemed necessary, or at least preferable, to balance the relative level of detail in the representation and the level of parameterization within each component of the climate system (cf. McGuffie and Henderson-Sellers, 1997) (Figure 8, cf. Figure 6).

EMICs are designed to bridge the gap between the three-dimensional global climate models and the demands for policy evaluations related to climate change. As described in Section 4.2, the main characteristics of EMICs are that they describe most of the processes implicit in GCMs including biogeochemical feedbacks (Figure 7) and socio-economic trends (e.g. Figure 9). They are designed to be computationally efficient enough to allow for long-term climate simulations over several tens of thousands of years or for a broader range of sensitivity experiments spanning millennia. EMICS pay a price for this design. This is most obvious in the atmospheric modelling area, where the dynamical basis of EMICs is very weak, commonly relying on parameterizations of zonally averaged fluxes in terms of zonally averaged winds and temperatures, or on very coarse resolutions east–west. Such methods rely heavily on tuning of parameters to fit the current climate, and thus, may not be reliable in the climate prediction context.

Simplifications or enhancements can be made so that any climate model has the appropriate complexity for each task (Table IV). The comparison between the domains of the EMIC and GCM shown in Figure 9 illustrates the creative tension which still pervades climate modelling: the conflict between the drive towards increased complexity and the pull-back to simpler schemes which can be applied over longer time-scales (cf. Covey, 2000). Simpler models allow exploration of the potential sensitivity of the climate to a particular process over a wide range of parameters. For example, Wigley (1998) used a modified version of the Wigley and Raper (1987, 1992) upwelling–diffusion energy budget climate model (see Kattenberg *et al*., 1996) to evaluate Kyoto Protocol implications for increases in global mean temperatures and sea-level. While such a simple climate model relies on climate sensitivity and ice-melt parameters obtained from a full GCM, it allows a first-order analysis of various post-Kyoto emission reduction scenarios.

The perceived and desired importance of various processes captured in climate models and the basis for parameterizations employed in their incorporation into different types of simulation can be discussed using the 'climate simulation pyramids' (Figure 10). The edges of the two square pyramids represent one series of elements in today's climate models, while integration is shown increasing upwards as far as the top of the lower pyramid and then decreasing as the edges of the upper pyramid diverge. Around the base of the lower pyramid are the simpler 'science-based' climate models, which incorporate only one primary process. As indicated on the lower pyramid in Figure 10(a), there are four basic types of physical climate model:

1. EBMs are one-dimensional models predicting the variation of the surface temperature with latitude. Simplified relationships are used to calculate the terms contributing to the energy balance in each latitude zone.
2. One-dimensional RC models compute the globally averaged vertical temperature profile by explicit modelling of radiative processes and a 'convective adjustment', which re-establishes a predefined environmental lapse rate.

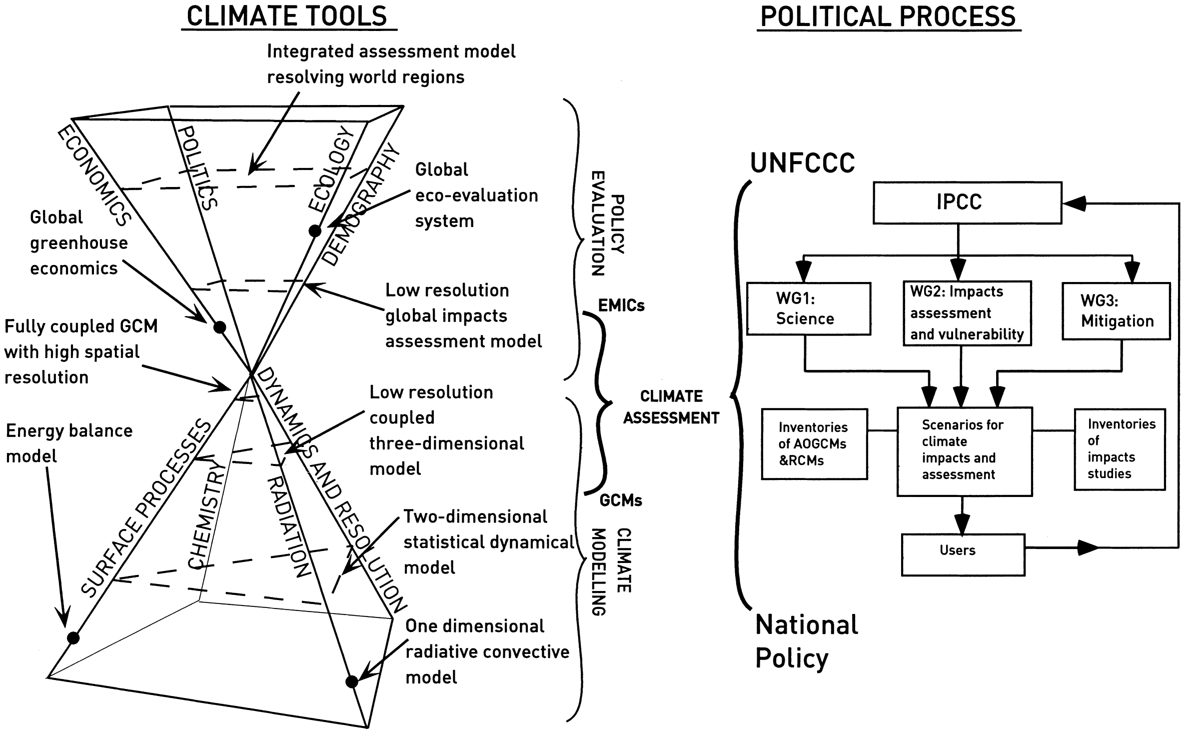

Figure 10. (a) The 'balancing climate pyramids' illustrates the range of modelling tools. The lower pyramid represents aspects of climate modelling (here shown as surface processes, radiation, dynamic and resolution and chemistry). Models can include one or more aspects of the climate system. Typically as more are included the resolution of the model increases (i.e. moves nearer the apex). The upper pyramid represents aspects of climate policy evaluation (here, shown as economics, politics, ecology and demography). Models can include one or more aspect of climate policy. Typically, as more are included, the resolution of the model increases (i.e. moves further from the lower apex). The schematic illustrates that increasing complexity does not stop at the apex of the climate modelling pyramid, but 'blossoms' into many diverging aspects and strands of climate policy evaluation. (b) The policy associated with climate assessment spans a range similar to that of the balancing pyramids. International treaties, such as the UNFCCC create the need for intercomparisons and international reports (e.g. the IPCC), which flow finally to users at national and local levels. These users' interpretations of the current status influences national climate policy, and so the loop is closed. The main areas of modelling input to climate assessment (i.e. from (a) to (b)) today lie between and including EMICs through to GCMs. (modified from *A Climate Modelling Primer*, by K McGuffie and A Henderson-Sellers, 1997, reproduced by permission of John Wiley & Sons, Ltd)

3. Two-dimensional SD models deal explicitly with surface processes and dynamics in a zonally averaged framework, and have a vertically resolved atmosphere. These models were the starting point for the incorporation of atmospheric chemistry in global models about 15 years ago.
4. GCMs. The three-dimensional nature of the atmosphere and the ocean is incorporated.

Moving vertically upwards in the lower pyramid of Figure 10(a) shows increasing integration (i.e. more processes included and linked together). The vertical dimension indicates increasing resolution on both pyramids: models appearing higher in the balanced pyramid pair have higher spatial and temporal resolutions. Thus, for example, as the synthesis diminishes among the themes of economics, politics, ecology and demography, the detail (or resolution) increases, leading to complex models of, for example, ecological generation and transition without any other links to themes at the top of the upper pyramid. Thus, integration is a maximum at the balancing tips of the pyramid and decreases as the upper pyramid is traversed.

The components considered to be important in constructing or understanding a modern numerical model of the climate system are those captured in this pyramid pair. The lower pyramid encompasses the science-based model types calculating:

- Radiation—the input and absorption of solar radiation and the emission of infrared radiation;
- Dynamics—the movement of energy around the planet by winds and ocean and vertical movements (from small scale turbulence to deep-water formation);
- Surface processes—inclusion of the effects of sea- and land-ice, snow, biota and rock cycles and their resultant changes;
- Chemistry—the chemical composition of the atmosphere, ocean and surface interfaces and the interactions with other components (e.g. carbon exchanges between ocean, atmosphere, biota, rocks and soil).

To these characteristics, the demands of policy makers have added the characterization of human enterprises and endeavours to form the upper climate pyramid in Figure 10(a) which includes:

- Economics—the way in which global markets will demand and fulfil technological and other human requirements (e.g. the viability of fossil cf. 'alternative' fuel resources);
- Politics—a representation of the effects of international treaties, trade and other political instruments designed to curb or restrain human impact on the climate system;
- Ecology—the way in which ecosystems (plants, animals, insects, microbes etc.) respond to and affect climate;
- Demography—human societal response to, and interaction with, the climate system.

It is not accidental that the cartoon of the variety of ways of simulating the climate system depicted in Figure 10(a) suggests an unstable balancing of social and biological aspects on top of calculations of physical and chemical processes. There is an interesting debate about the way in which policy relating to climate change draws on different model types and model outputs (e.g. Shackley *et al*., 1998, cf. Wigley, 1998; Covey, 2000). However, the main region of this pyramidal structure in which climate assessment is currently conducted lies between the GCMs and the EMICs. Above the EMICs, a myriad of policy components fan out achieving ever increasing resolution. Below the GCMs, the fundamental components of the climate system (cf. WMO, 1975) separate into the traditional disciplines of science. Climate policies, say with regard to the protocols to limit damage to stratospheric ozone or reduce human-induced greenhouse gas emissions, tend to be developed from models lying between and particularly emphasizing GCMs and EMICs. EMICs have recently been instrumental in increasing understanding of the climate system as it is affected by socio-economic trends and very valuable for determining the sensitivity of more complex climate models to geological and astrophysical forcings (Rothman, 2000).

5. MODELLING ISSUES FOR TODAY

5.1. Parameterization in climate models

As the climate system depends upon scales of motion and interactions ranging from molecular to planetary dimensions, and from time-scales of nanoseconds to geological eras, parameterizations are a necessary part of the modelling process (Table I and Figure 8). A decision is generally made very early in the model construction phase about the range of space- and time-scales which will be modelled explicitly (Table IV). Outside this range, there are 'frozen' boundary conditions at longer time-scales and 'random variability' at shorter time-scales. Thus, the two examples shown in Figure 9 illustrate the range of prognostic computations for a fully coupled GCM used as an example, say, for El Niño–Southern Oscillation (ENSO) simulations and an EMIC, again as an example, say, examining the effect of Milankovitch variations on the 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. The same process occurs for space scales with sub-grid processes being either parameterized or ignored (Harvey, 2000a and Table IV).

The interactions between processes in any model of the climate system are crucially important. So-called 'wiring diagrams', which show these interactions, are often used to illustrate the complexity of incorporating them all adequately. A most important concept in climate modelling is that both the relative importance of processes and the interlinking of different processes are a strong function of the time-scale being modelled. All aspects of parameterization are subsumed in this statement. 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 (Figure 7). 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. Understanding and correctly developing and using parameterizations is a fundamental and evolving component of the art of climate modelling (e.g. Figure 6).

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 while the other is neglected (e.g. Figure 7). 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 6 depicts the evolution over the last 40 years or so of two important sub-components of a full global climate model: the land-surface and the ocean. As was noted with reference to Table II, the history of climate model development is not always a straightforward move from simple to more complex, although there is a tendency for more processes to be included and more complex parameterizations to be employed when data and computer power permit (cf. Figure 5).

In the case of ocean modelling, two philosophies have been evident throughout the 40-year lifetime of climate modelling as a discipline. These might be characterized as the 'go for global and three-dimensional' and the 'ensure eddy resolving capability' schools of thought. When climate modelling was young, in the 1960s and 1970s, achieving both was impossible because of the severe limitations of computational power (e.g. Bryan and Cox, 1967).

As a consequence, the 'real' oceanographers maintained models that captured mesoscale eddies but could not encompass the globe (e.g. Pedlosky, 1979). At the same time, the 'early' climate modellers gave up resolution for global coverage (Semtner and Chervin, 1988). It is only in the last 10 years or so that computational power and agreement on types of parameterizations have permitted some convergence among these two ocean modelling schools. Just as this convergence was beginning, a new group of ocean models arose in response to the need to try to understand and subsequently predict ENSO events (e.g. Zebiak and Cane, 1987).

The development of land-surface schemes has moved somewhat more directly from simplicity to complexity (e.g. Sellers *et al*., 1997), but even within this community, there is great diversity of model types and serious debates rage about how best to represent sub-grid heterogeneity and include new fluxes such as those of carbon (see e.g. Henderson-Sellers and Hopkins, 1998) (Figure 6).

An interesting twist on the evolution from the very simple 'bucket' land model of Manabe (1969) to the complex schemes of today (e.g. Schlosser *et al*., 2000) is that recent research seems to indicate that most of the achieved complexity in land surface schemes is redundant. Desborough (1999) has shown that a very large proportion of the range in results given by all land-surface schemes can be captured by a bucket model with the addition of only a single extra parameter: stomatal resistance. However, it is still necessary to get the stomatal and aerodynamic resistances correct and the former requires correct ground hydrology, itself not an easy task (e.g. Milly and Dunne, 1994). The net radiation must also be correct, which requires correct albedos and correct incident radiation. The difficulties which persist in simulating clouds and thus the downward solar radiation are currently obscuring real improvements in land surface schemes (Slater *et al*., 2001).

A traditional view of parameterization would be that the simplest approximations to the climate system (models) lie at the base of the lower, climate modelling, pyramid, (Figure 10(a)) with increasing parameter numbers being synonymous with increasing (and perhaps more desirable) complexity on ascent through the lower pyramid. The apex of this pyramid would be presumably that all necessary processes would be parameterized at the appropriate level in GCMs. Indeed, some GCM modellers claim that to include 'all other processes' is relatively easy because they are perceived to require very much less computational power than the demanding calculations associated with dynamics and radiation. Others would contest this, pointing to chemistry, which has the potential to be enormously demanding, and human systems which may require very many realizations to try to envelope likely social and technological adjustments.

While a goal of full representation of *all* physical, chemical and biological processes may have seemed absurd in the 1980s, it could be argued that the desire for increasing computer power and parameterization complexity sought by some GCM modelling groups at that time was visionary (Table II). The foresight behind GCM development had, by the end of the 1990s, led to models which did, as fully as computing power allowed, include many of the processes envisaged in the modelling plans of the early 1990s (ocean and land carbon cycles, vegetation growth and decay, chemistry, aerosols (generation, scavenging, radiative effects etc.) and probably more, as chemistry is now being included to a greater extent than anticipated (cf. Shackley *et al*., 1998). An alternative view might be that some of the more sophisticated lower-resolution EMIC models on the upper, policy evaluation, pyramid contain the maximum information currently available/verifiable for very long-term climate integration periods (e.g. Rodhe *et al*., 2000). These models, although greatly simplifying atmospheric and ocean dynamics, can be tuned to the current climate. Thus, while such tunings hold, they are adequate and appropriate parameterizations. In such an EMIC application, the climate system over long time-scales would be deemed to be insensitive to higher-resolution features and be assumed to be valid within today's climate sensitivity. The key elements in all numerical climate models are: (i) the choice of computational engine (Figure 5), (ii) the parameterization choice, through which processes that cannot be treated explicitly are instead related to variables that are considered directly in the model (Table IV), and (iii) the time and space scales to be included (Figure 9).

5.2. Climate model evaluation

It is now well established that human activities are impacting on the climate system (e.g. Houghton *et al*., 1996). Over the last few years, models and observations have combined to confirm this fact. As a result, treaties and protocols are being developed and agreed which aim to reduce, and perhaps ultimately reverse, these human-induced disturbances to the climate system (e.g. Taplin, 1996). The tools with which future climates are to be predicted are climate models, but first these must be tested against observed climates. This process is now termed model 'evaluation', although many researchers still use the term 'validation'. The former has been chosen over the latter by Working Group I of the IPCC for its Third Assessment Report because it is argued that 'evaluation' denotes a comparison, while 'validation' appears to offer some form of approval (AJ Pitman, personal communication, February 2000).

There is now a plethora of climate model intercomparison exercises, prompting new demands for high quality observations (e.g. Boer *et al*., 1992; Chen *et al*., 1997). One of the oldest and most influential of these intercomparisons is the Atmospheric Model Intercomparison Project (AMIP, Gates *et al*., 1999; Phillips *et al*., 2000). This project has even produced a new model intercomparison and evaluation diagram (Figure 11). In these polar plots, the angular displacement represents the correlation between the model result and the validation data (with the horizontal = 1.0 being agreement). The distance from the origin represents the normalized standard deviation of the model result over that of the validation data (so that a radius of 1.0 is exact agreement).

Thus, a 'perfect' model would generate a result with zero angular displacement and at a distance of 1 from the origin. Unfortunately, the scatter among observational datasets (the 'truth' or reference point) is often as large as that among the AGCMs participating in AMIP.

5.2.1. Climate of the mid-Holocene. The mid-Holocene ($\sim$ 6000 BP) has been chosen to test the response of climate models to orbital forcing with CO_2 at pre-industrial levels, but a cryospheric extent and state similar to that of today. The Earth's orbital configuration intensifies (weakens) the seasonal distribution of the incoming solar radiation in the northern (southern) hemisphere by about 5%. Climate models of different complexities have been used to try to 'hindcast' this palaeoclimate (e.g. Joussaume and Taylor, 1995). The degree to which models can correctly capture known characteristics of this era is one means of verifying their performance prior to their use for future climate prediction.

In the Northern Hemisphere summer, the vast majority of modern climate models simulate an increase and northward expansion of the African monsoon; warmer than present conditions in high northern latitudes; and drier than present conditions in the interior of the northern continents. Palaeoclimatic data support the expanded monsoon in northern Africa (Street-Perrott and Perrott, 1993; Hoelzmann *et al*., 1998), but the simulated drying in central Eurasia (e.g. Harrison *et al*., 1996) is not supported by palaeodata. The modelled warming in the Arctic (Texier *et al*., 1997) and drying in interior North America (Webb *et al*., 1993) are in agreement with palaeodata. However, while climate model simulations produce a northward shift of the Arctic treeline in agreement with observed shifts (Tarasov *et al*., 1998), all appear to underestimate greatly its extent (TEMPO, 1996; Texier *et al*., 1997; Harrison *et al*., 1998).

Regional simulations of mid-Holocene climates are highly variable. For example, Masson *et al*. (1999) show considerable differences among model representations of European conditions during this period. Some key features of palaeoclimate reconstructions, from pollen and lake level data, such as warmer than present winters in north eastern Europe and colder and/or moister than present growing seasons in southern Europe (Cheddadi *et al*., 1997; Prentice *et al*., 1998) are not well reproduced by the numerical models. Similarly, in northern Africa, the northward displacement of the desert-steppe transition, whilst consistent with palaeoreconstructions, underestimates the extent (Harrison *et al*., 1998).

Mid-Holocene climate system simulations have also been conducted using models which include ocean subcomponents (Kutzbach and Liu, 1997; Hewitt and Mitchell, 1998). While these successfully produce a larger enhancement of the African monsoon than the atmosphere only experiments, the simulated sea surface temperature changes are small, and probably insufficient to generate the observed changes in biome shifts in northern Africa.

A few studies have attempted a comparison of coupled climate model variability with observed variability derived from Holocene palaeodata. As compared with the levels of decadal-scale variability in summer palaeo temperature proxies from 1600–1950 (Bradley and Jones, 1993 and Figure 2(e)), analysis of long control integrations from three global climate models (Barnett *et al*., 1996) found that the models underestimate climate variability with increasing disparity with observations at lower frequencies.

The issue of changes in interannual to interdecadal variability under climatic forcing conditions different from the present day has also been examined using coupled models. Some palaeoenvironmental evidence has suggested that short-term climate variability associated with the ENSO was reduced during the early to mid-Holocene (Sandweiss *et al*., 1996; Rodbell, 1999). Up to now, only one coupled model simulation has analysed ENSO variability and did not find any significant change at the mid-Holocene (Otto-Bliesner, 1999).

Continental surface changes, particularly vegetation, standing surface water and soil moisture differences, are believed to have provided additional important climate system feedbacks during the mid-Holocene. For example, vegetation changes in northern Africa seem to have favoured greater monsoon precipitation (e.g. Kutzbach *et al*., 1996; Claussen and Gayler, 1997; Pollard *et al*., 1998; Claussen *et al*., 1999). In addition, the occurrence of lakes, rivers and marshes (Coe and Bonan, 1997) helps to intensify the monsoon in model simulations. These land-surface feedbacks amplify the effects of orbital (Milankovitch) forcing at high latitudes where they lead to greater and more realistic shifts of vegetation cover (Foley *et al*., 1994; Texier *et al*., 1997).

Despite these successful simulations of feedbacks in the climate system, it has not yet proved possible to explain the observed biome shifts in the Sahara in the mid-Holocene. Combining feedbacks between land and ocean achieves an improved agreement with palaeodata because the ocean feedback increases the supply of water vapour, while the vegetation feedback increases local moisture recycling. The combined effect is to lengthen the monsoon season in Africa.

5.2.2. Climate of the LGM. At the height of the LGM (Figure 2(c)) about 20000 years ago, the climate system differed very markedly from today. Massive ice sheets extended to great heights across the northern continents; sea-levels were lower; sea-surface temperatures cooler; and CO_2 levels reduced (200 ppmv cf. today's level of 360 ppmv). The ambient solar radiation was, in contrast, very similar to that of today. Macrofossil evidence have been used to estimate the annual mean global cooling of the oceans as being about $-4°C$ (CLIMAP, 1981). The importance of this period is linked to assessment of the interactions between different forcings (e.g. reduced CO_2 but similar solar radiation as compared with today's climate) and the responsiveness of feedback effects, especially the ice–albedo feedback (e.g. Figure 7).

As well as physical forcings and feedbacks on the climate of the LGM, there are likely to have been pronounced biogenic feedbacks prompted by reduced CO_2 levels causing changes in vegetation structure (Street-Perrott *et al*., 1997) and CO_2-induced changes in leaf conductance (Jolly and Haxeltine, 1997), as well as vegetation and biome responses to climate change itself and possibly even bio-geochemical feedbacks (e.g. Figure 7(b)). In Eurasia, forests were replaced by tundra or steppe which may also have contributed to the observed cooling (Kubatzki and Claussen, 1998). In the tropics, it has been hypothesized that large areas of warming may have been caused by climate-related deforestation (Crowley and Baum, 1997). Atmospheric chemistry feedbacks may also have resulted from known land-surface changes. For example, mineral aerosol (dust) concentrations significantly increased during the LGM, possibly inducing further cooling in the tropics.

Assessing the agreement (or its lack) between model-simulated climates and those derived from palaeodata can offer an independent check on climate model validity. Estimates for the LGM for global climate sensitivity are now being derived from a variety of simulations specifically to compare with climate sensitivity estimates for a doubling of CO_2.

5.2.3. Attribution of the greenhouse signal. It may be considered somewhat paradoxical to choose the greenhouse signal as an example of evaluation of climate models. Nonetheless, the very careful and detailed work undertaken by climate modellers to unravel the forcings and feedbacks that have controlled the climate of the twentieth century is indeed a robust evaluation of the models used (Santer *et al*., 1996).

The IPCC Second Assessment Report concluded that 'the balance of evidence suggests that there is a discernible human influence on global climate' (Houghton *et al*., 1996, p. 5). This statement was based on painstaking analysis of both observed and simulated climate features, primarily associated with the warming over the twentieth century. The two aspects of greenhouse evaluation: detection and attribution, were carefully separated and analysed by the IPCC. Detection is the process of showing that an observed change is significantly different than can be explained by natural variability (a signal-to-noise problem). Clearly, the mere detection of a change in climate does not mean that its cause is known. Attribution of climate change, the determining of cause and effect, requires systematic and repeatable experimentation with the climate system.

The difficulty in attributing climate change once it is detected, to a single, or indeed multiple causes, results from the fact that the climate system cannot be tested with different experimental forcings until a match is found. However, climate models offer just this possibility. This is why the detection of climate change, demanding information about climate variability, provided in part by GCMs and the attribution of the detected change to greenhouse gas increases, which also requires multiple model simulations, are a demanding test for these models. Since 1994, global climate models have been used to:

- evaluate whether observed changes could be due solely to natural causes;
- weigh the evidence for climate change in the observations;
- generate simulations which share the characteristics of the observed changes and are a consistent response to the imposed natural and anthropogenic forcings.

Reconstructions of solar and volcanic forcings have been used in coupled ocean–atmosphere models to estimate the contribution of natural forcing to climate variability and change (e.g. Cubasch *et al*., 1997; Hegerl *et al*., 1997; Tett *et al*., 1999). The forcings, based mostly on proxy data, are uncertain, but including their effects produces an increase in variance at multidecadal time-scales. One consequence of this is to bring the low frequency variability closer to that deduced from palaeo-reconstructions (e.g. Rind *et al*., 1999). Assessments based on the results of climate models indicate that the recent global warming, which is now evident, is unlikely to be explained by natural forcing alone. (Lean and Rind, 1998). That the warming in the latter part of the twentieth century is unlikely to be explained by natural variability is also confirmed by statistical assessments. However, there is evidence that the climate is influenced to a discernible extent by volcanoes and, at least in the early part of the century, by solar variability (Crowley and Kim, 1999). Even if the magnitude of the response to these latter two forcings is underestimated by the models, the spatial and temporal patterns are such that they alone cannot explain the temperature changes in the twentieth century (Stott *et al*., 2001).

Sophisticated analysis techniques have been developed using GCM simulations in order to try to differentiate between proposed explanations of observed climate change. Confidence in the results follows from the use of different assumptions and many different model simulations (Santer *et al*., 1996; Tett *et al*., 1999). The conclusion is that recent changes in the mean global surface temperature cannot be simply explained by natural causes. In the majority of cases, climate model estimates of anthropogenic temperature changes are consistent with observed changes. Models differ in their estimates of the magnitudes of the main forcing factors and their relative contributions. Overall however, the evidence is that the forced response is dominated by the increases in greenhouse gases (Allen and Tett, 1999).

In summary, there are three periods of the Earth's climate history that are now being simulated specifically to permit evaluation of current climate models. These are the mid-Holocene, the LGM and the climate of the twentieth century. For the last of these, the models perform fairly well (Figures 3 and 11). The former two periods offer situations in which either solar irradiation or surface and atmospheric conditions (but not both at once) differ very significantly from those of today. For both periods, some climate models generate plausible simulations but full validation is constrained by known parameterization and initialization weaknesses and by the inadequacy and incompleteness of the palaeodata, especially for the case of the LGM.

6. CLIMATE MODELS FOR THE FUTURE

6.1. Global warming

The natural greenhouse effect maintains the Earth's climate at temperatures hospitable to life. Human activities have been recognized as contributing radiatively active trace gases to the atmosphere for over a century (Henderson-Sellers and Jones, 1990). The potential impacts of human-induced global warming prompted the WMO and the United Nations Environment Programme (UNEP) to establish the IPCC in 1988 (Taplin, 1996). Open to all member nations of the UNEP and WMO, the IPCC has a mandate to assess the scientific, technical and socio-economic information relevant for the understanding of the risk of human-induced climate change. It bases its assessment on published and peer reviewed scientific technical literature. Working Group I assesses the scientific aspects of the climate system and climate change. Working Group II addresses the vulnerability of socio-economic and natural systems to climate change, negative and positive consequences of climate change and options for adapting to it. Working Group III assesses options for limiting greenhouse gas emissions and otherwise mitigating climate change. The IPCC also includes a Task Force on National Greenhouse Gas Inventories which oversees the members' national efforts to account for and measure greenhouse gas sources and sinks. This policy framework co-exists with the tools: numerical climate models (Figure 10).

The IPCC's First Assessment Report published in 1990 states (Houghton *et al*., 1990, p. xiii):

> There is concern that human activities may be inadvertently changing the climate of the globe through the enhanced greenhouse effect, by past and continuing emissions of carbon dioxide and other gases which will cause the temperature of the Earth's surface to increase—popularly termed the 'global warming'. If this occurs, consequent changes may have a significant impact on society.

This assertion had a role in establishing the Intergovernmental Negotiating Committee for a UNFCCC by the UN General Assembly (Taplin, 1996). The UNFCCC provides the overall policy framework for addressing this aspect of climate change.

The IPCC Second Assessment Report was able to conclude that (Houghton *et al*., 1996, p. 5):

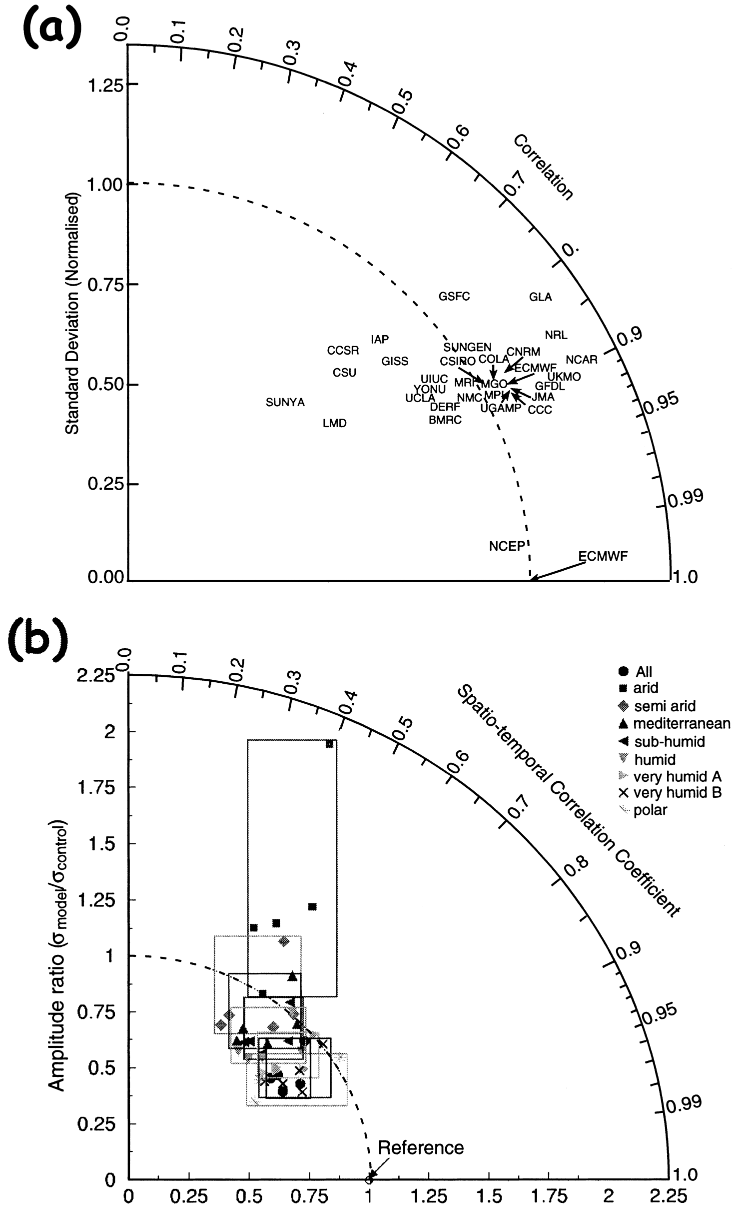

Figure 11. (a) Climate model 'evaluation' diagrams showing the total space-time pattern variability of the AMIP models' mean sea-level pressure, in terms of the standard deviation of the modelled monthly means (proportional to the distance from the origin), the rms difference between the simulated and observed monthly means (proportional to the distance from the reference point), and the correlation between the simulated and observed monthly means over the simulated period. The standard deviations and rms differences have been normalized by the observed standard deviation. (b) The position of the National Center for Environmental Prediction (NCEP) reanalysis relative to the reference ECMWF reanalysis is also indicated (from Gates *et al.*, 1999). (a) sea-level pressure simulated in the AMIP I experiment over the years 1979-1988 compared to a reference from ECMWF. (Note that the other reference NCEP is distinct from ECMWF). (b) Climate regimes calculated using the simulated continental surface climates of five AMIP II models over the years 1979–1996. (Note that arid climate regimes compare least well with the reanalysis results (the reference) while the polar regimes perform fairly well (after Irannejad *et al.*, 2000. Part (a) of this figure is reproduced from PCMDI Report No. 45, by Gates *et al.*, 1998. The illustration was creatd at LLNL for U.S. DOE, neither LLNL or the U.S. Government makes any warranty, express or implied, or assumes any legal liability or responsibility for the accuracy, completeness, or usefulness of any information, product or process disclosed in this figure. The U.S. Government's right to retain non-exclusive, royalty-free licence in and to any copyright covering this figure is acknowledged. Credit is given to the University of California, Lawerence Livermore National Laboratory, and the Department of Energy under whose auspices the work was performed.)

> Our ability to quantify the human influence on global climate is currently limited because the expected signal is still emerging from the noise of natural variability, and because there are uncertainties in key factors. These include the magnitude and patterns of long term natural variability and the time-evolving pattern of forcing by, and response to, changes in concentrations of greenhouse gases and aerosols, and land surface changes. Nevertheless, the balance of evidence suggests that there is a discernible human influence on global climate.

This key input to international negotiations led to the adoption of the Kyoto Protocol to the UNFCCC in 1997. The fact of human-induced greenhouse warming is now widely accepted. The IPCC has continued to provide scientific, technical and socio-economic advice to the world community and, in particular, to the 170-plus parties to the UNFCCC through its periodic assessment reports on the state of knowledge of causes of climate change, its potential impacts and options for response strategies (e.g. Houghton *et al*., 1990, 1992, 1996). The Third Assessment Report (preliminary findings published in January 2001) currently under preparation will offer a comprehensive and up-to-date (mid-2000) assessment of the policy-relevant scientific, technical and socio-economic dimensions of climate change. However, it is already well known that many aspects of modelling of global warming will remain unresolved by this review (e.g. Harvey, 2000b; O'Neil, 2000).

The IPCC process aims to determine the current level of confidence in understanding of the forcings and mechanisms of climate change and climate models are a significant component of the IPCC process. Models are used as tools to predict future changes and as tools to assess our understanding of the climate system. Since the first IPCC assessment report, the science of climate modelling has progressed considerably. The systematic evaluation of climate models has emerged as a major focus of modelling groups and significant advances have been made in the modelling of individual components of the climate system and their interactions (e.g. Gates *et al*., 1999, Figure 11). Many of these advances have been a direct response to the questions raised during the IPCC process (cf. Smith and Wigley, 2000a,b).

Through an exhaustive review process, the IPCC aims to provide assessments which discuss climate change on a global scale and represent international consensus of current understanding. The aim is to include only information which has been subjected to rigorous review, although this is balanced by a desire to include the latest information in order that the best possible assessment can be made. These two competing goals mean that the development of the IPCC documents is an extremely time-consuming process but ensure that the final result is a consensus statement of the state of current knowledge of the climate system (e.g. Santer and Wigley, 2000).

Corrective adjustments are not a new concept in numerical modelling but the acknowledgement by the IPCC in its Second Assessment Report (e.g Kattenberg *et al*., 1996) that the vast majority of the global climate models from which their predictions of future climate were derived were either 'flux adjusted', or unable to achieve or maintain the present day climate caused some sceptical responses in the mid 1990s (cf. Shackley *et al*., 1998). Since then, considerable efforts have been made to reduce or even remove the need for flux corrections. However, while there have been great improvements in the ability of GCMs to simulate large scale oceanic and atmospheric heat balances (e.g. Gordon *et al*., 2000), some studies have described differences in the climate change responses of flux adjusted and non-flux adjusted models (e.g. Gregory and Mitchell, 1997) and other effects, such as cooling by atmospheric aerosols, remain unagreed (Harvey, 2000b).

The inclusion of a negative radiative forcing effect owing to increasing atmospheric aerosols has a significant influence on the interpretation of greenhouse predictions from global climate models (e.g. Hansen *et al*., 1993; Andreae, 1995). However, the compensatory effect of aerosol cooling as compared to greenhouse warming is not an argument for 'business as usual' in global industrialization (Houghton *et al*., 1996). The much shorter atmospheric lifespan of atmospheric aerosols compared with the greenhouse gases means that the longer the industrially-induced compensation is allowed to persist, the greater is the rapid greenhouse-only temperature rise when the 'aerosol mask' is finally removed (Mitchell *et al*., 1995). Furthermore, while on a global mean basis, aerosol-induced cooling does offset greenhouse heating, invoking this as a global climate management policy would lead to an ever-widening difference in the climatic forcing between the Northern and the Southern Hemispheres. This is potentially even more disruptive to the climate system than a uniformly distributed greenhouse effect (e.g. Wigley, 1991; Giambelluca and Henderson-Sellers, 1996).

Predicted enhanced greenhouse warming is likely not only to modify the atmospheric circulation (e.g. McAvaney and Holland, 1995), but could also strengthen the thermohaline circulation in the Atlantic by increasing the rate of water vapour loss from the Atlantic Basin. If this were to occur, the effect could be to sustain the current warm conditions in the Atlantic. On the other hand, enhanced greenhouse warming coupled with changes in land-use (Henderson-Sellers, 1995) could increase the flow of freshwater into the northern Atlantic tending to decrease salinity in northern Atlantic surface waters which could weaken the oceanic conveyor belt. Continuing reduction in salinity might eventually halt the formation of North Atlantic Deep Water and could shut down the ocean conveyor. Maier-Reimer and Mikolwajewicz (1989) have demonstrated, using an oceanic GCM, that addition of excess freshwater to this region can terminate the model's thermohaline circulation on a time-scale of a few decades. This would plunge the countries around the North Atlantic into a climate about 5°C colder than the present day, and would have climatic consequences for other parts of the globe. In summary, we do not know whether human-induced global warming will cause a small climate hiccup or a massive climate catastrophe (cf. Figure 2).

6.2. Land-surface forcing and its effects

People are beginning to make regional-scale changes to the character of the Earth's surface; the most important of which are desertification, re- and de-forestation and urbanization. Desertification is a problem that affects millions of people (Verstraete and Schwartz, 1991). The sparse vegetation natural to arid and semi-arid areas can be easily removed or destroyed by the direct impact of human activity, such as overgrazing or poor agricultural practices, and as a result of relatively minor changes in the climate. Removal of vegetation and exposure of bare soil decreases soil water storage and capacity, increases runoff and increases the albedo (e.g. Charney, 1975; Mintz, 1984; Cunnington and Rowntree, 1986; Henderson-Sellers, 1996). Less moisture available at the surface means decreased latent heat flux leading to an increase in surface temperature. On the other hand, the increased albedo produces a net radiative loss. In climate model calculations, the latter effect appears to dominate in arid and semi-arid regions and the radiation deficit causes large scale subsidence. In this descending air, cloud and precipitation formation would be very difficult and aridity would tend to increase. This is therefore a positive feedback loop tending to further augment a detrimental human impact on climate (cf. Charney, 1975; Charney *et al*., 1975).

At present, around 30% of the land surface of the Earth is forested and about a third as much is cultivated. However, the amount of forest land, particularly in the tropics, is rapidly being reduced, while reforesting is prevalent in mid-latitudes. As a consequence, the surface characteristics of large areas are being greatly modified. Climate modellers have attempted to examine the climatic effects of forest planting and clearance: particularly noticeable when forests are replaced by cropland (e.g. O'Brien, 2000). One area which is undergoing deforestation is the Amazon Basin in South America (Henderson-Sellers *et al*., 1996). The important change in deforestation is in the surface hydrological characteristics since the evapotranspiration from a forested area can be many times greater than from adjacent open ground. Most climate model simulations of Amazonian deforestation show a reduction in moisture recycling (because of the lack of the moist forest canopy) which reduces precipitation markedly (Henderson-Sellers and McGuffie, 1998)). However, the available global model experiments do not agree on outcomes, i.e. whether an increase in surface temperature occurs or on reasons (cf. Table V). Indeed, one of the longest experiment sets undertaken on this subject found that albedo increases were the dominant cause of a reduction in rainfall (Lean and Rowntree, 1997). The largest impacts are the local and regional effects on the climate, which could exacerbate the effects of soil impoverishment and reduced biodiversity accompanying the deforestation. Recently, some modellers have detected impacts resulting from tropical deforestation propagating to areas distant from the sites of deforestation underlining that model simulations have the potential to uncover the full effect of forest removal (Zhang *et al*., 1996a,b). As indicated by the last entry in Table V, GCMs are now being used to try to forecast the synergistic effects of the combined anthropogenic forcings of greenhouse warming and tropical deforestation (Zhang *et al*., 2001).

Table V. Annual regional response of temperature, precipitation, evaporation and moisture convergence to Amazon tropical deforestation from various GCM studies from the first in 1984 to date

Study	Albedo change	Roughness change	ΔT (°C)	ΔP (mm)	ΔE (mm)	Moisture convergence change
Henderson-Sellers and Gornitz (1984)	0.11/0.19	N/A	0	−220	−164	+
Dickinson and Henderson-Sellers (1988)	0.12/0.19	2.00/0.05	+3.0	0	−200	+
Lean and Warrilow (1989)	0.136/0.188	0.79/0.04	+2.4	−490	−310	−
Nobre *et al.* (1991)	0.13/0.20	2.65/0.08	+2.5	−643	−496	−
Dickinson and Kennedy (1992)	0.12/0.19	2.00/0.05	+0.6	−511	−256	−
Mylne and Rowntree (1992)	0.135/0.200	no change	−0.1	−335	−176	−
Dirmeyer (1992)	+0.03	2.65/0.08	N/A	+33	−146	+
Lean and Rowntree (1992)	0.136/0.188	0.79/0.04	+2.1	−296	−201	−
Henderson-Sellers *et al.* (1993)	0.12/0.19	2.0/0.2	+0.6	−588	−232	−
Pitman *et al.* (1993)	0.12/0.19	2.00/0.05	+0.7	−603	−207	−
Manzi (1993)	0.13/0.20	2.00/0.06	+1.3	−15	−113	+
Polcher and Laval (1994a)	0.098/0.177	2.30/0.06	+3.8	+394	−985	−
Polcher and Laval (1994b)	0.135/0.216	2.30/0.06	−0.1	−186	−128	−
Sud *et al.* (1996)	0.092/0.142	2.65/0.08	+2.0	−540	−445	−
McGuffie *et al.* (1995)	0.12/0.19	2.0/0.2	+0.3	−437	−231	−
Manzi and Planton (1996)	0.13/0.20	2.00/0.06	−0.5	−146	−113	−
Zhang *et al.* (1996a)	0.12/0.19	2.0/0.2	+0.3	−402	−222	−
Lean and Rowntree (1997)	0.13/0.18	2.10/0.03	+2.3	−157	−296	+
Hahmann and Dickinson (1997)	0.12/0.19	2.00/0.05	+1.0	−363	−149	−
McGuffie *et al.* (1998)	0.15/0.21	1.1/0.1	+0.9	+445	+248	+
Zhang *et al.* (2001)	0.12/0.19	2.0/0.2	+0.3	−403	−221	−
(with greenhouse warming)	(0.12/0.19)	(2.0/0.2)	(+0.4)	(−424)	(−215)	(−)

NA, non-applicable.

As well as the differences among models and among the model experiments (Table V), intrinsic errors are known to exist in complex numerical climate models. The latter can be revealed by careful comparison with observations as in the case of simulations of incoming energy fluxes at the Earth's surface (e.g. Wild *et al*., 1997). Once identified, it is often possible to improve parameterizations so that these errors are reduced or removed. Changes to the land-surface, vegetation and soil state can also add aerosols to the atmosphere in the form of dust and particulates and it is sometimes the case that inclusion of additional components of the climate system may reduce the observed errors (cf. Figure 11(b)). For example, Garratt *et al*. (1998) found that inclusion of atmospheric aerosols in models tends to reduce the differences between computer-observed incoming surface fluxes.

6.3. What is the future of climate modelling?

An historical view of numerical climate models illustrates that calculations of climate states have been under development for at least the last 40 years (Table II and Figure 6). The developments which we see now in coupled models are the result of a long history of simpler studies which provided the basis for the components which are now included in EMICs and in GCMs. As our understanding of the climate system improves through observation and analysis, the use of these models and the computational resources devoted to climate modelling continue to increase (Figure 5). Thus, continuing sophistication and improved validation can be anticipated. One aspect of future climate models will be continuing synergy and tension between simpler (e.g. EMIC) and more complex (e.g. GCM) model types and their respective capabilities to deliver predictions of value for policy development (Figure 10).

Numerical climate modellers have yet to tackle some aspects of their science. For example, the climate system is currently modelled by systems of coupled, non-linear differential equations. Chaotic behaviour is the prime characteristic of all such systems. This results in unpredictable fluctuations at many time-scales and a tendency for the system to jump between highly disparate states. It is not yet known if chaos is the primary characteristic of the climate system but the Earth's climate has been documented as undergoing very rapid transitions on time-scales of decades to centuries (Peng, 1995 and Figure 2). There is no reason to believe that this characteristic will disappear in the future. Similarly, it is now well-established that the Earth's climatic history has included catastrophic events induced by the impacts of comets and asteroids (e.g. Rampino, 1995). A large body (~ 2 km in diameter) impacting on the Earth is estimated as having a 1 in 10000 chance of occurrence in the next 100 years. Catastrophic climatic shifts including very rapid cooling and a massive reduction in incident solar radiation at the surface will follow such an impact and will persist for, at the least, hundreds of years.

Numerical models of long-period climatic evolution indicate that, in the absence of human-induced climate warming, the Earth would tend to move into cooler climatic conditions culminating in a full glacial epoch (Figure 2). Quasi-oscillatory cooling would be expected with progressively colder episodes occurring around 5000, 23000 and 60000 years into the future. The culminating glaciation occurring 60000 years in the future is predicted as having a similar intensity to the LGM. Based on astronomical forcing alone, the Earth would not be expected to return to conditions similar to the current Holocene thermal optimum any earlier than 120000 years from now. One possible result of anthopogenerated global warming is that enhanced greenhouse warming will so greatly weaken the positive feedback mechanisms, which are believed to transform the relatively weak orbital forcing signal into global interglacial–glacial cycles, that the initiation of any future glaciations will be prevented indefinitely. The possibility of chaotic characteristics and the proposal that anthropogenic effects might shift the Earth's climate into a new state are topics that could benefit from probing by future climate models.

Human beings are curious: we seek to understand, and hence, to predict. Although we cannot, yet, predict future climates, we often behave as if we can. Policy, development, business, financial and even personal decisions are made every day around the world as if we knew what climates people will face in the future. While local-scale climatic dependencies may remain weak in many places, technology and engineering, international trade and aid, food and water resources are likely to become increasingly dependent on, and even an integral part of, the climate system (Figure 10). Human infrastructure and well-being are dependent upon the climate and so the desire to predict future climates is not driven solely by curiosity but by a need to plan for the possible future system states. So far, our predictive skills are rather poor.

Improved understanding by policy makers, and by those who vote them into government, of all aspects of the climate system is one way of increasing the chances of sustaining Earth's climate in an hospitable state. For example, while technology and the harnessing of natural resources appear to have decreased the need to predict the future climate of industrial, architectural or even agricultural and water resource developments, oil and gas pipelines laid across permafrost, airport operations, floods, droughts and air pollution incidents cost lives and revenue every year. International policies regarding the global climate have been successfully negotiated and some, for example, the Montréal Protocol, implemented while others such as the Kyoto Protocol of the FCCC, which calls for the reduction in emission of greenhouse gases, have yet to be ratified by many nations. Climate assessment concerns the nature of climatic changes and also the validity of climate models (Figure 11). While researchers can continue to debate the acceptability of conclusions drawn from climate models (Rodhe *et al.*, 2000; Santer and Wigley, 2000) and the reasons for preferences and funding support (Shackley *et al.*, 1998), there is still work to be done to improve numerical climate models.

This review has presented a rather personalized summary of the last 40 years of climate modelling. Models have not developed in clear-cut ways in direct response to needs. Rather, they have advanced, and sometimes retreated, when new observations, new ideas, increased computational power and the failures and successes in evaluation and intercomparison exercises have become known. There is no one 'right' climate model, or even one 'best' climate model type. All have the potential to add value if they are honestly evaluated and appropriately applied. Climate modelling has reached its fortieth birthday with some glory, but with great endeavours still to be achieved.

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