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FOR CLIMATE PREDICTION AND RESEARCH

Climate Change Prediction

by

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Abstract.

A brief description of climate processes is given. Numerical models of climate are described and an assessment of their ability to simulate climate and climate change is given. Results from simulations with increased atmospheric CO_2 concentrations are given, including the temporal evolution of the warming and estimates of the geographical distribution of changes. The main sources of uncertainty are listed , and possible shortcomings in current research programs are identified.

1 Introduction: The climate system.

In order to predict changes in climate, one must first identify and understand the various components of the climate system . These include the atmosphere, the oceans, the cryosphere (land-ice and sea-ice), the biosphere and (over geological timescales) the geosphere. The fundamental process driving climate is radiation. The climate system is heating by solar radiation, which is generally strongest in low latitudes, and cooled by long (thermal or infra red) radiation to space, which is more uniformly distributed with latitude. The resulting temperature contrast between low and high latitudes drives atmospheric and oceanic motions which transport heat polewards and thus tend to reduce the equator-to-pole temperature gradient. The resulting circulations and their interactions with orography and the biosphere determine the earth's climate.

Changes in the radiative heating of the earth will produce changes in it's climate. Changes in solar heating can arise from external factors such as variations in incident solar radiation due to changes in the earth's orbit, or in the earth's reflectivity, for example, due to addition of volcanic aerosols to the atmosphere, and to changes in the reflectivity of the surface such as that caused by deforestation. The longwave cooling can be modified by changes in atmospheric composition, notably by changes in the concentration of greenhouse gases. Other non-radiative factors such as changes in orography can also affect climate, but these occur on geological timescales, and so will not be considered further here. Note that changes in climate will themselves produce changes in the earth's radiation budget: these are known as radiative feedbacks and are discussed later in this section.

In this paper, we are primarily interested in predicting changes in climates over the next century or so, hence it may be possible to neglect changes in some of the more slowly varying components of the system. The atmosphere and land surface, including seasonal snow cover, adjust to changes in heating on timescales of a year or less. The seasonal mixed layer of the ocean (typically up to 100 metres deep) and sea-ice respond over periods of up to a decade or so. The oceanic warm water sphere (down to the permanent thermocline, typically at half a kilometre) has a thermal response time of several decades, whereas the deep ocean, extending to several kilometres, has a timescale of centuries to millenia. The major ice-sheets vary over millennia, so apart from the substantial contributions from changes in accumulation and melting on sea-level, they can be regarded as fixed in the present context.

In predicting climate change, it is generally assumed that for each specification of the factors affecting climate (solar heating, atmospheric conditions, orography), the earth's climate will move to a long term steady state, known as the equilibrium climate, in which there is no net heating of the system. Note that this does not preclude considerable variability on interannual or even interdecadal timescales, but there are no longer term trends. If the factors affecting climate are changed (for example, by increasing the concentration of greenhouse gases), then the system will adjust slowly towards a new equilibrium state. The transition period is sometimes referred to as a "transient" or "time-dependent" climate change: the difference between the initial and final states is the "equilibrium climate change".

One possible method of predicting climate changes is to look for periods in the past when the factors affecting climate (solar parameters, greenhouse gas concentrations) were similar to those expected in the near future. One could then use reconstructions of the relevant past climates as a forecast for the future climate. Unfortunately, in the recent geological past, there are no close analogies to an increase in greenhouse gases, and so this method cannot provide reliable forecasts of climate over the next century or so.

The only alternative to the climate analogy approach is the use of physical-mathematical models of climate. Among such models,the most highly developed are general circulation models (GCMs). These are numerical models in which the equations of classical physics (including the laws of motion, and requirements for conservation of heat and mass) are solved in a three-dimensional grid, with a horizontal spacing of 250 to 800 kms, depending on the model concerned.

Many processes (for example, those associated with cloud or vertical mixing in the oceans) occur on much smaller scales than can be resolved by the model grid, and so are represented in a simplified or idealized way (parameterisations). These may be based on approximations of the underlying equations , data from observational studies or laboratory experiments, results from numerical experiments using a finer grid, or as a last resort, sensitivity experiments using the GCM itself. In a climate model, an atmospheric component (essentially the same as a weather prediction model) is coupled to a model of the oceans and sea-ice, which may be equally complex.

Climate forecasts are derived in a different way from weather forecasts. A weather prediction model gives a description of the atmosphere's state up to 10 days or so ahead, starting from a detailed description of an initial state of the atmosphere at a given time. Such forecasts describe the movement and development of large weather systems, though they cannot represent very small scale phenomena; for example, individual shower clouds.

To make a climate forecast, the climate model is first run for a few (simulated) decades. The statistics of the model's output will be a description of the model's simulated climate which, if the model is a good one, will bear a close resemblance to the climate of the real atmosphere and ocean. The above exercise is then repeated with increasing concentrations of the greenhouse gases in the model. The differences between the statistics of the two simulations (for example in mean temperature and interannual variability) provide an estimate of the accompanying climate change.

In this paper, reference will be made to atmospheric models run with prescribed sea surface temperature (referred to henceforth as AGCMs), atmospheric models coupled to an oceanic mixed layer, with or without allowance for ocean heat transport (A/MLMs) and atmospheric models coupled to full dynamical ocean models (CGCMs).

As alluded to above, changes in climate can lead to changes in the components of the earth's radiation budget which may be amplified (positive feedback) or reduced (negative feedback) . The simulated strength of feedbacks (such as those due to cloud) vary from model to model. The relative strengths of the radiative feedbacks in different models have been analysed by substituting the globally averaged changes in the simple energy balance equation.

$$\Delta Ts \quad = \quad \Delta Q/LAMBDA$$

where ΔQ is the applied radiative perturbation (Wm^{-2}), ΔTs is the equilibrium change in globally averaged surface temperature, and LAMBDA is the climate sensitivity parameter ($Wm^{-2} K^{-1}$). For example, doubling CO_2 produces an increase of 4.4 Wm^{-2} in the radiative heating of the troposphere and surface. If the system responded as a radiative black body, this would produce an equilibrium warming of 1.2^o C. Warming the atmosphere leads to an increase in atmospheric water content - water vapour is also a greenhouse gas, and estimates based on both observations (Raval and Ramanathan,1989) and models indicate that produces a positive feedback, increasing the warming to 1.7^o C.

Snow and sea-ice reflect solar radiation back to space. The global warming associated with increases in greenhouse gases would reduce the aerial extent of snow and ice, increasing solar absorption, leading to a further increase in temperature. The magnitude of this feedback is generally much smaller than that due to water vapour, though estimates of the strength vary from model to model.

Clouds cool climate through reflecting solar radiation back to space, and warm climate through their greenhouse effect. In our present climate, it appears that the solar effect dominates. Any change in cloud amount, height or cloud radiative properties will alter the net effect of clouds on the earth's radiation budget.

Numerical studies suggest that doubling CO_2 amounts would lead to an equilibrium warming of 2 to 5^o C , much of the uncertainty being associated with uncertainties in the strength of cloud radiation feedbacks and associated processes (for example, Mitchell et al,1989, Cess et al, 1989).

Some scientists have suggested that the climate sensitivity is much smaller than indicated by models. A few, including Idso, and Newell and Dopplick considered the energy balance at the surface only or neglected the water vapour feedback (see Luther and Cess,1985, for references and a detailed explanation why such approaches are misleading). More recently, Lindzen (1990) and Ellsaesser (1989) have questioned the strength of the water vapour feedback in climate models. Lindzen argues that cumulus convection "short circuits " much of the potential water vapour feedback by transporting heat from the boundary layer to the upper troposphere where it is more readily radiated to space. This process is undoubtedly represented in current general circulation models and provides, as expected, a strong negative feedback through changes in lapse rate.(see for example Schlesinger and Mitchell, 1987). Lindzen and Ellsaesser also argue that as convection penetrates to higher levels in a warmer climate, the compensating subsidence will start from higher and therefore colder and drier levels. This would lead to a reduction in the absolute humidity in the upper troposphere in the decending branch of the Hadley circulation, and hence a local negative water vapour feedback. Since Lindzen has yet not published the details of his argument, it is not possible to assess it quantitatively. Nevertheless, this mechanism is also included in current models (see for example Gregory and Rowntree,1990). Indeed, on doubling CO_2 concentrations, the Met Office high resolution model produces decreases in absolute humidity in this region (Senior appears that other processes including the detrainment of warmer, moister air and lateral mixing of moistened air from the inner tropics and mid-latitudes reduce the extent of the drying through deeper subsidence. The increased drying through subsidence appears to be responsible for the decreases in relative humidity in mid-latitudes and the tropics which leads to the reduction in cloud in the upper troposphere and a strong positive cloud feedback found in many models (Mitchell and Ingram,1990). Thus, if models have exaggerated the positive water vapour feedback, it is also likely that they have underestimated the strength of the positive feedback associate with reductions in cloud amount. In summary, although tropical convection remains one of the major sources of uncertainty, it is unlikely that errors in the parametrization of convection lead to a gross overestimate of climate sensitivity (Even removing the water vapour feedback completely and globally in a model with a sensitivity of 2.5° C due to doubling CO_2 - the IPCC "best guess" - would only reduce the equilibrium warming to about 1.4° C)

2. Validation of numerical climate models

In this section, we attempt to answer the question "To what extent should one believe results from climate models?" First, the models are based on a firm physical basis with a minimum of adjustable parameters, as discussed in the previous section. Second, they show considerable skill in reproducing the large-scale features of current climate. Third, they have been shown to be capable of reproducing many features of contemporary climate change, and some of the main features in more recent paleoclimate. These last two points are expanded on below.

2.1 Simulation of present climate

The simplest way of validating a climate model is to prescribe present day "boundary conditions" and compare the long term statistics of the resulting simulation with observational climatologies. This shows that AGCMs and A/MLMs have considerable skill in the portrayal of the large scale distribution of pressure, temperature, wind and precipitation in both summer and winter, although this success is due in part to the constraints on sea surface temperature and sea-ice. There has been a general reduction in the errors in more recent AGCMS as a result of increased horizontal resolution, improvements to the parameterisation of convection, cloudiness and surface processes and the introduction of parameterisations of gravity wave drag. For example, the three most recent simulations considered in the IPCC Working Group I Report show a marked improvement in the simulation of the depth and position of the Antarctic circumpolar surface pressure trough (Figure 1).

Changes in variability of climate are as important as changes in mean climate in the assessment of climate impacts. Hence, the ability of models to simulate the variability of current climate should also be assessed. The daily and interannual variability of temperature and precipitation have been examined but only to a limited extent. There is evidence that variability is overestimated in some models, especially in summer. The daily variability of sea-level pressure can be well simulated , but the eddy kinetic energy in the upper troposphere (indication of the variability of the flow) tends to be underestimated. The level of interannual variability of global mean surface temperature in CGCMs (for example, Figure 2) is comparable to that observed over similar timescales if allowances is made for the estimated trend in the observed..

On regional scales, there are significant errors in all models. A validation of A/MLMs for five selected regions (typically 4 x 10^6 Km^2) showed errors in area average surface temperature of 2 to 3^o C (IPCC,1990). This is small compared with the average seasonal range of temperature of 15^o C. Errors in mean precipitation for the same five regions ranged from 20 to 50% of the observed average. All the recent models reproduce the northern summer monsoon rainfall maximum over South East Asia (Figure 3), but in the example shown, the mean rainfall over South East Asia is substantially less than observed as the simulated rainband does not extend far enough to the north and east. Given the large errors in the simulation of present day regional climate, confidence in the simulated changes in regional climate must be low.

Models of the oceanic circulation simulate many of the observed large-scale features of ocean climate, especially in low latitudes, although their solutions are sensitive to resolution and to the parameterisation of sub-grid scale processes such as mixing and convective overturning. It is particularly important that vertical mixing in the ocean is reproduced correctly as this determines the rate at which the ocean responds to increases in greenhouse gases. This can be validated to some extent by simulating the spread of a passive tracers, such as that of tritium following the atomic bomb tests in the 1950s and 1960s. Current models are capable of reproducing the main features of the spread of passive tracers (for example, Figure 4.), though there may be errors in the detailed changes (for example, in Figure 4b, underestimating the strength of penetration near 30 to 50°N). In CGCMs, it has proved necessary to add empirical adjustments to the ocean surface fluxes in order to reduce errors in the simulated climate.

2.2 Simulation of contemporary climate change and recent paleoclimate.

AGCMs have been used to simulate the atmospheric response to prescribed anomalies in sea surface temperature (SST), and have been notably successful in simulating the response to tropical SST anomalies. Large scale positive SST anomalies occur in eastern tropical Pacific during the occurrence of El Nino , and models have simulated successfully the associated observed anomalies in precipitation and circulation (for example, Fenessy and Shukla, 1988). Other AGCMs have reproduced the relationship between large scale SST distribution and summer rainfall in the Sahel (for example, Folland et al,1990).

A/MLMs have reproduced some of the large scale features of the mid-Holocene climate, only the changes in the earth's orbital parameters being specified, and also the last glacial maximum, the distribution of land ice and the prevailing CO_2 concentration being the only prescribed changes (for example, COHMAP members, 1988).

2.3 Summary

The most recent models are able to simulate the large scale features but not the regional (4000 - 2000 km) scale details of present climate. AGCMs reproduce the atmospheric response to contemporary SST anomalies, and A/MLMs are capable of reproducing the major changes in climate over the last 18000 years provided certain boundary conditions (CO_2 concentrations, orbital parameters, major ice sheets) are prescribed. This, combined with the physical basis on which the models are developed, gives us some confidence in their ability to predict future changes in climate, particularly on larger scales. The least reliable aspects of models are their treatment of sub-grid scale processes, and their ability to simulate variations in climate on a regional scale. Aspects of the simulation of present day climate by CGCMs may be inadequate unless corrective measures, such as adjustments to the surface fluxes, are made.

3. Simulation of climate change

We now consider the simulation of climate change due to increases in greenhouse gases. Most of our understanding of greenhouse gas induced climate change is based on equilibrium experiments based on mixed layer models. In the last year or so, several studies have been made of the time-dependent response to increases in greenhouse gases using CGCMs. However, results presented here will be based largely on the equilibrium experiments which have been analysed in much greater detail. Furthermore, the rates of increase in the time-dependent coupled experiments do not correspond to the IPCC scenarios of increases in greenhouse gases. Hence, in order to calculate the evolution of temperature resulting from the IPCC scenarios, we have used results from one-dimensional upwelling diffusion models calibrated using the more complex A/MLMs and CGCMs.

3.1 Equilibrium climate sensitivity

The equilibrium sensitivity of mixed layer models to doubling CO_2 ranges from 2 to 5°C. The range is due to uncertainties associated with sub grid scale parametrification, particularly those associated with cloud. Some of the more recent studies attempt to allow for changes in the microphysical properties of cloud - these more detailed, but not necessarily more accurate, models give an equilibrium sensitivity of 2 to 4 °C.

The observed change in temperature (1860-1990) seems to be consistent with an equilibrium warming of 1 to 3° C, allowing as far as possible for natural variability and assuming other factors such as the effect of sulphate emissions on cloud albedo have not affected the warming due to increases in greenhouse gases. On the basis of modelling studies, the sensitivity of climate due to doubling CO_2 is most likely to lie between 1.5 and 4.5° C, and, in view of the observational evidence, a "best guess" of 2.5° C was chosen to illustrate the IPCC scenarios.

3.2 Equilibrium and transient climate change

When the radiative heating of the earth-atmosphere system is changed by increases in greenhouse gases, the atmosphere will respond (by warming) immediately. The atmosphere is closely coupled to the oceans so the oceans also have to be warmed; because of their thermal capacity this takes decades or centuries. This exchange of heat between the atmosphere and ocean will act to slow down the temperature rise forced by the greenhouse effect..

Consider the concentration of greenhouse gases in the atmosphere rising to a new level and remaining constant thereafter . The radiative forcing would also rise rapidly to a new level. This increased radiative forcing would cause the atmosphere and oceans to warm, and tend towards a new equilibrium temperature . Commitment to this equilibrium temperature rise would occur as soon as the greenhouse gas concentration had changed. But at any time before equilibrium is reached, the actual temperature will only have risen by part of the equilibrium temperature change - known as the realised temperature change..

One CGCM (Stouffer et al, 1990) predicts that, for the case of a steady increase in radiative forcing similar to that currently occuring, the realised temperature at any time is about 60 - 80% of the committed temperature. If the forcing was to be held constant, temperatures would continue to rise slowly, but it is not certain whether it would take decades or centuries for most of the remaining rise to equilibrium to occur.

3.3 Changes in global mean temperature based on the IPCC scenarios

The following results are derived from an upwelling -diffusing model,* assuming that the temperature of surface water sinking in high latitudes does not change. The IPCC "business as usual scenario", in which effective CO_2 concentrations double over pre-industrial levels by 2020, this gives a warming of 1.3 to 2.6^o C above pre-industrial levels at 2030, corresponding to a prescribed climate sensitivity of 1.5 to 4.5^o C (Figure 5). The "best guess" sensitivity gives a warming of 1.8^o C at the time of doubling, which is reduced to about 1.5^o C in the other scenarios (Figure 6).Scenario B assumes an effective doubling of CO2 by 2040, scenario C by about 2050 Note that the changes in emission scenarios take some time to produce an.effect - this is a result of the slowness with which both the gas concentrations and the oceans respond- much of the warming immediately after 1990 can be attributed to the system "catching up" the effect of previous emissions.

3.4 Patterns of climatic change due to an effective doubling of CO_2

Only one AGCM has been used in both an equilibrium experiment coupled to a mixed layer ocean, and in a long time-dependent experiment coupled to a deep ocean model (a CGCM, Stouffer et al 1989). (Figure 7). The equilibrium sensitivity of this model to doubling CO_2 is 4^o C. In most regions heat is mixed down to the main thermocline, at about 500m; and the surface temperature change at the time of doubling CO_2 in the time-dependent experiment (Figure 7b) is similar to the equilibrium response (Figure 7a), but reduced by 20 to 40% (Figure 7c). The exceptions are round Antarctica, where there is little or no warming in the

* Upwelling velocity 4 ms^{-1}, Diffusivity 0.63 $cm^2\ s^{-1}$.

time-dependent experiment, and over the northern North Atlantic and North Western Europe,where the reduction is 40 to 60%. In these regions, a deep wind driven circulation (circumpolar ocean) or deep convection (N. Atlantic) mix heating down to several kilometres. The surface warming is greatest in high northern latitudes, and north of 30°S, is generally greater over land than over the ocean at the same latitudes, as found in equilibrium experiments. The warming in northern high latitudes is generally greatest around the sea-ice margins in late autumn and early winter, and around the snow-line over the continents in spring. Over the low latitude and moist areas of the low latitude continents, the warming is small relative to global mean .

The changes in the hydrological cycle are qualitatively similar in these equilibrium and transient experiments. Hence, the remainder of this section is based on results from equilibrium experiments. Precipitation generally increases in high latitudes throughout the year and in mid-latitudes in summer (Figure 8). Precipitation generally increases in the tropics, these changes are accompanied by shifts in the main tropical rain bands which vary from model to model, so there is little consistency in results for a particular region.

The warming produces an increase in global mean precipitation and evaporation .Enhanced precipitation increases soil moisture levels in high northern latitudes in winter. Most models produce a drying of the land surface over the northern mid-latitude continents in summer (for example, Figure 9) as a result of earlier snow melt, enhanced evaporation, (and in some regions, reduced precipitation)

Predicted changes in the day to day variability of weather are uncertain. However, simulated episodes of high temperature become more frequent simply due to the substantial increases in mean temperature. There is some evidence of an increase in convective precipitation. Numerical experiments show a reduction in mid-latitude storms, but as current models only resolve the larger disturbances, this does not rule out the possibility of smaller but more intense storms. An increase in the maximum intensity of tropical cyclones might be expected on theoretical grounds due to the increased availability of latent heat, but changes in the intensity of tropical disturbances simulated in different A/MLMs are inconsistent.

3.5 Estimating regional climate change: Pre-industrial to 2030 (IPCC "business as usual" scenario)

The reader is reminded of the limited ability of current climate models to simulate regional climate change. In deriving the results below, the following assumptions have been made:-

i) The "best guess" of the magnitude of the global mean equilibrium increase in surface temperature at 2030 is 1.8° C (this is consistent with a climate sensitivity of 2.5° C).

ii) The patterns of equilibrium and transient climate change are similar (this may be approximately true for the regions considered, but not for the high latitude Southern Ocean, see Figure 7).

iii) The regional changes in temperature, precipitation and soil moisture are proportional to the global mean changes in surface temperature (this will be approximately valid except in regions where the changes are associated with a shift in the position of steep gradients, for example where the snowline retreats, or on the edge of a rainbelt which is displaced).

Although it is hard to justify some of these assumptions on rigorous scientific grounds, the errors involved are substantially smaller than the uncertainties arising from the threefold range in climate sensitivity. The results below (IPCC,1990) are derived using equilibrium results from three high resolution models (see caption to Figure 1) and scaling them to give the appropriate change in global mean temperature. For a climate sensitivity of 1.5° C, the changes should be reduced by 30%, and for a climate sensitivity of 4.5° C they should be increased by 50%. All the changes below are averages over the region- the range arises because of the use of different models.

Central North America (35-50 N, 85-105 W)

The warming varies from 2 to 4° C in winter and 2 to 3° C in summer. Precipitation increases range from 0 to 15% in winter whereas there are deceases of 5 to 10% in summer. Soil moisture decreases in summer by 15 to 20%.

South East Asia (5-30 N, 70-105 E)

The warming varies from 1 to 2° C throughout the year. Precipitation changes little in winter and generally increases throughout the region by 5 to 15% in summer. Summer soil moisture increases by 5 to 10%.

Sahel (10-20 N, 20 W-40 E)

The warming ranges from 1 to 2° C. Area mean precipitation increases and area mean soil moisture decreases marginally in summer. However there are areas of both increase and decrease in both parameters throughout the region which differ from model to model.

Southern Europe. (35-50 N, 10 W-45 E)
The warming is about 2° C in winter and varies from 2 to 3° C in summer. There is some indication of increased precipitation in winter, but summer precipitation decreases by 5 to 15%, and summer soil moisture by 15 to 25%.

Australia (10-45 S, 110-155 E)
The warming ranges from 1 to 2° C in summer and is about 2° C in winter. Summer precipitation increases by around 10% , but the models do not produce consistent estimates of the changes in soil moisture. The area averages hide large variations at the sub-continental level.

Evidence from both modelling and observational studies suggest that the equilibrium global mean warming due to doubling atmospheric CO_2 is most likely to lie in the range 1.5 to 4.5° C Assuming an effective doubling of CO_2 by 2020 and an equilibrium sensitivity of 2.5° C, simple models calibrated using the more complex GCMs predict a warming of just under 2° C above pre-industrial levels by 2030. The warming is expected to be greatest over the higher northern latitudes in winter and least over the southern ocean throughout the year. Simulated precipitation increases in middle and high latitudes in winter, but soil moisture generally decreases in northern mid-latitudes in summer. The magnitude of changes in precipitation is generally 0 to 15%.

3.6 Summary

Evidence from both modelling and observational studies suggest that the equilibrium global mean warming due to doubling atmospheric CO_2 is most likely to lie in the range 1.5 to 4.5° C Assuming an effective doubling of CO_2 by 2020 and an equilibrium sensitivity of 2.5° C, simple models calibrated using the more complex GCMs predict a warming of just under 2° C above pre-industrial levels by 2030. The warming is expected to be greatest over the higher northern latitudes in winter and least over the southern ocean throughout the year. Simulated precipitation increases in middle and high latitudes in winter, but soil moisture generally decreases in northern mid-latitudes in summer. The magnitude of changes in precipitation is generally 0 to 15%. Little confidence can be placed in the variations in the changes on a regional scale (less than 2000 km)

4. Reducing uncertainties: current research programs and their possible shortcomings.

The assessment of recent results by IPCC gives an opportunity to review existing research programs and to identify areas where further initiatives are required. The major sources of uncertainty in the simulation of climate change arise from:

(i). lack of knowledge concerning the processes leading to the formation and dissipation of clouds, and determining their radiative properties. Furthermore, an improved understanding of the relevant microphysical processes needs to be matched with improved parametrization of these processes in large scale models of climate if the current uncertainties in climate sensitivity are to be reduced. This issue is addressed to some extent by the World Climate Research Programme (WCRP) through its Global Energy and Water Cycle Experiment (GEWEX) and the ongoing International Satellite Cloud Climatology Project (ISCCP), though it is not obvious, for example, that clouds and related processes are the top priority in GEWEX. Thus, it may be necessary to review widely spread programmes such as GEWEX to ensure that the major sources of uncertainty are given due priority.

(ii). lack of knowledge concerning the processes leading to the vertical mixing of heat into the deep ocean, particularly in high latitudes. The World Ocean Circulation Experiment (WOCE) of the WCRP was set up to describe the ocean circulation at all depths during a five year period (1990-1995). The area under consideration was belatedly extended to cover the high latitude southern oceans and does not include the Norwegian Sea. In view of the apparent importance of high latitude oceanic mixing in recent CGCM experiments, there may be a need for additional field experiments in high latitudes to supplement WOCE. The activities of the Joint Global Flux Study (JGOFS, part of the International Geosphere- Biosphere Programme of the International Council of Scientific Unions) is also relevant to the problem of ocean mixing.

(iii). lack of knowledge concerning the parametrization of tropical convection, which appears to be associated with an uncertainty of a factor of two in the magnitude of the sensitivity in the tropics. This is presumably addressed as part of GEWEX, but may need to be made more explicit.

(iv). lack of knowledge of land surface processes.and the interaction between climate and ecosystems. This gives rise to uncertainties in the.simulation important hydrological characteristics of the land surface such as soil wetness and evaporation This is addressed explicitly in GEWEX and in the International Sateillite Land Surface Climatology Programme.(ISLSCP).

Other factors which limit progress are insufficient computing power and lack of scientists with experience in the relevant topics. Note that in this paper we have assumed that changes in greenhouse gases are prescribed, and so additional uncertainties due to the.effect of changes in climate on greenhouse gas concentrations are ignored.

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7.Figure Captions

Figure 1 Zonally averaged sea-level pressure (mb) for observed (Schutz and Gates, 1971, 72) and models.

(a) December, January, February
(b) June, July, August

NCAR National Center for Atmospheric Research (15 spectral waves)
GISS Goddard Institute for Space Studies (8^{o} x 10^{o} latitude x longitude)
GFLO Geophysical Fluids Dynamics Laboratory (15 spectral waves)
UKLO United Kingdom Meteorological Office (5^{o} x 7.5^{o})
GFHI Geophysical Fluid Dynamics Laboratory (30 spectral waves)
UKHI United Kingdom Meteorological Office (2.5^{o} x 3.75^{o})
CCC Canadian Climate Centre (32 spectral waves)

The more recent high resolution runs referred to in the text are GFHI,UKHI and CCC.

Figure 2 The temporal variation of the deviation of global mean surface air temperature (^{o}C) of the GFDL coupled ocean atmosphere model from its long term average. (Manabe, 1990; personal communication).

Figure 3 Precipitation (mm/day) for June,July, August: Contours at 1, 2, 5 and 10 mm/day, stippled where greater than 5mm/day.

(a) Observed (Jaeger, 1976)
(b) Simulated (United Kingdom Meteorological Office high (2.5^{o} x 3.75^{o}) resolution model.

Figure 4 Tritium in the GEOSECS section in the Western North Atlantic approximately one decade after major bomb tests

(a) GEOSECS observations
(b) as predicted by a 12 level model (Sarmiento, 1983) in tritium units.

Figure 5 Evolution of global mean warming (above pre-industrial temperatures) assuming IPCC Business as Usual Scenario. The curves correspond to an equilibrium sensitivity to doubling CO_2 of 4.5, 2.5 and 1.5^{o}C respectively.(IPPC,1990)

Figure 6 Evolution of global mean warming (above pre-industrial temperatures) assuming an equilibrium sensitivity to doubling CO_2 of 2.5^{o} for the IPCC scenarios (top to bottom), "Business as Usual", B, C and D.(IPCC,1990).

Figure 7 (a) The geographical distribution of the time-dependent response of surface air temperature (^{o}C) in the GFDL coupled ocean-atmosphere model to a 1% per year increase of atmospheric CO_2. Shown is the difference between the 1% per year perturbation and the control run for the 60th-80th year period when the CO_2 approximately doubles.(Stouffer et al 1989)

(b) The geographical distribution of the equilibrium response of surface air temperature (^{o}C) in the atmosphere-mixed layer ocean model to a doubling of atmospheric carbon dioxide.(Stouffer et al,1989)

(c) The geographical distribution of the ratio of time dependent to equilibrium responses shown above. (Manabe, personal communication)

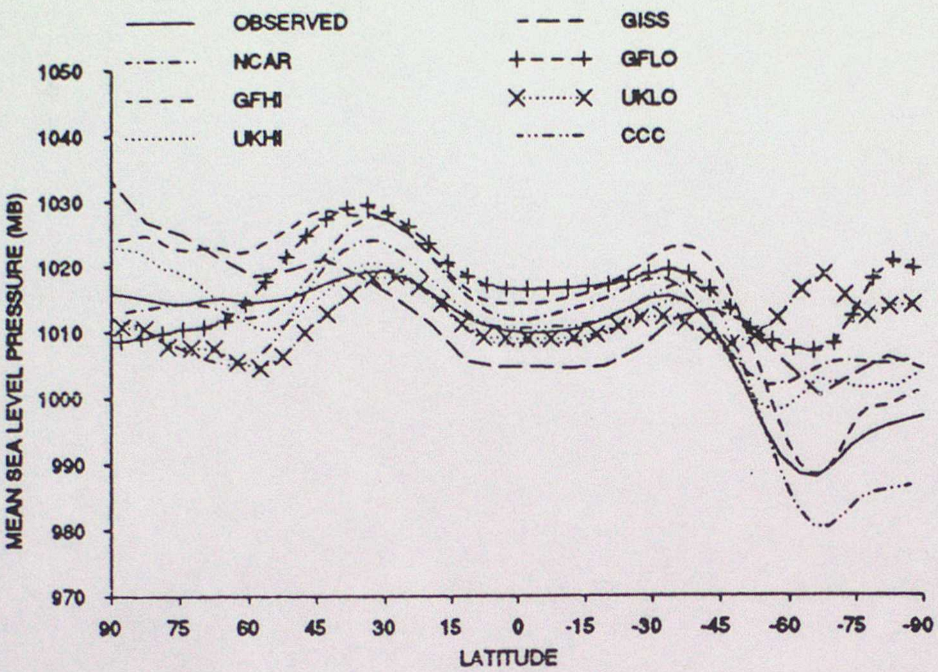
Figure 8 Simulated changes in precipitation rate equilibrium following a doubling of atmospheric CO_2 concentrations , in the United Kingdom Meteorological Office $2.5^{o} \times 3.75^{o}$ resolution model.(mm/day, contours at 0, +/- 1, 2, 5mm/day) Areas of decrease stippled

(a) December, January, February

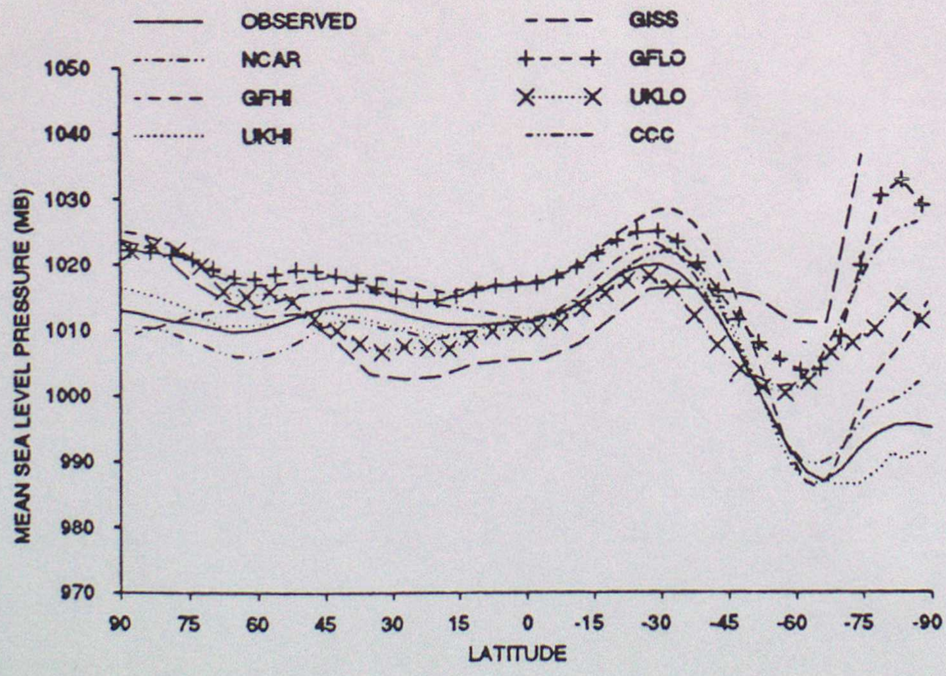
(b) June, July and August

Figure 9 Simulated changes in soil moisture at equilibrium due to doubling atmospheric CO_2 in June, July, August contours every cm, areas of decrease are stippled. (Canadian Climate Centre model with a horizontal resolution of 32 spectral waves.Boer, personal communication,1990)

(a)


OBSERVED
GISS
NCAR
GFLO
GFHI
UKLO
UKHI
CCC
MEAN SEA LEVEL PRESSURE (MB)
1050
1040
1030
1020
1010
1000
990
980
970
90
75
60
45
30
15
0
-15
-30
-45
-60
-75
-90
LATITUDE

(b)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1010
1000
990
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-90
LATITUDE

Figure 1

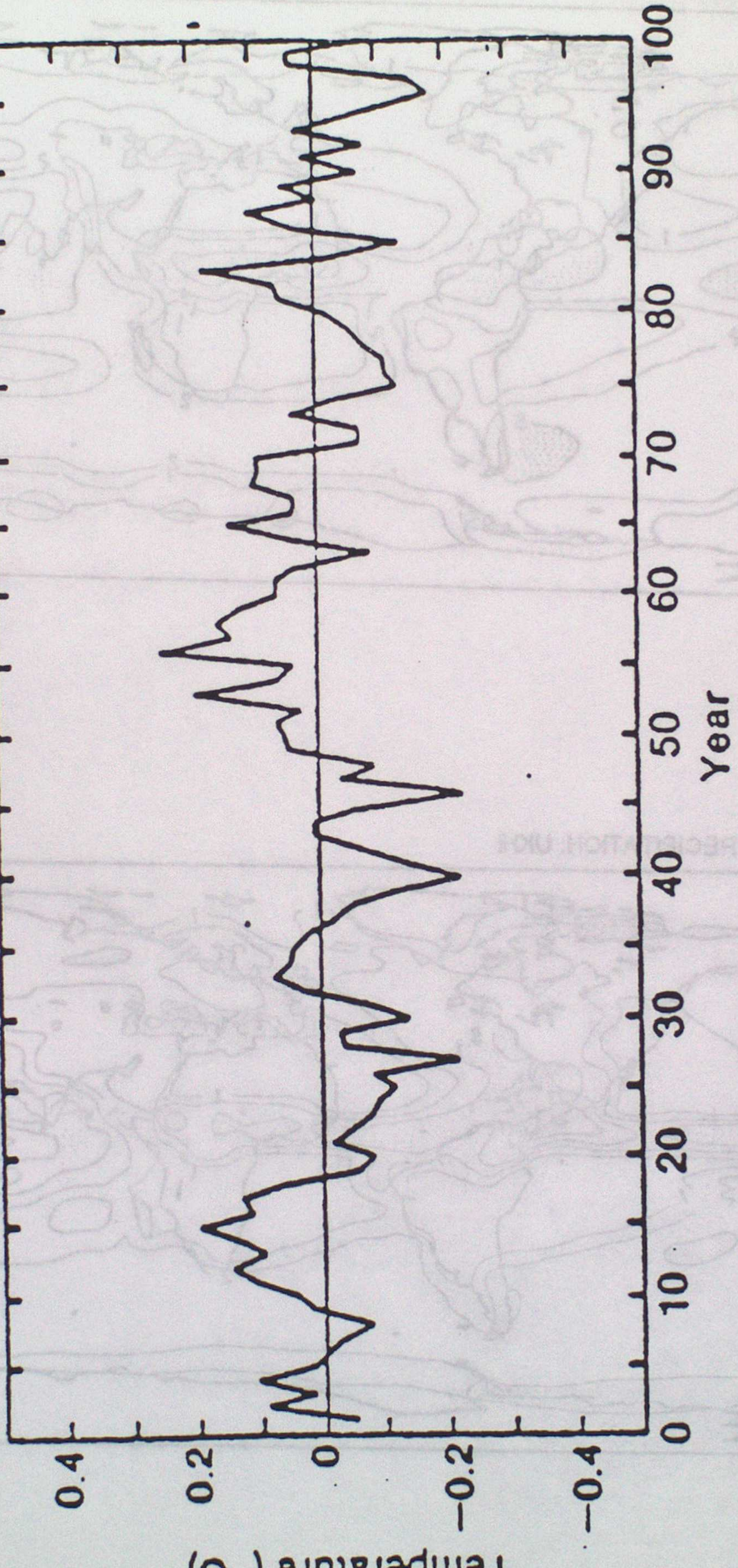
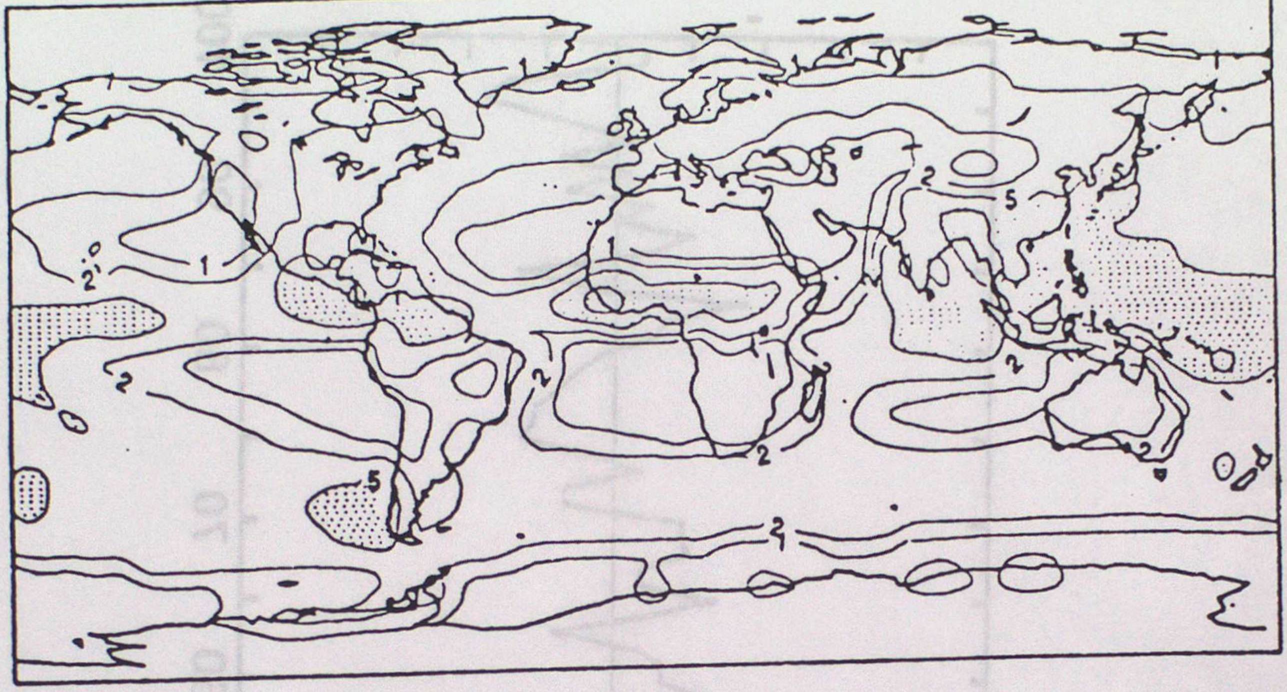


Figure 2

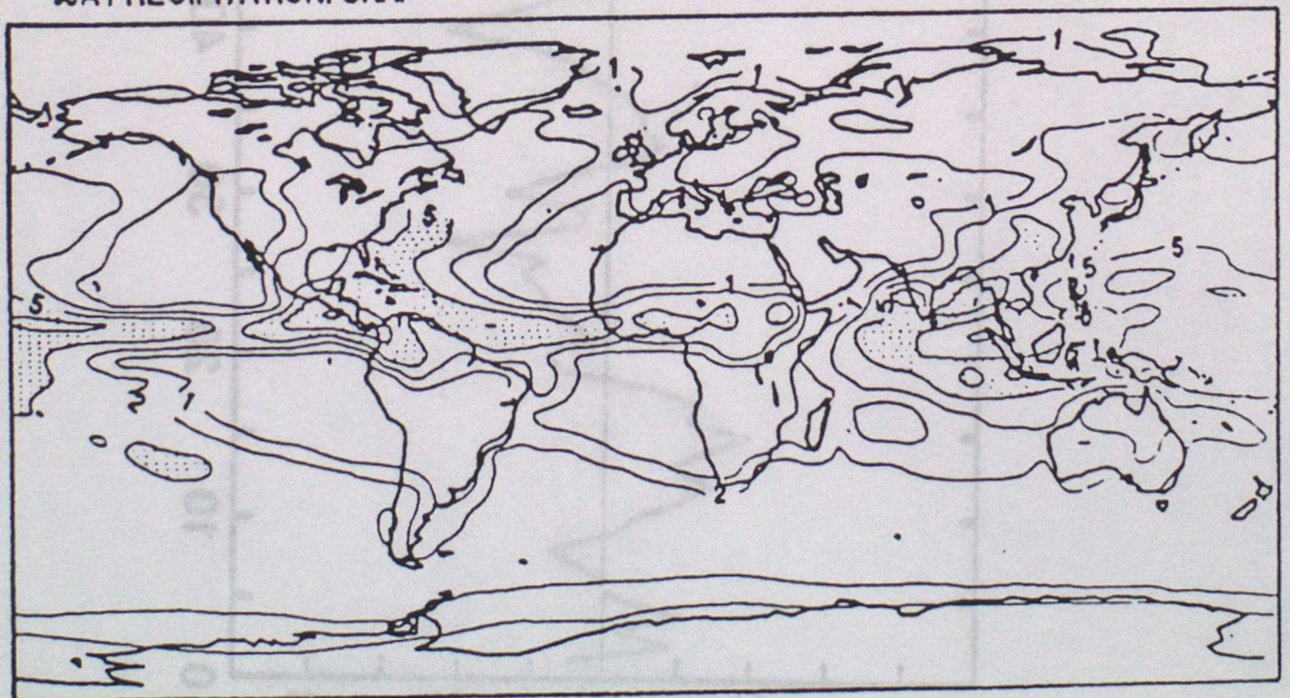
(a)

JJA PRECIPITATION: OBSERVED



(b)

JJA PRECIPITATION: UKHI



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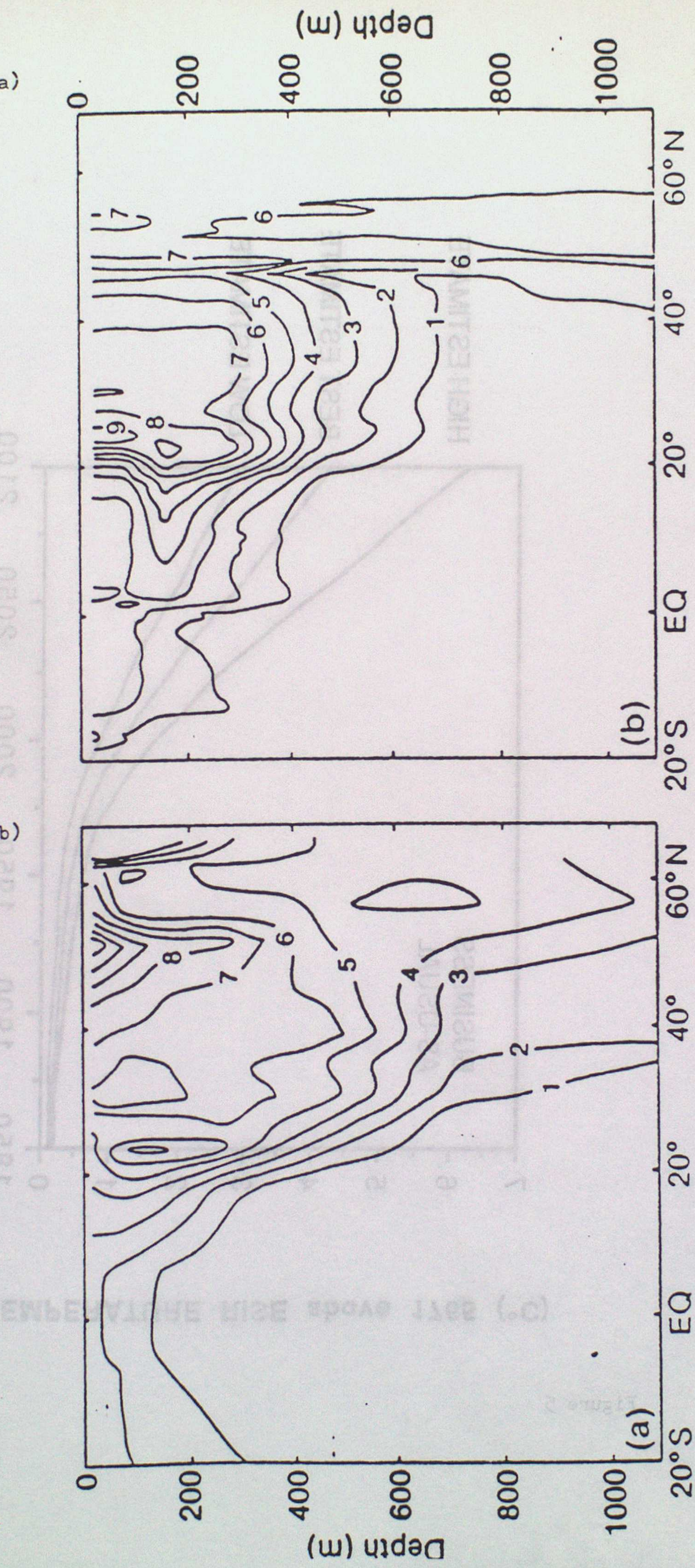

Figure 4

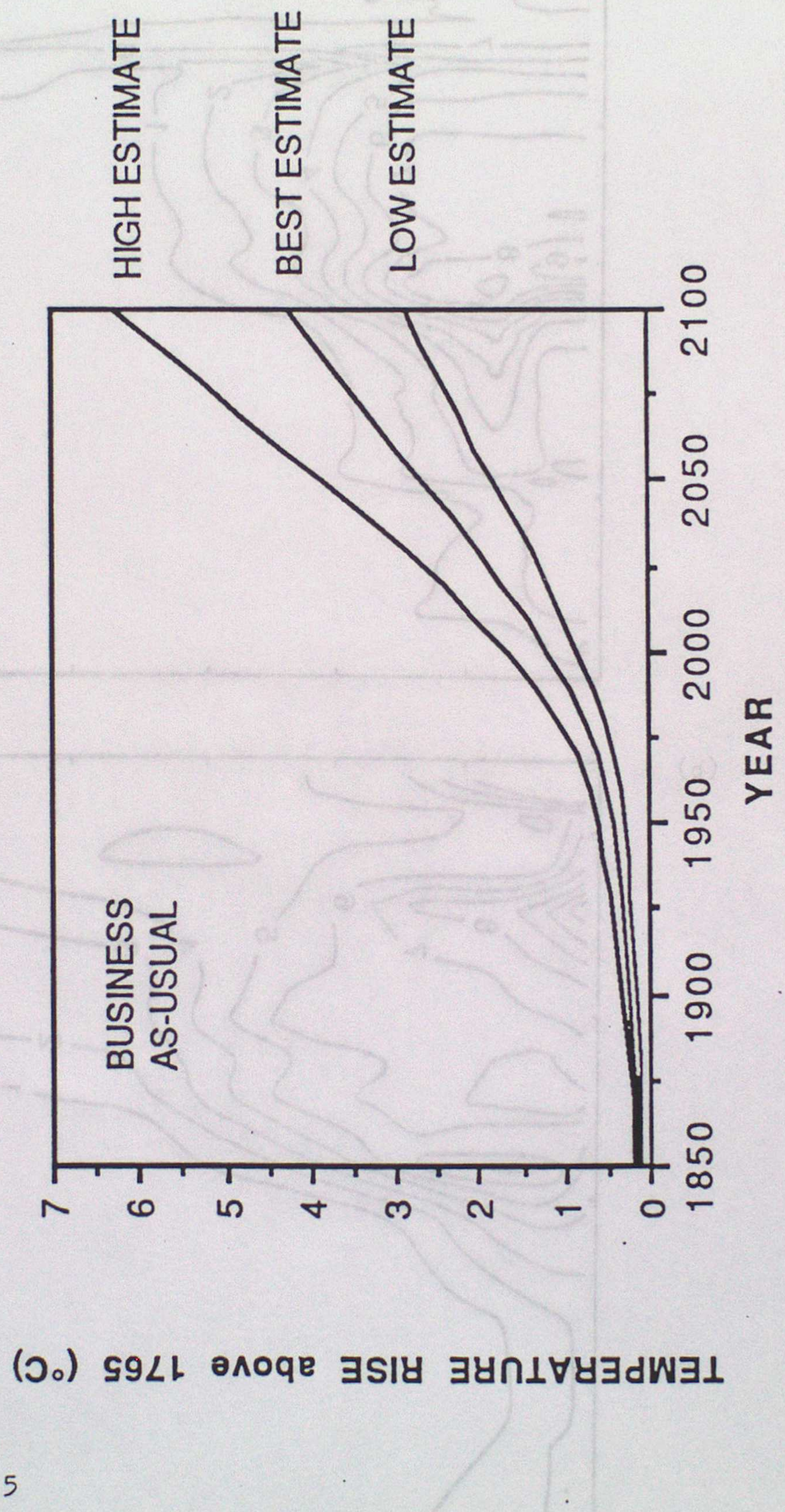
TEMPERATURE RISE above 1765 (°C)
YEAR
BUSINESS AS-USUAL
HIGH ESTIMATE
BEST ESTIMATE
LOW ESTIMATE
0
1
2
3
4
5
6
7
1850
1900
1950
2000
2050
2100

Figure 5

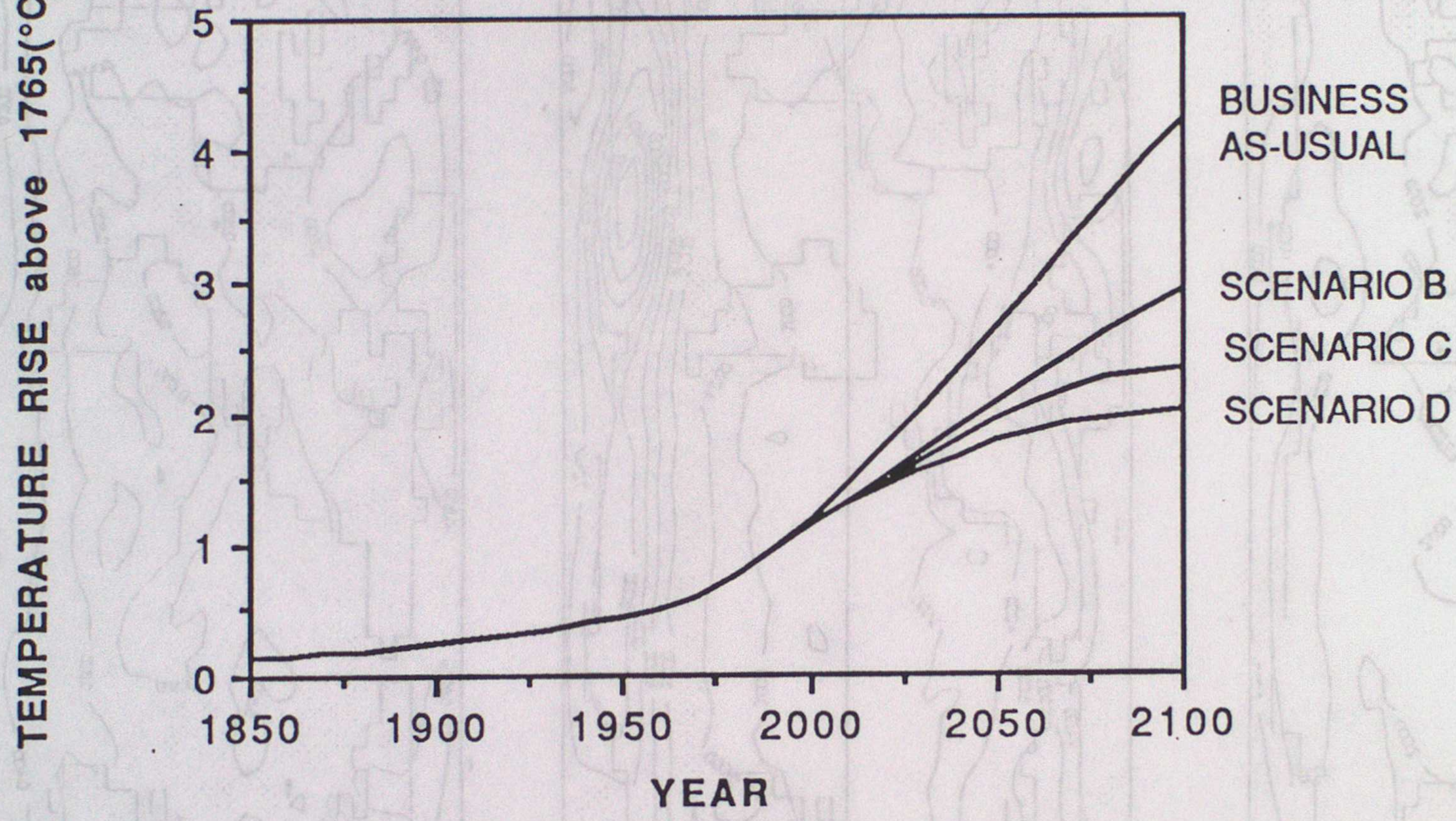
TEMPERATURE RISE above 1765(°C)
0
1
2
3
4
5
1850
1900
1950
2000
2050
2100
YEAR
BUSINESS AS-USUAL
SCENARIO B
SCENARIO C
SCENARIO D

Figure 6

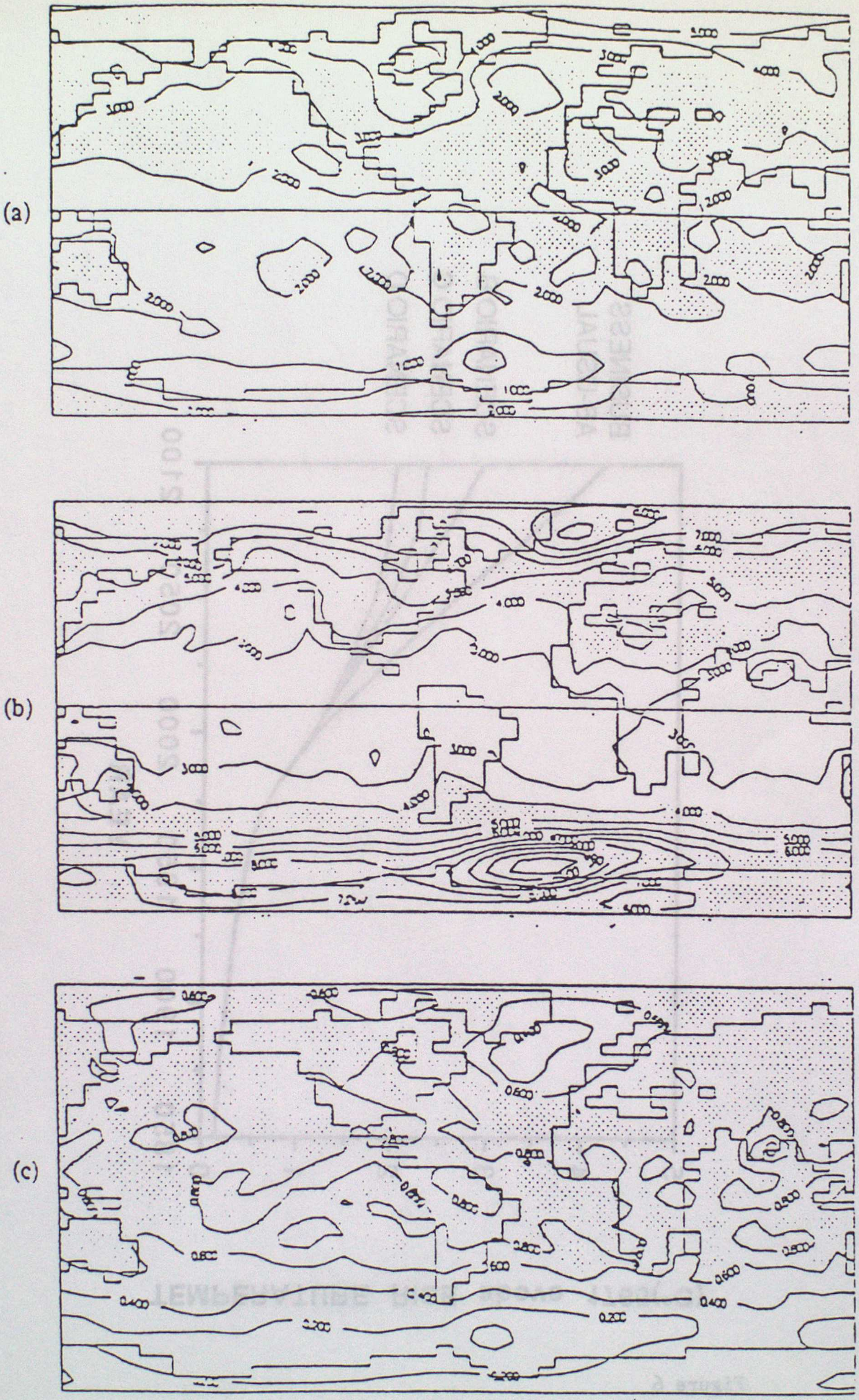

Figure 7

(a)

DJF 2 X CO2 - 1 X CO2 PRECIPITATION - UKMO HIGH

(b)

JJA 2 X CO2 - 1 X CO2 PRECIPITATION - UKMO HIGH

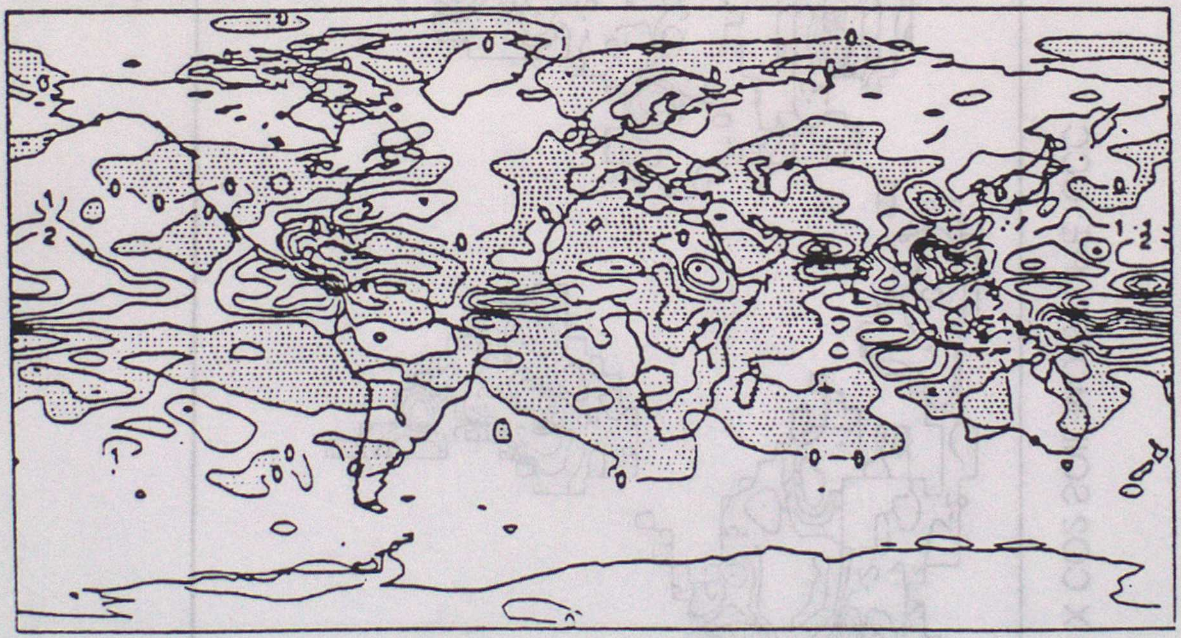

Figure 8

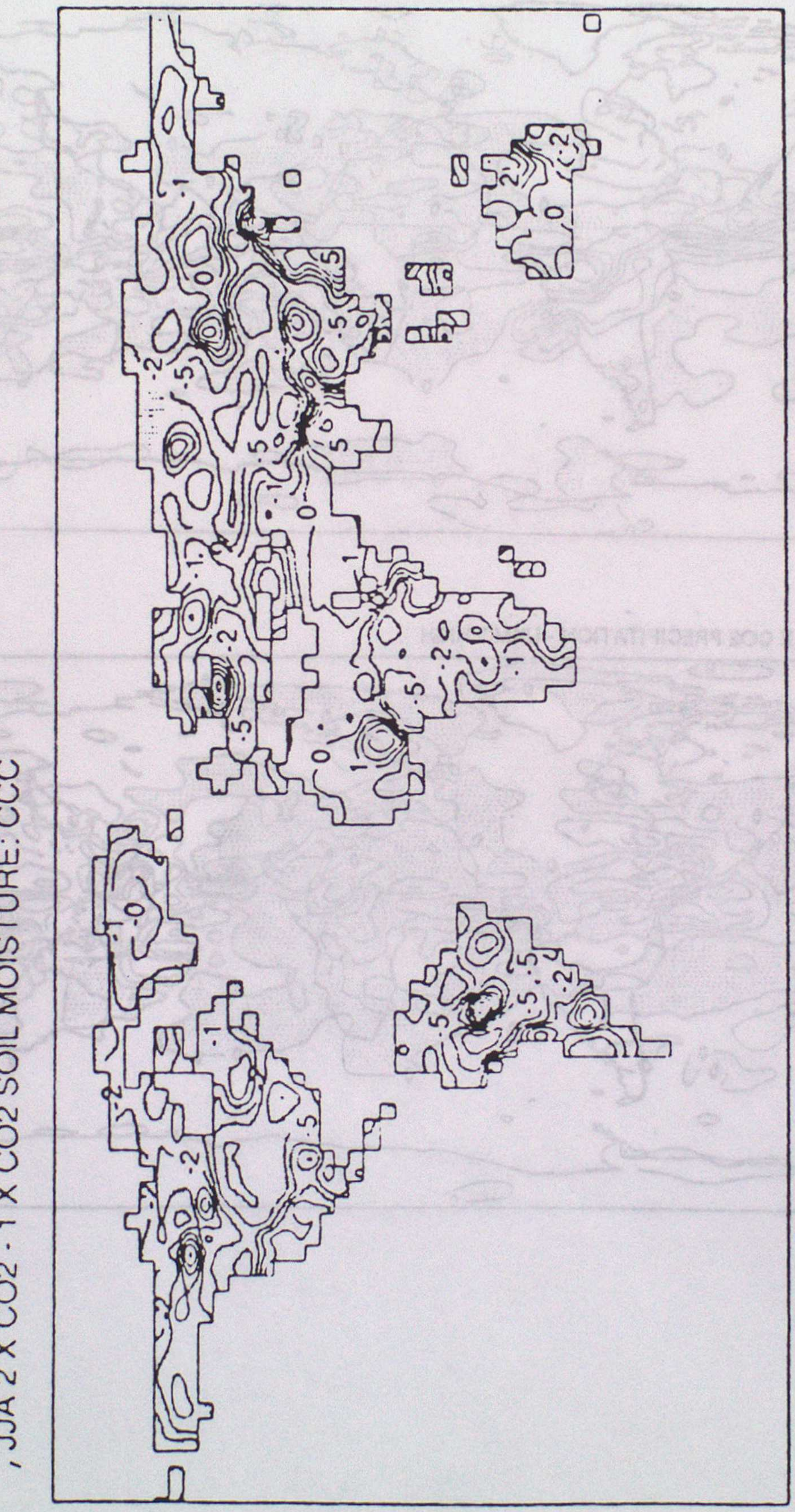

Figure 9

CLIMATE RESEARCH TECHNICAL NOTES

CRTN 1	Oct 1990	Estimates of the sensitivity of climate to vegetation changes using the Penman-Monteith equation. P R Rowntree
CRTN 2	Oct 1990	An ocean general circulation model of the Indian Ocean for hindcasting studies. D J Carrington
CRTN 3	Oct 1990	Simulation of the tropical diurnal cycle in a climate model. D P Rowell
CRTN 4	Oct 1990	Low frequency variability of the oceans. C K Folland, A Colman, D E Parker and A Bevan
CRTN 5	Dec 1990	A comparison of 11-level General Circulation Model Simulations with observations in the East Sahel. K Maskell
CRTN 6	Dec 1990	Climate Change Prediction. J F B Mitchell and Qing-cun Zeng