PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

The Bakerian Lecture, 1991 The Predictability of Weather and Climate

John Houghton

Phil. Trans. R. Soc. Lond. A 1991 **337**, 521-572
doi: 10.1098/rsta.1991.0136

The Bakerian Lecture, 1991
The predictability of weather and climate

By John Houghton
Meteorological Office, London Road, Bracknell, Berkshire RG12 2SZ, U.K.

Contents

There is large public and political interest in the predictability of weather and climate, in particular in the influence of human activities on the likely climate change during the next century. Numerical models are the main tools which enable the nonlinear processes involved in the dynamics and physics of the atmosphere and other components of the climate system to be integrated in an effective way. The performance of such models used for weather forecasting has continued to improve as more accurate data with better coverage has become available, as improved descriptions of the physics and dynamics have been incorporated and as computing capacity and speed have increased. Studies of the predictability with models suggest that with further improvements in data and models deterministic forecasting of detailed weather may ultimately have useful skill up to 2–3 weeks ahead.

Beyond the limit of deterministic forecasting, some skill remains for the forecasting of general weather patterns which can be pursued by studying ensembles of model forecasts from slightly varying initial conditions. The largest difficulty with further

Phil. Trans. R. Soc. Lond. A (1991) **337**, 521–572
Printed in Great Britain

improvements of numerical models lies in their inadequate treatment of the motions too small to be explicitly resolved.

Interactions between the atmosphere and the ocean are responsible for substantial variations on seasonal, interannual and longer timescales. Forecasts are being provided of seasonal precipitation in the Sahel region of Africa based on a knowledge of global sea surface temperature (SST) anomalies together with the assumption that such anomalies tend to persist from one season to the next. Attempts to forecast SST anomalies have centred on tropical regions in particular on the El Niño. Simple models show some skill in forecasting El Niño events 3–9 months in advance. Studies with more elaborate models which as yet only show partial success in simulating these events demonstrate the complex nature of the interactions involved.

Turning to the likely changes in climate next century: if no changes occur in the atmosphere other than the increase in CO_2 and other greenhouse gases due to human activities, the increase in radiative forcing due to a doubling of atmospheric CO_2 concentration would lead to an increase of about 1.2 °C in global average temperature. Water vapour and ice–albedo feedbacks raise this to a figure of about 2.5 °C (with an uncertainty range of 1.5–4.5 °C) as estimated by the Intergovernmental Panel for Climate Change. Such a change would dominate over forcing likely to arise from other factors, and this estimated rate of change next century is probably greater than any which has occurred on earth during the past 10000 years.

The main uncertainties in climate change predictions arise from the inadequacies of the models in their descriptions of cloud-radiation and ocean circulation feedbacks. Until there is more confidence in the treatment of these feedbacks there are bound to be large uncertainties associated with any predictions of regional climate change. To reduce the uncertainties there need to be improvements in computer power, in model formulation and in our understanding of climate processes together with a large programme of observations of climate parameters to provide early detection of climate change and to provide validation of climate models and to provide data for initialization of model integrations.

An important question is whether changes in climate due to changes in radiative forcing are predictable. It is pointed out that the response to climate over the past half million years to changes in forcing due to the variations in the Earth's orbit (Milankovitch cycles) is a regular one; some 60% of variations in the global temperature as established from the palaeontological record occur near frequencies of the Milankovitch cycles. We can, therefore, expect the changes in climate due to increasing greenhouse gases to be a largely predictable response. Large, but probably predictable, changes in the circulation of the deep ocean have modified climate change during past epochs and could have significant influence on future climate change.

1. Introduction

In a country like the United Kingdom the variability of the weather is a common topic of conversation. So too is the accuracy of weather forecasts. In the past year or two this interest has been even further encouraged because politicians and the general public have become aware of the possibility that the climate may be changing because of human activity. Human industry, in particular our insatiable demand for energy, is causing an increase in the concentrations of the so-called

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

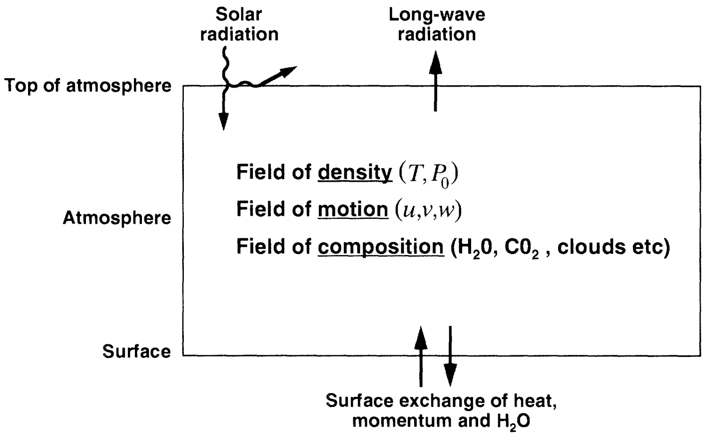


Figure 1. Schematic illustrating the parameters and physical processes involved in atmospheric models.

greenhouse gases in the atmosphere, which is likely to lead to substantial climate change. Because there is concern regarding the impact of climate change on the survivability and the future development of the human species the question is being asked as to the extent to which climate change as a result of human activity is predictable.

In this paper I review our current understanding of the predictability of weather and climate on timescales from a few days up to a century or so. For our purposes here it will be adequate to define weather as the detailed behaviour of the atmosphere on appropriate space and time scales (say 10^4 m, 10^3 s) expressed in terms of the distribution of variables such as temperature, pressure, wind and precipitation. Climate is defined as the average weather over much longer timescales and is expressed in terms of statistical quantities related to the above variables.

The large advances which have occurred during the past 20 years or so in both weather and climate forecasting have been possible because of the use of mathematical models. These models are essential tools for the study of the atmosphere and the climate because they are the only way we know of describing and properly taking into account the complex nonlinear processes involved in a system as large and complicated as that of the atmospheric circulation. Even more is this the case for the much more complex climate system. We are continually comparing real weather or climate situations with the simulations or predictions of models. The detailed behaviour of the models can readily be diagnosed and to the extent that the models accurately reproduce the real situations, an understanding of the processes occurring in the real atmosphere can be achieved. We begin therefore with a description of weather forecasting models and what is known of their predictability.

2. Weather forecasting models

We first take a look at the atmospheric models used for weather forecasting at the Meteorological Office (for a description of the dynamics and physics involved in numerical modelling see Houghton (1986), for a detailed description of the Meteorological Office forecast models see Bell & Dickinson (1987)). Within them the atmosphere's behaviour is represented by values of appropriate parameters at a grid of points. The models in operational use are one covering the globe with a grid

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

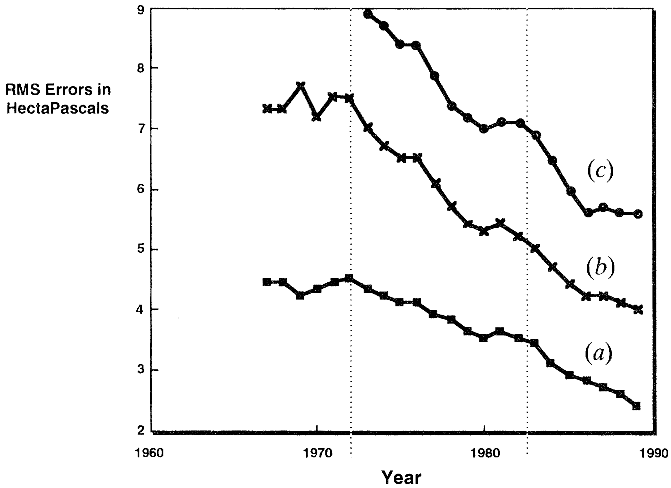


Figure 2. Errors (r.m.s. differences of forecasts of surface pressure compared with analyses) of Meteorological Office forecasting models since 1966 for (*a*) 24 h, (*b*) 48 h and (*c*) 72 h forecasts. Dotted lines show where major changes of model or computer have occurred.

spacing in the horizontal of about 90 km and a limited area model with a horizontal grid spacing of about 40 km. In both there are about 20 levels in the vertical.

The dynamics which is included in the model (figure 1) description consists of the horizontal momentum equations, the hydrostatic equation (vertical accelerations are neglected), and the continuity equation. The other physics in the model consists of the equation of state, the thermodynamic equation and parametric descriptions of moist processes (evaporation, condensation, formation and dispersal of clouds, etc.) of the radiative and convective processes within the atmosphere, and of the exchange of momentum, heat and water vapour with the underlying surface. Further allowance has to be made through suitable parametrizations of the motions which occur on scales smaller than the model's grid size. To generate a forecast from a given initial state, the equations are integrated forward in time to provide new descriptions of the atmospheric circulation and structure at times up to six or more days ahead. Because the model equations are nonlinear, integrations must be performed numerically by using electronic computers – hence the description 'numerical models'.

In the Meteorological Office the first model integrations for operational use were made in 1965. Although the skill of the early models was limited they provided useful guidance to the human forecaster. Their potential was rapidly developed and within 10 years the models were providing forecasts of the basic motion field with greater skill than could be achieved by an unaided human forecaster. The forecast skill of the models has continued to improve to an extent beyond any which was envisaged by those involved in the early model developments. As improvements have been made in the dynamical or physical descriptions in the model formulation, in the accuracy or coverage of the data used for initialization or in the resolution of the model the resulting forecast skill has increased. For instance, as can be seen from figure 2, three-day forecasts of surface pressure today are as skilful on average as two-day forecasts of five years ago. Many of these improvements have become possible through the continued development of computers of higher and higher speed and storage capacity (figure 3).

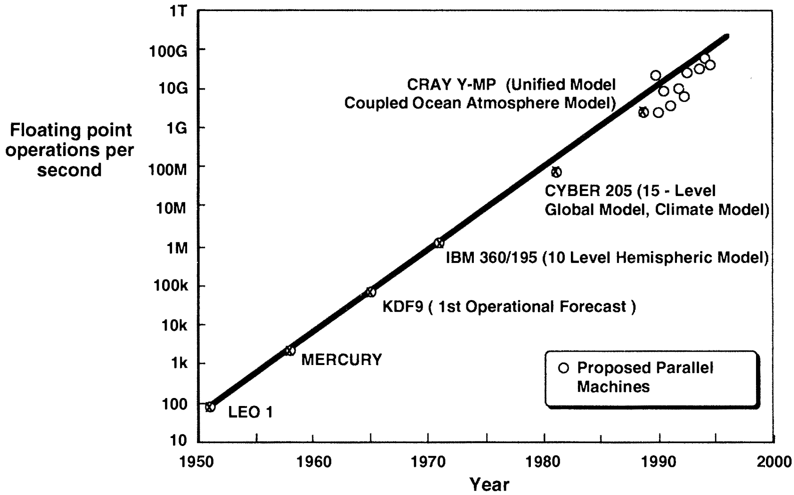


Figure 3. Computers used by the Meteorological Office for numerical weather prediction (NWP) Research and since 1965 for operational forecasting.

3. Data and data assimilation

The initial description of the atmosphere from which a model integration begins is generated from data coming from a wide range of sources (figure 4). Of particular importance to forecast accuracy is not only the volume and accuracy of these data used for initialization, but also the method used for the assimilation of those data into the model. The assimilation process is a gradual one which proceeds during a run of the model for a period preceding the analysis time (Atkins & Woodage 1985). This assimilation run starts from the forecast situation predicted for the appropriate time from the model's previous run. By the gradual introduction of new observations, it is ensured that the process is computationally stable; it is also ensured that useful information contained in the previous model run is retained. Since previous model runs include information from observations made at earlier times, the process of running model forecasts interspersed with the assimilation of observations means that at any one time the model has essentially integrated information from all the observations over the period during which the model has forecast skill, i.e. the period of about the previous five days.

A good illustration of the importance of adequate data coverage and of a good assimilation scheme is provided by the Meteorological Office forecasts for the storm which hit Southeast England in the early morning of Friday, 16 October 1987. During this storm a number of stations in Southeast England recorded record gusts of over 90 knots (167 km h^{-1}), and approximately 15 million trees were blown over in the area. Although forecasts on the previous Sunday had given good early warning of a storm of unusual severity, the model forecasts available during the day of the 15 October gave much poorer guidance than earlier forecasts and failed to predict the intensity or the correct track of the storm. The question was raised at the time as to whether the numerical models are capable of the accurate prediction of such an exceptional event. Extensive studies have been carried out (Meteorological Office (1987) and Shutts (1990)) of the effect on the forecasts of different mixes of initial data and different methods of assimilation. Because of the very rapid development of the storm it was an occasion when the model predictions were particularly

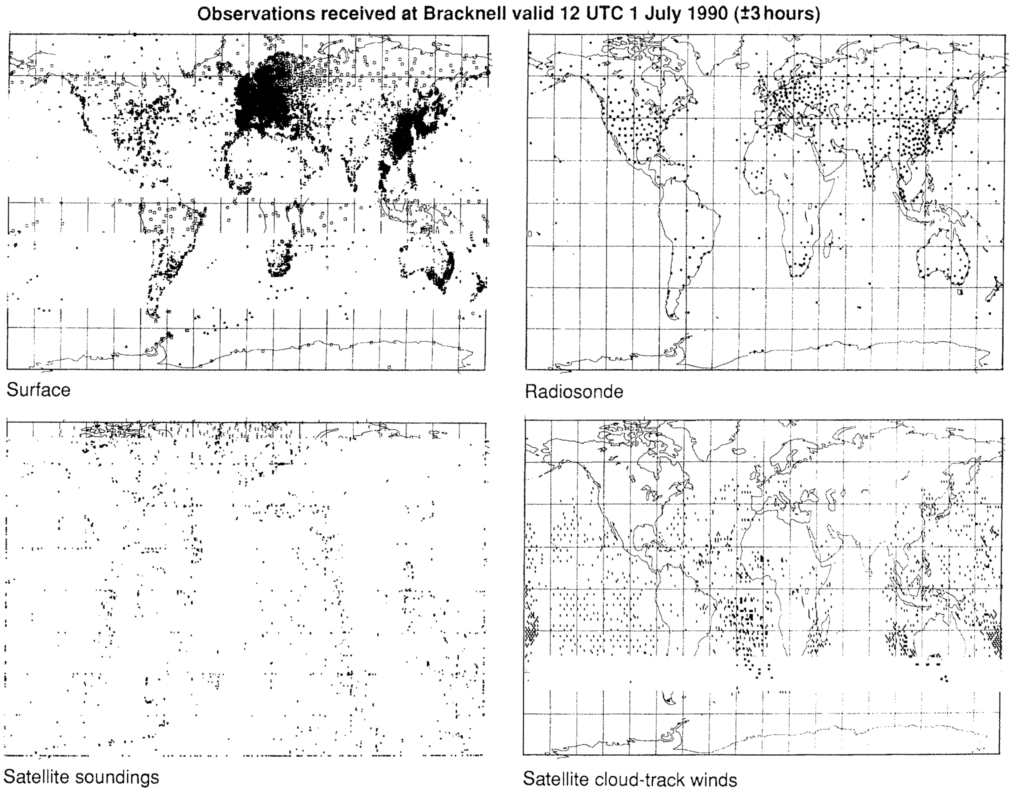


Figure 4. Some of the sources of data for input into the Meteorological Office weather forecasting numerical model on a typical day.

sensitive to the accuracy of the analyses and to the detailed data assimilation scheme. The studies demonstrate that if all the data available for that day, especially data from aircraft reports, are included (some data was not available early enough to be included in the operational runs at the time) and if an improved method of data assimilation is used, a good forecast of the storm's development and progress can be achieved (figure 5).

4. The atmosphere as a chaotic system

Before describing work on the predictability of complex atmospheric models, it will be helpful to introduce the idea of 'chaos' and the way in which the study of relatively simple atmospheric models has assisted in the development of our understanding of the behaviour of so-called chaotic systems over the past 30 years. It was a meteorologist, E. Lorenz, who was one of the pioneers of the modern study of chaos; last year he was elected to Foreign Membership of the Royal Society in recognition of his contributions to this field. In a classic paper, Lorenz (1963) considered the variation in the behaviour of a simple three-parameter model – the simplest numerical model he could formulate to describe convection in the atmosphere – with variations in the input conditions from which the model integrations began. He thereby developed a description of what has become known as the Lorenz attractor, which describes the evolution of values of the three variables

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

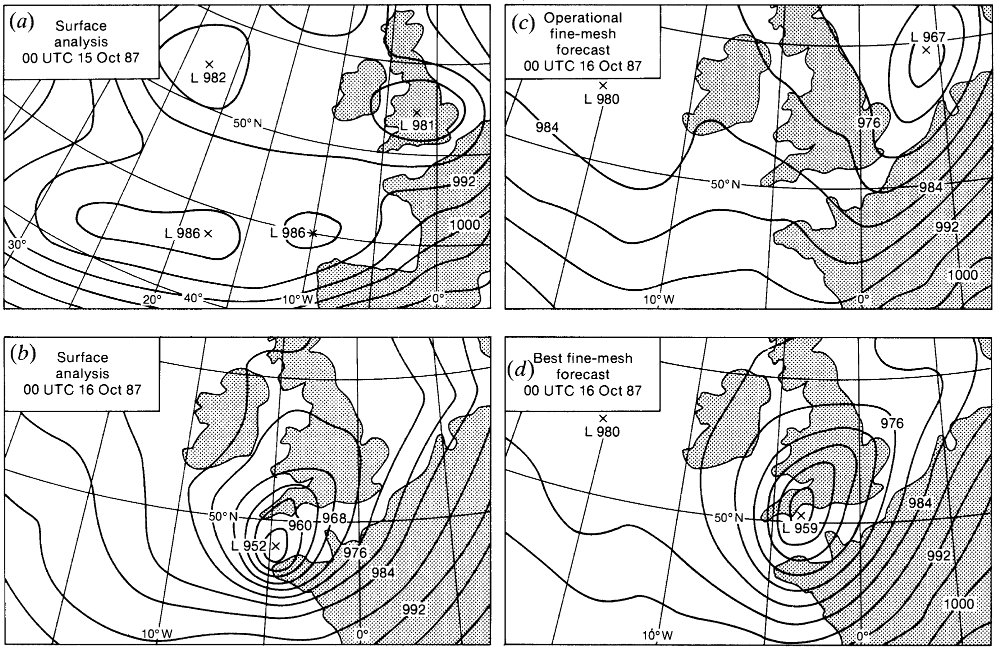


Figure 5. Analyses (*a*, *b*) of surface pressure in millibars (1 mbar = 100 Pa) and forecasts (*c*, *d*) for the storm which passed over southern England on 15/16 October 1987. (*c*) The operational fine-mesh forecast at the time and (*d*) the best fine-mesh forecast produced after the event using better assimilation procedures and more complete data (from Meteorological Office 1987).

from different starting conditions. It shows the regions of phase space in which the system's behaviour is most likely to lie and also shows how solutions from very close starting conditions can diverge into very different regions. For instance, for the Lorenz attractor, we can identify evolving trajectories that remain close together for a long time, other trajectories that diverge initially but end up in the same overall region of the phase space. Yet other neighbouring trajectories diverge and end up in the very different regions of phase space (figure 6). It is this critical dependence of the system's behaviour on the precise starting conditions which is known as chaotic behaviour. It is sometimes described colourfully as the 'butterfly effect', taken from the title of a lecture given by Lorenz in 1979: 'Predictability: does the flap of a butterfly's wings in Brazil set off a tornado in Texas?'.

One condition for a system to be chaotic is that it be nonlinear. Read (1991) points out that for useful studies of the behaviour of a chaotic system to be possible a further crucial property of such a system is that it should possess few degrees of freedom or that for it there should exist an attractor of low dimension governing the system's dynamical behaviour which would indicate an underlying simplicity which belies apparent complexity. Now the atmosphere undoubtedly possesses a very large number of degrees of freedom and its behaviour is in principle extremely complex. A measure of the complexity of an aspect of a system such as the atmosphere is the dimension of the attractor which describes it. This can in principle be determined from a study of long time series of data but, as Read (1991) points out, data series for atmospheric parameters are neither sufficiently long nor sufficiently detailed for good estimates of the attractor dimension to be made. However, studies of the behaviour of models and of the atmosphere itself show that aspects of the motion of

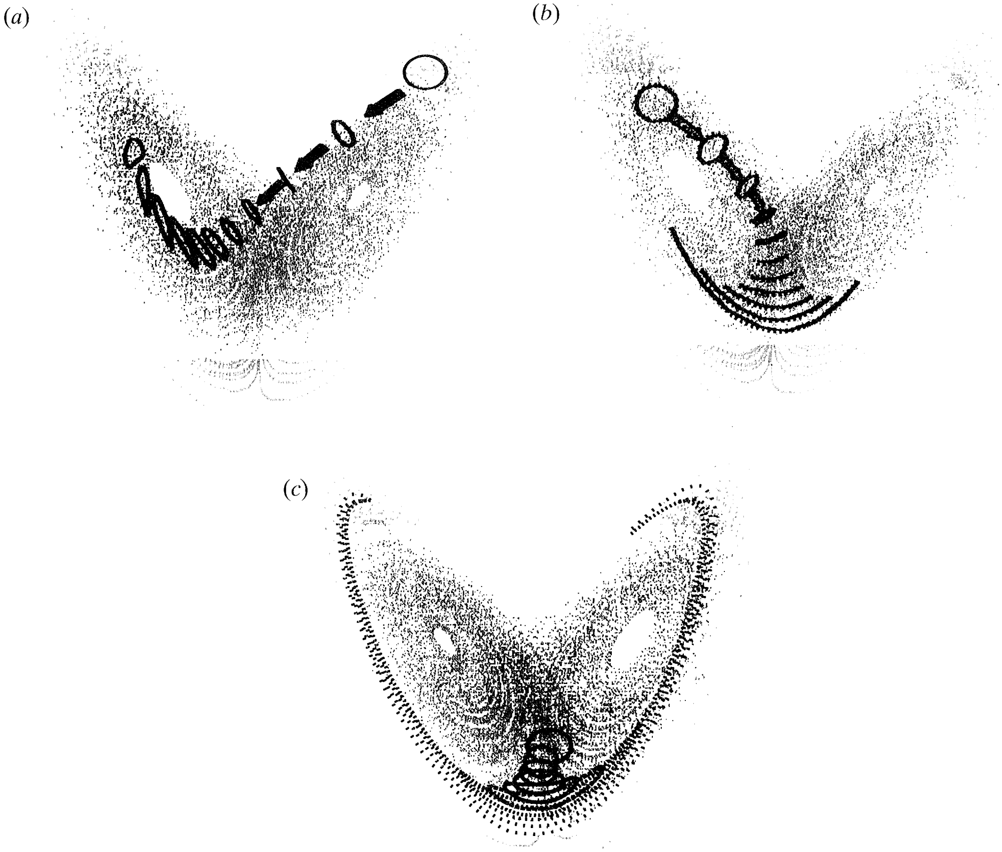


Figure 6. Cross sections across the Lorenz attractor showing the development with time of solutions to the equations beginning from different ensembles of initial states. Parts (a)–(c) illustrate increasing divergence of solutions (after T. N. Palmer, personal communication).

the atmosphere on the large scale appear to exhibit that required degree of simplicity for useful studies of the behaviour of the atmosphere as a chaotic system to be made.

It is clear from atmospheric studies that the atmosphere is not equally sensitive to all small perturbations nor are all parts of the atmosphere equally affected by perturbations. What we need to know in any consideration of predictability are the conditions under which particular chaotic character is present and the degree of sensitivity to initial conditions.

In particular, for our consideration here, we want to know how forecasts of detailed weather depend on the initial conditions and what limit that places for a given region of the globe, on deterministic forecasting. We also wish to know what sort of predictability remains for longer periods. For instance, although it may not be possible to forecast detailed weather, e.g. the timing of particular events, can forecasts of the general character of the weather be given for a month or so ahead?

5. Model predictability

To investigate the sensitivity of the evolution of the atmosphere's circulation to any given initial state, we can first investigate the sensitivity of the behaviour of an atmospheric model to its initial conditions. Such experiments have been carried out

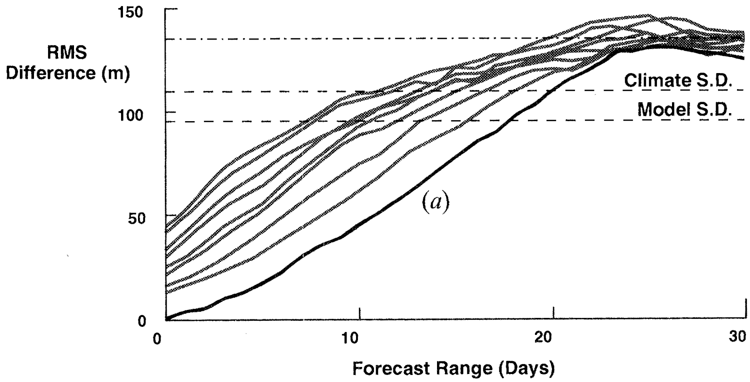


Figure 7. Upper bound on predictability. Curves show the r.m.s. differences between Meteorological Office model forecasts of the Northern Hemisphere 500 hPa height field separated by 6, 12, 18, ..., 48 h. Curve (*a*) is an estimate of the growth of the r.m.s. differences for infinitesimal initial data error. Also plotted are the standard deviations of model fields (model s.d.) and of real analyses of the 500 hPa height field (climate s.d.) and also the saturation level determined from the average r.m.s. difference between random forecasts (after Milton, personal communication).

from the early days of numerical models (Charney 1966; Smagorinsky 1969; Lorenz 1982; Chen 1989). A series of experiments has recently been carried out by Milton (personal communication) in which he studied a set of 27 integrations over 30 days during the winters (DJF) of 1985–86 to 1988–89 with a Meteorological Office global model (for these integrations the horizontal resolution was about 200 km and there were 15 levels in the vertical). Consider two integrations of the model from analyses of the atmospheric state 6 h apart. If we make the assumption that the later analysis is the true atmospheric state at its initial time, then the six-hour forecast from the earlier analysis can be considered as this true state plus a small error. A comparison of the integrations from these two initial states provides information on the growth of this error with time. Similar comparisons can be made between runs 12 h, 24 h up to 48 h apart to find out about the growth of larger initial error fields. As the forecast period increases the r.m.s. difference (error) between any pair of runs increases until it reaches a saturation value which is a typical r.m.s. difference between any pair of model forecasts chosen at random. This represents the theoretical limit of predictability where error growth arises solely from model internal dynamics and error growth due to model imperfections is ignored. In figure 7 are shown curves for the growth of error of the geopotential height of the 500 hPa surface averaged over the 27 model runs. Since there appears to be approximately a linear relationship between the initial error and the limit of predictability, extrapolation can be made to give a limit of predictability appropriate to infinitesimal initial error. This is also shown in figure 7; for this particular model it turns out to be about 26 days.

It is now interesting to compare the model atmosphere against the real atmosphere by comparing atmospheric analyses with model forecasts. Such comparisons for the same 27 cases are shown in figure 8. If the limit of useful forecasts (Chen 1989; Murphy 1990) is defined as that when the r.m.s. error in the forecast is equal to the standard deviation of model analyses (just under 100 m) then current forecasts are useful at about six days. If we assume that the limit of predictability of the model is a useful guide to the upper limit of the predictability of the atmosphere then, following Chen (1989), the other curves in figure 8 suggest that the existence of a perfect model would improve the range of useful forecasts from six to about 12 days;

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

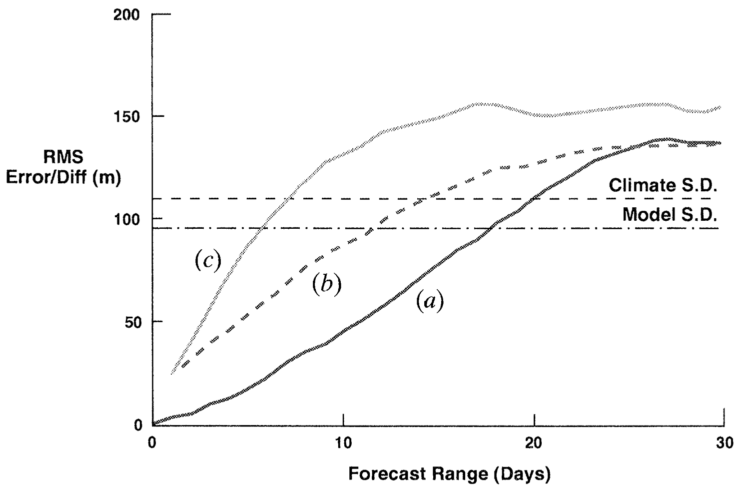


Figure 8. Potential skill improvements. Curve (*c*) is the r.m.s. error of forecasts of the 500 hPa height field from a global model of the Meteorological Office. Curve (*a*) is the same as curve (*a*) of figure 7. Curve (*b*) is an estimate (using the data of figure 7) of the growth of r.m.s. difference for a day 1 r.m.s. difference equal to the day 1 r.m.s. error (after Milton, personal communication).

the provision of near-perfect data would move the range out to about 18 days. The ultimate range of useful deterministic forecasts appears to be therefore between 2 and 3 weeks, a figure in good agreement with estimates made in a similar way using other models (Lorenz 1982; Chen 1989).

6. Improvements in initial data

The patterns of atmospheric circulation on the largest scale determine developments on much smaller scales so that for good forecasts more than one day ahead over the British Isles, data coverage is required over the whole hemisphere and, for more than two or three days ahead, over the whole globe. Better coverage of the data sparse areas, such as the oceans, is particularly important.

We have mentioned the importance of data coverage over the Atlantic Ocean if reliable forecasts for the British Isles and Western Europe of exceptional events such as the October 1987 storm are to be achieved. However, for European forecasts the acquisition of better observations over the Pacific is of even greater importance. This is because disturbances that are well observed over North America generally take about three days to cover the North Atlantic during which time their behaviour is reasonably well forecast by the models. The models in their forecasts for Europe are able to make good use of the observations over North America taken three days or more earlier. The Pacific Ocean, however, is wider than the Atlantic; after the six days which would be a typical transit time for a disturbance, the model forecast description of the disturbance is much poorer than after three days. Similar considerations emphasize the importance of improved data coverage in the Southern Hemisphere where vastly improved coverage during the past 10–15 years have been provided from satellite observations.

Satellite observations, particularly those of atmospheric temperature structure, are still, however, not of the accuracy which is really required nor are they in their operational use realizing their full potential in terms of accuracy. A lot of effort is

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

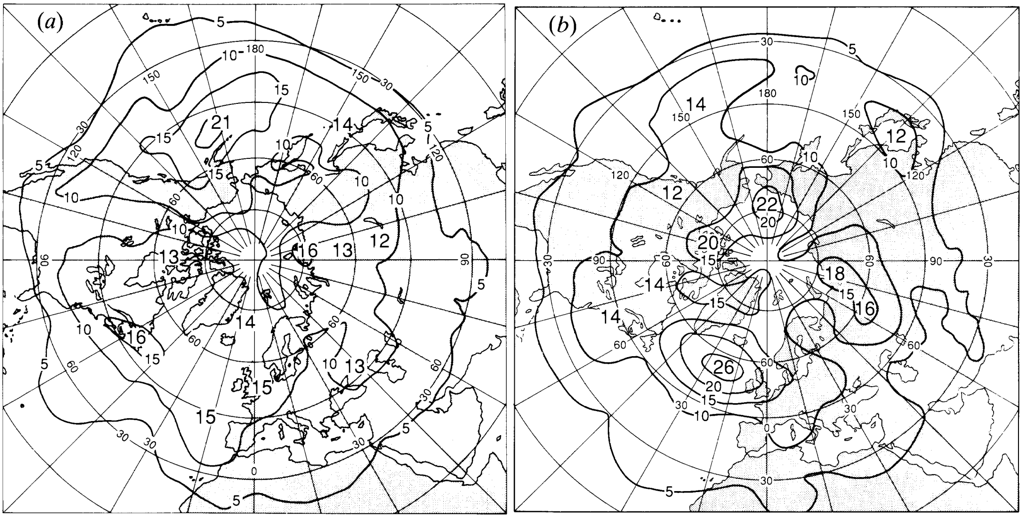


Figure 9. Root mean square variability in daily values of 500 hPa height in dm averaged over (*a*) January 1988, (*b*) February 1988.

currently being put into the improvement of schemes for satellite data retrieval and quality control. Further significant improvement should occur around 1993, when improved instrumentation for microwave observations of the atmosphere's temperature and humidity structure, the Advanced Microwave Sounding Unit (AMSU) will be available on the *Tiros* polar orbiting satellites. Later on for the polar platforms being planned to be launched at the end of the 1990s and early next century plans are underway for improved infrared soundings together with wind observations throughout the troposphere (except in regions obscured by cloud) from measurements of the Doppler shift of lidar signals scattered back to the satellite from the atmosphere. These improvements can be expected to increase the range of useful forecasts at mid-latitudes by a significant fraction of the six days which appears possible according to our discussion in the last section.

7. Variations in forecast skill

The results shown in figure 7 and 8 are averages over the winter Northern Hemisphere north of latitude 30. But individual forecasts vary a great deal in their skill. For instance, because the large-scale flow is predictable over a longer period than the small scale, forecast skill at mid latitudes is highest during winter when the large-scale quasi-stationary waves are of high amplitude and lowest during the summer when smaller-scale flows prevail (Dalcher & Kalnay 1987). That some regions over particular periods show much more variability than others is easily illustrated by comparing the variability over two successive months for different regions of the Northern Hemisphere (figure 9). Some of this difference in variability arises because of large variations in space and time of the incidence of large quasi-stationary waves known as 'blocking' situations associated with persistent areas of high pressure. Considerable attention has been given to the forecasting of the onset and the ending of such situations (Hollingsworth 1987; Tibaldi & Molteni 1991) and in recent years there have been some notable successes and failures in such forecasts. (Miyakoda *et al.* 1983; Sirutis & Miyakoda 1990).

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

Since some situations lead to more rapid error growth than others, the question has been asked as to whether regions of rapid error growth can be identified. Palmer (1988) compared the skill of medium range forecasts over the Northern Hemisphere and showed that their skill is strongly correlated with the sign and amplitude of the Pacific North America (PNA) upper air height anomaly pattern present in the initial conditions. In his study occasions with negative PNA index (positive height anomaly over the Aleutian Islands) showed less skill than occasions with positive PNA index. Using a simple barotropic model Palmer showed that this behaviour matched the different rates of growth of perturbations to basic states having opposite signs of PNA Index. Other such situations have been identified by O'Lenic & Livezey (1988) and by Palmer *et al.* (1990).

Further variations in forecast skill arise because of the influence of tropical disturbances at mid-latitudes. Ferranti *et al.* (1990) have studied the impact of the tropical flow field on mid-latitude dynamics especially during periods when the 30–60-day oscillation which occurs particularly at tropical latitudes was active. They showed that improvements of forecast skill at mid-latitudes could be achieved if a better description of the tropical circulation could be included in the model.

By studying these various influences on forecast skill, the possibility arises of being able to add to a forecast an estimate of its likely skill (Palmer & Tibaldi 1988; Molteni *et al.* 1991). To add an estimate of skill to a forecast would make it much more valuable especially for longer range forecasts where the average skill is small and where some forecasts show virtually no skill at all.

8. Ensemble forecasting

We saw in §5 that although the limit of useful deterministic forecasts is currently about six days, on average some skill is present in model forecasts at somewhat longer ranges. We have also seen that the large-scale features of the atmosphere possess more predictability than the small scale. This means that although detailed weather patterns (e.g. the exact timing and location of systems such as depressions) cannot be forecast for more than say six days in advance, some skill in the prediction of the general pattern of weather (e.g. cyclonic, anticyclonic, blocked, etc.) might be present for longer periods.

To explore the dependence of such extended range forecasts (10–30 days) on the initial conditions, and to achieve some improvement in forecast skill at this range, the method of ensemble forecasting has been used. An ensemble of forecasts is run from a cluster of initial states, generated by adding a series of perturbations to the analysed initial states consistent with observation and analysis errors. To avoid problems early in the integrations it is also necessary that the initial states be balanced and dynamically consistent. In practice the simplest way to ensure this, and also to minimize the use of computer time, is to use as the ensemble, forecasts from a number of consecutive analyses each separated by a constant time interval (e.g. 6 h as for the data from which figure 7 was generated) with the latest corresponding to the start of the forecast period. The forecast from the mean of such an ensemble shows a significant improvement in skill compared with the individual forecasts from members of the ensemble. For a perfect model the theoretical improvement in skill is a factor of two (Leith 1974; Murphy 1988).

It might be expected that the spread of the forecasts (figure 10) in an ensemble would provide a good indication of the skill of that particular forecast; for instance

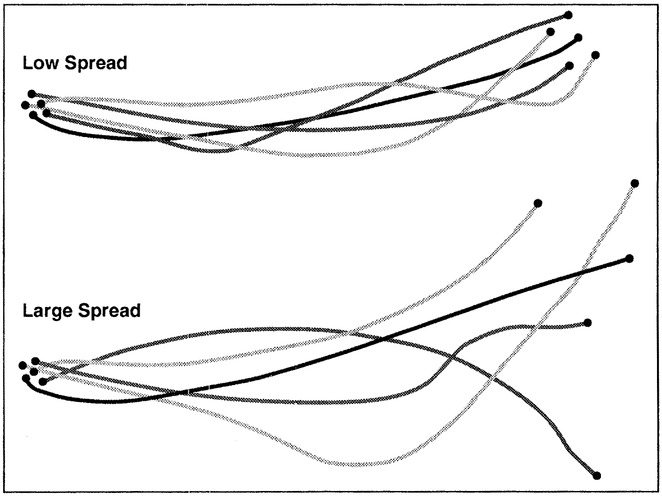


Figure 10. Schematic projections of ensembles of long-range forecasts. From close initial states, some ensemble forecasts after 30 days show low overall spread, whereas others show large spread.

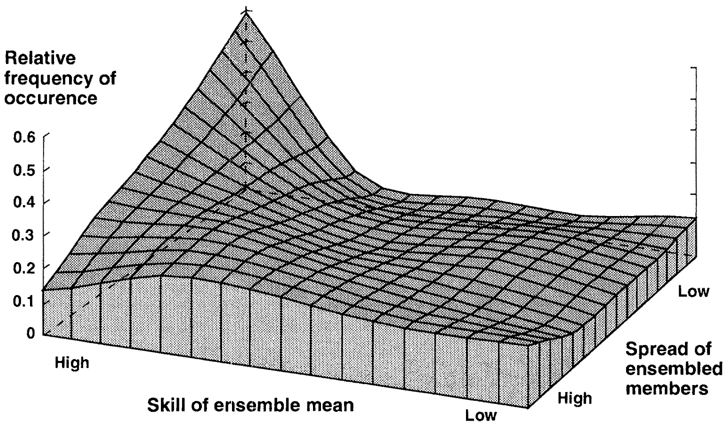


Figure 11. Relative frequency of occurrence of ensemble forecasts with different spread between ensemble members and with different skill (from Meteorological Office Annual report 1988, p. 34).

if all the forecasts fall close together, a forecast of high skill might be expected. There is some correlation between spread and skill (Murphy 1990; Brankovic *et al.* 1990). Forecasts having low spread (high agreement) are more likely to be skilful than unskilful whereas with those with high spread (low agreement) there is no discrimination between skilful and unskilful cases (Murphy 1990 and figure 11). Similar results were found by Kistler *et al.* (1988) at the National Meteorological Centre in the U.S.A. Ensemble techniques are already providing on a regular basis at the Meteorological Office forecasts of time-averaged weather in the time range between 1 and 4 weeks into the future. Although these forecasts do not possess high skill they are useful in a wide range of applications especially in the water and energy industries in which planning up to a month ahead is dependent on estimates of many factors one of which is the weather. As models improve and are able to simulate more accurately the range of atmospheric flow régimes and transitions between them (e.g. zonal to blocked flow) then it is expected that ensembles will provide a range of

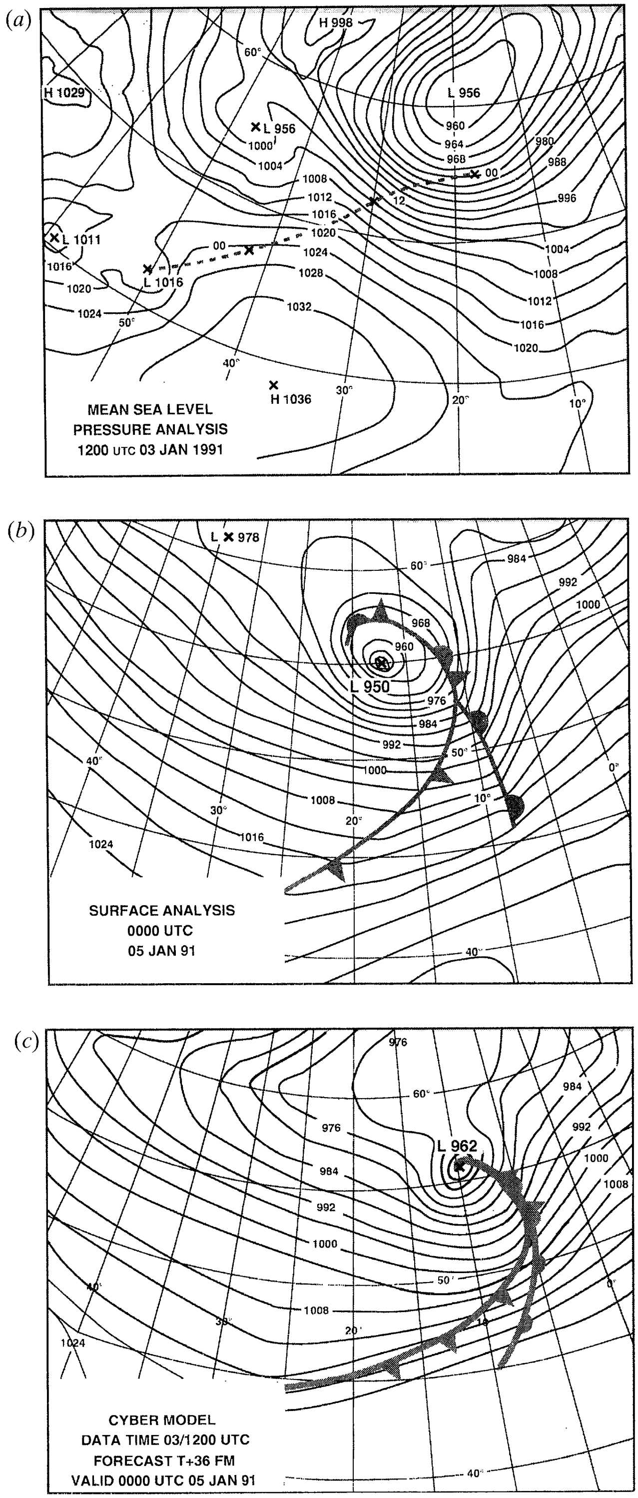


Figure 12*a*–*c*. For description see opposite.

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

(*d*)

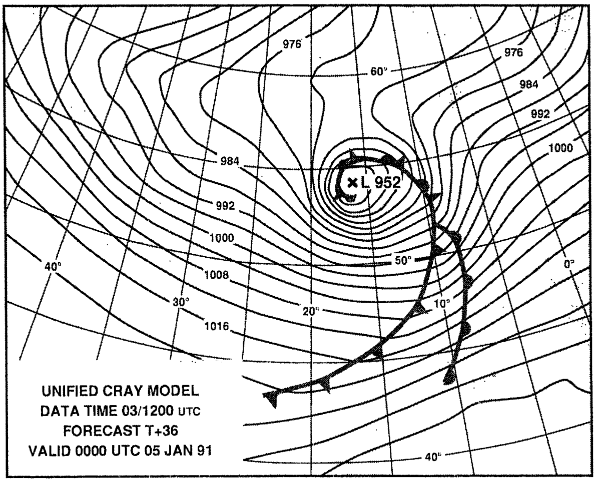


Figure 12. Illustrating improvement in 36 h forecast made on 3 January 1991 resulting from increased model resolution. (*a*), (*b*) Surface pressure analyses for initial and verification times. (*c*) Forecast with the Cyber model (150 km resolution and 15 levels) and (*d*) with the Cray model (90 km resolution and 20 levels).

solutions each with an associated probability of occurrence. This is the most natural framework in which to express the uncertainty in the forecasts beyond the deterministic limit.

9. Model developments

We have mentioned the importance of improvements in the accuracy and coverage of data. Improvement in the models is also important. If, for instance, the variability in the model is less than the observed variability, our estimates of predictability based on model results may be over optimistic. Substantial progress has been made over the past few years in the removal of systematic errors from the models (Palmer *et al.* 1990). One further improvement which is clearly possible is that, as higher performance computers become available, the model resolution can be increased. The impact of resolution on model performance in the medium range has been studied by Tibaldi *et al.* (1990). They show a very large improvement in forecasts for the 5–20 day period when the effective resolution is increased from about 500 km grid spacing in the model to about 250 km but a smaller gain on increasing the effective resolution from about 250 km to about 100 km. The Meteorological Office global model has recently been increased in resolution from a grid spacing of about 150 km to about 90 km and from 15 levels in the vertical to 20. A number of examples of improved forecasts attributable to the increased resolution have been identified; one example for a three-day forecast is shown in figure 12.

The parametrization schemes in the models are also continually being improved. A good example of the benefit of an improved description of physical processes is that of the better representation of surface drag in the models which was introduced in 1986 (Palmer *et al.* 1986) and which includes a description of the generation of gravity waves due to variations in topography and of the transport of momentum by those gravity waves to higher levels in the atmosphere. The introduction of this gravity wave scheme reduced the tendency of the model to develop excessive westerly flow at mid-latitudes. More recently in the new model just being introduced

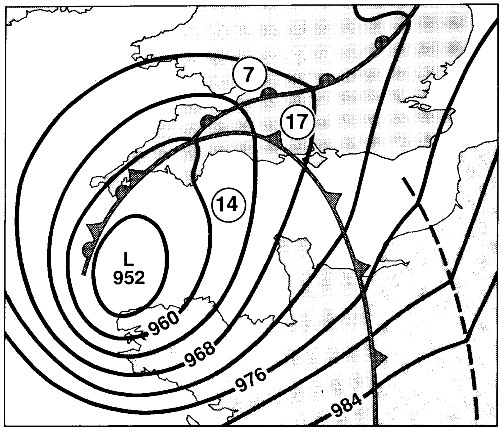


Figure 13. Surface pressure synoptic situation at 0000Z 16 October 1987 as an intense storm approached the U.K. Figures in circles are observations of surface temperature; they show the very large gradient across the warm front.

on the new Cray YMP computer at the Meteorological Office, substantially better parametrization of radiation, of cloud (including cloud water and ice and cloud optical properties as variables), and of the boundary layer are included, benefits from which are already being realized.

A basic problem, however, remains with models of the kind we have described which is concerned with the simple way in which the sub-grid scale motions are parametrized. This is done by assuming that the transport of momentum heat and water vapour by these small scale motions follows fickian diffusion (i.e. diffusion down the gradient proportional to its magnitude) with diffusion coefficients chosen to give the best results; different values may be chosen in the horizontal and vertical, they may also vary with position and season. Further, in order to ensure numerical stability for the integrations the value of coefficients which need to be used are typically significantly larger than are physically realistic. The question arises as to the influence of this parametrization on the performance of the models. In particular, since the dissipation of energy and momentum through smaller-scale motions is crucial to longer-term predictability, we want to know whether the estimates we have made of predictability are significantly affected by the inadequacy of the parametrization of small-scale processes.

The inadequacy of this simple representation is illustrated by the inability of the model to create or maintain sharp discontinuities in the way that is common in the real atmosphere. Warm and cold fronts associated with mid-latitude depressions are examples of such discontinuities (figure 13). Another example is the sharp boundary which exists at the edge of the polar vortex in the Southern Hemisphere in the winter; air in the polar vortex does not mix significantly during the winter months with air from other parts so creating the conditions necessary for the development of the so-called ozone hole in the spring (McIntyre 1990). In both these cases the model descriptions would tend to blur the boundaries and would generate unreal amounts of diffusion across them.

Further understanding of this problem can come from experiments with very different model formulations. The kinematics in the model we have described is set up in a eulerian framework. Cullen and his co-workers at the Meteorological Office

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

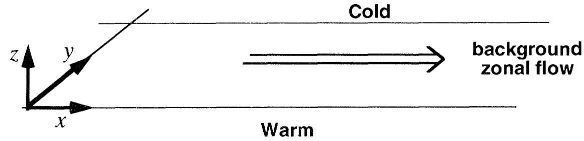


Figure 14. Formulation of Eady problem. Baroclinic instability and the development of baroclinic waves occur in the presence of a horizontal temperature gradient.

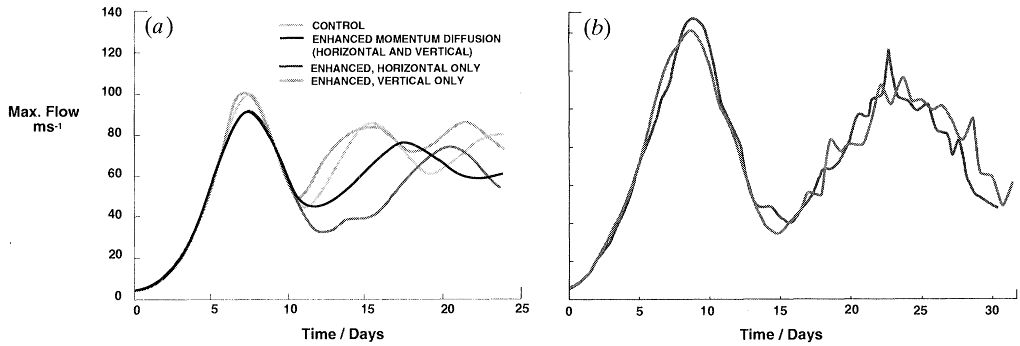


Figure 15. Growth and decay of the maximum flow in baroclinic waves as described (*a*) by a eulerian primitive equation grid point model (from Nakamura & Held 1988) with various diffusion coefficients, (*b*) by a lagrangian semi-geostrophic model (from Cullen & Roulstone 1991); the two curves show the results of two different model resolutions.

(Cullen 1983) have been experimenting with models whose kinematics is lagrangian (i.e. the coordinate system follows the flow as opposed to eulerian in which the coordinate system is fixed with respect to the Earth's surface). The models are constructed with semi-geostrophic dynamics (Hoskins 1982) in which an approximation is made regarding the balanced nature of the motion. They have been particularly used to model strong discontinuities such as frontal regions to which conventional models are not suited. Further, since they do not include explicit representation of diffusion in the equations of motion, a comparison of results from them with those from primitive equation models could shed some light on the sensitivity to the representation of small-scale processes.

Because of the very different formulations of the model, comparisons for the same situation are not easy to carry out. However, Cullen & Roulstone (1991) have applied their model to a study of the life-cycle of an unstable Eady wave (figure 14) and compared their results with those of Nakamura & Held (1989) who applied a two-dimensional eulerian primitive equation model to the same problem. In both cases integrations were carried out through two cycles of growth and decay of the waves. The results show (figure 15) that the development of the wave in Nakamura & Held's model, after the first growth phase, is strongly dependent on the values of diffusion coefficients used whereas in Cullen & Roulstone's model the development of the wave is comparative unaffected by changes in the model's resolution.

Such work is at an early stage, but there is at least a suggestion that because of the substantial degree of organization in small-scale atmospheric motions which is not described by grid-point primitive equation models, there may be less predictability present in the models than in the real atmosphere.

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

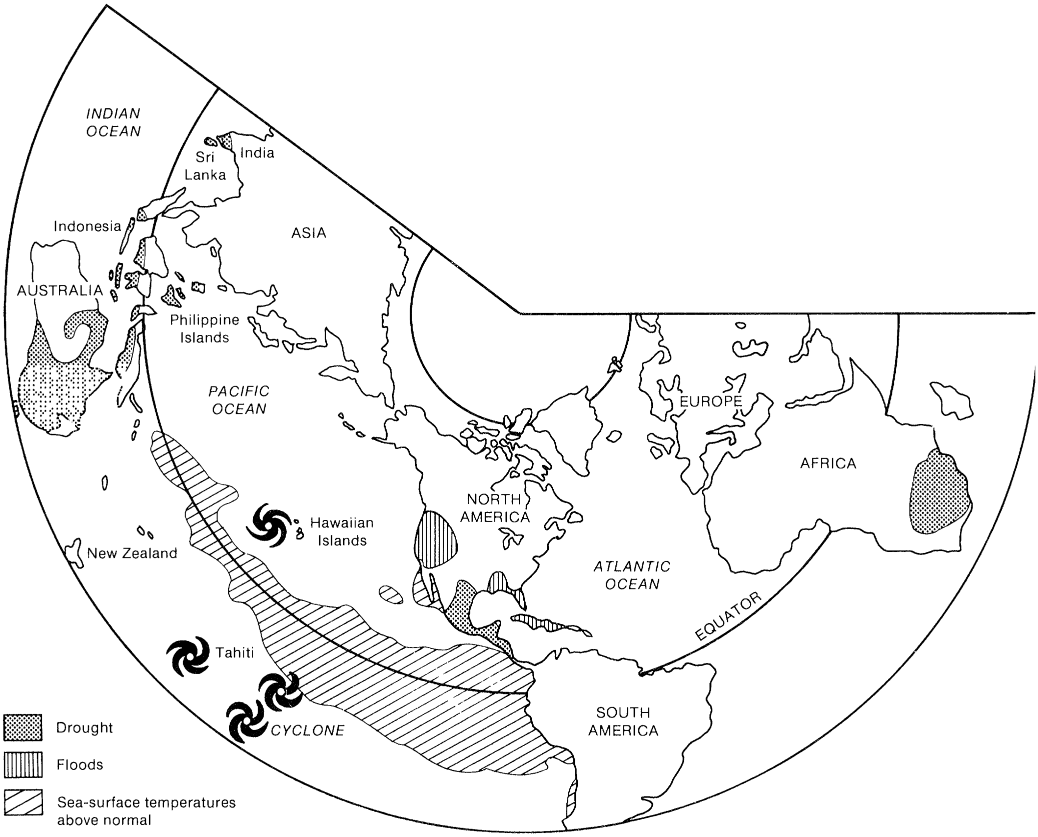


Figure 16. Regions where droughts and floods occurred associated with the 1982–83 El Niño (after Canby 1984).

10. Influence of the ocean boundary

So far we have been considering the performance of an atmospheric model when run over periods of up to around 30 days. Suppose such a model is run for a much longer period. Although the model possesses no skill so far as forecasts of detailed weather are concerned, it would be instructive to know first of all if it is able to reproduce adequately the atmosphere's seasonal variations and secondly if it is able to reproduce the variability of the atmosphere which occurs on longer timescales from monthly to interannual.

For reproduction of the seasonal variations it is clearly necessary to vary the boundary conditions in particular the input of solar radiation at the top of the atmosphere and the temperature of the sea surface at the lower boundary. Lau (1981) ran such a model for 15 years using average climatological values of sea surface temperature. Although the average seasonal cycle was well represented, the observed atmospheric variability on monthly or longer timescales was not reproduced. Lau (1985) repeated the experiment with, at the lower boundary of the model, actual patterns of sea surface temperature (SST) observed in the tropical Pacific during the 15-year period 1962–76. Much of the missing variance now appeared demonstrating that much of the atmosphere's variability especially in tropical regions is closely related to fluctuations in SST.

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

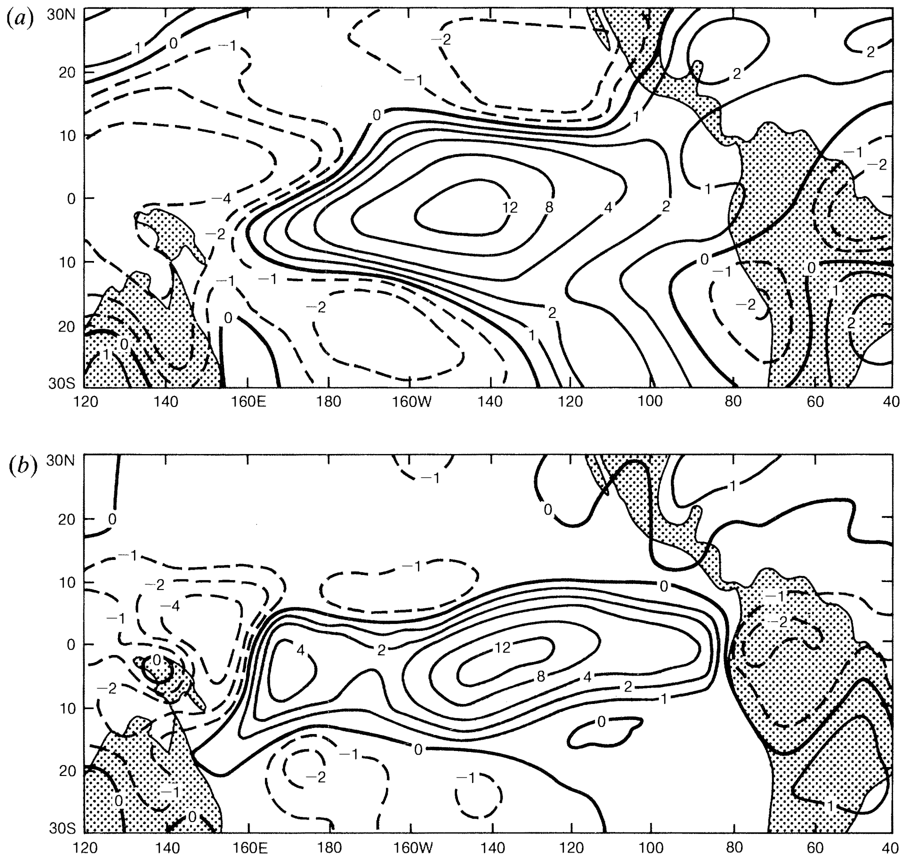


Figure 17. Maps of the precipitation anomaly over the Pacific Ocean in millimetres per day averaged over the period December, January, February 1982–83 during the large El Niño event, (*a*) as inferred from the observed anomaly in outgoing long-wave radiation; (*b*) from a general circulation model (from Shukla & Fennessy 1988).

It is not surprising that the atmosphere is so influenced by tropical SST. The largest contribution to the heat input to the atmosphere is due to evaporation from the ocean surface which because of the very rapid variation of the saturation water vapour pressure with temperature, increases rapidly with the SST.

The largest variations in tropical SST occur in the east tropical Pacific during periods of the El Niño when changes of SST of up to 7 K from its normal climatological average can occur (Enfield 1989). Associated with these El Niño events are anomalies in the circulation and rainfall in all tropical regions and also to a lesser extent at mid-latitudes. A particularly intense El Niño event occurred in 1982–83 associated with which were extreme events (droughts and floods) somewhere in almost all the continents (figure 16). Shukla & Fennessy (1988) ran an atmospheric model with the observed sequence of Pacific SST at the lower boundary and showed that the model is capable of simulating may of the weather anomalies at that time especially those in the tropics (figure 17).

Because of the large heat capacity of the oceans, SST anomalies tend to persist for some months. The possibility therefore exists, for regions where there is a strong correlation between weather and patterns of SST, of making forecasts some weeks or

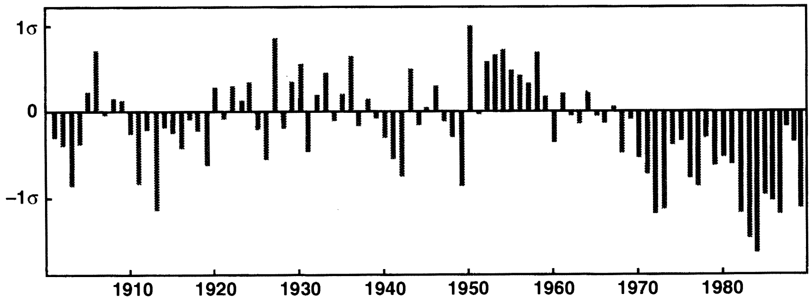


Figure 18. Standardized annual rainfall anomalies for the Sahel 1901–90. Values to 1984 after Nicholson (1985): 1985–89 established from CLIMAT reports (after Folland *et al.* 1986).

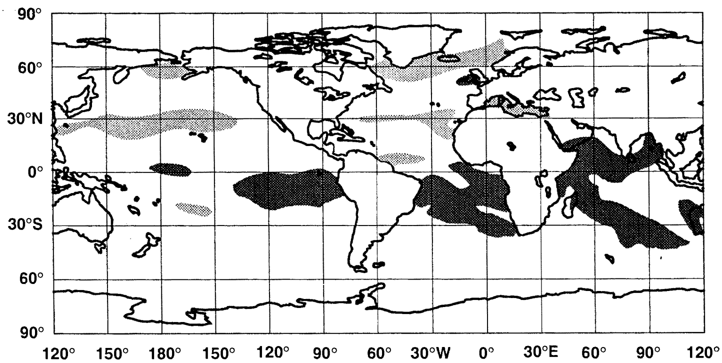


Figure 19. Sea surface temperatures: difference between Sahel dry years (average of 1972–73, 1982–84) and Sahel wet years (average of 1950, 1952–4, 1958) (after Folland *et al.* (1986)). Dark shading shows regions with differences greater than 0.5 °C, light shading shows regions with differences less than −0.5 °C.

months in advance. Such seasonal forecasts have been attempted at the U.K. Meteorological Office (Folland *et al.* 1986, 1990, 1991) for the Sahel Region of sub-Saharan Africa, a region where human survival is very dependent on the marginal rainfall (figure 18). The difference in the pattern of SST between wet and dry years in the Sahel shows as its largest feature a difference in the average SST between the Northern and Southern Hemispheres although detailed features in all the oceans are also included (figure 19). Folland *et al.* (1990) found that runs of the model in which the observed SSTs were included simulated well the observed rainfall during the important months of July to September. Simulations for six particular years, some wetter and some drier than average, are shown in figure 20. Having gained confidence from such simulations, correlations have been established between global SST patterns and rainfall which enable forecasts of rainfall in these months to be made. The success of the method can be judged from the hindcasts shown in figure 20 from which it will be seen that the main difficulty in providing accurate forecasts some months in advance is that the global SST pattern can vary substantially between April and June. Such forecasts are, however, proving valuable in the planning of agriculture and resources for the Sahel countries. Similar attempts have also been made with some success to use anomalous SST patterns to forecast rainfall in the North Nordeste region of Brazil (Ward & Folland 1991), and to simulate winter rainfall in Australia (Nicholls 1989).

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

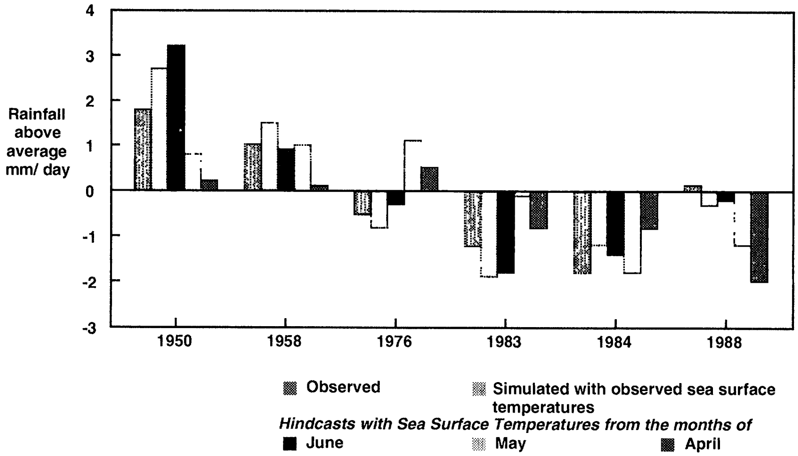


Figure 20. Observed and simulated rainfall in the Sahel (after Hulme *et al.* 1991).

A further step forward could be taken if it were possible to forecast changes in SST. To do that requires understanding of, and the ability to model, the ocean circulation and the way it is coupled to the atmospheric circulation. Because the largest changes in SST occur in the tropics and because there are reasons to suppose that the ocean may be more predictable in the tropics than elsewhere, most emphasis on the prediction of SST has been placed in tropical regions, in particular on the prediction of El Niño events themselves.

A large international project, the Tropical Ocean Global Atmosphere (TOGA), is addressing these phenomena. Of particular importance to the predictability of the variations associated with El Niño events is the filtering role of the tropical oceans both in space and time which appears to make the coupled tropical-ocean–global-atmosphere system less sensitive to the unpredictable components of the atmosphere's behaviour on these timescales of a month or more. In addition to this filtering role the ocean dynamics also provides the memory which is a prerequisite for continual oscillation.

A large variety of models have been used in the simulation of El Niño events from relatively simple conceptual models to complete coupled atmosphere–ocean circulation models (these are described in more detail in §17). McCreary & Anderson (1991) have recently reviewed this hierarchy of models. By way of example the mechanism typical of a simple model (McCreary 1983) is illustrated in figure 21. In this model a Rossby wave in the ocean propagates westwards from a warm SST anomaly near to the equator, then it is reflected from the western boundary as an eastward travelling Kelvin wave which cancels and reverses the sign of the anomaly, so triggering a cold event. This half-cycle determined by the speed of propagation of the waves takes of the order of two years. It is essentially driven by ocean dynamics, the associated atmospheric processes being determined by the patterns of SST (and in turn reinforcing those patterns) resulting from the ocean dynamics. As such the process appears essentially predictable in nature.

Some success in the prediction of El Niño events has been achieved with relatively simple models that couple together these essential features of the coupled atmosphere and ocean circulation. Barnett *et al.* (1988) report the successful prediction of the occurrence of the El Niño event of 1986–87 three to nine months in advance.

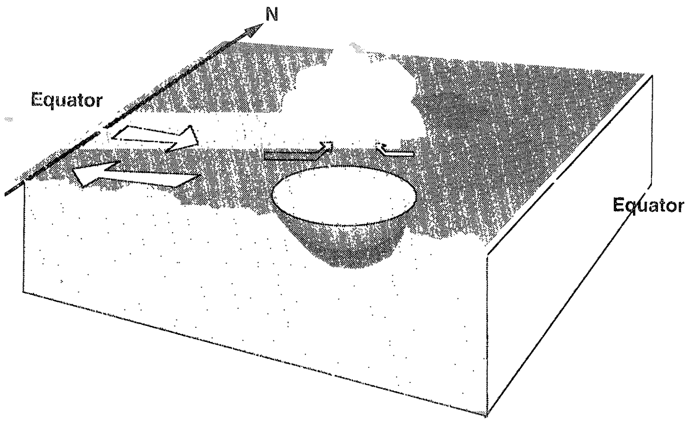


Figure 21. Schematic to illustrate El Niño oscillation.

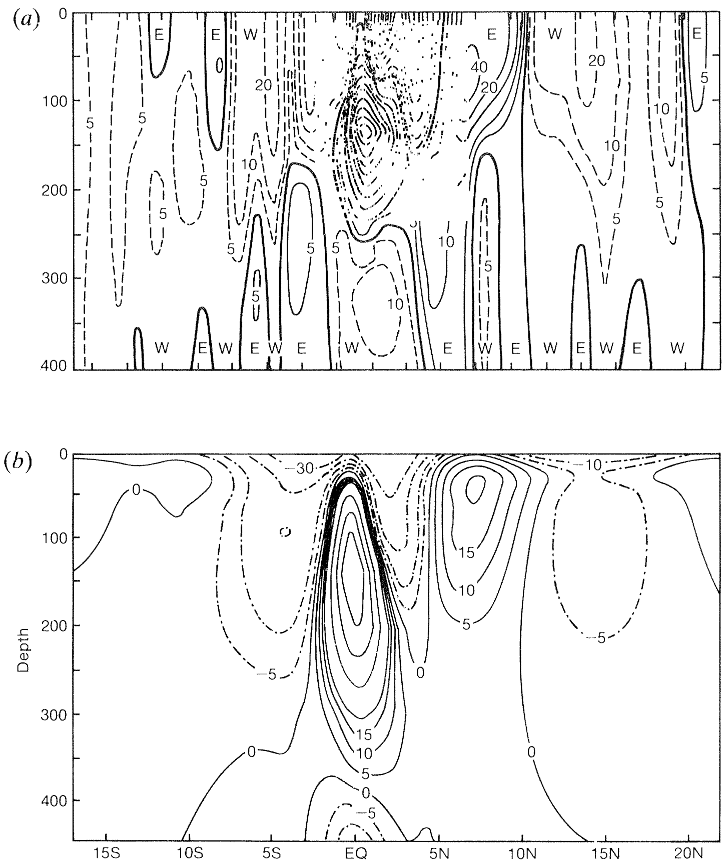


Figure 22. Annual mean zonal currents in centimetres per second along 155 °W as observed (a) and modelled (b). The depth and speed of the undercurrent is well simulated but the counter current and the north equatorial current are too weak (after Gordon & Corry 1990).

More complete coupled atmosphere–ocean models of the kind which are described in more detail in §17 have had only partial success in the simulation of El Niño events. An ocean model driven by forcing from the atmosphere has been developed

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

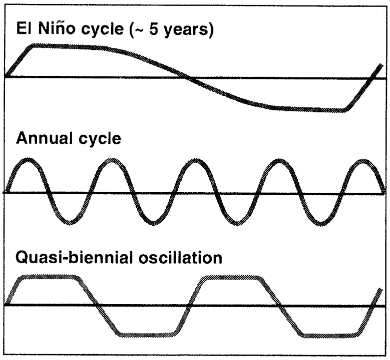


Figure 23. Illustrating cycles in the tropical atmosphere which can interact with the El Niño cycle.

by Gordon & Corry (1991). In this model only the tropical Pacific region of the oceans is fully interactive with the atmosphere, the remaining ocean temperatures being specified from climatology. The model (figure 22) captures well the main dynamical features in the ocean, for instance the North Equatorial Counter Current which flows in the opposite direction to the surface wind and the equatorial undercurrent, which is a strong jet-like feature only a few hundred kilometres wide but stretching over nearly the full span of the Pacific basin. When this ocean model is coupled to an atmospheric model it simulates well the seasonal cycle in both atmosphere and ocean but it fails to simulate El Niño events. Philander *et al.* (1989) have reported a successful simulation of El Niño type events with a coupled atmosphere–ocean model which included climatological cloud but no seasonal cycle.

While, therefore, it is clear that some elements of El Niño events are predictable and some success in their prediction 3–9 months in advance has been achieved we have yet to elucidate all their complex features. Further, the typical period of the El Niño cycle of about five years almost certainly interacts in a partly chaotic way with the seasonal cycle and with other cyclic events of importance in tropical regions, for instance, the quasi-biennial oscillation (figure 23) so limiting the degree of predictability which we can ultimately expect to achieve for El Niño events.

11. The climate system

So far we have considered the forecasting of detailed weather over a few days and of average weather for a month, perhaps up to a season ahead. Climate is concerned with substantially longer periods of time from a few years to perhaps a decade. A description of the climate over a period involves the averages of appropriate components of the weather over that period together with the statistical variations of those components.

The climate variables which are commonly used are concerned mainly with the atmosphere. But, in considering the climate we cannot look at the atmosphere alone. We have already seen that processes in the atmosphere are strongly coupled to the oceans; they are also coupled to the land surface and to those parts of the Earth covered with ice (the cryosphere). There is also strong coupling to the biosphere (the vegetation and other living systems on the land and in the ocean). These five components (atmosphere, land, ocean ice and biosphere) together make up the climate system (figure 24).

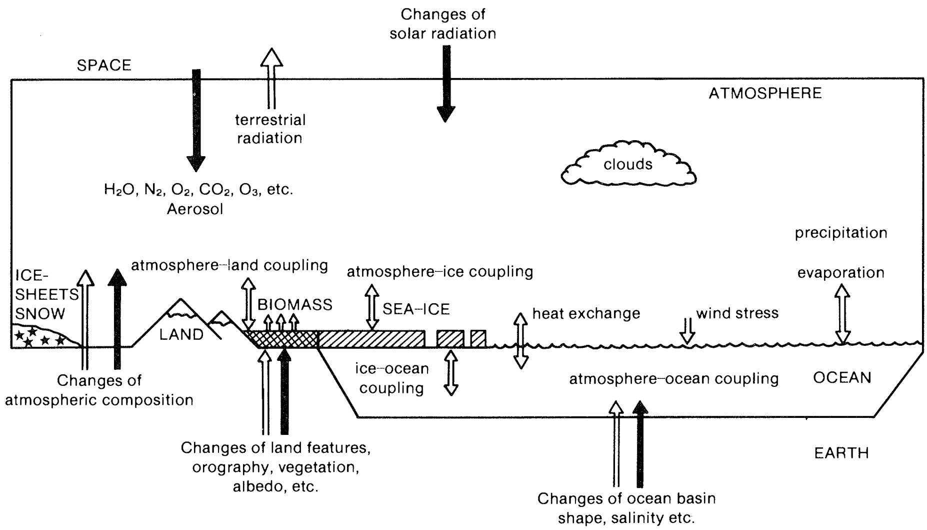


Figure 24. Schematic of the climate system.

Fluctuations of climate occur on many timescales. These may occur because of internal interactions within the total climate system (e.g. interchanges between its various components) or because of variations in the external forcing (e.g. variations in the distribution of solar radiation or in volcanic activity). Such changes are often referred to as natural climate variability. Changes may also occur as a result of human activities; the particular climate changes which I consider in this paper are those which may occur over the next century as a result of human activities.

12. The Greenhouse Effect

The driving force for weather and climate is energy from the Sun. The atmosphere and the surface of the Earth intercept the incident solar radiation; about 30 % of it is reflected back out to space, the rest is absorbed. The energy absorbed from solar radiation must on average be balanced by outgoing infrared radiation from the Earth. As the outgoing radiation is determined by the Earth's temperature, this temperature will adjust until there is a balance between incoming and outgoing radiation.

Short-wave solar radiation passes through a clear atmosphere relatively unimpeded, but the infrared radiation emitted by the warm surface is partly absorbed and then re-emitted by a number of gases present in the cooler atmosphere above. Together these processes add to the net energy input to the lower atmosphere and the underlying surface thereby increasing their temperature. This is the basic Greenhouse Effect; by transmitting the short-wave and absorbing the infrared radiation the gases act in an analogous way to the glass in a greenhouse.

The main greenhouse gases are not the major constituents, nitrogen and oxygen, but water vapour (the largest contributor), carbon dioxide, methane, nitrous oxide and ozone. The presence of these gases maintains the mean temperature of the Earth's surface 32 K warmer than it would otherwise be. Since the beginning of the industrial revolution, due to human activities, the concentration of carbon dioxide in the atmosphere has risen by about 25 % (figure 25), that of methane has about doubled. Further greenhouse gases such as the chlorofluorocarbons have appeared in

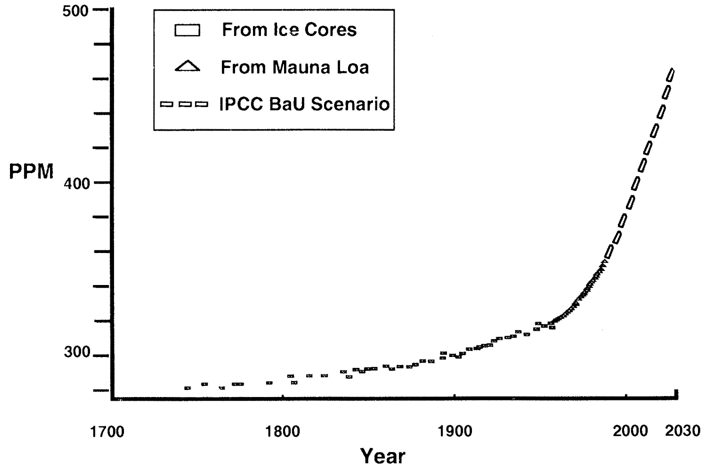


Figure 25. Atmospheric CO_2 concentration in parts per million since 1700 and projected to 2030.

significant quantities for the first time. These increases are leading to an enhanced Greenhouse Effect.

Increased human activity in the future especially the burning of fossil fuels and deforestation will cause an acceleration in the rate of increase in the concentration of many of these gases in particular of carbon dioxide (figure 25). Substantial global warming of the climate is therefore expected next century. Estimates by the Intergovernmental Panel on Climate Change (IPCC 1990) are that the increased concentrations of greenhouse gases since pre-industrial times could already have caused a global warming of about 0.8 K. The Panel further estimates that if the emissions of greenhouse gases follow a 'business as usual' scenario (i.e. no control of emissions), the mean global temperature is likely to rise by a further 1 K by the year 2025 and a further 3 K by the end of next century (figure 26). So far as we can judge these represent faster rates of change in the global average than have occurred on Earth at any time since the end of the last ice age over 10000 years ago.

Many careful studies have been carried out of the observational record of global temperature over the past 100 years. Although an increase of about 0.6 °C has been observed overall (Folland *et al.* 1990), the temporal pattern of change over the period does not fit in well with that expected due to the enhanced Greenhouse Effect (figure 26). Natural variability can also be seen in the record. Although therefore the record over this period is consistent with the predictions of the Intergovernmental Panel it cannot be used as unequivocal evidence for them. The Panel suggests that it may be a decade or more before such unequivocal detection of the enhanced greenhouse signal is possible from the observations.

The predictions of change by the Intergovernmental Panel have been made largely on the basis of climate models. It is a matter of considerable debate and concern as to whether these predictions are well founded, what uncertainties they contain and indeed to what extent the effect of human activities on the climate over this period are predictable. In the rest of this paper I consider these questions. Firstly I describe other known forcing factors on the climate and their likely effect on the century timescale. I then give a brief description of the present generation of climate models and how they are used for climate prediction. I then address the major uncertainties in the models and the degree of confidence we can have in their predictions.

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

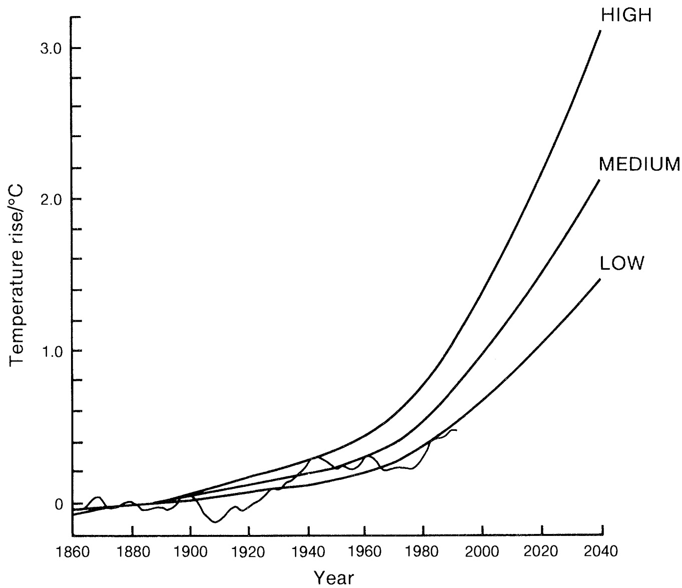


Figure 26. Temperature rise as predicted by the IPCC (1990) as a result of the increase in greenhouse gases under the IPCC 'business-as-usual' scenario. Also shown is the observed global average temperature from 1860–1990 (from Folland *et al.* 1990). The three curves show the estimated range of uncertainty.

13. Other forcing factors

We have already seen how the enhanced Greenhouse Effect disturbs the radiative balance of the planet. In the absence of any change in other parameters (such as the water vapour content, the cloud cover or the ocean circulation) it is relatively easy to calculate the change in net radiation at the tropopause (the top of the lower atmosphere) which would occur if the carbon dioxide concentration were, for instance, doubled from its pre-industrial value. The change is about 4 W m^{-2}, which can be compared with the average solar radiation absorbed by the Earth of about 236 W m^{-2} (figure 27).

The most obvious place to look for external variations in radiative forcing is in the solar radiation incident on the Earth. There are two ways in which this can vary. The first arises from changes in the orbit of the Earth around the Sun and in the Earth's rotation which are known to occur on timescales of 10^4–10^5 years. The second is due to physical changes of the Sun itself. We briefly consider these types of variations in turn.

Three regular variations occur in the Sun–Earth orbital parameters. The eccentricity of the earth's orbit varies with a period of around 10^5 years, the obliquity of the Earth's axis with a period of about 41000 years and the longitude of perihelion varies with a period of about 23000 years. As the Earth's orbit changes in its relationship to the Sun, although the total quantity of solar radiation reaching the Earth varies very little, the distribution of that radiation with latitude and season over the Earth's surface varies considerably especially in polar regions (figure 28). It was first pointed out by James Croll (1867), whose ideas were later developed by Milankovitch (1920), that the major glacial–interglacial cycles over the past few hundred thousand years might be linked with these regular variations in the

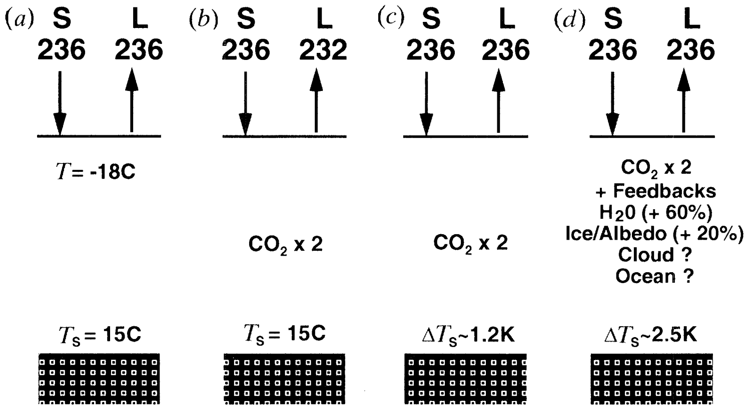


Figure 27. The enhanced Greenhouse Effect due to doubled CO_2 showing solar (S) and longwave (L) radiation in watts per square metre at the top of the atmosphere. (*a*) The present situation with an effective radiating temperature of the Earth of -18 °C and an average surface temperature of 15 °C; (*b*) with doubled atmospheric CO_2 and no other change; (*c*) with doubled atmospheric CO_2 and temperature change to bring the system back into radiation balance but with no other changes (average surface temperature increases by 1.2 K); (*d*) with doubled atmospheric CO_2 and other changes (including feedbacks) as estimated by the IPCC (1990) (average surface temperature increases by about 2.5 K).

distribution of solar radiation reaching the Earth. The covariation of the orbital parameters and the record of the Earth's climate over this period as established particularly from data from ice cores and ocean sediments provides compelling evidence for the Milankovitch theory (Berger 1988; figure 28)). However, the magnitude of the observed climate changes are larger than might be expected from radiation forcing alone especially for the dominant period of 10^5 years. Internal feedback processes have therefore to be invoked in addition to Milankovitch forcing to explain the observed climate variations. The strong correlation observed in the climatic record between average atmospheric temperature and CO_2 concentration (figure 29) suggests that one such feedback arises from changes in CO_2 influencing atmospheric temperature through the Greenhouse Effect (Lorius *et al.* 1990). Not that such a correlation proves the existence of the greenhouse feedback; indeed there is evidence that at least part of the correlation arises because the atmospheric CO_2 concentration is itself influenced, through biological feedbacks, by factors which are related to the average global temperature. However, the successful simulation by models (see §15) of past climatic conditions is not possible without allowing for changes in the Greenhouse Effect due to changes in CO_2 concentration.

Changes in forcing because of the Milankovitch effect occur on timescales of thousands of years. Over much shorter periods the changes are small. For instance, over the past 10^4 years the incident solar radiation at 60° N in July has decreased by about 35 W m^{-2} (Rind *et al.* 1989). The average change over 100 years would be about 0.35 W m^{-2}, which is small compared with that estimated for the forcing due to changes in greenhouse gas concentrations over the same period. I refer again to Milankovitch forcing in §20.

We now consider whether there are physical changes in the sun itself which might lead to changes in solar output. Accurate measurements of the radiation output of the Sun have been made from spacecraft only during the past decade. These show that variations occur during the 11-year solar cycle. The largest of these are in the

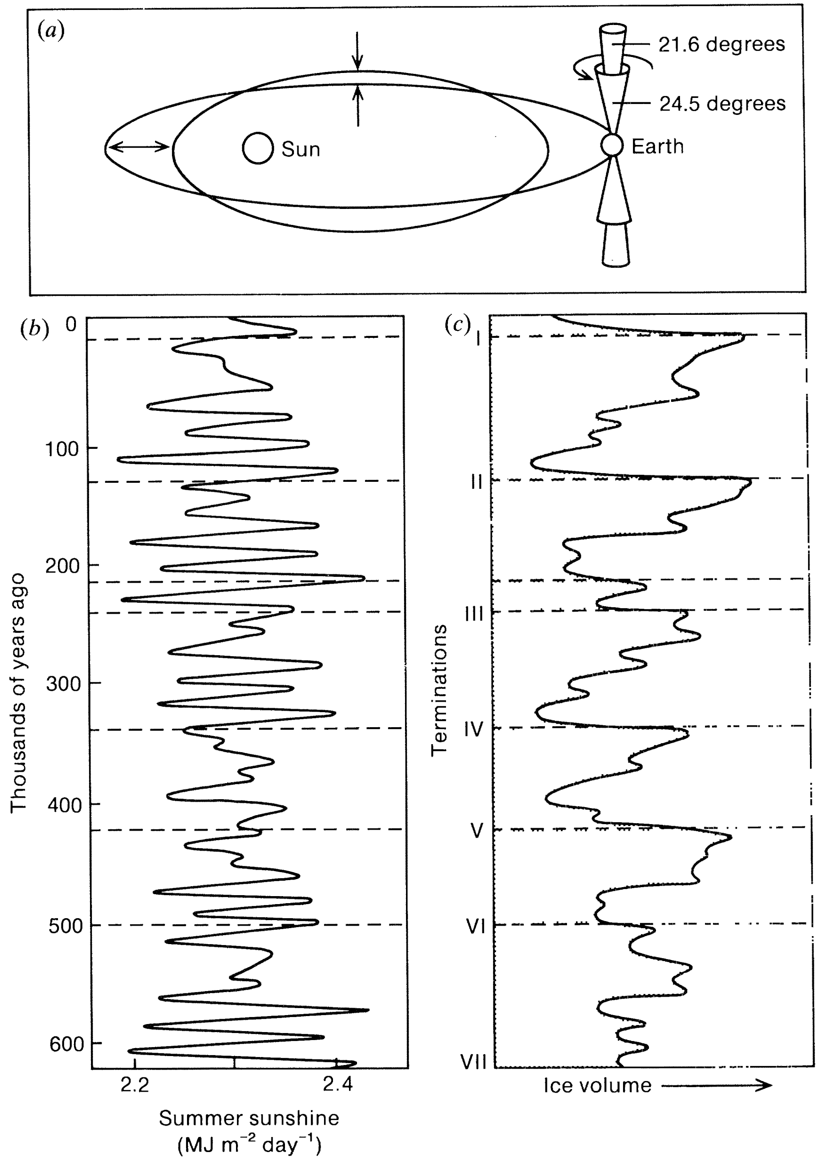


Figure 28. Variations in the Earth's orbit (*a*) in its eccentricity, the orientation of its spin axis (between 21.6° and 24.5°) and the longitude of perihelion cause changes in the incidence of solar radiation at the poles (*b*) which appear as cycles in the climate record (*c*), a mechanism for the triggering of climate change suggested by Milankovitch (1920) (after Broecker & Denton 1990).

far ultraviolet part of the spectrum. At wavelengths below 300 nm the variations during a solar cycle amount to about 1%. Since, however, the energy at these wavelengths is only about 1% of the total solar output, the ultraviolet variations are very small in energy terms. At other wavelengths evidence from several carefully calibrated spacecraft instruments indicate that total solar energy has varied by about 0.1% during the last solar cycle (Wilson & Hudson 1988). Foukal & Lean (1990) have studied the spacecraft measurements and have shown that the detailed changes in solar irradiance correlate closely with changes in the area of faculae on the

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

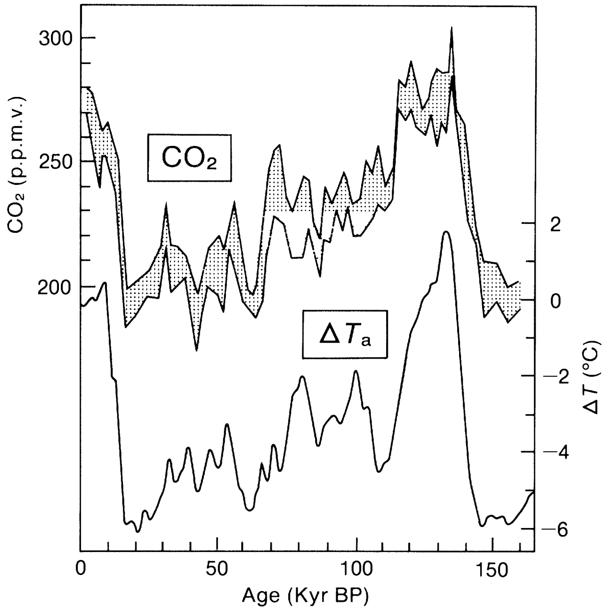


Figure 29. Variations during the last climate cycle as derived from measurements along the Vostok ice core of the CO_2 atmospheric concentration (showing also the range of uncertainty) and the atmospheric temperature over Antarctica (from Lorius *et al.* 1990).

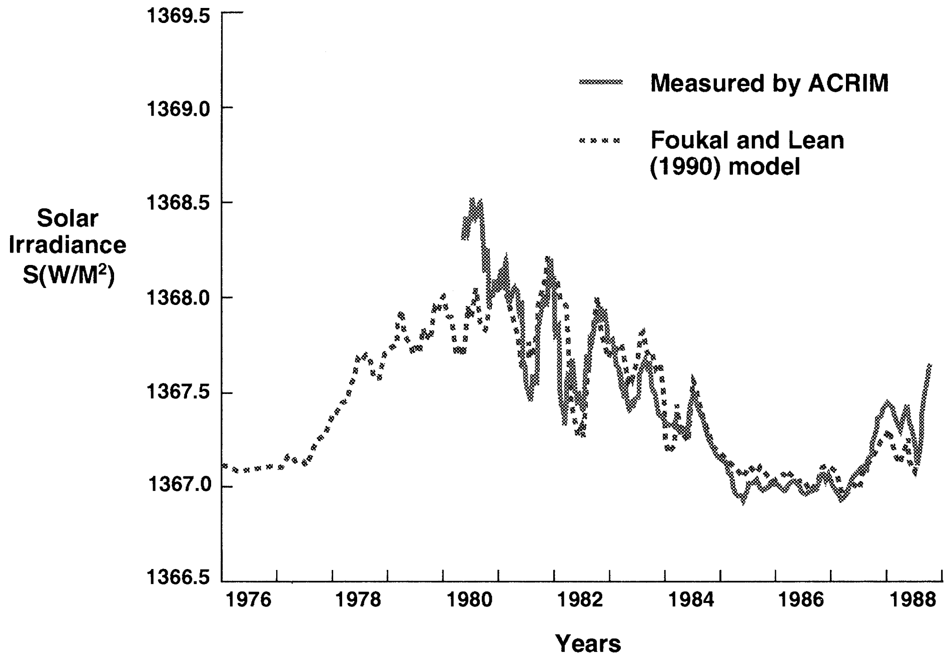


Figure 30. Changes in solar irradiance since 1976 as observed from satellites and as simulated by a model (after Foukal & Lean 1990).

Sun's surface as indicated by the flux from the sun observed at the Lyman α wavelength (figure 30). Using sunspot number *Rz* as a proxy for facular area, and allowing also for the dimming factor which depends on total sunspot area, Foukal &

MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES
THE ROYAL SOCIETY A
PHILOSOPHICAL TRANSACTIONS OF

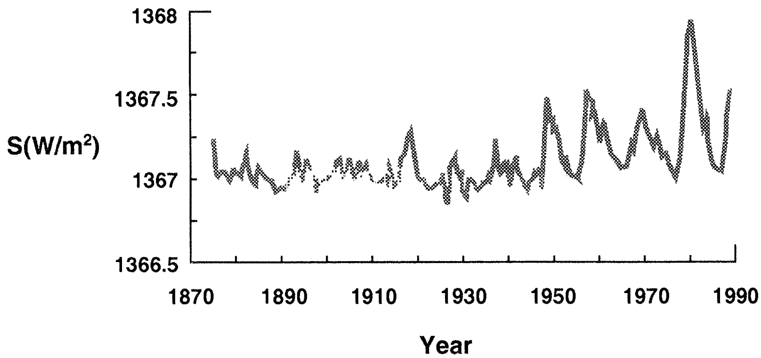


Figure 31. Changes in solar irradiance 1870–1990 as estimated by Foukal & Lean (1990).

Lean have reconstructed a curve of total solar irradiance form 1874 to 1988 (figure 31). The largest variation of around 0.1 % during a cycle occurred during the last solar cycle in the 1980s, largely because during this particular cycle the variations in sunspot number *Rz* were of different amplitude to the sunspot dimming factor.

In recent years some have speculated that there might be larger changes in the solar output over longer periods (see for instance Eddy 1977). These speculations have largely been based on possible correlations between climatic events and changes in solar activity as evidenced by the atmospheric radiocarbon record (Wigley & Kelly 1990). Although there are some notable coincidences, for instance the 'Little Ice Age' occurred at about the same time as the Maunder minimum in frequency of sunspot occurrences, the overall correlations are far from convincing and since they are not supported by any direct measurements, it must be concluded that there is no strong evidence for the existence of such larger changes in solar output.

Other radiative forcing of a magnitude comparable with that due to the increase of greenhouse gases could arise from changes in the aerosol content of the atmosphere either through volcanic activity, changes in land use or from emissions of aerosol because of human activities. The effects of the first two are direct, the aerosol either absorb or reflect solar radiation directly. The resulting forcing has been shown by Shine *et al.* (1990) as likely to be much smaller than that due to increasing greenhouse gases. The effects of aerosol from human activities are more likely to be indirect. Emission of sulphur compounds create sulphate aerosol which act as effective nuclei for the condensation of water drops in clouds. Over polluted areas therefore clouds will contain a greater number of droplets; they will possess a higher albedo (Twomey *et al.* 1984) and, providing the cloud extent is unchanged, less energy from the Sun will be absorbed. Observations from spacecraft show that clouds in the wake of ships possess higher albedo than in surrounding areas (Coakley *et al.* 1987); also there is evidence that clouds near to the eastern seaboard of major continents possess higher albedo than those near the western seaboard (Shine *et al.* 1990), so supporting the hypothesis. However, again Shine *et al.* (1990) (see also Hansen & Lacis 1990) conclude that changes in forcing arising from this aspect of human activity are likely to be much less than those from the expected increase in greenhouse gases.

14. Feedbacks in the climate system

At the beginning of the last section we noted that in the absence of any other change the increase in radiative forcing which would occur as a result of a doubling

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

of atmospheric CO_2 concentration would be about 4 W m^{-2} (figure 27). Again in the absence of any other changes apart from that of atmospheric temperature, a relatively simple radiative transfer calculation enables us to calculate that the resulting average increase in surface temperature would be 1.2 °C (figure 27). Other changes, however, will occur because of the many feedbacks between atmospheric temperature and other parameters. The most important and direct of these are the following.

1. *Water vapour feedback*. A warmer atmosphere is expected on average to possess a higher water vapour content. Since water vapour is a powerful greenhouse gas, on average a positive feedback results (Raval & Ramanathan 1989).

2. *Ice albedo feedback*. As some ice melts at the warmer surface, solar radiation which had been reflected back to space by the ice or snow surface, is absorbed leading to further increased warming – another positive feedback.

3. *Cloud radiation feedback*. This is described in more detail below in §16; it can be of either sign.

Taking into account these feedbacks, the IPCC (1990) estimated an increase in global average temperature under equilibrium conditions if CO_2 were doubled of about 2.5 °C with a range of uncertainty (largely arising from model uncertainties) from 1.5 to 4.5 °C.

Further feedbacks which do not influence the radiative régimes so directly are concerned with changes which may occur in the ocean circulation (see §17 below) and changes which may occur in the biosphere which I do not address in any detail in this paper.

15. Climate modelling

In earlier sections we have seen that an atmospheric model with prescribed conditions at the lower boundary (specified sea surface temperature, for instance) can be run over a period to provide a description of the atmosphere's climate. However, to simulate adequately the change in climate due to increasing greenhouse gases including all the relevant feedbacks comprehensive models are required which take account of all components of the climate system (figure 24) and the interactions between them. The difficulty with the development of such models is that they are bound to be highly complex and they are extremely demanding on computer resources. It is only during the past decade that computers have been sufficiently powerful to enable climate models to be developed which couple together, even in an elementary way, the circulations of the atmosphere and the oceans.

Important tests of climate models are their ability to simulate current climate. In §10 we described the performance of atmospheric models (see also the summary by Gates *et al.* (1990)) and showed that they possess significant skill in the portrayal of the large-scale flow including its seasonal variations and the response of that flow to anomalous changes in the distribution of sea surface temperature. On the regional scale, however, atmospheric models show significant errors especially in precipitation (typical mean errors of 20–50%) with the more recent higher-resolution models giving the better results.

The particular problems of modelling the ocean circulation and of atmosphere–ocean coupled models will be discussed in §17. Suffice it here to say that ocean models driven by mean fluxes of heat, momentum and water at the ocean–atmosphere interface also show considerable skill in the simulation of the ocean circulation (Gates *et al.* 1990), although far less data exist with which to compare the

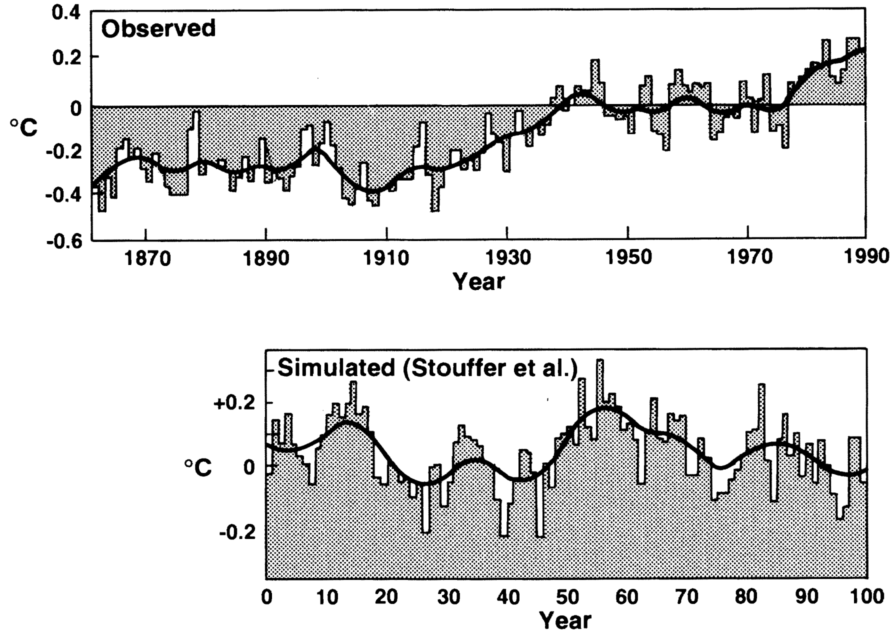


Figure 32. Observed and simulated variability in global average surface temperature (*a*) observed (from IPCC 1990) (*b*) simulated by a coupled ocean–atmosphere model (Stouffer *et al.* 1989).

model simulations than is the case with the atmospheric models. Further, coupled ocean–atmosphere models are capable of similar skill to the atmospheric models in their simulation of atmospheric circulation when run with the same resolution (see §17). That atmospheric models with constant boundary conditions can generate variability on all timescales (even very long timescales) has been demonstrated by James & James (1991). However, more variability is shown on decadal timescales by atmosphere–ocean coupled models – variability similar to that in the real climate system (figure 32) – presumably because interactions between the atmospheric and the oceanic components of the climate system are important in generating the variability.

To establish the credibility of climate models it is clearly necessary that they provide satisfactory simulations of the current climate and its variability. The models have now gone a long way towards meeting that requirement. A further and potentially more stringent test is whether they are able to simulate past climatic conditions when the radiative forcing possessed a substantially different distribution. There are two difficulties with such tests. Firstly there is only limited information about the details of such past climates against which to test the model simulations and secondly it is necessary to add as a constraint to the model simulations the distribution of the ice sheets and sea level at the time of the simulations. Ideally the models might be expected to simulate the behaviour of the ice sheets in the same way as they simulate other components of the system. However, to do that would mean being able to run a model with confidence over tens of thousands of years, something which is not possible at present. But given that limitation, Manabe & Broccoli (1985), by using an atmospheric model coupled to a mixed-layer ocean model, have successfully simulated the distributions of sea surface temperature and sea ice during the last glacial maximum around 18000 years ago (figure 33). This

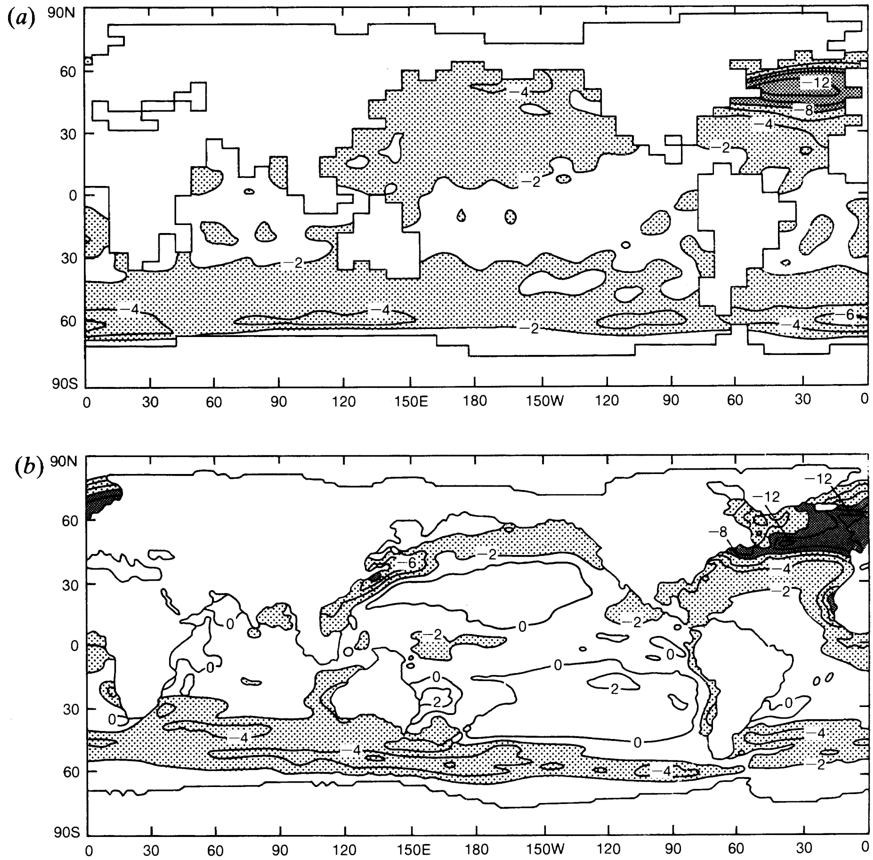


Figure 33. Distribution (in kelvins) of differences in sea surface temperature between the last glacial maximum and the present (*a*) as simulated by a model (*b*) as estimated from CLIMAP data (after Manabe & Broccoli 1985).

simulation also demonstrated the importance of the reduced Greenhouse Effect at that time due to the lower level of atmospheric CO_2 (Broccoli & Manabe 1987). Other simulations of past climate are summarized by Gates *et al.* (1990).

Having given a brief résumé of the present status of climate modelling we now return to address the problems of incorporating within the models particular feedbacks which we summarized in §14. The main uncertainties in model predictions at the present time arise from two of these feedbacks, namely the cloud–radiation feedback and feedbacks due to the effects of ocean-circulation. These feedbacks are discussed in turn.

16. Cloud–radiation feedback

Clouds interfere with the transfer of radiation in the atmosphere in two ways (figure 34 and see Fouquart *et al.* (1990) for a review). Firstly they reflect a certain proportion of solar radiation back to space so reducing the total energy available to the system; secondly they absorb and reradiate infrared radiation emitted by the Earth's surface below in a similar way to the greenhouse gases, so reducing the long wave heat loss to space at the surface. Which effect dominates for any particular cloud depends on the cloud temperature (and hence on the cloud height) and on its

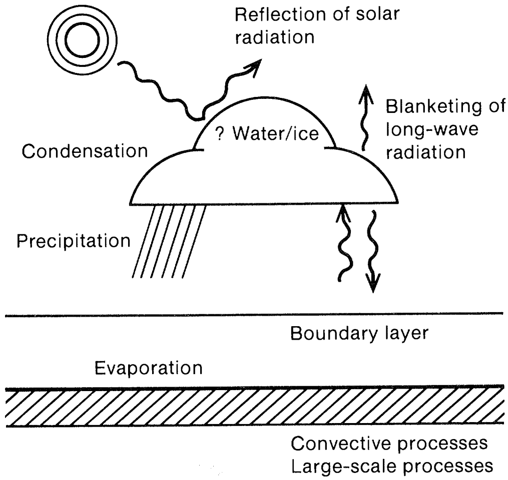


Figure 34. Schematic of the physical processes associated with clouds.

detailed optical properties. The latter depend on whether the cloud is of water or ice, on the liquid water content and on the average droplet size in the cloud. In general low clouds tend to cool the Earth–atmosphere system and high clouds tend to warm it.

A concept which is helpful in distinguishing between the two effects we have mentioned is that of cloud radiative forcing (CRF) which can be divided into a short-wave (F_s) and a long-wave (F_l) component. It is defined as:

$$F_s = S_0 - S = Q(a_0 - a)$$

for the shortwave component and

$$F_l = L_0 - L$$

for the long-wave component, where S and L are respectively the short-wave and long-wave radiant fluxes leaving the top of the atmosphere and S_0 and L_0 are the values of these fluxes in the absence of cloud. The short-wave fluxes may also be written in terms of the insolation at the top of the atmosphere Q and the albedo a.

Representative values of CRF as inferred from satellite instruments during the Earth Radiation Budget Experiment (ERBE) in 1985 and 1986 are shown in figure 35. These data and other more detailed data available from ERBE (Ramanathan *et al.* 1989; Kiehl & Ramanathan 1990; Harrison *et al.* 1990; Hartmann & Doelling 1991) show that (1) on average clouds act to cool the Earth–atmosphere system, (2) tropical cloud systems show near cancellation of short-wave and long-wave forcing so that for them the net CRF is about zero (Slingo & Jones 1991, and figure 36), (3) the largest magnitudes of net CRF are over the mid-latitude Atlantic and Pacific oceans, especially where there are regions of extended stratus. In table 1, by using these results and simple radiative calculations, the estimates of the effect on global average surface temperature of clouds and of greenhouse gases are compared (see also Slingo 1990). From this table we can see that a small percentage change in cloud cover (high cloud or low cloud) can have an effect on the surface temperature of similar magnitude to that due to a doubling of CO_2. It is therefore very important that changes which might occur in cloud cover are adequately understood.

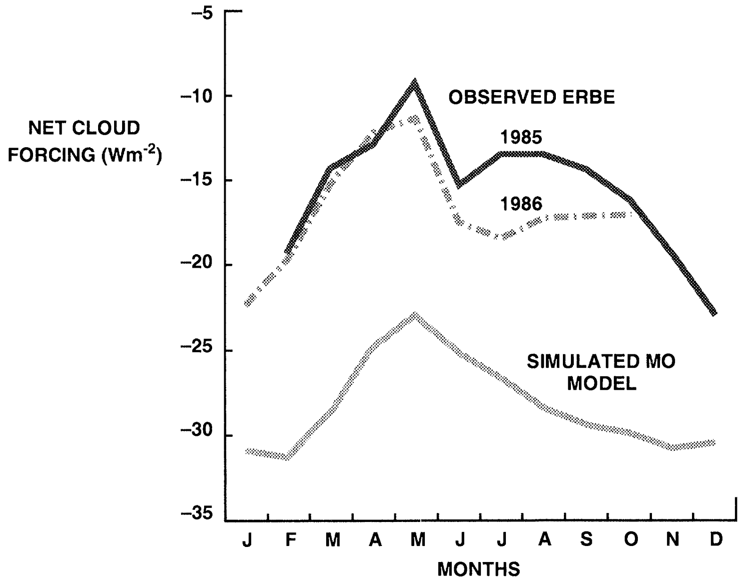


Figure 35. Net globally averaged cloud forcing as observed from satellites in two years of the Earth Radiation Budget Experiment and as simulated in a climate model of the Meteorological Office (after Slingo, personal communication). The model broadly simulates the seasonal cycle in agreement with the observations; the systematic difference between model and observations is a feature resulting from the way in which the cloud parametrization in the particular model is 'tuned'.

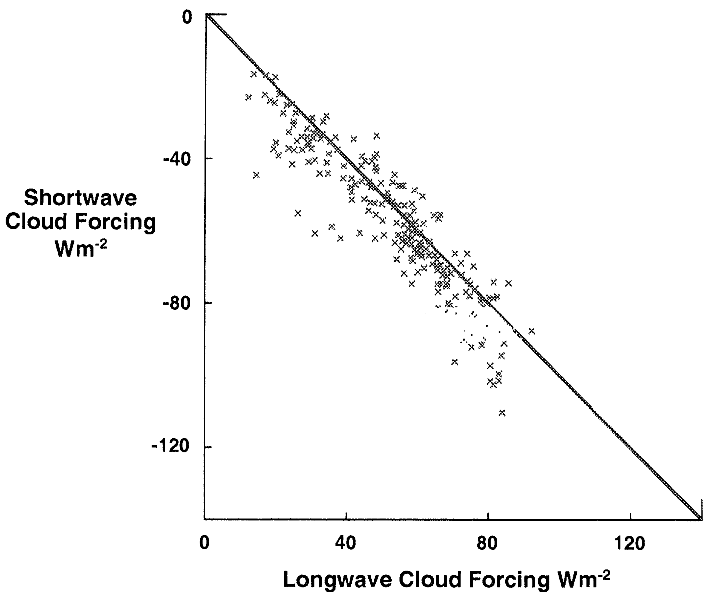


Figure 36. Shortwave and longwave forcing of tropical clouds showing near cancellation (after Slingo & Jones 1991).

Now an atmosphere with increased temperature due to the enhanced Greenhouse Effect will also possess on average increased water vapour. It might be argued that an atmosphere containing more water vapour would also possess increased cloud

Table 1. *Global average temperature changes under different assumptions about greenhouse gases and clouds*

greenhouse gases	clouds	(change from current average global surface temperature of 288 K)/K
as now	as now	0
none	as now	−32
none	none	−21
as now	none	+4
as now	as now but +3 % high cloud	0.3
as now	as now but +3 % low cloud	−1.0
doubled CO_2 concentration; otherwise as now	as now i.e. no additional cloud feedback	1.1
doubled CO_2 concentration + increased H_2O + ice/albedo feedback	as now, i.e. no additional cloud feedback	2.2

cover. Then, since clouds have an overall cooling effect on the climate it would be expected that the feedback effect of clouds on the warming due to the enhanced greenhouse effect would be on average slightly negative, but with little effect in the tropics. However, such an argument is too simple. The development of clouds is a complex process and cloud cover is not necessarily related linearly to the fields of temperature and water vapour content. To study the effects in more detail we need to turn to the models which as we have seen are the most effective way of taking proper account of all the nonlinear processes involved.

Most climate models to date have used comparatively simple schemes for the parametrization of cloud. In the Meteorological Office model, for instance, three levels of cloud have been permitted each with prescribed optical properties. For dynamically created cloud, the presence or absence of cloud has been determined by a given value of relative humidity, the actual value of threshold humidity for each cloud layer being 'tuned' so that the average cloud cover in the model is approximately that observed in the real atmosphere. For convective cloud, a particular prescription of cloud cover has been incorporated as part of the convection scheme. Comparison of the cloud–climate feedback in 14 atmospheric models show large differences between them (Cess *et al.* 1989) even though the cloud parametrization schemes in many of them are similar. It is clear from this and other considerations that cloud feedback is dependent on many aspects of a model's formulation and not just on the parametrization of cloud formation.

It is only recently that more sophisticated schemes have been introduced into models. The most extensive experiments with more elaborate schemes have been carried out by Mitchell *et al.* (1989). They compared the results from the model when three different cloud parametrization schemes were used. The first was the simple relative humidity scheme described in the last paragraph. The second explicitly included a cloud water variable and distinguished between the properties and the precipitation processes associated with water and ice clouds. The third scheme varied the radiative properties of clouds dependent on the cloud water content. For each scheme runs of the model were carried out both with normal and with doubled CO_2

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

concentration in the atmosphere. The results show on average substantial positive cloud–radiation feedback for the first scheme (global average temperature rise about 5 °C), less positive feedback for the second scheme (global average temperature rise about 3 °C) and slight negative feedback for the third scheme (global average temperature rise about 2 °C). Although the second and third schemes are more elaborate and include descriptions of more of the physical processes, because they are at an early stage of development, Mitchell *et al.* (1989) point out that they are not necessarily more accurate. The large differences shown up in such comparisons effectively demonstrate the sensitivity of both the sign and the magnitude of the feedback to the details of the parametrization scheme used.

How can the large uncertainty arising from our inadequate knowledge of cloud–radiation feedback be reduced? More studies, carefully carried out, with the more elaborate parametrization schemes are clearly required. But model studies alone will not provide the confidence needed. Accurate observations are also required both of clouds and the relevant radiative parameters (as provided by the ERBE) with extended coverage in space and time. Careful comparisons then need to be made with model results to verify that the models are capable not only of generating the average observed cloud amounts and radiative fluxes, but also that they can generate a similar range and frequency structures in the variability of these quantities as are observed in the real atmosphere (Smith & Vonder Haar 1991). Only after such stringent comparisons have been successfully carried out on a number of models will there be sufficient confidence to narrow the uncertainty in the predictions of climate change due to increased greenhouse gas emissions next century.

17. Ocean–circulation feedback

We now turn to address the influence of the oceans on climate and climate change. The oceans play a large part in determining the existing climate of the Earth; they are therefore also likely to be an important influence on climate change due to human activities. The oceans act on the climate in three important ways: (1) through the flux of water vapour with its subsequent release of latent heat they provide the largest single heat source for the atmosphere; (2) because of their large thermal capacity compared with the atmosphere (the entire heat capacity of the atmosphere is equivalent to less than 3 m depth of water) they control the rate of change of atmospheric parameters; in this they act on all timescales from the diurnal to the century or more; (3) through the internal circulation of the ocean they redistribute heat throughout all components of the climate system. Estimates of the transport of heat by the oceans (for instance by Oort & Vonder Haar 1976) are of similar overall magnitude to that by the atmosphere. However, the regional distribution (see figure 37) is very different. Even small changes in the regional heat transport by the ocean could have large implications for climate change. For instance, if we consider the region of the North Atlantic ocean west of the British Isles, the heat input (figure 37) carried by the ocean circulation is on average about 100 W m^{-2}, of similar magnitude to that reaching the ocean surface there from the incident solar radiation. Any accurate simulation of likely climate change especially of its regional variations, therefore, must include a description of the ocean structure and dynamics.

Although the general features of the main water masses in the ocean and of the ocean currents which transport heat and salt are know from a large number of soundings from ships over the years (figures 37 and 38), very little is known about

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

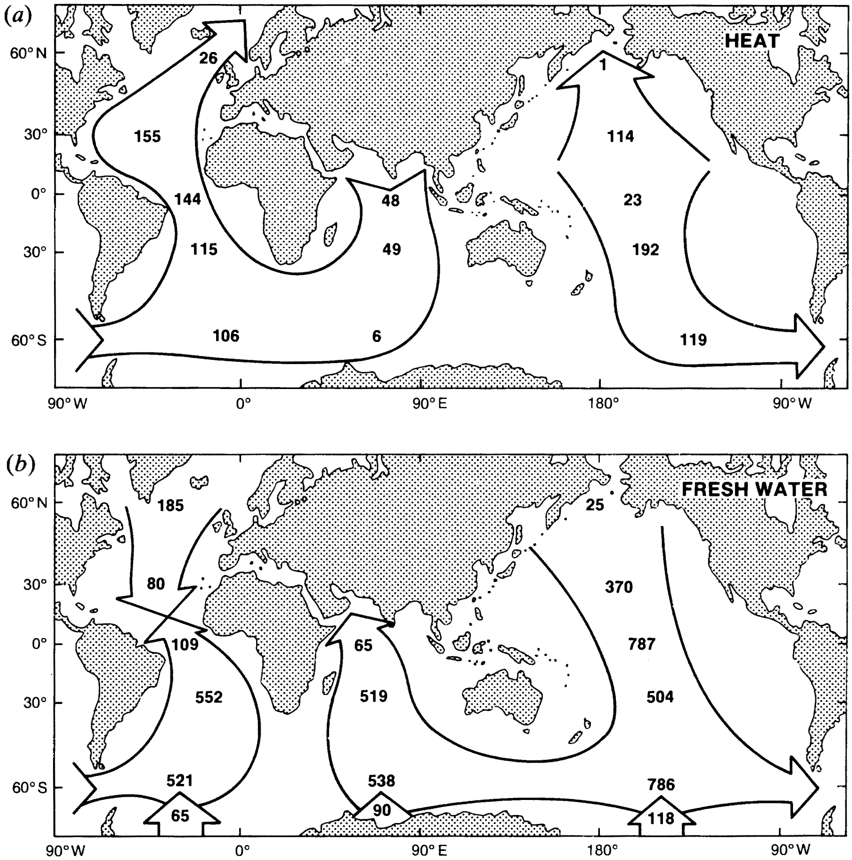


Figure 37. Estimates of transport by the oceans of heat and fresh water (after Woods 1984). Units are 10^{13} W for heat and kt s^{-1} for fresh water.

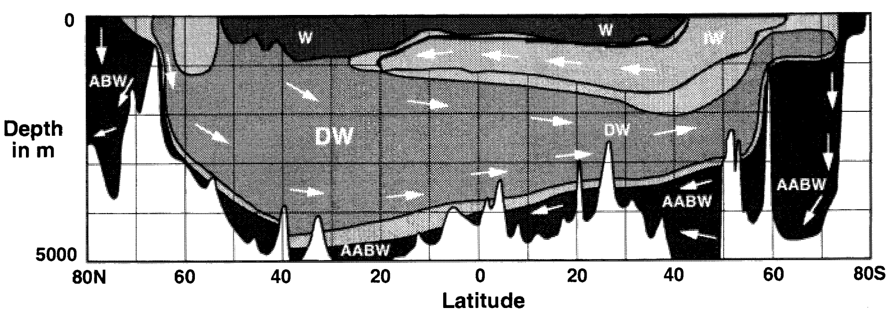


Figure 38. Water masses in the Atlantic ocean showing warm water sphere (W), deep water (DW), intermediate water (IW) and Arctic and Antarctic bottom water (ABW and AABW) (adapted from Wunsch 1984).

the variability of these currents. Even less well known from observations are details of the dynamics of the ocean on smaller scales (less than 1000 km) and of the interactions which occur between motions on different scales. Of particular importance are the mechanisms which link together motions at different ocean depths, for instance the links between the slow deep ocean dynamics and the

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

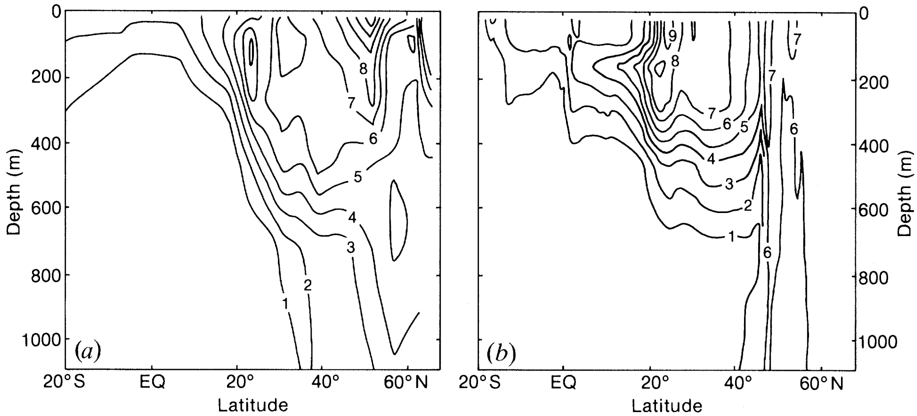


Figure 39. Tritium in the GEOSECS section in the western North Atlantic approximately one decade after the major bomb tests; (*a*) GEOSECS observations (*b*) as predicted by a 12-level model (Sarmiento 1983). In tritium units.

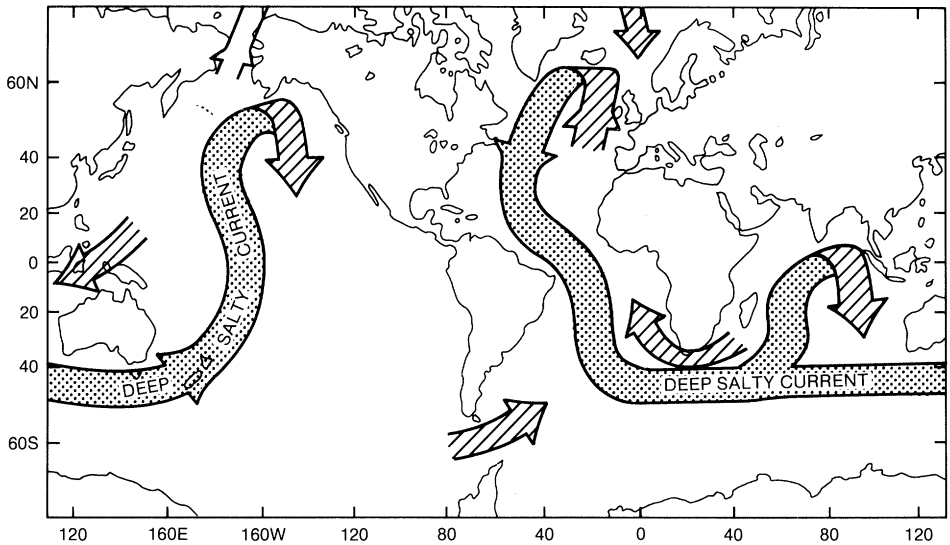


Figure 40. Salt-laden deep water formed in the North Atlantic contributes to a deep ocean circulation that involves all the oceans (after Broecker & Denton 1989).

dynamics of the upper layers. Tracer experiments have been helpful in indicating the regions of the ocean where strong coupling to the deep ocean occurs (figure 39). It seems that there are two important regions of deep water formation (i.e. where cold, salty and hence dense water sinks down to the deep ocean), namely in the North Atlantic ocean between Scandinavia and Greenland and in the region of Antarctica (figures 38 and 40). It is one of the objectives of the World Ocean Circulation Experiment (WOCE) to elucidate the variety of ocean processes (Woods 1985). Both observational and modelling projects are included within the WOCE. One of the British NERC-funded modelling projects, the Fine Resolution Antarctic Model, is already demonstrating how ocean eddies cause the Antarctic circumpolar current to interact with the bottom topography.

A central challenge therefore in climate modelling is that of providing an adequate

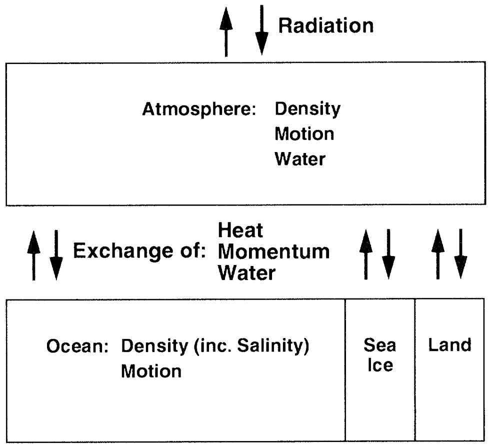


Figure 41. Component elements and parameters of a coupled atmosphere–ocean model including exchange at the atmosphere–ocean interface.

representation of the ocean. Many climate models, in fact most of those used so far for the simulation of future climate, represent the ocean by a simple slab of 50 or 100 m deep which is the approximate depth of the 'mixed layer' of ocean which responds to the seasonal heating and cooling at the ocean surface. In such a model the surface fluxes which describe for instance the exchange of heat between the atmosphere and the ocean can be adjusted by prescribed amounts to make allowance for the transport of heat by ocean currents. The best values for the adjustments can be determined by finding through appropriate numerical experiments those values which give the best representation of the current ocean temperature. For simulating the effect of changing human activities, for instance increasing greenhouse gases, on the climate such a description allows for some of the effects of the ocean and its circulation but it also clearly has its limitations as it takes no account of changes which might occur in ocean transport as a result of changes in the climate.

A more complete description of the climate system is provided by models which couple together an atmospheric model and an oceanic model (figure 41 and Foreman (1989)). At the ocean–atmosphere interface the fluxes from the ocean which influence the atmospheric circulation are those of momentum, heat and water vapour while those from the atmosphere which influence the ocean are those of momentum, heat and liquid water. The model state is very sensitive to the distribution of these fluxes. But the development of algorithms to describe these fluxes with adequate accuracy for use in the coupled models is proving difficult; furthermore, these fluxes are not easy to estimate or measure (Woods 1984). Therefore, if present coupled models are run until steady climate states have been achieved, those states are likely to be somewhat different from the current climate.

There are two possibilities for using such a model for the simulation of climate change. The first is to run the model to its steady state, calling that the model's control climate. The perturbation due to human activities to the model (e.g. an increase in the greenhouse gas concentrations) is then applied and the difference noted between the two model runs. The other method is, by numerical experiment, to make an artificial adjustment to the fluxes in the control run so that the control climate is a good representation of the current climate. The perturbed simulation is then run while keeping the same adjustments to the fluxes. Because the first method

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

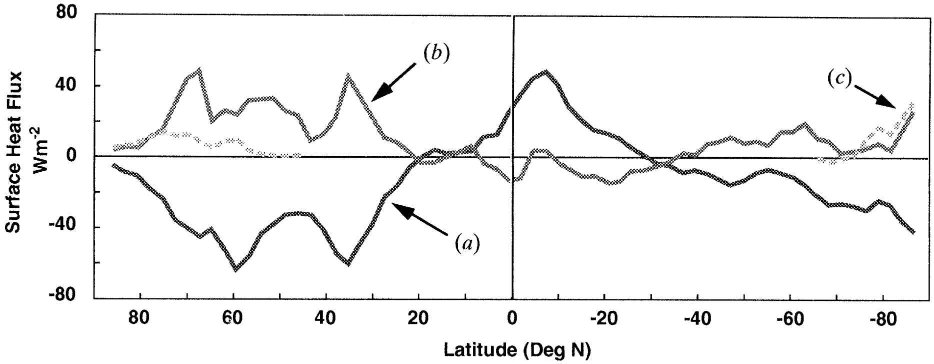


Figure 42. Averages around latitude circles for an atmosphere–ocean coupled model of net heat flux from atmosphere into ocean (*a*), flux correction (*b*) and sea ice contribution to flux correction (*c*) (after Murphy, personal communication).

provides a simulation of current climate that is highly distorted, especially in high latitudes, the second method has been preferred, it being argued that it is important to maintain the climatic state as near as possible to the current climate if credible simulations of perturbed climates are to be achieved.

The flux adjustments are allowed to be seasonally dependent, they also vary a great deal from place to place (figure 42). However, despite the large magnitude of these adjustments (they are similar in many places to the magnitudes of the fluxes themselves), runs of a model including them produce a stable consistent climate (Stouffer *et al.* 1989). We showed earlier in §15 an example of the variability of the climate of such a model which compared favourably with the observed climate variability.

At the Geophysics Fluid Dynamics Laboratory (GFDL) at Princeton experiments with a coupled atmosphere–ocean model have been carried out for a 175-year period (results presented in Bretherton *et al.* (1990)) first with atmospheric CO_2 concentration kept constant at its current level (the control run) and secondly with the CO_2 concentration both increasing and decreasing by 1% (compound) per annum, the increasing CO_2 scenario being similar to the 'business as usual' scenario of the IPCC Scientific Assessment. The most important conclusion from these runs is that the results from them are similar in respect of global averages to those from experiments (often called equilibrium experiments) with models having relatively simple descriptions of the ocean in which runs with doubled atmospheric CO_2 concentration are compared with control runs with the current CO_2 concentration. (figures 43 and 44). But the time-dependent coupled runs differ from the equilibrium runs in two respects. First at any given time the average response can be described as a warming only about 60% as large as in the equilibrium case and with a lag of about 10 years (figure 43). The lag arises from the delaying effect of the oceans, the upper layers of which need time to warm up to match the forcing. The reduced response occurs because water is continually being removed from the ocean's upper layers to form the deep water (figure 38) where it will remain for many centuries and replaced by intermediate water which has not been subject to the warming process. Secondly, the regional distribution of warming is different in the time-dependent runs (figure 44). This is largely because of the further effect of the process of deep water formation that in the two regions where this is occurring, namely, the region around Antarctic south of latitude 50° S and the northern North Atlantic there is

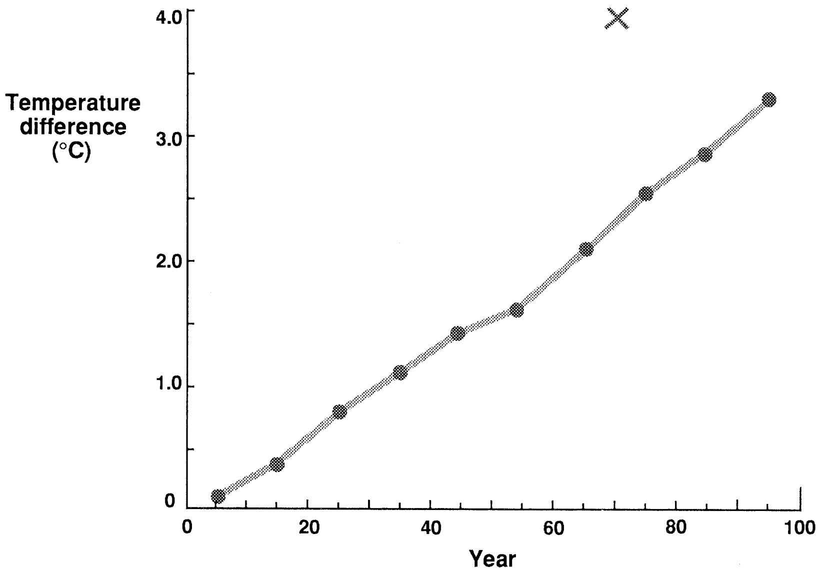


Figure 43. The difference (in kelvins) in globally averaged surface air temperature between the perturbation run (with 1% p.a. increase of atmospheric CO_2) and the control run of the coupled ocean–atmosphere model of Manabe and his colleagues at the Geophysics Fluid Dynamics Laboratory (GFDL), Princeton, U.S.A. For comparison, the equilibrium response of the GFDL atmosphere–mixed-layer-ocean model to a doubling of atmospheric CO_2 is illustrated by X at year 70 when the gradually increasing CO_2 doubles from its initial concentration (after Bretherton *et al.* 1990).

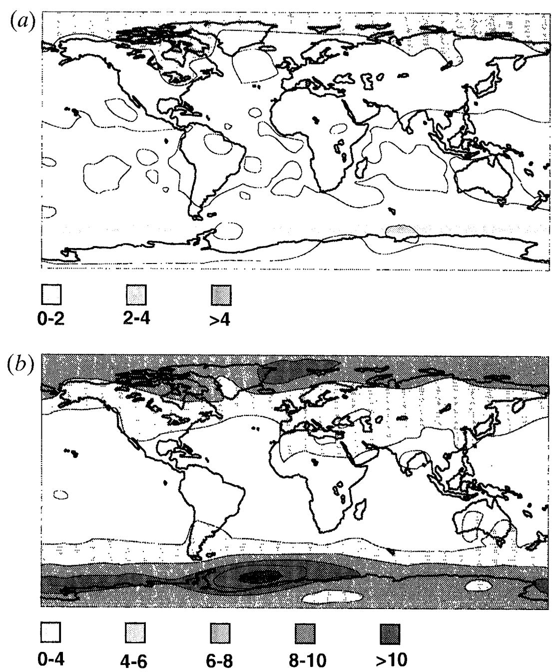


Figure 44. The distribution of the increase in surface air temperature for the coupled ocean–atmosphere model around year 70 (*a*) and the atmosphere–mixed-layer-ocean model for doubled CO_2 (*b*): further details of the models in the captions to figure 43.

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

substantially less warming than elsewhere. Similar experiments with coupled atmosphere–ocean models showing results possessing the same characteristics have recently been carried out at the Meteorological Office (Murphy, personal communication) and at the Max-Planck Institute at Hamburg (Cubasch *et al.* 1991).

Despite the limitations of such atmosphere–ocean coupled models, as illustrated by the large flux adjustments required and by the crude way in which the smaller-scale motions are treated, confidence in these results comes from (1) the consistency of the control simulation and its simulation of the current climate and of climate variability (figure 32) on a range of timescales and (2) the reasonable agreement between the simulated movement of radioactive tracers in the ocean over a 20-year period and the observed movement (figure 39).

18. How can coupled atmosphere–ocean models be improved?

Although it is a significant achievement to have set up coupled atmosphere–ocean models which can describe the current climate and which can be used for future climate prediction, they are as yet at an early stage of development and they suffer from some major deficiencies. Major improvements are required particularly in the ocean component of the coupled models.

The first and most obvious improvement which is becoming possible is for the models to be run with higher resolution particularly in the ocean. With the current typical horizontal resolution of about 300 km, important features in the ocean such as the major boundary currents cannot be resolved with much accuracy. Further, with such coarse resolution very high values of the diffusion coefficients which describe small-scale processes need to be included to maintain numerical stability. As a result the ocean circulation in the model is far too sluggish; as we have seen in §17 large corrections have to be made to the surface fluxes to compensate. Running the models at higher resolution, however, requires much increased computer resources. Because the models are four dimensional (three space and one time dimension), to increase the resolution by a factor of 2 requires 16-fold increase in computer power and to increase by a further factor of 10 (which still may not be adequate) requires a 10^4-fold increase in computer power. Computers having a speed of a teraflop (10^{12} floating point operations per second) or greater which could begin to fulfil this need are currently being planned and should be available before the end of the century.

In addition to better resolution in the models, more understanding of the processes which are included in the models is required. We identified in §16 the need for a better description of atmospheric processes in particular those concerned with cloud–radiation feedback. More understanding is also needed of ocean processes particularly the influence of the ocean eddies as mentioned in §17 which are being addressed by the WOCE. Further, algorithms need to be formulated which can provide a much better description of the fluxes of heat, momentum and water vapour at the ocean–atmosphere interface. Also sea ice formation, dynamics and decay needs much better description in the models.

Climate models require to be tested and validated against a wide range of variables describing the current climate for both atmosphere and ocean. Continuous observations over the globe of the necessary variables are required with adequate accuracy both to provide the earliest possible detection of climate change and to validate the models. For the atmosphere many of the required observations are already made for weather forecasting purposes through the World Weather Watch;

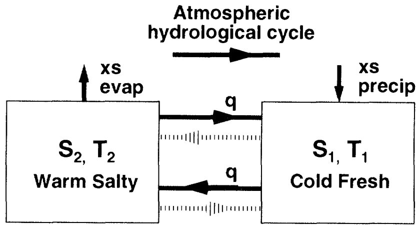
PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

other observations required on a continuous basis and which can be observed from satellites are of the radiation budget at the top of the atmosphere (as was provided by ERBE from 1985 to 1988) and more accurate observations of the atmospheric humidity distribution. For the oceans relevant observations of the ocean surface which can be obtained from space are of surface temperature, of surface stress (through scatterometer observations of radar back scatter from small surface waves), of surface topography (through radar altimetry) which provides information on the surface geostrophic component of ocean circulation and of ocean colour which provides information of biological activity relevant to elucidation of the global carbon cycle. A major challenge is also to provide for observations of the structure and circulation within the ocean which is not accessible to remote sensing from space. For this, to provide for adequate coverage in both space and time, new systems need to be developed. One possibility is the development of acoustic tomography; Munk & Forbes (1989) have recently completed a trial in which signals from an acoustic source located at Heard Island in the southern Indian Ocean were detected at receivers at various locations around the world after an accurately timed delay, so providing continuous information about the ocean temperature structure. Another possibility is the use of instrumented unmanned automated deep submersibles (Autosubs) which can be programmed to provide many profiles within the ocean of parameters such as temperature and salinity (Woods, personal communication). For the international organization of such observations, for which very substantial resources will be required, a Global Climate Observing System (GCOS) (to include a Global Ocean Observing System (GOOS)) is being set up which will not only define what is required but will also coordinate its implementation and ensure the timely and effective dissemination and exchange of data.

Improvements in computer power, in model formulation and in climate observations will all contribute to an ability to give more accurate estimates of the likely climate change associated with human activities, not only in its global coverage but also in the details of the expected regional change.

19. Is the ocean circulation stable?

An interesting feature of the coupled model experiment reported by Stouffer *et al.* (1989) (further details in Bretherton *et al.* 1990) is the weakening, as the CO_2 content of the atmosphere increases, of the rate of formation of deep water in the North Atlantic. About 20 years into the integration, when atmospheric CO_2 has doubled, the source is weaker by about 30%; after 150 years of integration, when the CO_2 has quadrupled, the source has ceased altogether. The reason for this change is the increased precipitation minus evaporation ($P-E$) over the North Atlantic which leads to gradually decreasing salinity of the surface water so that its density eventually becomes inadequate for it to sink to the ocean bottom. This feature was noticed in earlier runs of the GFDL coupled model by Manabe & Stouffer (1988) in which they reported two apparently stable equilibria of the model, one with an active interhemispheric thermohaline circulation originating in the North Atlantic ocean and one in which the interhemispheric circulation was absent. The difference in the climates of the two equilibria was largely confined to the North Atlantic region, but there the climatic differences were large. Manabe & Stouffer (1988) point out that in both of the states the net $P-E$ is small and not adequate to drive any rapid transition from one state to the other.



Present Circulation	Ice Age Circulation
Small $S_2 - S_1$ Large flow q	Large $S_2 - S_1$ Small reversed flow

Figure 45. Schematic of box model of North Atlantic circulation showing transport under the present circulation of warm salty water from low latitudes to higher latitudes with flow in the reverse direction at greater depths. Transport of water through the hydrological cycle in the atmosphere is also illustrated (after R. Wood, personal communication).

That two different stable circulations, driven either by temperature or salinity gradients, might exist in an ocean was shown by Stommel (1961) using a simple box model (figure 45). The current circulation in the Atlantic with a large flow and small salinity difference can be slowed and eventually changed into the other stable circulation with a small reversed flow through the effects of changes in the fluxes in the hydrological cycle. Using such a box model, R. Wood (personal communication) has estimated that an increase in the high-latitude freshwater flux into the ocean of between about 10 and 40% would turn off the current thermohaline circulation, an estimate in line with the model results of Stouffer *et al.* (1989) reported above. Since current models estimate that in a world with doubled CO_2 freshwater flux at high latitudes would increase by about 20%, a major change in the thermohaline circulation as a result of global warming through the increase of greenhouse gases is a possibility that requires further careful study through more sophisticated models.

There is substantial evidence from palaeoclimate records that changes in deep ocean circulation have accompanied climate changes in the past. Broecker *et al.* (1985), Ghil *et al.* (1987) and Broecker & Denton (1989) suggest that such changes in the deep mean circulation could have assisted in some of the relatively rapid changes in climate which occurred during climatic history in particular the rapid transitions from glacial to interglacial conditions, the last of which occurred about 18000 years ago. The palaeoclimate evidence suggests that during glacial periods deep water formation was cut off but that it began again early in the approach to interglacial periods. We can see how this could occur at those times because the possibility of an interaction with the hydrological cycle again exists through the increased evaporation which would accompany the melting of sea ice and the warming of the surface waters through absorption of solar radiation. Because at the cold temperatures of the north Atlantic the density of sea water is more dependent on salinity than on temperature, the effect of continued evaporation is eventually to increase the salinity of the water and hence its density even though it might decrease at first due to the melting of ice. As the density increases deep water formation can restart.

We can see therefore that studies of past records provide information which enables us to speculate regarding the changes which may occur in the future. However, reliable estimates of the influence of ocean circulation on future changes of

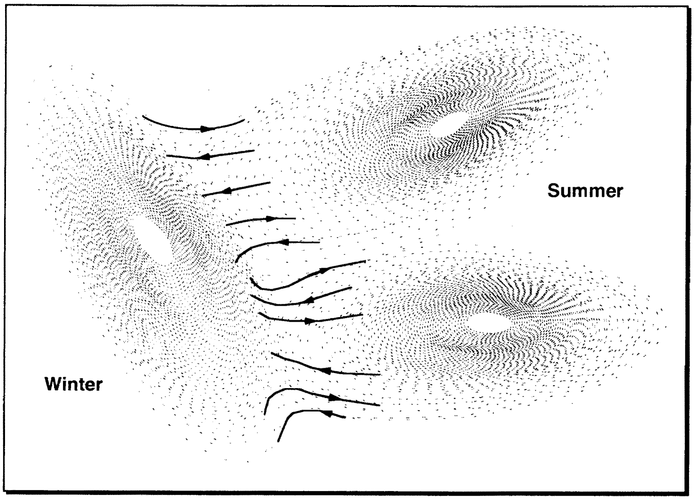


Figure 46. Schematic, using part of the Lorenz attractor to represent a 'climate', of transitions between two different summer 'climates' and one winter 'climate' to illustrate Lorenz's argument that the atmospheric system is unlikely to be intransitive.

climate must await elucidation and satisfactory modelling of the processes which are responsible for the complex feedbacks between the ocean circulation, the radiative régime and the hydrological cycle.

20. How chaotic is the climate?

I finally turn, as I attempt to summarize, to consider the question as to what extent the climate is chaotic. That is, given a description of the present climate and the future variations of forcing, to what extent can we expect to predict its future state? A related question is whether the climate is intransitive, that is whether there are two or more possible 'climates' any one of which once established will persist indefinitely. We consider the problem as it concerns different timescales, firstly on short timescales up to a month or so ahead. We saw in §5 that the evolution of the atmospheric state over a few days is highly sensitive to the initial conditions and that it cannot be expected that much predictability is left so far as the detailed state is concerned after about 14 days. Some small predictability remains for the average atmospheric state for somewhat longer, perhaps for 30 days. But on longer timescales the atmosphere responds to seasonal changes and changes in the state of the lower boundary for instance the sea surface temperature (§10). Otherwise the atmosphere can be considered largely chaotic although still confined to a limited region of climate space which describes the average current climate.

The problem of transitivity over short timescales has been considered by Lorenz (1975) and (1990) who has argued that because seasonal variations are large compared with day-to-day variations, the atmospheric system is unlikely to be intransitive. With the use of a simple model Lorenz (1990) simulated an atmosphere which possessed two different but stable summer 'climates' either one of which if set up in the model in the absence of seasonal variation would persist indefinitely (figure 46). However, when the seasonal cycle is introduced both types of 'summer' occur in a random manner, the intervening 'winter' conditions ensuring that conditions at the beginning of the 'summer' could initiate either type of 'summer'. Because of the large seasonal cycle therefore and the chaotic variations which occur during different seasons, in the absence of changes in forcing external to itself the atmosphere is unlikely to find itself in a completely unexplored part of climate space.

On seasonal or interannual timescales covering a few years, the oceans as well as the atmosphere come into play. We have seen (§10) that given information about the distribution of ocean surface temperature a useful degree of prediction can be achieved of some climate variables, e.g. Sahel precipitation. Further we have also seen (§10) that it is expected that some degree of prediction can be expected for tropical sea-surface temperature anomalies.

For consideration of predictability over timescales of a century or longer, it is necessary to inspect the climate record over the past million years. As we saw in §13 the discovery of the correlation between the Milankovitch cycles in the Earth's orbital parameters and the cycles of climatic change provide strong evidence to substantiate the Earth's orbital variations as the main factor which triggered climate change (figure 28). Some $60 \pm 10\%$ of the variance (Imbrie *et al.* 1984) in the global average temperature record from palaeontological sources over the past million years occurs close to frequencies identified in the Milankovitch theory. The existence of this surprising amount of regularity suggests that the climate system is not strongly chaotic so far as these large changes are concerned but shows a largely predictable response to the radiation forcing. Also over this large range of variation the climate system does not show intransitivity.

The mechanisms of the climatic response to the radiative forcing of the Earth's orbital changes, however, appear to be complex. We have seen that major changes in ocean circulation are involved but with the strong suggestion that these major changes are predictable responses to forcing and are not random changes of the régimes of ocean circulation. Further, there is evidence that biological-climate feedbacks are involved (figure 29). Feedbacks between the growth and decay of ice sheets and other climate parameters are also important. Simple models, for instance the three-parameter model studied by Saltzman (1987) can provide some insights but our ability to model these factors more completely is as yet far from satisfactory (Rind *et al.* 1989). However, the evidence of the compelling regularity of the climatic cycles is a powerful spur to seek out all the mechanisms involved.

Where does this leave us so far as the prediction of climate change in the future is concerned? First, natural variations of weather and climate occur on all timescales which have predictability associated with them over a range comparable with the timescale involved. Day-to-day variations of weather are predictable a few days ahead. Some major seasonal variations have some predictability a season in advance and we may expect some predictability in interannual variations a year or two in advance. Secondly, the response of the climate system to the forcing provided by the increase of greenhouse gases arising from human activities is almost certainly a largely predictable response. However, to make the prediction within acceptable accuracy a great deal more understanding is required especially of the cloud–radiation and ocean circulation feedbacks. The generation of such understanding will entail large efforts in modelling, in the acquisition of observations of all components of the climate system, and in addressing the fundamental science of predictability and so ascertaining to what extent various aspects of climate change can be predicted. Climate change research presents an enormous challenge to the world scientific community. The international scientific programmes underway or planned to address climate change are the largest ever undertaken. The challenge is not only to provide predictions of future change with adequate accuracy but also to apply that information for the benefit of the whole world community.

I have discussed the material in this paper with many colleagues, especially at the Meteorological Office, in particular with David Anderson, Howard Cattle, Mike Cullen, Alan Dickinson, Colin Flood, Chris Folland, Chris Gordon, Raymond Hide, Brian Hoskins, Ian James, Peter Killworth, Sean Milton, John Mitchell, James Murphy, Tim Palmer, David Parker, Peter Read, Peter Rowntree, Tony Slingo, Richard Wood and John Woods, and I gratefully acknowledge their help in the paper's preparation.

References

Atkins, M. J. & Woodage, M. 1985 Observations and data assimilation. *Met. Mag.* **114**, 227–233.

Barnett, T., Graham, N., Cane, M., Zebick, S., Dolan, S., O'Brien, J. & Legler, D. 1988 On the prediction of the El Niño of 1986–87. *Science, Wash.* **241**, 192–196.

Bell, R. S. & Dickinson, A. 1987 The Meteorological Office numerical weather prediction system. Meteorological Office Scientific Paper no. 41.

Berger, A. 1988 Milankovitch theory and climate. *Rev. Geophys.* **26**, 624–657.

Brankovic, C., Palmer, T. N., Molteni, F., Tibaldi, S. & Cubasch, U. 1990 Extended range prediction with ECMWF models: time-lagged ensemble forecasting. *Q. Jl R. met. Soc.* **116**, 867–912.

Bretherton, F. P., Bryan, K. & Woods, J. D. 1990 Time-dependent greenhouse-gas-induced climate change. In *Climate change: The IPCC scientific assessment* (ed. J. T. Houghton, G. J. Jenkins & J. J. Ephraums), pp. 173–193. Cambridge University Press.

Broccoli, A. J. & Manabe, S. 1987 The influence of continental ice, atmospheric CO_2 and land albedo on the climate of the last glacial maximum. *Clim. Dyn.* **1**, 87–99.

Broecker, W. S. & Denton, G. H. 1990 What drives glacial cycles? *Sci. Am.* **262**, 43–50.

Broecker, W. S., Peteet, D. M. & Rind, D. 1985 Does the ocean–atmosphere system have more than one stable mode of operation? *Nature, Lond.* **315**, 21–26.

Broecker, W. S. & Denton, G. H. 1989 The role of ocean–atmosphere reorganisations in glacial cycles. *Geochim. cosmochim. Acta* **53**, 2465–2501.

Canby, T. Y. 1984 El Niño's ill wind. *Natn. geogr. Mag.*, pp. 144–183.

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

Charney, J. G. 1966 The feasibility of a global observation and analysis experiment. *Bull. Am. met. Soc.* **47**, 200–220.

Chen, W. Y. 1989 Estimate of dynamical predictability from NMC DERF experiments. *Mon. Wea. Rev.* **117**, 1227–1236.

Coakley, J. A. Jr, Bernstein, R. L. & Durkee, P. A. 1987 Effect of ship-stack effluents on cloud reflectivity. *Science, Wash.* **237**, 1020–1022.

Croll, J. 1867 On the change in the obliquity of the ecliptic, its influence on the climate of the polar regions and on the level of the sea. *Phil. Mag.* **33**, 426–445.

Cubasch, U., Sausen, R. & Boüttinger, M. 1991 Simulation of the greenhouse effect with coupled ocean–atmosphere models. *Cray Channels*, Winter 1991, pp. 6–9. Cray Research Incorporated, Minnesota, U.S.A.

Cullen, M. J. P. 1983 Solutions to a model of a front forced by deformation. *Q. Jl R. met. Soc.* **109**, 565–573.

Cullen, M. J. P. & Roulstone, I. 1991 A geometric model of the nonlinear equilibration of two-dimensional Eady waves. *J. atmos. Sci.* (In the press.)

Dalcher, A. & Kalnay, E. 1987 Error growth and predictability in operational ECMWF forecasts. *Tellus* A **39**, 474–491.

Eddy, J. A. 1977 Climate and the changing sun. *Climate Change* **1**, 173–190.

Enfield, D. B. 1989 El Niño, past and present. *Rev. Geophys.* **27**, 159–187.

Ferranti, L., Palmer, T. N., Molteni, F. & Klinker, E. 1990 Tropical–extratropical interaction associated with the 30–60 day oscillation and its impact on medium and extended range prediction. *J. atmos. Sci.* **47**, 2177–2199.

Folland, C. K., Karl, T. R. & Vinnikov, K. Ya. 1990 Observed climate variations and change. In *Climate change: the IPCC scientific assessment* (ed. J. T. Houghton, G. J. Jenkins & J. J. Ephraums), pp. 195–238. Cambridge University Press.

Folland, C. K., Owen, J., Ward, M. N. & Colman, A. 1991 Prediction of seasonal rainfall in the Sahel region using empirical and dynamical methods. *J. Forecasting* **10**, 21–56.

Folland, C. K., Palmer, T. N. & Parker, D. E. 1986 Sahel rainfall and world-wide sea temperature 1901–85. *Nature, Lond.* **320**, 602–607.

Foreman, S. J. 1989 Experiences with a coupled global model. *Phil. Trans. R. Soc. Lond.* A **329**, 275–288.

Foukal, P. & Lean, J. 1990 A empirical model of total solar irradiance variation between 1874 and 1988. *Science, Wash.* **247**, 556–558.

Fouquart, V., Buriez, J. C., Herman, M. & Kandel, R. A. 1990 The influence of clouds on radiation: a climate-modelling perspective. *Rev. Geophys.* **28**, 145–166.

Gates, W. L., Rowntree, P. R. & Zeng, Q.-C. 1990 Validation of climate models. In *Climate change: the IPCC scientific assessment* (ed. J. T. Houghton, G. J. Jenkins & J. J. Ephraums), pp. 41–68. Cambridge University Press.

Ghil, M., Mullhaupt, A. & Pestiaux, P. 1987 Deep water formation and quaternary glaciations. *Climate Dynam.* **2**, 1–10.

Gordon, C. & Corry, R. A. 1991 An ocean model simulation of the seasonal cycle in the tropical Pacific using climatological and modelled surface forcing. *J. geophys. Res.* **96**, 847–864.

Hansen, J. E. & Lacis, A. A. 1990 Sun and dust versus greenhouse gases; an assessment of their relative roles in global climate change. *Nature, Lond.* **346**, 713–719.

Harrison, E. F., Minnis, P., Barkstrom, B. R., Ramanathan, V., Cess, R. D. & Gibson, G. G. 1990 Seasonal variation of cloud radiative forcing derived from the Earth Radiation Budget Experiment. *J. geophys. Res.* **95**, 18687–18703.

Hartmann, D. L. & Doelling, D. 1991 On the net radiation effectiveness of clouds. *J. geophys. Res.* **96**, 869–891.

Hollingsworth, A., Cubasch, U., Tibaldi, S., Brankovic, C., Palmer, T. N. & Campbell, L. 1987 Mid-latitude atmospheric prediction on timescales of 10–30 days. In *atmospheric and oceanic variability* (ed. H. Cattle). *R. met. Soc. Lond.* 117–151.

Hoskins, B. J. 1982 The mathematical theory of frontogenesis. *A. Rev. Fluid. Mech.* **14**, 131–151.

Houghton, J. T. 1986 *The physics of atmospheres*, 2nd edn, pp. 271. Cambridge University Press.

Hulme, M., Biot, Y., Borton, J., Buchanan-Smith, M., Davies, S., Folland, C. K., Nicholls, N., Seddon, D. & Ward, N. 1991 Seasonal rainfall forecasting for Africa: Part 1. Current status and future developments. *Int. J. environ. Stud.* A **39**. (In the press.)

Imbrie, J., Hays, J., Martinson, D. G., McIntyre, A., Mix, A.-C., Morley, J. J., Pisias, N. G., Prell, W. L. & Shackleton, N. J. 1984 The orbital theory of Pleistocene climate: support from a revised chronology of the marine ^{18}O record. In *Milankovitch and climate* (ed. A. Berger *et al.*), pp. 269–305. Hingham, Massachusetts: D. Reidel.

Intergovernmental Panel on Climate Change 1990 *Climate change: the IPCC scientific assessment* (ed. J. T. Houghton, G. J. Jenkins & J. J. Ephraums), pp. 365. Cambridge University Press.

James, I. N. & James, P. M. 1991 Ultra low frequency variability of the flow in a simple atmospheric circulation model. (Submitted.)

Kiehl, J. T. & Ramanathan, V. 1990 Comparison of cloud forcing derived from the earth radiation budget experiment with that simulated by the NCAR Community Climate Model.

Kistler, R., Kalnay, E. & Tracton, M. S. 1988 Forecasting agreement, persistence and forecasting skill. *Eighth Conference on Numerical Weather Prediction, Baltimore, U.S.A.* American Meteorological Society.

Lau, N.-C. 1981 A diagnostic study of recurrent meteorological anomalies appearing in a 15 year simulation with a GFDL general circulation model. *Mon. Wea. Rev.* **109**, 2287–2311.

Lau, N.-C. 1985 Modelling the seasonal dependence of the atmospheric response to observed El Niños in 1962–76. *Mon. Wea. Rev.* **13**, 1970–1996.

Leith, C. E. 1974 Theoretical skill of Monte Carlo forecasting. *Mon. Wea. Rev.* **102**, 409–418.

Lorenz, E. 1963 Deterministic non-periodic flow. *J. atmos. Sci.* **20**, 130–141.

Lorenz, E. 1975 Nondeterministic theories of climate change. *Quat. Res.* **6**, 495–506.

Lorenz, E. 1982 Atmospheric predictability experiments with a large numerical model. *Tellus* **34**, 505–513.

Lorenz, E. N. 1990 Can chaos and intransitivity lead to interannual variability? *Tellus* A **42**, 378–389.

Lorius, C., Jouzel, J., Raynaud, D., Hansen, J. & Le Treut, H. 1990 The ice-core record: climate sensitivity and future greenhouse warming. *Nature, Lond.* **347**, 139–145.

Manabe, S. & Broccoli, A. J. 1985 A comparison of climate model sensitivity with data from the last glacial maximum. *J. atmos. Sci.* **42**, 2643–2651.

Manabe, S. & Stouffer, R. J. 1988 Two stable equilibria of a coupled ocean–atmosphere model. *J. Clim.* **1**, 841–866.

McCreary, J. P. 1983 A model of tropical ocean–atmospheric interaction. *Mon. Wea. Rev.* **11**, 370–387.

McCreary, J. P. Jr. & Anderson, D. L. T. 1991 An overview of coupled ocean–atmosphere models of El Niño and the Southern oscillation. *J. geophys. Res.* **96**, 3125–3150.

McIntyre, M. E. 1990 Middle atmosphere dynamics and transport: some current challenges to our understanding. In *Dynamics, transport and photochemistry in the middle atmosphere of the Southern Hemisphere* (*Proc. San Francisco NATO Workshop*) (ed. A. O'Neill), pp. 1–18. Dordrecht: Kluwer.

Meteorological Office 1987 The Storm 15/16 October 1987. Meteorological Office Report, Bracknell.

Milankovitch, M. M. 1920 *Theorie mathematique des phenomenes thermiques produits par la radiation solaire.* Academie Yougoslave des Sciences et des Arts de Zagreb. Paris: Gauthier-Villars.

Mitchell, J. F. B., Senior, C. A. & Ingram, W. J. 1989 CO_2 climate: a missing feedback. *Nature, Lond.* **341**, 132–134.

Miyakoda, K., Gordon, T., Caverley, R., Stern, W., Sirutis, J. & Bourke, W. 1983 Simulation of a blocking event in January 1977. *Mon. Wea. Rev.* **111**, 846–869.

Molteni, F., Mureau, R. & Palmer, T. N. 1991 Atmospheric instability and ensemble weather prediction. Tech. Memo no. 17, ECMWF, Reading, U.K.

Munk, W. H. & Forbes, A. M. G. 1989 Global warming: an acoustic measure? *J. Phys. Oceanogr.* **19**, 1765–1778.

Murphy, J. M. 1988 The impact of ensemble forecasts on predictability. *Q. Jl R. met. Soc.* **114**, 463–493.

Murphy, J. M. 1990 Assessment of the practical utility of extended range ensemble forecasts. *Q. Jl R. met. Soc.* **116**, 89–125.

Nakamura, N. & Held, I. M. 1989 Non-linear equilibration of two-dimensional Eady waves. *J. atmos. Sci.* **46**, 3055–3064.

Nicholls, N. 1989 Sea surface temperature and Australian winter rainfall. *J. Clim.* **2**, 965–973.

Nicholson, S. E. 1985 Sub-Saharan rainfall 1981–84. *J. Clim. appl. Met.* **24**, 1388–1399.

O'Lenic, E. A. & Livezey, R. E. 1988 Relationships between systematic errors in medium range numerical forecasts and some of the principal modes of low frequency variability in the Northern Hemisphere 700 mb circulation. *Mon. Wea. Rev.* **117**, 1262–1280.

Oort, A. & Vonder Haar, T. H. 1976 On the observed cycles in the ocean–atmosphere heat balance over the Northern hemisphere. *J. Phys. Oceanogr.* **6**, 781–800.

Palmer, T. N. 1988 Medium and extended range predictability and stability of the Pacific/North American mode. *Q. Jl R. met. Soc.* **114**, 691–713.

Palmer, T. N. & Tibaldi, S. 1988 On the prediction of forecast skill. *Mon. Wea. Rev.* **116**, 2453–2480.

Palmer, T. N., Brankovic, C., Molteni, F. & Tibaldi, S. 1990 Extended range predictions with

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

ECMWF models: interannual variability in operational model integrations. *Q. Jl R. met. Soc.* **116**, 799–834.

Palmer, T. N., Schutts, G. J. & Swinbank, R. 1986 Alleviation of a systematic westerly bias in general circulation and numerical prediction models through an orographic gravity wave drag parametrization. *Q. Jl R. met. Soc.* **112**, 1001–1039.

Philander, S. G. H., Lau, N. C., Pacanowski, R. C. & North, M. J. 1989 Two different simulations of the southern oscillation and El Niño with coupled ocean–atmosphere general circulation modes. *Phil. Trans. R. Soc. Lond.* A **329**, 167–178.

Ramanathan, V., Cess, R. D., Harrison, E. F., Minnis, P., Barkstrom, B. R., Ahmad, E. & Hartmann, D. 1989 Cloud-radiative forcing and climate: results from the Earth Radiation Budget Experiment. *Science, Wash.* **243**, 57–63.

Raval, A. & Ramanathan, V. 1989 Observational determination of the greenhouse effect. *Nature, Lond.* **342**, 758–761.

Read, P. L. 1991 Applications of chaos to meteorology and climate. In *Chaos and related non-linear phenomena* (ed. T. Mullin). Oxford University Press.

Rind, D., Peteet, D. & Kukla, G. 1989 Can Milankovitch orbital variations initiate the growth of ice sheets in a general circulation model. *J. geophys. Res.* **94**, 12851–12871.

Saltzman, B. 1987 Modelling the ^{18}O-derived record of quaternary climate changes with low order dynamical systems. In *Irreversible phenomena and dynamical systems analysis in geosciences* (ed. C. Nicolis & G. Nicolis), pp. 355–380. R. D. Reidel.

Sarmiento, J. L. 1983 A simulation of bomb-tritium entry into the Atlantic Ocean. *J. Phys. Oceanogr.* **13**, 1924–1939.

Shine, K. P., Derwent, R. G., Wuebbles, D. T. & Morcrette, J. J. 1990 Radiative forcing of climate. In *Climate change: The IPCC scientific assessment* (ed. J. T. Houghton, G. J. Jenkins & J. J. Ephraums), pp. 41–68. Cambridge University Press.

Shukla, J. & Fennessy, M. J. 1988 Prediction of time-mean atmospheric circulation and rainfall: influence of Pacific sea surface temperature anomaly. *J. atmos. Sci.* **45**, 9–28.

Shutts, G. J. 1990 Dynamical aspects of the October Storm 1987: a study of a successful fine-mesh simulation. *Q. Jl R. met. Soc.* **116**, 1315–1348.

Sirutis, J. & Miyakoda, K. 1990 Subgrid scale physics in 1-month forecasts. Part II. Systematic error and blocking forecasts. *Mon. Wea. Rev.* **118**, 1065–1081.

Slingo, A. 1990 Sensitivity of the earth's radiation budget to changes in low clouds. *Nature, Lond.* **343**, 49–51.

Slingo, A. & Jones, A. 1991 Cancellation of ERBE cloud radiative forcing components over tropical convection. (Submitted.)

Smagorinsky, J. 1969 Problems and promise of deterministic extended range forecasting. *Bull. Am. met. Soc.* **50**, 285–311.

Smith, L. D. & Vonder Haar, T. H. 1991 Cloud–radiation interactions in a general circulation model: impact upon the planetary radiation balance. *J. geophys. Res.* **96**, 893–914.

Stommel, H. 1961 Thermohaline circulation with two stable regimes of flow. *Tellus* **13**, 224–230.

Stouffer, R. J., Manabe, S. & Bryan, K. 1989 Interhemispheric asymmetry in climate response to a gradual increase of atmospheric CO_2. *Nature, Lond.* **342**, 660–662.

Thompson, P. D. 1957 Uncertainty of initial state as a factor in the predictability of large scale atmospheric flow problems. *Tellus* **9**, 275–295.

Tibaldi, S., Palmer, T. N., Brankovic, C. & Cubasch, U. 1990 Extended range predictions with ECMWF models; influence of horizontal resolution on systematic error and forecast skill. *Q. Jl R. met. Soc.* **116**, 835–866.

Tibaldi, S. & Molteni, F. 1990 On the operational predictability of blocking. *Tellus* A **42**, 343–365.

Twomey, S. A., Piepgrass, M. & Wolfe, T. L. 1984 An assessment of the impact of pollution on global cloud albedo. *Tellus*, B **36**, 356–366.

Ward, M. N. & Folland, C. K. 1991 Prediction of seasonal rainfall in the North of Brazil using eigenvalues of surface temperature. *Int. J. Climatology* **11**. (In the press.)

Wigley, T. M. L. & Kelly, P. M. 1990 Holocene climatic change, ^{14}C wiggles and variations in solar irradiance. In *The Earth's climate and variability of the Sun over recent millenia: geophysical,* *astronomical and archaeological aspects*. (*Phil. Trans. R. Soc. Lond.* A **330**), pp. 547–560. London: The Royal Society.

Willson, R. C. & Hudson, H. S. 1988 Solarluminosity variations in solar cycle 21. *Nature, Lond.* **332**, 810–812.

Woods, J. D. 1984 The upper ocean and air sea interaction in global climate. In *The global climate* (ed. J. T. Houghton), pp. 141–187. Cambridge University Press.

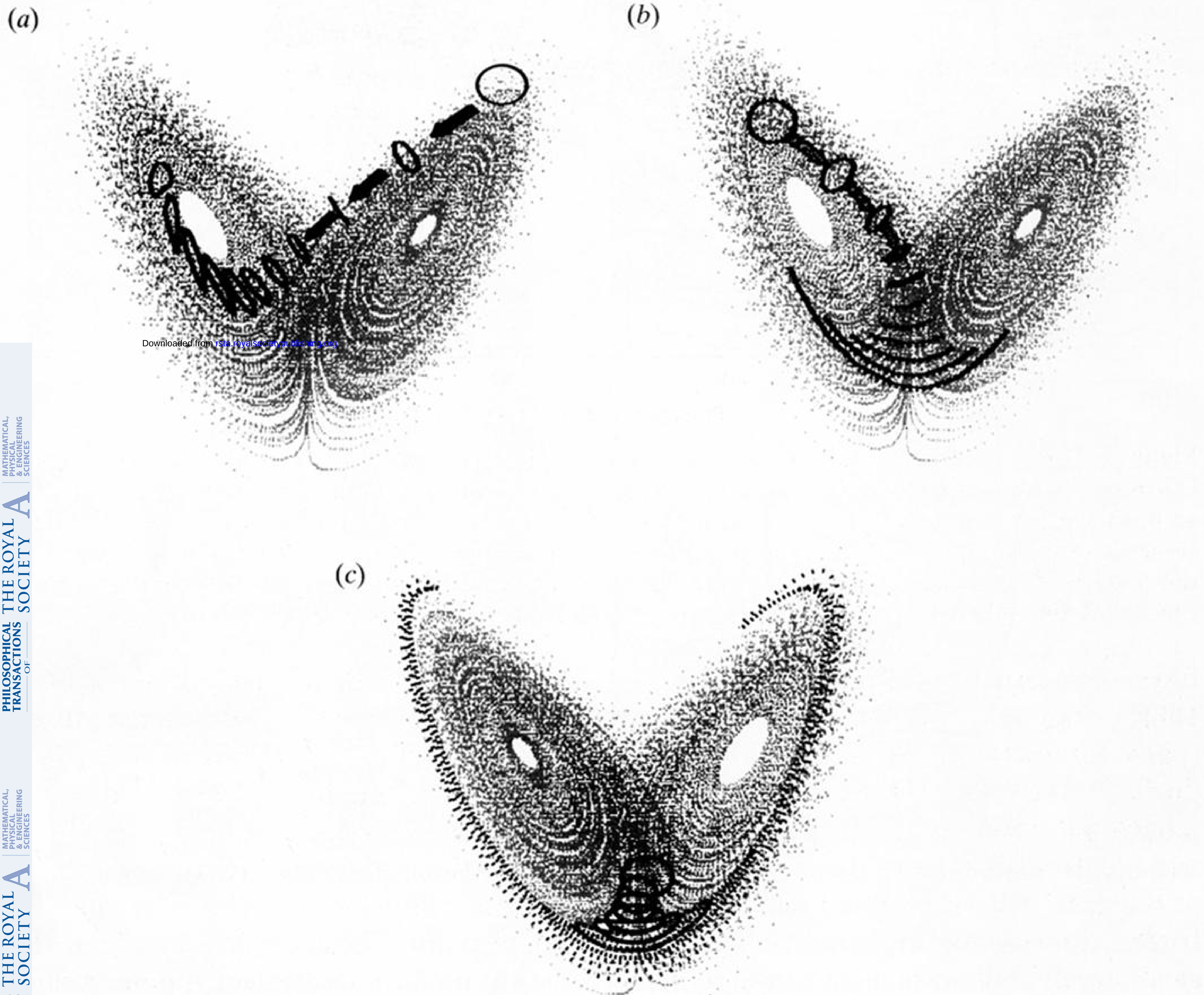
Woods, J. D. 1985 The World Ocean Circulation Experiment. *Nature, Lond.* **314**, 501–511.

Wunsch, C. 1984 The ocean circulation in climate. In *The global climate* (ed. J. T. Houghton), pp. 189–203. Cambridge University Press.

Wyrtki, K. & Kilansky, B. 1984 Mean water and current structure during the Hawaii-to-Tahiti shuttle experiment. *J. Phys. Oceanogr.* **14**, 242–254.

Lecture delivered 13 March 1991; typescript received 24 July 1991

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES



igure 6. Cross sections across the Lorenz attractor showing the development with time of solutions
) the equations beginning from different ensembles of initial states. Parts (a)–(c) illustrate
ncreasing divergence of solutions (after T. N. Palmer, personal communication).

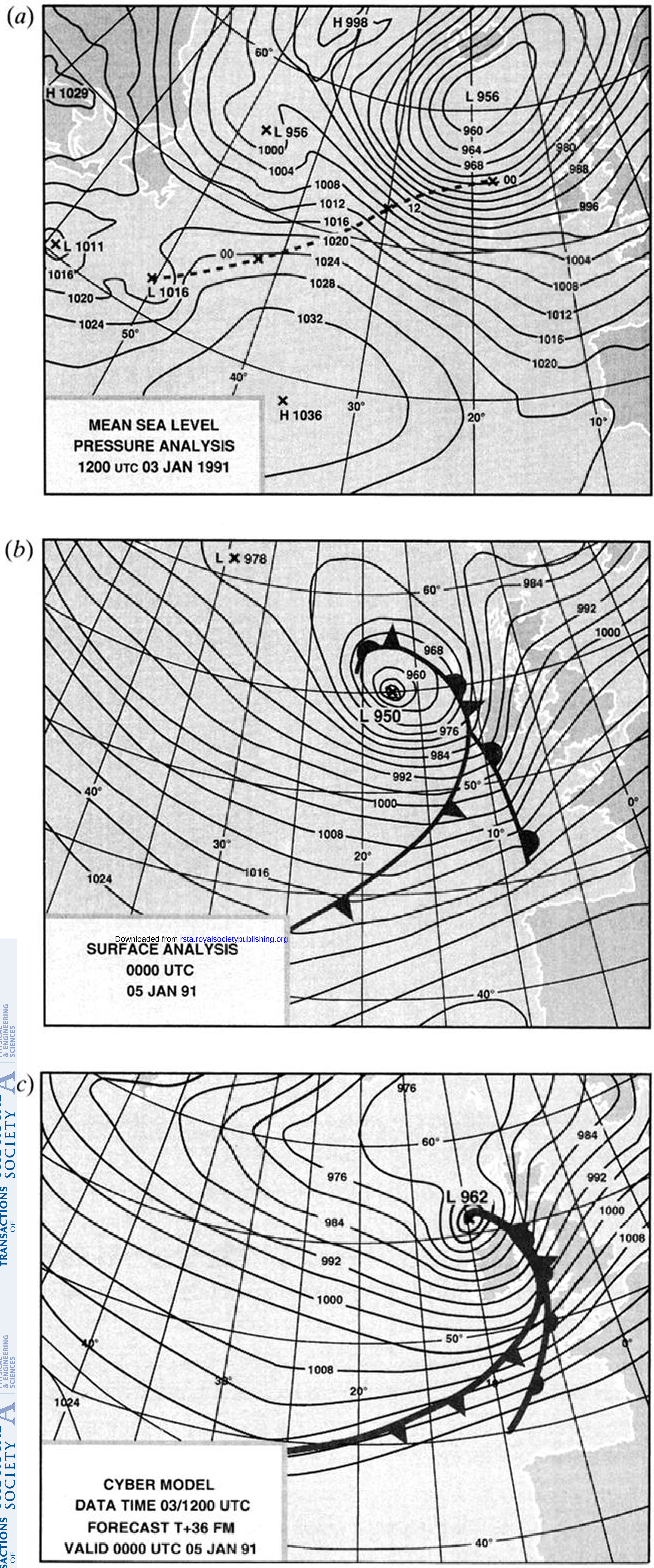
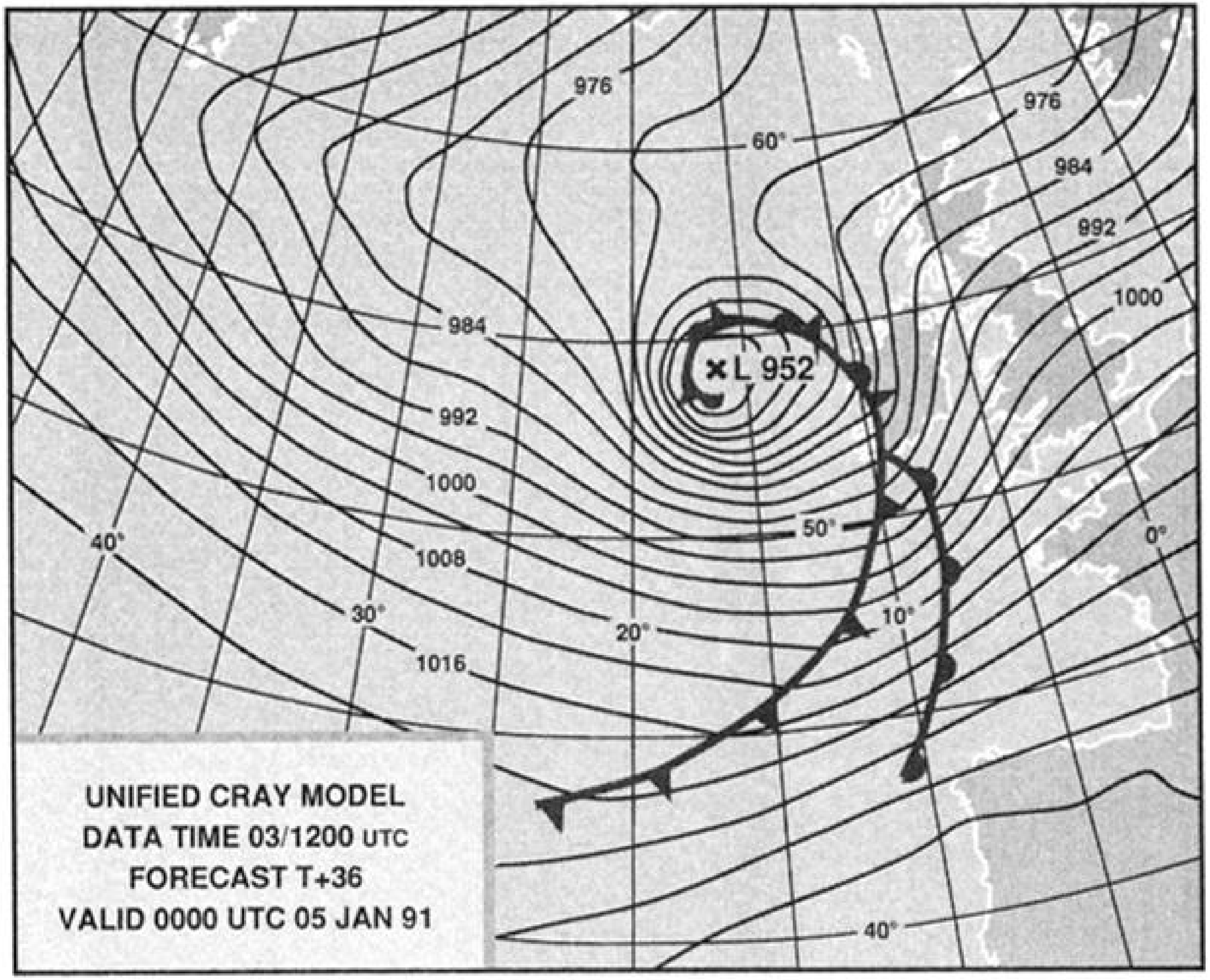


Figure 12*a*–*c*. For description see opposite.

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

(*d*)



'igure 12. Illustrating improvement in 36 h forecast made on 3 January 1991 resulting from ıcreased model resolution. (*a*), (*b*) Surface pressure analyses for initial and verification times. (*c*) 'orecast with the Cyber model (150 km resolution and 15 levels) and (*d*) with the Cray model)0 km resolution and 20 levels).

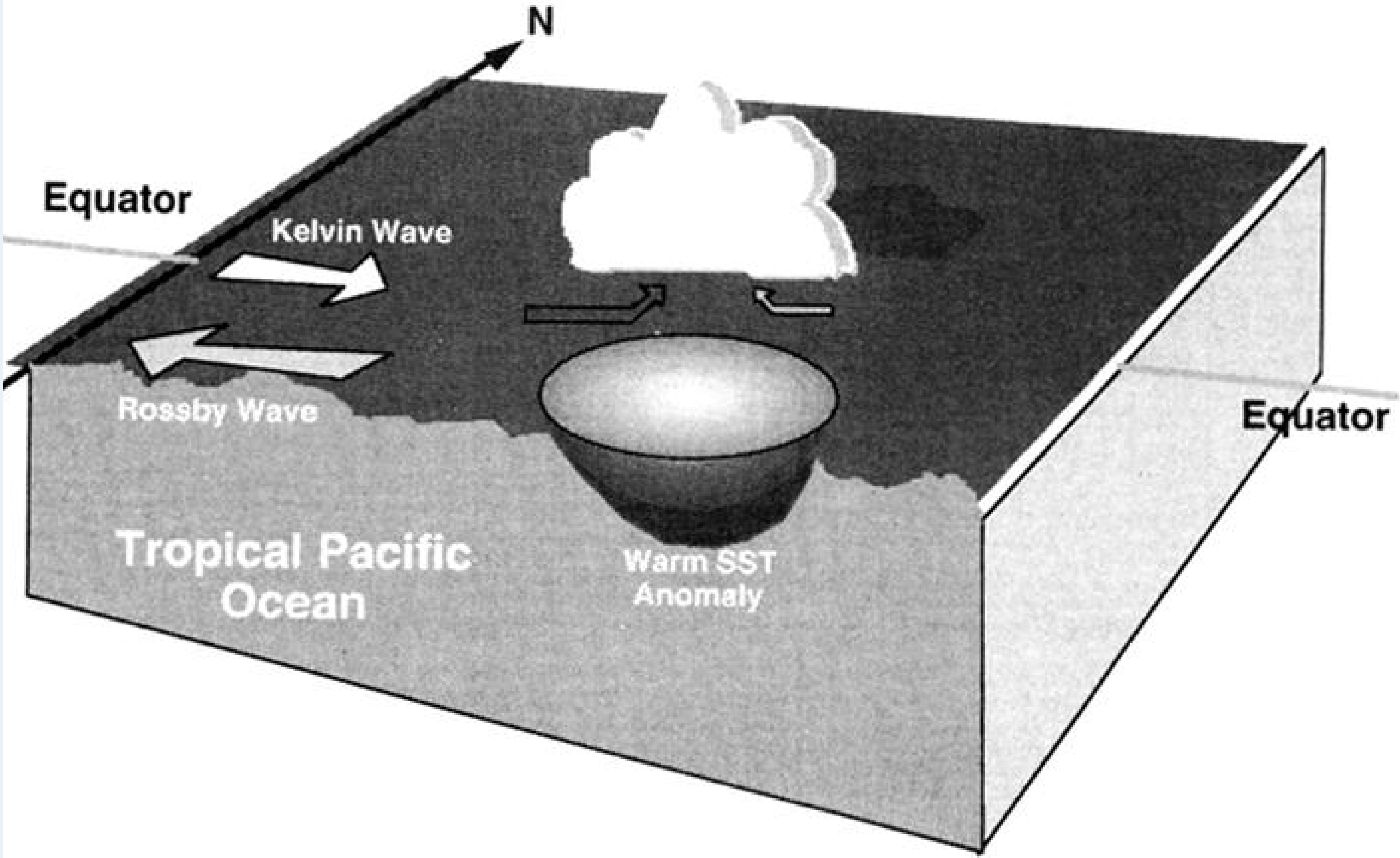
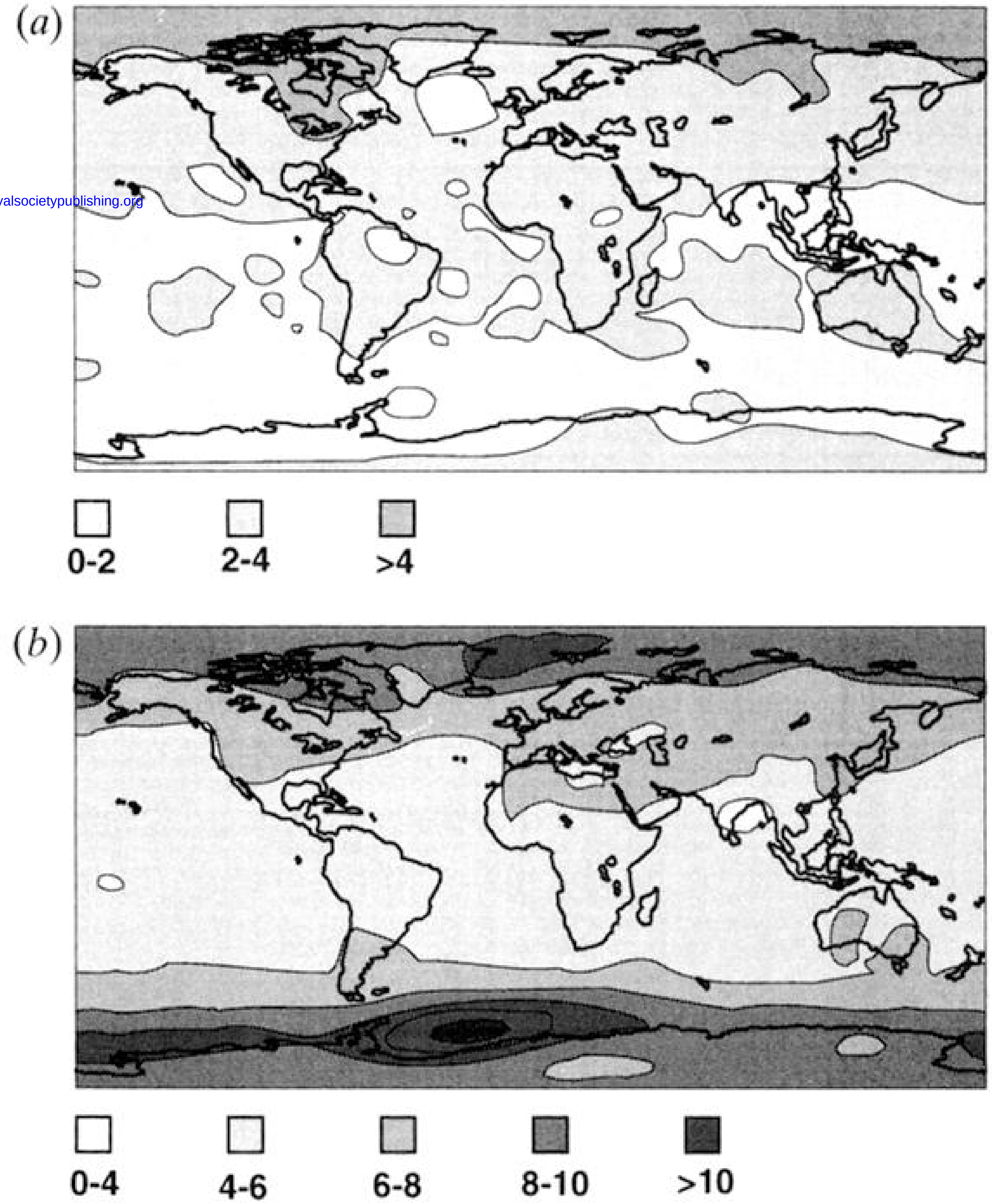


Figure 21. Schematic to illustrate El Niño oscillation.



igure 44. The distribution of the increase in surface air temperature for the coupled ocean–tmosphere model around year 70 (*a*) and the atmosphere–mixed-layer-ocean model for doubled O_2 (*b*): further details of the models in the captions to figure 43.

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES