B

Climate Modelling, Climate Prediction and Model Validation

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CONTENTS

EXECUTIVE SUMMARY

Coupled Model Experiments

- The new transient climate simulations with coupled atmosphere-ocean general circulation models (GCMs) generally confirm the findings of Section 6 in the 1990 Intergovernmental Panel on Climate Change report (IPCC, 1990), although the number of coupled model simulations is still small.
- There is broad agreement among the four current models in the simulated large-scale patterns of change and in their temporal evolution.
- The large-scale patterns of temperature change remain essentially fixed with time, and they become more evident with longer averaging intervals and as the simulations progress.
- The large-scale patterns of change are similar to those obtained in comparable equilibrium experiments except that the warming is retarded in the southern high latitude ocean and the northern North Atlantic Ocean where deep water is formed.
- All but one of the models show slow initial warming (which may be an artefact of the experimental design) followed by a nearly linear trend of approximately 0.3°C per decade.
- All models simulate a peak-to-trough natural variability of about 0.3°C in global surface air temperature on decadal time-scales.
- The rate of sea level rise due to thermal expansion increases with time to between 2 and 4cm/decade at the time of doubling of equivalent CO_2.

Regional Changes

- Although confidence in the regional changes simulated by GCMs remains low, progress in the simulation of regional climate is being obtained with both statistical and one-way nested model techniques. In both cases the quality of the large-scale flow provided by the GCM is critical.

Climate Feedbacks and Sensitivity

- There is no compelling new evidence to warrant changing the equilibrium sensitivity to doubled CO_2 from the range of 1.5 to 4.5°C as given by IPCC 1990.
- There is no compelling evidence that water vapour feedback is anything other than the positive feedback it has been conventionally thought to be, although there may be difficulties with the treatments of upper-level water vapour in current models.
- The effects of clouds remain a major area of uncertainty in the modelling of climate change. While the treatment of clouds in GCMs is becoming more complex, a clear understanding of the consequences of different cloud parametrizations has not yet emerged.

Atmospheric Variability

- Model experiments with doubled CO_2 give no clear indication of a systematic change in the variability of temperature on daily to interannual time-scales, while changes of variability for other climate features appear to be regionally (and possibly model) dependent.

Ocean Modelling

- Results from eddy-resolving ocean models show a broadly realistic portrayal of oceanic variability, although the climatic role of eddies remains unclear.
- Ocean-only GCMs show considerable sensitivity of the thermohaline circulation on decadal and longer time-scales to changes in the surface fresh-water flux, although coupled models may be less sensitive.
- Sea-ice dynamics may play an important role in the freezing process, and should therefore be included in models for the simulation of climate change under increased CO_2.

Model Validation

- There have been improvements in the accuracy of individual atmospheric and oceanic GCMs, although the ranges of intermodel error and sensitivity remain large.
- The lack of adequate observational data remains a serious impediment to climate model improvement.

B1 Introduction

Since the publication of the first IPCC Scientific Assessment of Climate Change (IPCC, 1990) there have been significant advances in many areas of climate research as part of a continuing worldwide acceleration of interest in the assessment of possible anthropogenic climate changes. In this section we concentrate on advances in the modelling of climate change due to increased greenhouse gases, improvements in the analysis of climate processes and feedbacks, and on advances in climate model validation that were not available to the IPCC in early 1990. This section is thus intended as an update to selected portions of the 1990 assessment rather than as a comprehensive revision, and an effort has been made to keep the discussion both concise and focussed in accordance with the stringent space limitations placed upon this supplementary report.

In Section B2 recent simulations of climate change are assessed, with emphasis on results from coupled ocean-atmosphere models. In Section B3, recent research on modelling climate feedbacks is discussed, and the IPCC 1990 estimates of climate sensitivity are reviewed. The simulation of atmospheric variability and its changes due to increased atmospheric CO_2 are discussed in Section B4 while developments in ocean and sea-ice modelling are presented in Section B5. Finally, Section B6 discusses advances in climate model validation.

B2 Advances in Modelling Climate Change due to Increased Greenhouse Gases

B2.1 Introduction

Simulation of the climatic response to increases in atmospheric greenhouse gases has continued to dominate climate modelling. Preliminary results from new integrations of coupled global atmosphere-ocean models with progressive increases of CO_2 show that the patterns of the transient response are similar to those in an equilibrium response, except over the high-latitude southern ocean and northern North Atlantic ocean; here the delayed warming has highlighted the critical role of the oceanic thermohaline circulation. Computing limitations have continued to restrict the resolution that can be used in GCM simulations of the climate changes due to increased greenhouse gases, although progress is being made in the simulation of regional climate by both statistical techniques and by locally nesting a higher-resolution model within a global GCM.

B2.2 New Transient Results from Coupled Atmosphere-Ocean GCMs

At the time of the 1990 IPCC report, preliminary results were available from only two coupled model integrations with transient CO_2, namely those made at NCAR (Washington and Meehl, 1989) and at GFDL (Stouffer *et al.*, 1989). Of these only the GFDL integration had been carried to the point of CO_2 doubling (which occurred after

***Table B1**: Summary of transient CO_2 experiments with coupled ocean-atmosphere GCMs*

	GFDL	MPI	NCAR	UKMO
AGCM	R15 L9	T21 L19	R15 L9	2.5° × 3.75° L11
OGCM	4.5° × 3.75° L12	4° L11	5° L4	2.5° × 3.75° L17
Features	no diurnal cycle, isopycnal ocean diffusion	prognostic CLW, quasi-geostrophic ocean		prognostic CLW, isopycnal ocean diffusion
Flux adjustment	seasonal, heat, fresh water	seasonal, heat, fresh water, wind stress	none	seasonal, heat, fresh water
Control CO_2 (ppm)	300	390 †	330	323
CO_2 (t)	1% yr^{-1} (compound)	IPCCa & d (1990 Scenarios), $2\times CO_2$	1% yr^{-1} (linear)	1% yr^{-1} (compound)
Length (yr)	100	100	60	75
CO_2 doubling time (yr)	70	60 (IPCCa)	100	70
Warming (°C) at CO_2 doubling	2.3	1.3	2.3 (projected)	1.7
$2\times CO_2$ sensitivity (°C) (with mixed layer ocean)	4.0	2.6	4.5	2.7 ††

L - number of vertical levels; R - number of spectral waves (rhomboidal truncation); T - no spectral waves (triangular truncation).
† - equivalent CO_2 value for trace gases
†† - estimate from low resolution experiment

approximately 70 years as a result of a 1% per year compound increase of CO_2). Here we present further results from the transient CO_2 experiments with both the GFDL and NCAR coupled models, along with preliminary results from new transient CO_2 integrations recently completed at the Max-Planck-Institute for Meteorology (MPI) in Hamburg and at the Hadley Centre of the United Kingdom Meteorological Office (UKMO). A fuller description and intercomparison of these results is being prepared (WCRP, 1992).

A summary of the new transient results is presented in Table B1. Note that there are differences in the CO_2 equivalent doubling times (by a factor of almost two) and that the equilibrium sensitivity of the models is different. The UKMO model has the finest horizontal resolution while the MPI model has the finest vertical resolution in the atmosphere and includes explicitly the radiative effects of other trace gases. Three of the four coupled models (see Table B1) use a correction or adjustment of the air-sea fluxes so that the CO_2-induced changes should be interpreted as perturbations around a climate similar to that presently observed. Adjustment of the fluxes also prevents distortion of the CO_2 induced perturbation by rapid drift of the model climate (since the same corrections are applied to the control and anomaly simulations).

1. The new transient climate simulations with coupled atmosphere-ocean GCMs generally confirm the findings of Section 6 in the 1990 IPCC report, although the number of coupled model simulations is still small.

The globally averaged annual mean increase of surface air temperature at the time of effective CO_2 doubling (70 years for the GFDL and UKMO models, 60 years for the MPI model with IPCC Scenario A, and 100 years for the NCAR model) is between 1.3°C and 2.3°C. These values are approximately 60% of the models' equilibrium warming (where known) with doubled CO_2 when run with simple mixed-layer oceans. These results confirm those from the simple models used in IPCC 1990. The lower values of warming compared to the equilibrium experiments are partly due to the fact that the transient experiments take into account the thermal inertia of the deep ocean and are therefore not at equilibrium at the time of effective CO_2 doubling.

2. All but one of the models show a limited initial warming (which may be an artefact of the experimental design) followed by a nearly linear trend of approximately 0.3°C per decade. All models simulate a peak-to-trough natural variability of about 0.3 to 0.4°C in global surface air temperature on decadal time-scales.

The evolution of the change of globally averaged surface air temperature (sea surface temperature for the NCAR experiment) during the course of the various transient experiments with coupled ocean-atmosphere GCMs is shown in Figure B1. In spite of the differences in the models' parametrizations and in their experimental configurations, all of the models exhibit a number of similar overall features in their response. Firstly, three of the models show relatively little warming during the first few decades of the integration rather than a constant rate of warming throughout, despite the near constant rate of increase in radiative heating. This so-called "cold start " is barely noticeable in the GFDL simulation, but in the UKMO and MPI models the warming is negligible during the first 2 to 3 decades. This phenomenon is thought to be an artefact of the experimental design and can be reproduced qualitatively using simplified models (Hasselmann *et al.*, 1992; J.M. Murphy, personal communication; see also Hansen *et al.*, 1985) which indicate that the length of delay grows with increasing model sensitivity (the equilibrium warming for doubling CO_2) and with effective heat capacity in the ocean. Until this "cold start" phenomenon is investigated and understood, it is not meaningful to match "model time" with calendar dates. Secondly, as the increase in CO_2 progresses, each model approximates a constant rate of warming, in overall agreement with the 0.3°C/decade in the IPCC (1990) projections made for the "business-as-usual" CO_2 forcing scenario (A) with a simplified upwelling-diffusion ocean model. Thirdly, all models exhibit variability on interannual, decadal and even longer time-scales, and some models reproduce ENSO-like effects (Meehl,1990; Lau *et al.*, 1991; J.F.B. Mitchell, personal communication). The peak-to-trough range of global surface air temperature intra-decadal natural variability is 0.3 to 0.4°C (see, for example, Figure B2).

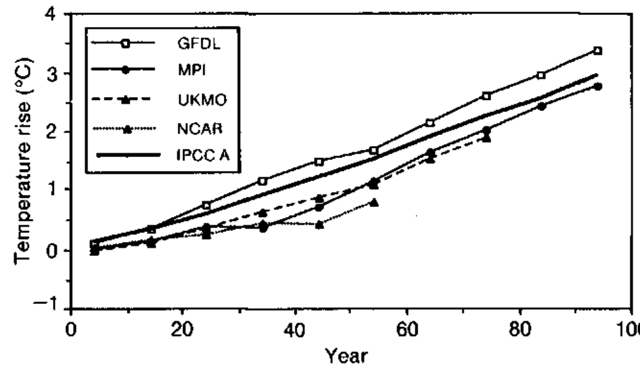

Figure B1: Decadal mean changes in globally averaged surface temperature (°C) in various coupled ocean-atmosphere experiments. (see Table B1). Note that the scenarios employed differ from model to model, and that the effect of temperature drift in the control simulation has been removed. Open boxes = GFDL; solid circles = MPI; triangles with dashed line = UKMO; triangles with dotted line = NCAR (sea temperatures only); solid line = IPCC 1990 Scenario A "best estimate".

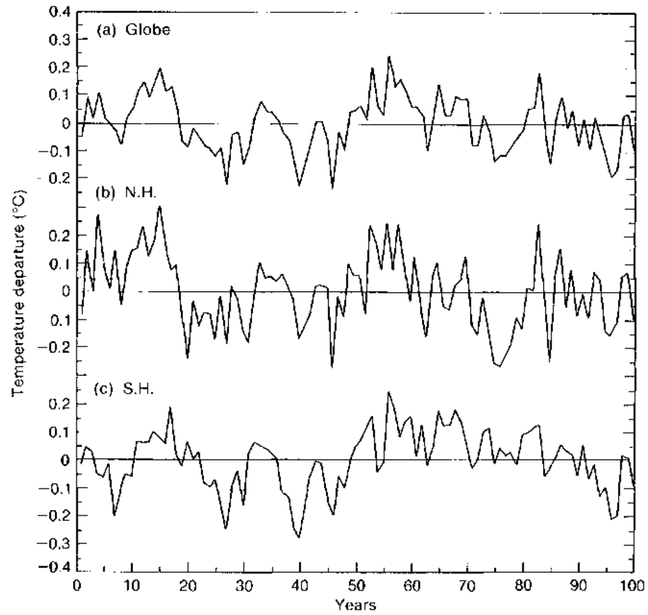

Figure B2: Temporal variations of the area-averaged deviation of annual mean surface air temperature (°C) from the corresponding 100-year average of the control as simulated by the GFDL coupled ocean-atmosphere model for: (a) the globe, (b) the Northern Hemisphere, and (c) the Southern Hemisphere. (From Manabe *et al.*, 1991.)

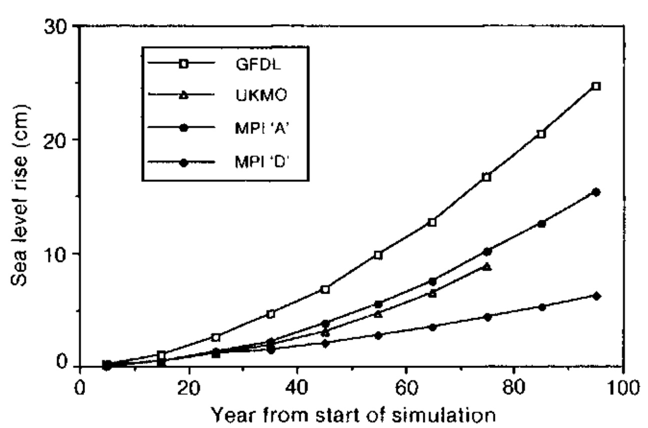

Figure B3: Decadal mean changes in globally averaged sea level change (cm) from various coupled ocean-atmosphere GCM experiments (see Table B1). Open squares = GFDL; solid circles = MPI (IPCC Scenario "A"); solid diamonds = MPI (IPCC Scenario "D"); open triangles = UKMO. Note the differences in forcing in Table B1, and that the effect of long-term drift in the models has been removed by differencing.

Analysis of the nature of this variability in the coupled integrations and of the implications for detecting global warming is currently in progress.

3. The rate of sea level rise due to thermal expansion increases with time to between 2 and 4cm/decade at the time of doubling of equivalent CO_2.

The "cold start" leads to a substantial delay in the associated change in global sea level rise due to thermal expansion (Figure B3), and is most pronounced in the UKMO and MPI models. The rate of sea level rise increases during the simulations to 2, 2.5 and 4cm/decade by the time of CO_2 doubling in the MPI A (Scenario A), UKMO and GFDL experiments, respectively. The simulated rate of sea level rise at 2030 (the time of doubling of CO_2) due to thermal expansion (Scenario "A" best estimate) in the previous IPCC assessment was about 2.5cm/decade. Note that we have not reassessed the contribution to sea level rise from changes in the snow and ice budgets over land.

4. The large-scale patterns of change are similar to those obtained in comparable equilibrium experiments except that the warming is retarded in the southern high latitude ocean and the northern North Atlantic Ocean where deep water is formed.

The geographical distribution of the change in surface air temperature at the approximate time of CO_2 doubling in the four transient experiments is shown in Figure B4. (For the NCAR model the CO_2 increase is only 60%) The same overall characteristics are seen in the distribution of the simulated change of surface air temperature. These characteristics are: (1) the largest warming occurs in the high latitudes of the Northern Hemisphere, (2) relatively uniform warming occurs over the tropical oceans, and (3) a minimum of warming or in some cases cooling occurs over the northern North Atlantic and over the Southern Ocean around Antarctica. Features (1) and (2) are familiar from the earlier studies of the equilibrium warming in response to doubled CO_2 with a mixed-layer ocean (see Section B2.3); the high latitude ocean warming minima, on the other hand, are due to the presence of upwelling and deep vertical mixing which increase the effective heat capacity and hence the thermal inertia of the ocean locally.

5. There is broad agreement among the four models in the simulated large-scale patterns of change and in their temporal evolution. The large-scale patterns of temperature change remain essentially fixed with time, and they become more evident with longer averaging intervals and as the simulations progress.

A larger warming over land areas compared to that over the oceans is common to all models in both winter and summer. This ocean-land asymmetry contributes to a more rapid warming in the Northern Hemisphere compared to that in the Southern Hemisphere. There is a maximum warming over the Arctic Ocean in winter (Figure B5a) and a minimum in summer (Figure B5b), similar to the equilibrium results. However, the warming and its seasonal

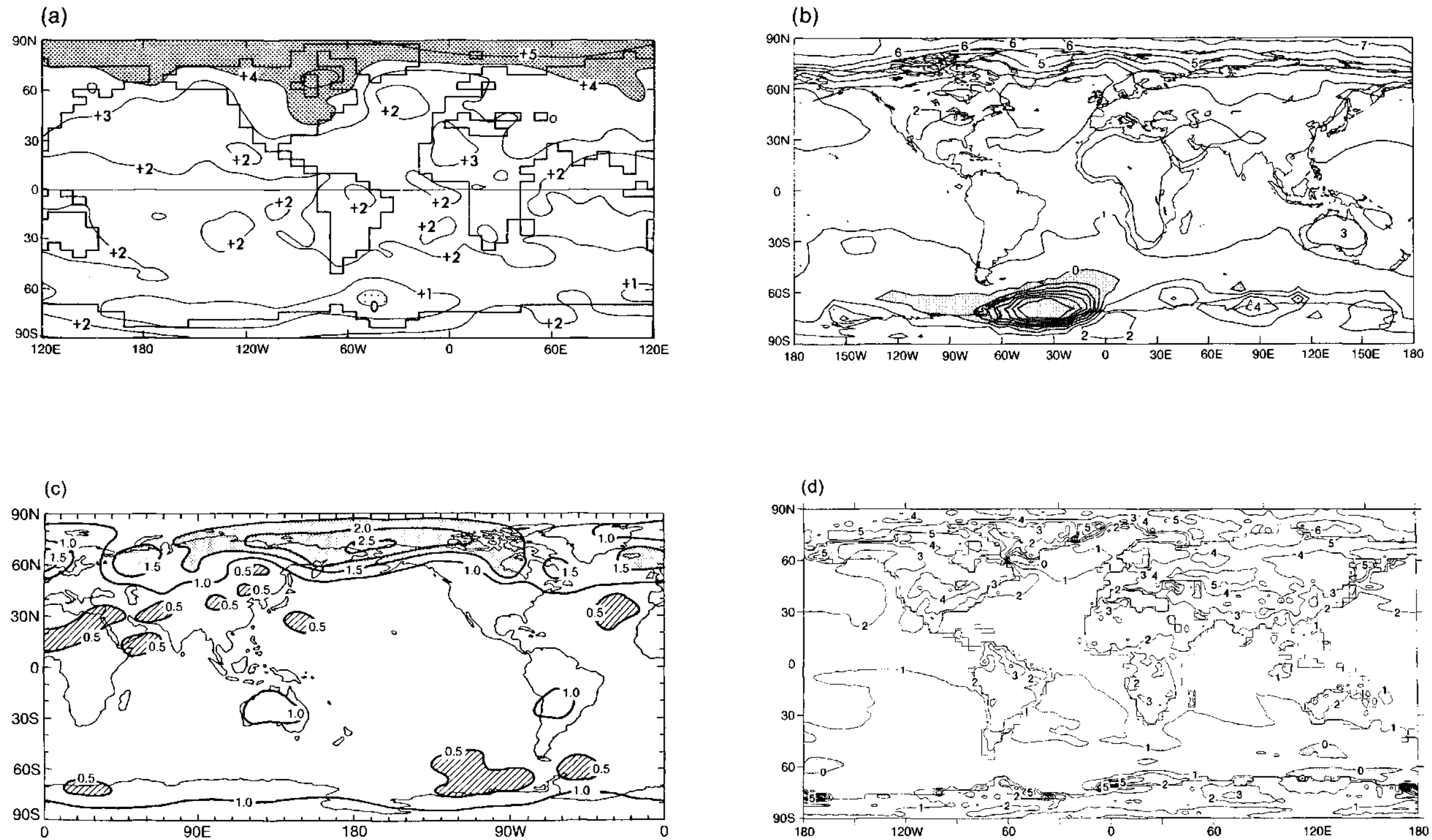

Figure B4: The distribution of the change of surface air temperature (°C) simulated near the time of CO_2 doubling by four coupled ocean-atmosphere GCMs in response to a transient CO_2 increase. (a) The GFDL results are averaged over years 60-80 and referenced to the 100-year average of a control; (b) the MPI results are averaged over years 56-65 and referenced to the corresponding years of a control; (c) the NCAR results are averaged over years 31-60 and referenced to the corresponding control years; (d) the UKMO results are averaged over years 65-75 and are referenced to the corresponding years of a control. (Manabe *et al.*, 1991; U. Cubasch, G.A. Meehl and J.F.B. Mitchell, all by personal communication.)

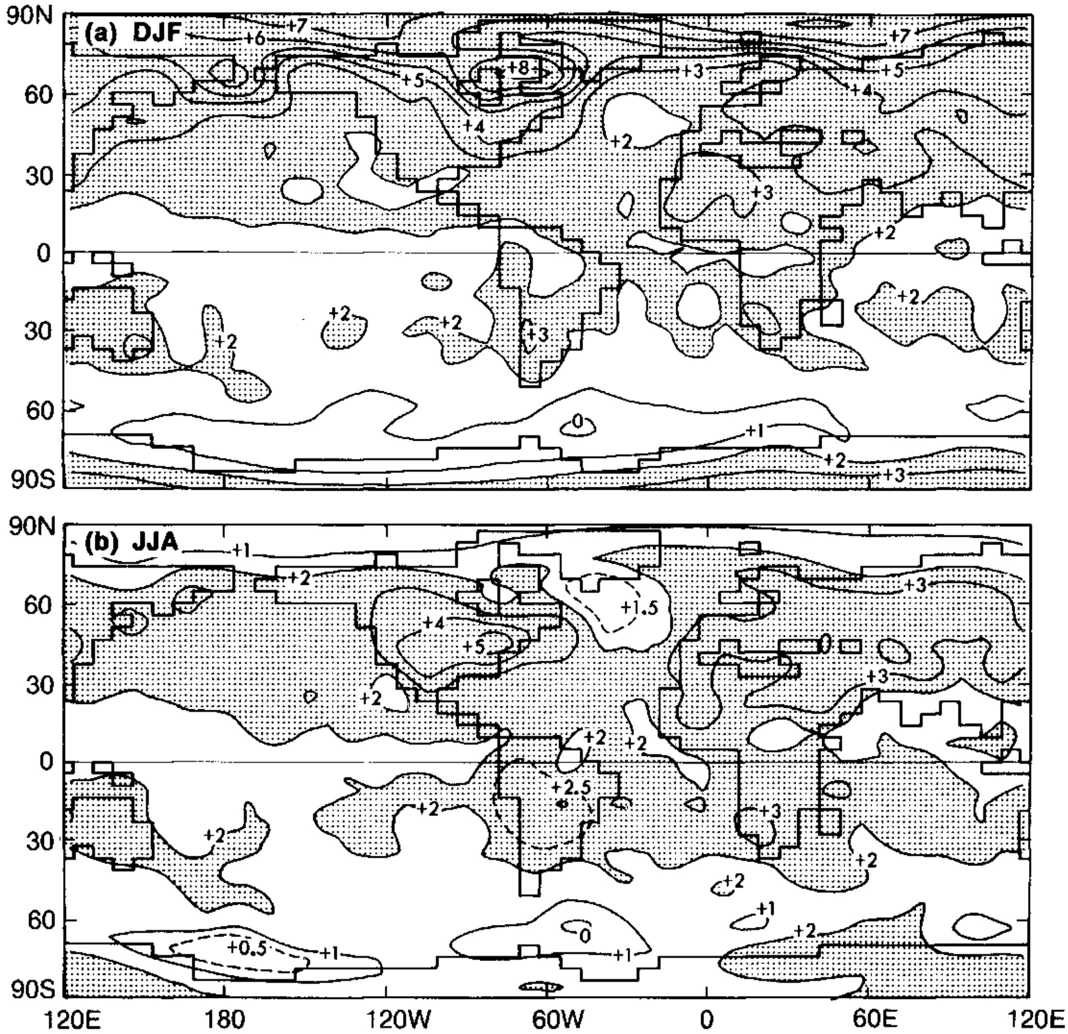

Figure B5: Distribution of the mean surface air temperature (°C) for (a) DJF, and (b) JJA during years 60-80 of a transient CO_2 simulation with the GFDL model, relative to the 100 year average of the control. (From Manabe *et al.*, 1992.)

variation over the circumpolar ocean in the Southern Hemisphere is considerably less than in the equilibrium models because of deep vertical mixing in the ocean. Precipitation increases in the Northern Hemisphere in high latitudes throughout the year, in much of mid-latitudes in winter, and in the southwest Asian monsoon (Figure B6a, b). In the Southern Hemisphere, there are precipitation increases along the mid-latitude storm tracks throughout the year. In the Northern Hemisphere, winter soil moisture increases over the mid-latitude continents, while in the summer there are many areas of drying; this is also similar to the results obtained by equilibrium models (Figure B7a, b).

In at least one simulation (Cubasch *et al.*, 1991) the progressive change in surface temperature is unrelated to the pattern of internal variability in the model; the first empirical orthogonal function (EOF) of the annual mean surface temperature in the control run is uncorrelated with the first EOF from the simulation with increasing greenhouse gas concentrations. The climate change pattern becomes more evident as the integration progresses, as indicated by the growth in the fraction of variance explained by the first EOF of the anomaly experiment (Figure B8). This feature is noticeable in other models (for example, Meehl *et al.*, 1991a) and in other fields such as soil moisture (Figure B9) and (to a lesser extent) in the precipitation (Santer *et al.*, 1991; J.F.B. Mitchell, personal communication). In IPCC 1990 it was assumed that the regional climate changes would scale linearly with the global mean temperature changes; while Figure B9 offers some support for this assumption, there is considerable interdecadal variability at regional scales in the coupled model simulations, so this scaling remains questionable.

There are a number of caveats to be kept in mind in assessing the results of the transient CO_2 simulations with coupled ocean-atmosphere models shown here. Chief among these is the use (in the GFDL, MPI and UKMO models) of adjustments to the oceanic surface fluxes so

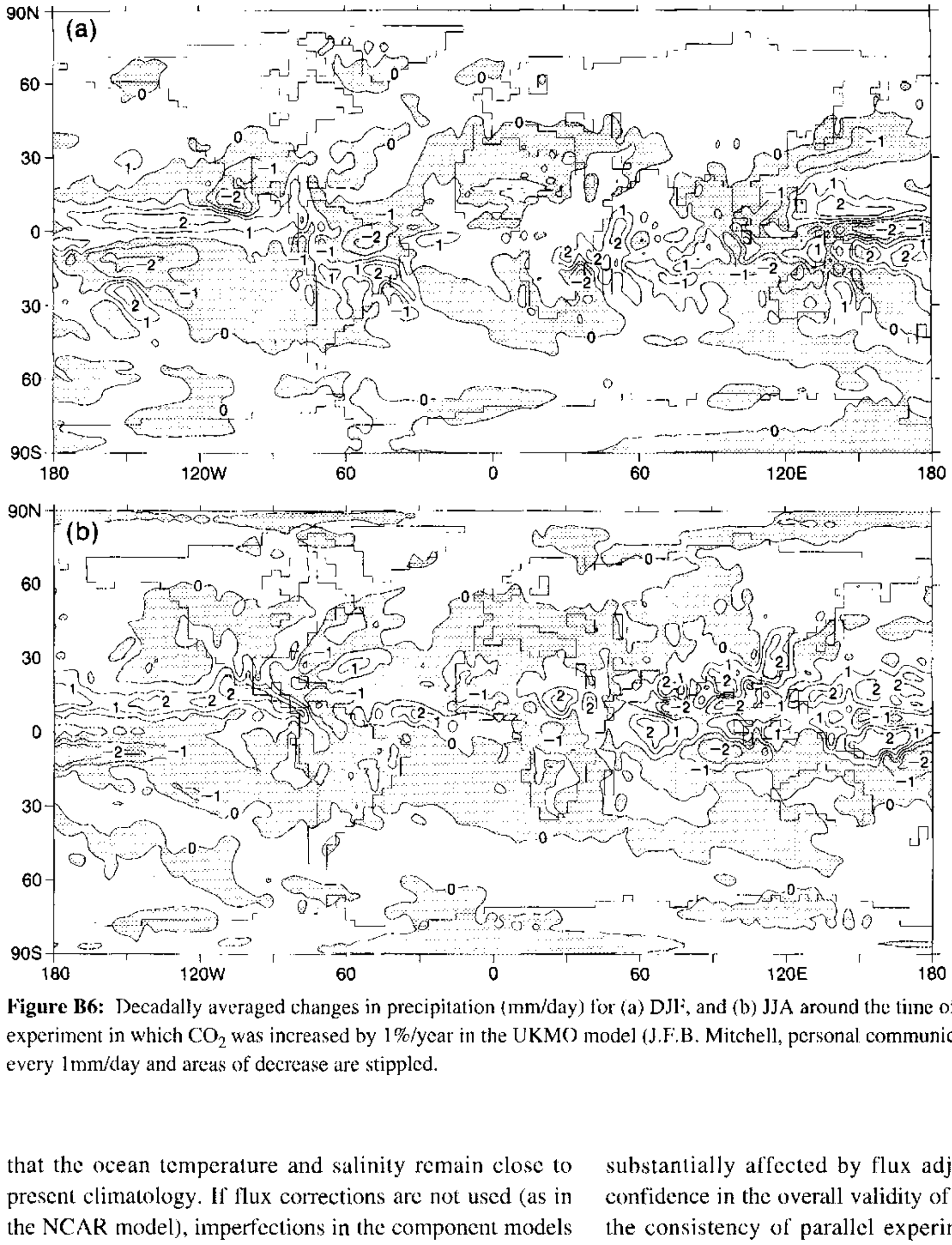

Figure B6: Decadally averaged changes in precipitation (mm/day) for (a) DJF, and (b) JJA around the time of doubling of CO_2 in an experiment in which CO_2 was increased by 1%/year in the UKMO model (J.F.B. Mitchell, personal communication). Contours are every 1mm/day and areas of decrease are stippled.

that the ocean temperature and salinity remain close to present climatology. If flux corrections are not used (as in the NCAR model), imperfections in the component models (and in their interaction) may introduce significant systematic errors in the coupled simulation. Such flux corrections or adjustments represent substantial changes of the fluxes that are exchanged between the component models and, as shown in Figure B10, are comparable in magnitude to the atmospheric fluxes. In the case shown, they tend to be of opposite sign, i.e., the flux correction substantially reduces the net effective flux to the ocean. The use of such adjustments may distort the models' response to small perturbations like those associated with increasing CO_2. On the other hand, the similarities in the responses of the NCAR model without flux corrections and in the three models that use flux corrections indicate that the simulated changes in the models may not be substantially affected by flux adjustment. In addition, confidence in the overall validity of the results is raised by the consistency of parallel experiments with the GFDL model in which the CO_2 undergoes a progressive transient reduction (Manabe *et al.*, 1991,1992).

B2.3 New Equilibrium Results from Atmospheric GCMs and Mixed-Layer Ocean Models

Although increasing attention is being paid to transient simulations with coupled atmosphere-ocean GCMs (see Section B2.2), doubled CO_2 experiments with atmospheric GCMs and mixed-layer ocean models continue to be of interest since they can be carried to statistical equilibrium relatively easily and they provide a benchmark for model sensitivity. The results of several new such experiments completed since the publication of the 1990 IPCC Scientific Assessment (IPCC, 1990) are summarized in

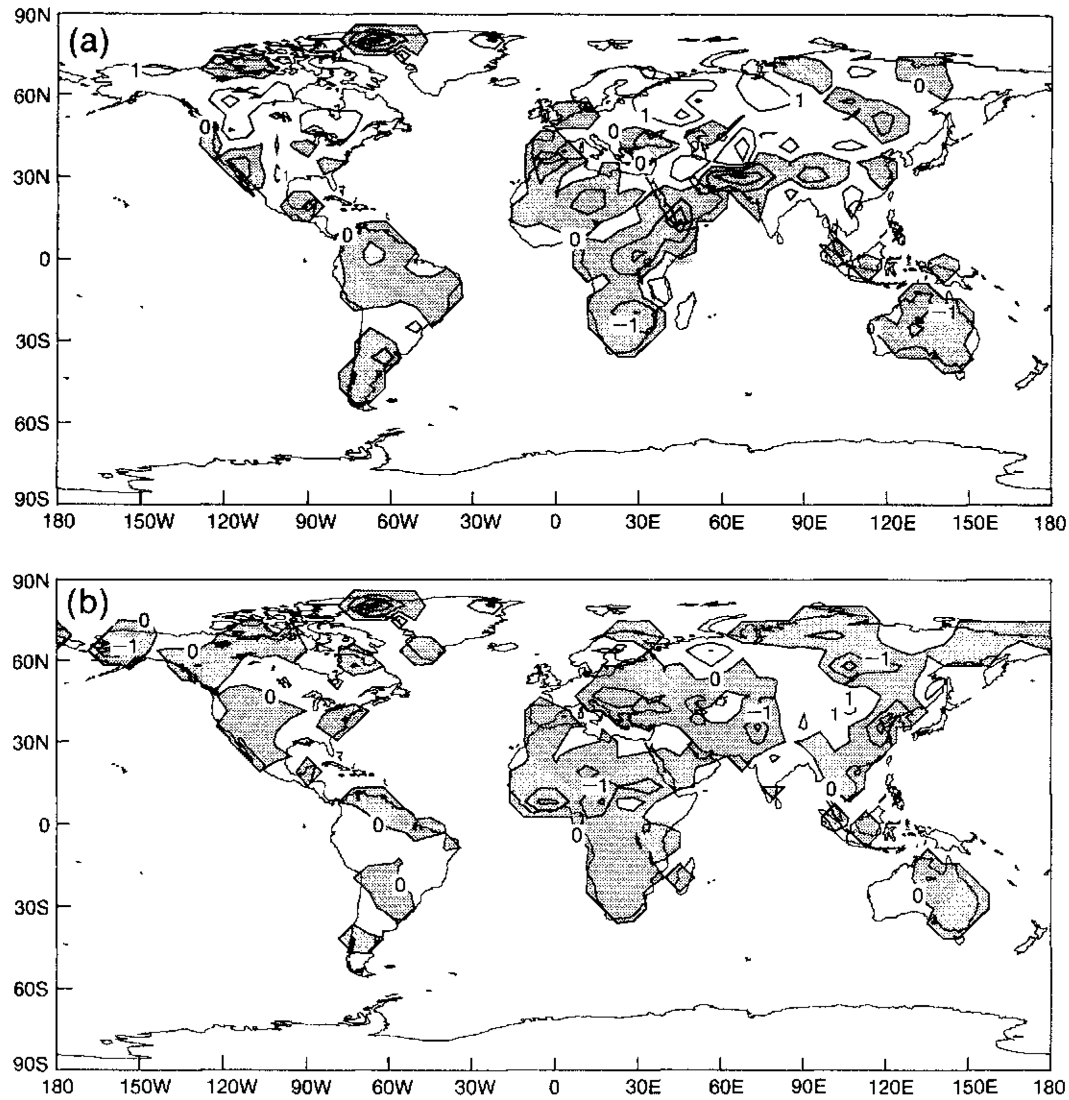

Figure B7: Decadally averaged changes in soil moisture in the MPI model for (a) DJF, and (b) JJA around the time of doubling of effective CO_2. Contour intervals are every cm. Areas of decrease are shaded. (Cubasch *et al.*, 1991.)

Table B2. In general the simulated globally-averaged increases in surface air temperature and precipitation are similar to those found earlier (see IPCC, 1990, Table 2.3 (a); the differences relative to earlier models can be related in some cases to the differences in cloud radiative feedback in the tropics (see Section B3.3)). The largest changes are found in the one new model (LMD) that includes prognostic cloud liquid water and variable cloud optical properties.

New model results suggest that the difference in the vertical distribution of the radiative forcing has implications for the simulations of the control climate as well as for the greenhouse gas-induced warmer climates. As demonstrated by Wang *et al.* (1991b), inclusion of individual trace gases (i.e., not as equivalent CO_2) in a control simulation produces a warmer atmosphere, especially in the tropical upper troposphere. Since this model (and most other GCMs without trace gases) show a cold bias relative to observations of the current climate (Boer *et al.*, 1991a; see Section B5.2), inclusion of the trace gases should tend to reduce this systematic error. Note, however, that the anticipated cooling effect of aerosols (Charlson *et al.*, 1991) is not included in current climate models.

The regional climate responses due to the inclusion of additional greenhouse gases have been examined by Wang *et al.* (1992). In one experiment the trace gases (CO_2, CFCs, methane and nitrous oxide) were represented explicitly, while in the second trace gases were represented by adding the amount of CO_2 which gives an equivalent increase in radiative heating at the tropopause, i.e., using CO_2 as a surrogate for the trace gases. When the trace gases and CO_2 are given values in 2050 according to the IPCC "business-as-usual" scenario (at which time the concentrations of CO_2, CH_4, N_2O, CFC-11 and CFC-12 would have increased by 52, 89, 19, 49 and 121%,

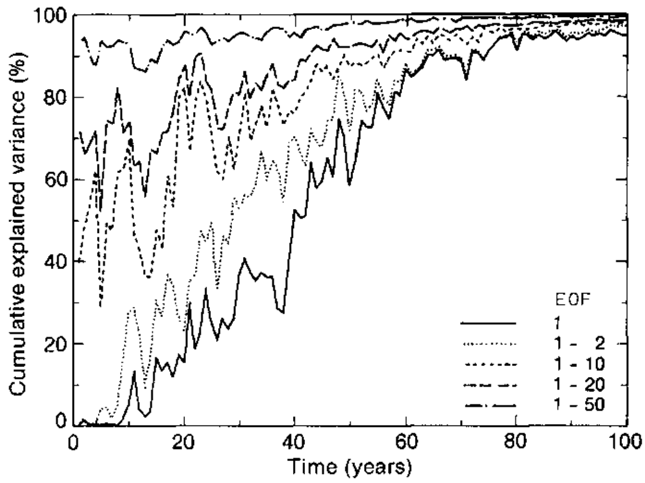

Figure B8: Cumulative spatial variance explained as a function of time and number of EOFs (1, 2, 10, 20 and 50) for Scenario "A" (see text for explanation). The signal is defined as the difference relative to the smoothed initial state (average over years 1-10) of the control run of the MPI model. Results are for annually averaged 2m temperature. (From Cubasch *et al.*, 1991.)

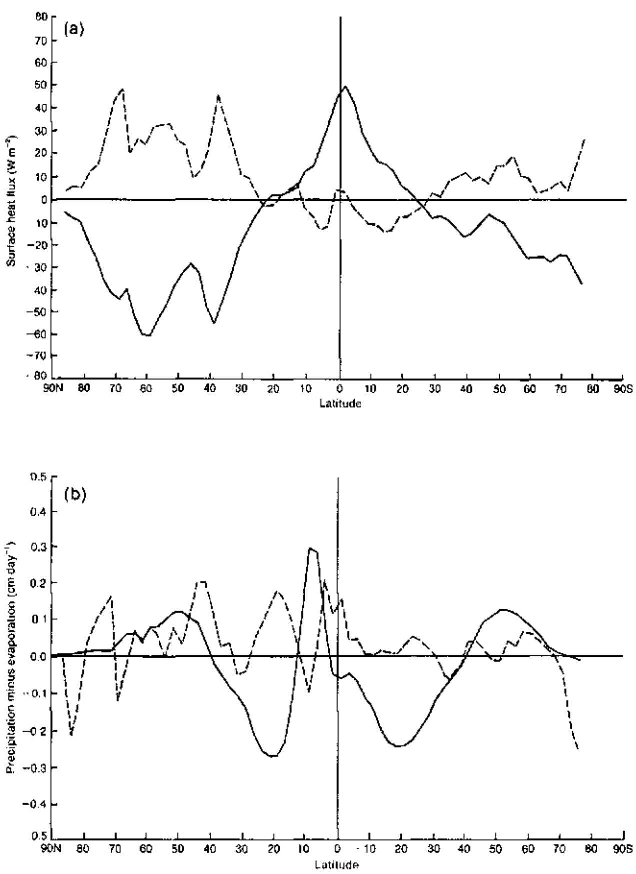

Figure B10: The zonal mean of the surface fluxes of (a) heat, and (b) precipitation minus evaporation, simulated by the atmosphere in the UKMO coupled GCM (solid line) and the flux corrections (dashed line) added during the course of the coupled model integration. (From Murphy, 1990.)

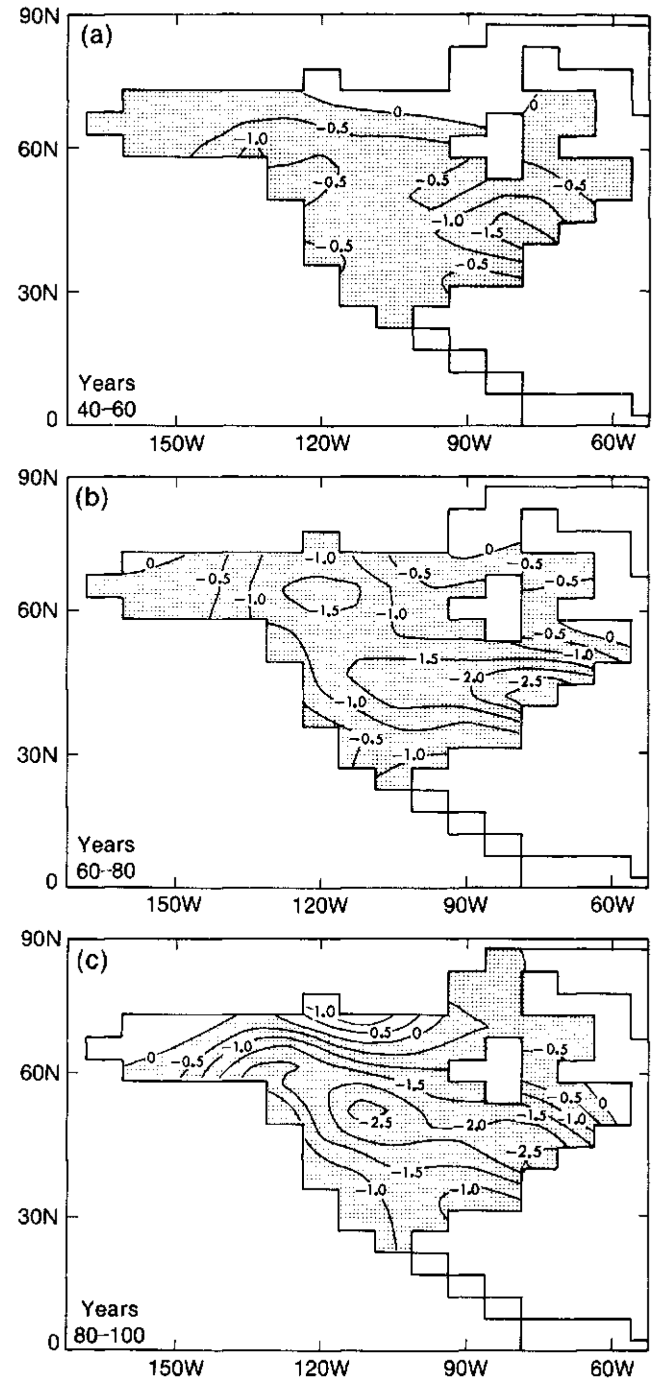

Figure B9: Distribution of the change in JJA soil moisture (cm) simulated during successive 20-year periods over North America relative to a control by the GFDL coupled model with a 1% per year (compound) increase of CO_2: (a) years 40 to 60; (b) years 60 to 80, and (c) years 80 to 100. (From Manabe *et al.*, 1992.)

***Table B2**: Summary of results from new equilibrium simulations for doubled CO_2 with atmospheric GCMs with a seasonal cycle and a mixed-layer ocean.*

Group	Investigators	Year	Resolution	Diurnal cycle	Convection	Cloud scheme	Cloud optical property	ML depth (m)	Ref (CO_2) (ppm)	ΔT (°C)	ΔP (%)	Simulation length (years)
BMRC	McAvaney *et al.*	1991	R21 L9	yes	PC	RH	fixed	50	330	2.2	3.0	15
YALE/ CCM1	Oglesby & Saltzman	1990	R15 L12	no	MCA	RH	fixed	50	330	3.8	n.a.	100
SUNY A/ CCM1	Wang *et al.*	1991b	R15 L12	no	MCA	RH	fixed	50	330	4.2	8.3	20
CSIRO	Dix *et al.*	1991	R21 L9	yes	PC	RH	fixed	50	330	4.8	10	30
NCAR/ CCM	Washington & Meehl	1991	R21 L9	no	MCA	RH	fixed	50	330	4.5	5	50
SUNY A/ CCM1	Wang *et al.*	1992	R15 L12	no	MCA	RH	fixed	50	354 †	4.0 †† 3.9 †††	7.1 †† 6.9 †††	100 20
LMD	Le Treut *et al.*	1992	5°x7.5° L11	no	MCA & KUO	CW	variable	50	320	5.3	8	20
IAP	Wang *et al.*	1991a	4°x5° L2	yes	MCA	RH	fixed	60	324	1.7	2.5	16

PC - Penetrative convection; MCA - Moist convective adjustment; RH - Relative humidity; CW - Cloud water; ML - Mixed layer.
L - number of vertical levels; R - number of spectral waves (rhomboidal truncation).
† - 1990 CO_2 concentrations in SA90, which also includes the explicit trace gas concentrations (CH_4) = 1.717 ppm, (N_2O) = 0.309 ppm, ($CFCl_3$) = 0.280 ppb and (CF_2Cl_2) = 0.484 ppb.
†† - Changes between equilibrium solutions at 2050 relative to 1990, for which the concentration of CO_2 increased by 52%, CH_4 by 89%, N_2O by 19%, $CFCl_3$ by 49% and CF_2Cl_2 by 121% according to SA90.
††† - As in †† except for equivalent CO_2. The concentration of equivalent CO_2, 660 ppm, is calculated from equating the increased global mean longwave radiative forcing for the troposphere-surface system.

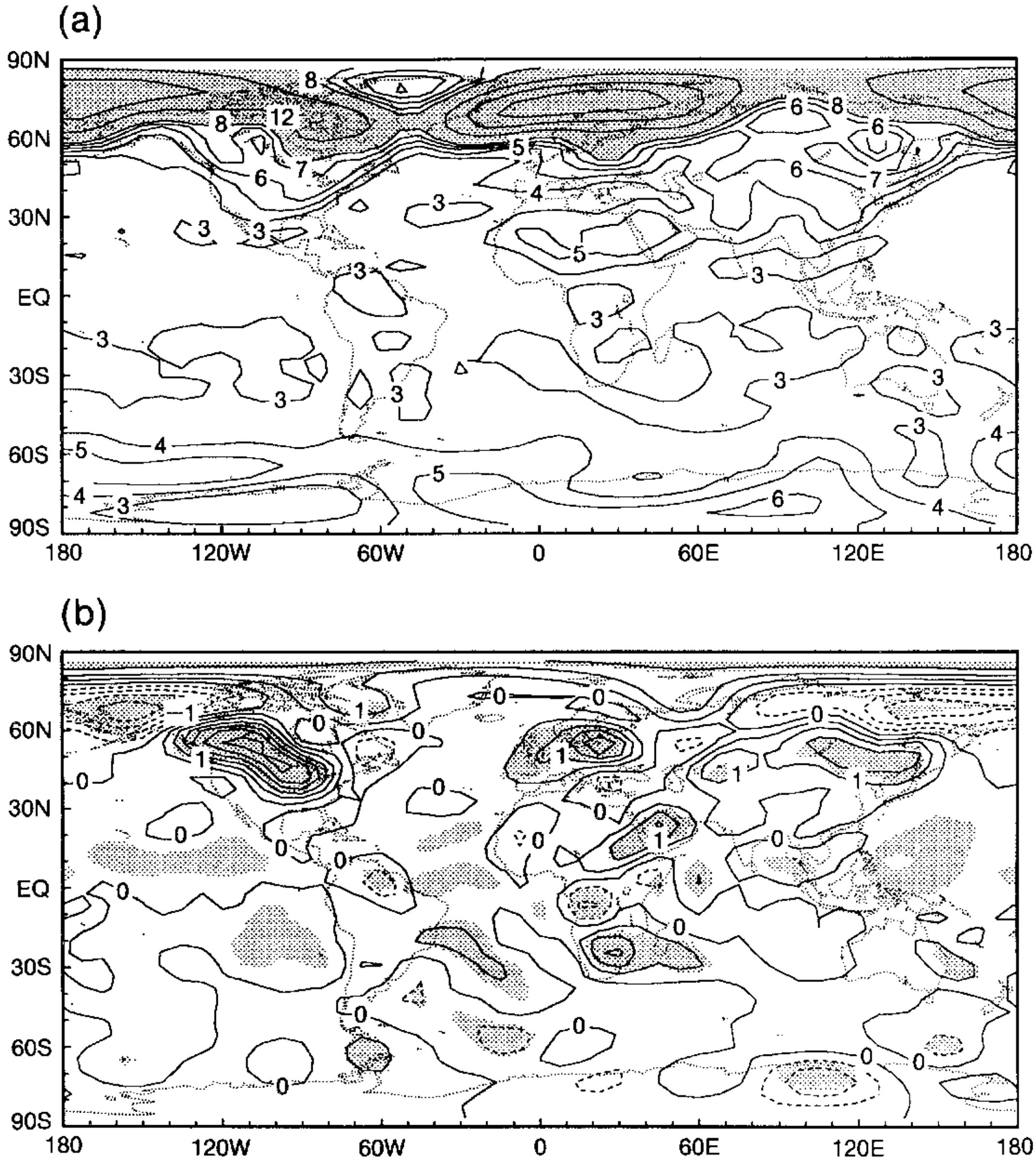

Figure B11: Effect on equilibrium surface temperature change (DJF) using actual trace gases (Scenario "A", 1990-2050) instead of radiatively equivalent increases in CO_2 (Wang *et al.*, 1992): (a) warming using actual trace gases. Contour interval is 1°C except in stippled region (>8°C) where the interval is 4°C; (b) difference due to use of actual trace gases rather than effective CO_2. Contour interval is 0.5°C, negative contours are dashed and areas of statistical significance are stippled.

respectively, relative to 1990), the equilibrium simulations show nearly identical annual and global mean surface warming and increased precipitation (see Table B2). However, the regional pattern of climate changes between the two experiments is different (Figure B11). Although the physical mechanisms for these differences have not yet been established, such results suggest that atmospheric trace gases other than CO_2 should be included explicitly in future GCM simulations of both the present and future climates.

B2.4 Regional Climate Simulations

Although confidence in the regional changes simulated by GCMs remains low, progress in the simulation of regional climate is being made with both statistical and one-way nested model techniques. In both cases the quality of the large-scale flow provided by the GCM is critical.

The horizontal resolution in current general circulation models is too coarse to provide the regional scale information required by many users of climate change simulations. The development of useful estimates of regional scale changes is dependent upon the reliability of GCM simulations on the large-scale. Recognizing that simple interpolation of coarse-grid GCM data to a finer grid is inadequate (Grotch and MacCracken, 1991), strategies have been developed that involve the projection

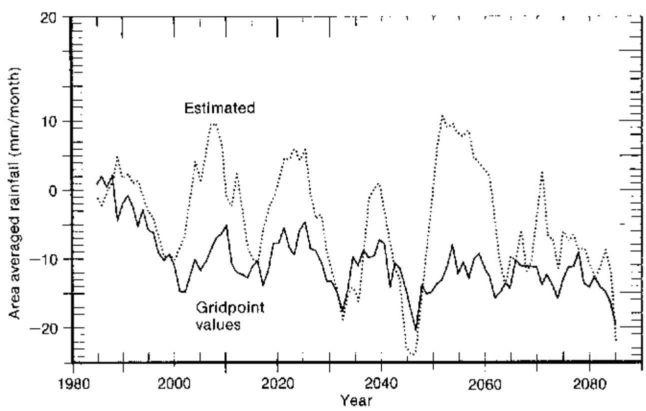

Figure B12: Projected 5-year mean changes in precipitation over the Iberian peninsula (mm/month) in a transient CO_2 experiment. Solid line = grid point values from a GCM; dashed lines = values from a statistical approach with large-scale patterns derived from the GCM. (From Von Storch *et al.*, 1992.)

of large-scale information from GCMs onto the regional scale either by using limited area models with boundary conditions obtained directly from the GCM (the one-way nested approach) or by using empirically-derived relationships between regional climate and the large-scale flow (the statistical approach). In both approaches, however, accuracy is limited by the accuracy of the large-scale flow generated in the GCM, and so they do not avoid the need for improvement in GCM simulations.

Since the preparation of the IPCC (1990) report there has been considerable research using statistical approaches to climate simulation; see, for example, Wigley *et al.* (1990), Karl *et al.* (1990) and Robock *et al.* (1991) amongst others. The statistical technique of Von Storch *et al.* (1991) has successfully reproduced observed rainfall patterns over the Iberian peninsula, and has been applied to an IPCC Scenario A model integration (Cubasch *et al.*, 1991). The resulting changes in the winter rainfall are shown in Figure B12 as compared to the original GCM

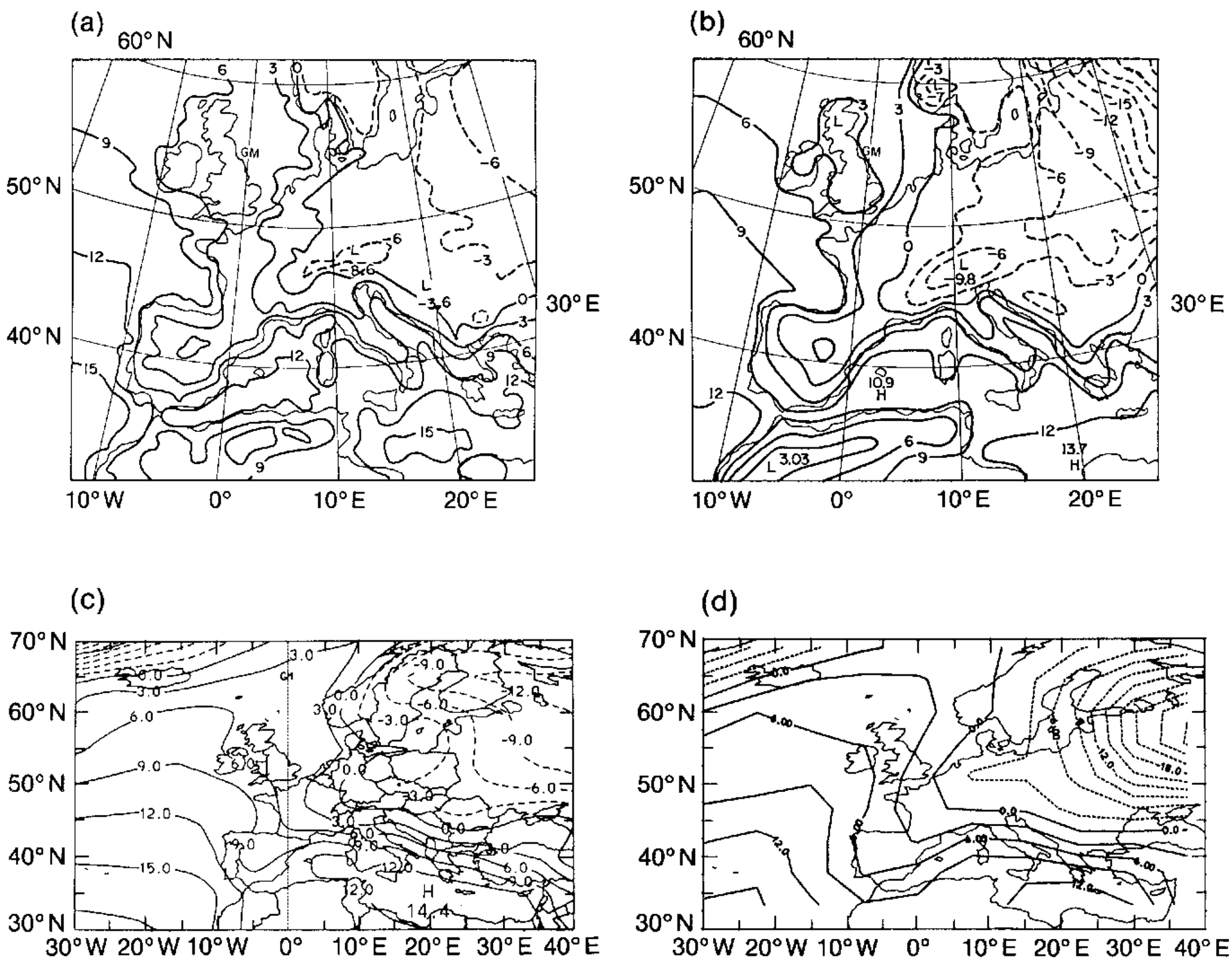

Figure B13: The average January surface air temperature (°C) over Europe: (a) as given by high-resolution observations; (b) as simulated by a mesoscale model nested within the NCAR CCM1 R15; (c) as given by large-scale observations, and (d) as simulated by the NCAR CCM1 alone. (From Giorgi *et al.*, 1990.)

grid-point rainfall amounts; here, however, only about half of the year-to-year variability is explained. Similar research is underway at other institutions (UKMO, BMRC). A recent review of research in these areas has been prepared by Giorgi and Mearns (1991).

The one-way nested modelling approach has been applied by Giorgi *et al.* (1990), whose regional scale climate simulations shown in Figure B13 compare well with high-resolution observations, and show a significant amount of detailed information that was not portrayed in the original GCM. (The limited area model is likely to produce significant improvements only in regions where topographic effects are substantial.) Similar results have been found by McGregor and Walsh (1991) with the CSIRO model and by other groups (MPI, UKMO). The application of the nesting approach to regional climate changes in increased CO_2 experiments is not quite so advanced, although Giorgi *et al.* (1991) have recently used the NCAR doubled CO_2 simulation of Washington and Meehl (1991) in this way; their simulated changes of January surface air temperature for both the original GCM and the GCM plus a one-way nested mesoscale model are shown in Figure B14.

In spite of this progress, the relatively low confidence attached to regional projections of any sort should be emphasized. Analysis by Karl *et al.* (1991) has shown that the observed changes of temperature and the winter-to-summer precipitation ratio over central North America are not consistent with the IPCC (1990) projections, though this inconsistency could be due to factors other than model errors.

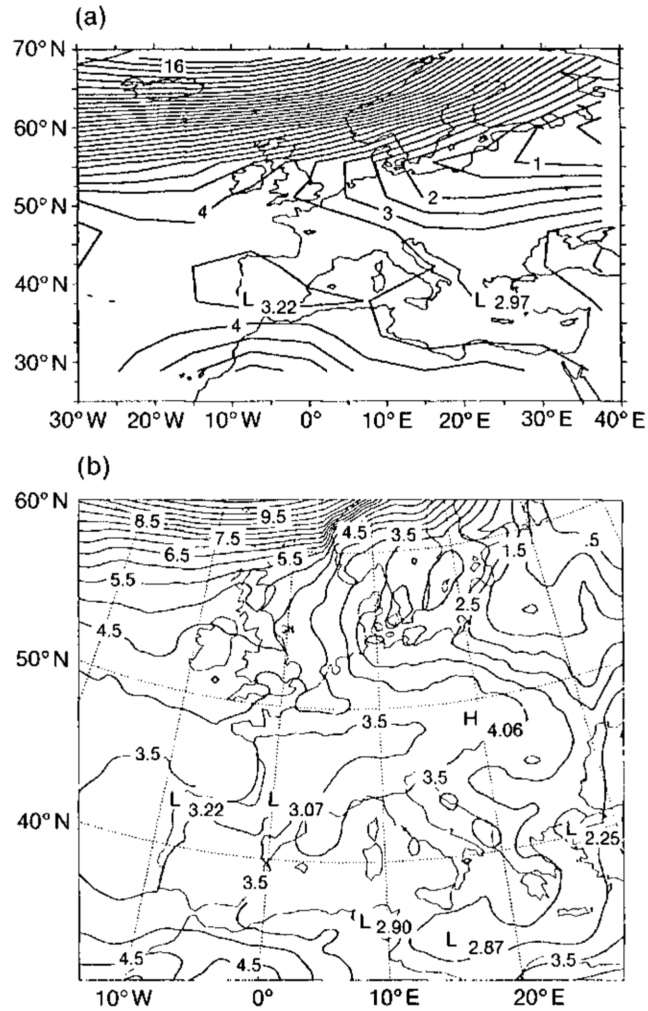

Figure B14: The distribution of changes in January surface air temperature over Europe as a result of doubled CO_2: (a) the temperature change (°C) given by the NCAR GCM relative to the GCM control; (b) the temperature change (°C) given by the nested mesoscale model relative to the nested model control. (From Giorgi *et al.*, 1991.)

B3 Advances in the Analysis of Climate Feedbacks and Sensitivity

B3.1 Introduction

While climate simulation continues apace, new attention has been given to the critical role of feedback processes in determining the climate's response to perturbations. In general, water vapour is expected to amplify global warming, while the effect of clouds remains uncertain. Overall, there is no compelling evidence to change earlier model estimates of the climate's sensitivity to increased greenhouse gases.

B3.2 Water Vapour Feedback

There is no compelling evidence that water vapour feedback is anything other than the positive feedback it has generally been considered to be, although there may be difficulties with the treatments of upper-level water vapour in current models.

The importance of water vapour feedback in climate change has long been recognized (Manabe and Wetherald, 1967) and the dependence of the current greenhouse effect on water vapour has been documented by Raval and Ramanathan (1989) using satellite data from regions of clear skies. The theoretical maximum concentration of water vapour is governed by the Clausius-Clapeyron relation, and increases rapidly with temperature (about 6%/°C); this is the physical basis for the strong positive water vapour feedback seen in present climate models (whereby increases in temperature produce increases in atmospheric water vapour which in turn enhance the greenhouse effect leading to a warmer climate). Increases in water vapour will be more effective at higher levels where temperatures are lower, and the change in emissivity (or effectiveness in absorbing thermal radiation) will be approximately proportional to the ***fractional*** change in water vapour. (This simple relationship is complicated by absorption by the water vapour continuum in the tropics, so that for a given percentage increase in water vapour, the

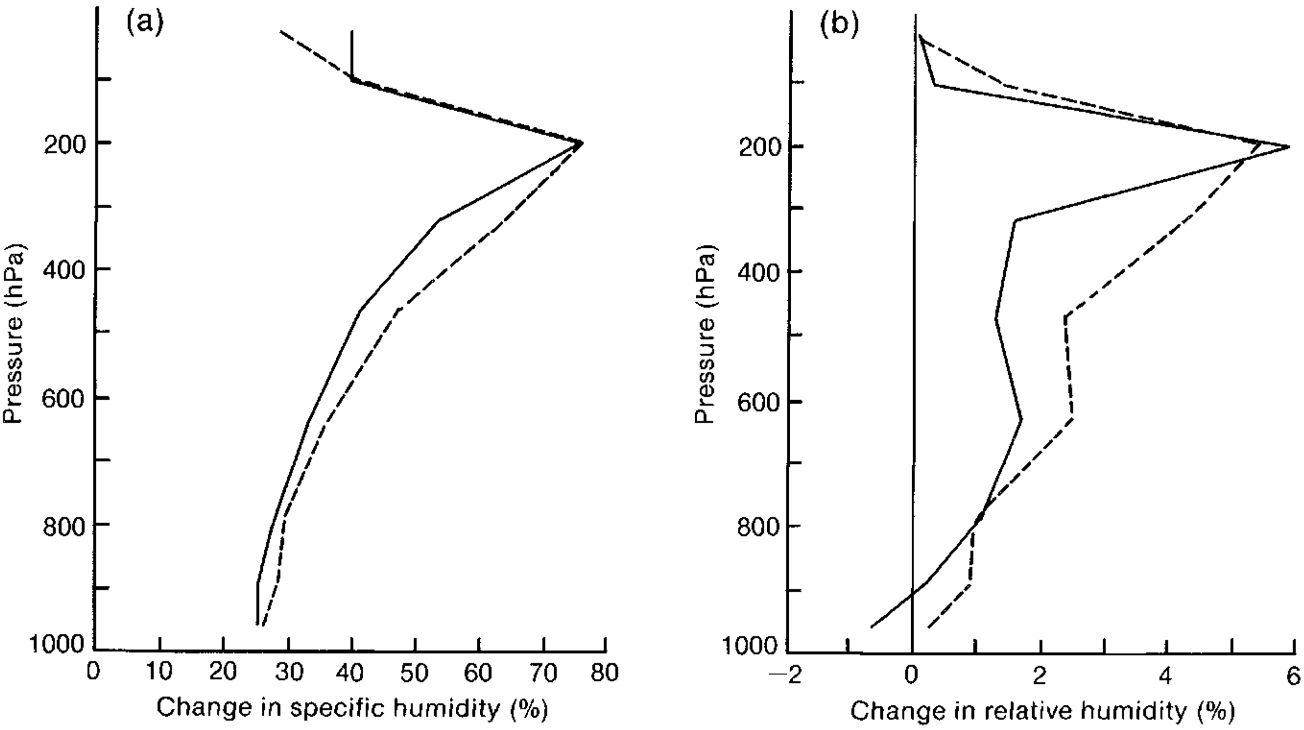

Figure B15: Vertical profiles of the global mean change in (a) specific humidity, and (b) relative humidity, simulated by the GISS GCM in a warm climate (+2°C SST change) relative to a cold climate (-2°C SST change). The full line denotes the standard version of the model and the dashed line denotes an improved version. (From Del Genio *et al.*, 1991.)

biggest contribution to increasing the greenhouse effect occurs from about 500 hPa in cold regions, but from about 800 hPa in the tropics (Shine and Sinha, 1991).)

The main source of water vapour is evaporation over the oceans, and humidities in the oceanic boundary layer are close to saturation, with the concentration decreasing rapidly with height. Both convective and large-scale motions and the evaporation of precipitation help to maintain the moisture distribution of the middle troposphere (Betts, 1990; Pierrehumbert, 1991; Kelly *et al.*, 1991). The generally strong vertical gradients of water vapour may, however, lead to systematic errors in models with parametrized vertical moisture diffusion or coarse vertical resolution. Consequently, the large-scale transport of moisture must be treated carefully in GCMs lest spurious sources and sinks of moisture occur (Rasch and Williamson, 1990, 1991; Kiehl and Williamson, 1991). Nevertheless, the overall realism of the outgoing clear-sky radiative fluxes and of the simulated mean tropospheric temperatures suggests that the models' simulated tropospheric water vapour concentrations are not greatly in error (Cess *et al.*, 1990; Boer *et al.*, 1991a).

Under global warming, there is general agreement that the atmospheric boundary layer would become moister, and as a result of both large-scale and convective mixing, the humidity in the upper troposphere is also likely to increase. All current GCMs simulate a strong positive water vapour feedback (Cess *et al.*, 1990) with ***relative*** humidity remaining constant to a first approximation (Mitchell and Ingram, 1992). Lindzen (1990) noted, however, that at warmer temperatures, tropical convective clouds might detrain moisture at higher, colder temperatures, and thus provide less water to the atmosphere directly. Del Genio *et al.* (1991) diagnosed results from two versions of the GISS GCM, and found that in a warmer climate, cumulus-induced subsidence indeed tends to increase drying as anticipated by Lindzen, but found that this effect is offset by increased moistening as a result of upward moisture transport by large-scale eddies and by the tropical mean meridional circulation (Figure B15).

Estimates of the strength of the water vapour feedback can be made from observations, although they are not independent of the effects of atmospheric motions. As noted earlier, Raval and Ramanathan (1989) computed the dependence of the normalized clear-sky greenhouse effect derived from ERBE satellite data against sea surface temperature from different locations and seasons. The enhancement due to ascent in the intertropical convergence zone (at temperatures around 303K) and the reduction due to subsidence (at temperatures around 295K) can be seen in Figure B16. The overall positive slope suggests that the greenhouse effect indeed increases with temperature, although work by Arking (1991) indicates that some of the

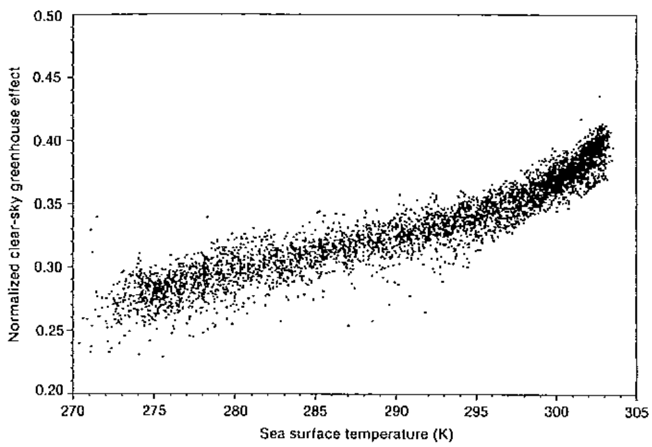

Figure B16: The dependence of the global clear-sky greenhouse effect normalized by infrared emission (g) as a function of sea surface temperature (T, in K) as given by ERBE data during April 1985.

contribution to the slope is due to changes in the vertical profile of temperature. Results from several GCMs have been found to agree with Raval and Ramanathan's results and with other satellite data. For example, Rind *et al.* (1991) compare the simulated July to January relative humidity difference simulated by the GISS model in cloud-free areas with the corresponding observations from the SAGE II instrument (which retrieves water vapour concentration between the stratopause and cloud top altitude). As shown in Figure B17, in the region above 700 hPa the model is in rather good agreement with the observations, especially in the tropical areas that are dominated (in the clear-sky areas) by subsidence from penetrating cumulus convection. (Although the simulated relative humidity was too high by about 30%, this need not necessarily exaggerate the water vapour feedback, as the feedback is approximately proportional to the fractional change in water vapour content.) The model also simulates the increased relative humidity in the middle and upper troposphere in summer and that in the convectively-dominated tropical western Pacific relative to the largely non-convective eastern Pacific.

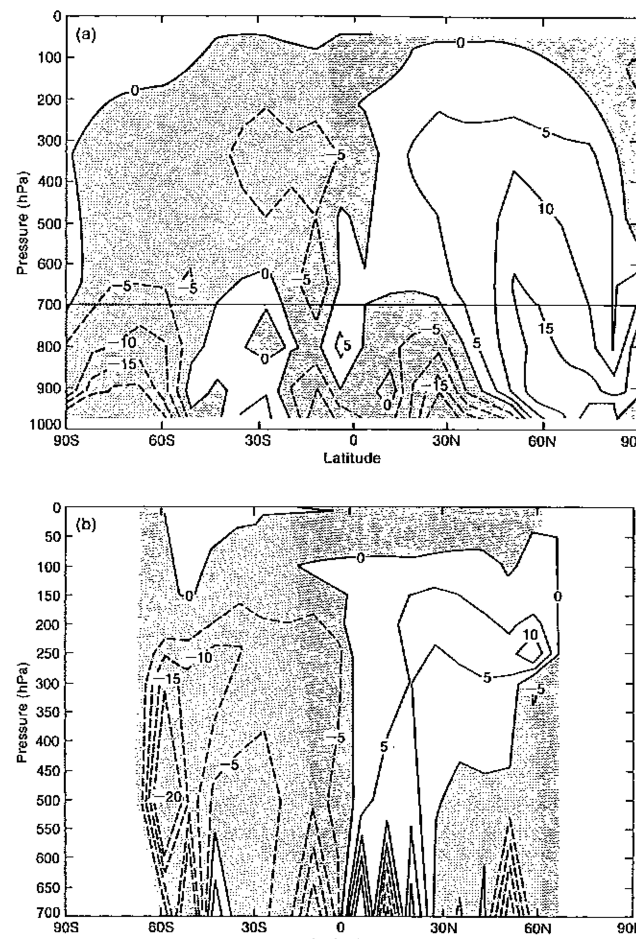

Figure B17: The change in zonally-averaged relative humidity (%) between July and January: (a) as simulated by the GISS GCM, and (b) as observed over a 5-year period by the satellite-borne SAGE II instrument. (From Rind *et al.*, 1991.)

B3.3 Cloud Feedback

The effects of clouds remain a major area of uncertainty in the modelling of climate change. While the treatment of clouds in GCMs is becoming more complex, a clear understanding of the consequences of different cloud parametrizations has not yet emerged.

Cloud feedback is the term used to encompass effects of changes in cloud and their associated radiation on a change of climate, and has been identified as a major source of uncertainty in climate models (see IPCC (1990) Sections 3 and 11). This feedback mechanism incorporates both changes in cloud distribution (horizontal and vertical) and changes in cloud radiative properties, although these cannot always be separated.

Observational data on the variation of cloud radiative properties in terms of other variables, such as cloud water content and temperature are relatively scarce. The observations of Feigelson (1978) over the Soviet Union, used by Somerville and Remer (1984), suggest a general increase of cloud optical thickness for temperatures below about 0°C, for which Betts and Harshvardhan (1987) provided plausible supporting thermodynamic arguments. The implication is that low clouds become more reflective as temperature increases, thereby introducing a negative feedback, while the feedback from high clouds depends upon their height and coverage and could be of either sign. The feedback effect of changes in temperature on low-cloud optical properties has recently been considered by Tselioudis *et al.* (1991) in a study based on ISCCP data. They infer a decrease in low-cloud optical depth with increasing temperature for warm continental and almost all maritime clouds. If this correlation holds for global warming, then it represents a positive feedback mechanism associated with low cloudiness in these regions. (Note, however, that other processes can lead to local changes in cloud optical depth, and the correlation with temperature may be fortuitous.) Feigelson's data, however, suggest little change in optical depth with increasing temperature for water clouds.

Ramanathan and Collins (1991) have discussed a negative feedback mechanism associated with the production of highly reflective cirrus clouds as a consequence of an increase in tropical convective activity associated with the 1987 El Niño. The importance of this mechanism for the case of greenhouse warming remains to be fully investigated, as the circulation changes in the tropics accompanying global warming are unlikely to be the same as those during an El Niño (Mitchell, 1991; Boer, 1991).

Cloud feedback effects have been investigated in GCM studies in order to understand some of the consequences of the different ways in which clouds and cloud optical properties are represented in models. Clouds are represented either diagnostically as a function of relative humidity, or prognostically by carrying an equation for cloud water. Additionally, the optical properties of the clouds may be specified to remain fixed or they may be parametrized to change as the climate changes (see Table B1; also Table 3.2a, in IPCC 1990). As noted in IPCC 1990, Cess *et al.* (1990) included examples of each of these possibilities in a study of 19 GCMs, and found that the simulated cloud feedback varied from slightly negative to strongly positive.

Cloud feedback due to changes in cloud amount and/or distribution may be different, depending on whether clouds are specified in terms of relative humidity or derived from an explicit cloud water prognostic equation. Differences may also arise from the differing treatment of clouds in explicit cloud water schemes. For example, in the UKMO model (Senior and Mitchell, 1992a) increasing the lifetime of water cloud relative to ice cloud weakens the cloud feedback (tending to reduce climate sensitivity), while in the LMD model (Li and Le Treut, 1991) lowering the temperature at which ice cloud forms strengthens the cloud feedback. In the UKMO model the parametrization of cloud properties (as a function of cloud water content) apparently results in a modest positive contribution to the total feedback in the tropics due to increased infrared longwave emissivities. In the CCC model, on the other hand, the parametrization of cloud properties (as a function of temperature) gives a negative contribution to the feedback in tropical regions due to the increase in cloud albedo. The reduced tropical solar input at the surface in the perturbed climate thus moderates both the warming and the increase in evaporation, so that the hydrological cycle does not strengthen as much as in other models (Boer, 1991).

B3.4 Surface Albedo Feedback

The conventional explanation of the amplification of global warming by snow feedback is that a warmer climate will have less snow cover, resulting in a darker surface which in turn absorbs more solar radiation. A recent analysis suggests that snow-albedo-temperature feedback processes in models are somewhat more complex than this view would indicate.

Using a methodology developed for an earlier study of GCM cloud feedbacks (Cess *et al.*, 1990), a perpetual April simulation was used with an imposed globally-uniform SST perturbation of 2°C relative to a prescribed climatological SST distribution. To isolate snow feedback, two such April climate change simulations were performed, one in which the snow cover was held fixed and one in which the snow line was allowed to retreat following the imposition of the SST anomaly. The results of these simulations are shown in Figure B18 in terms of the climate feedback or sensitivity parameter λ that was defined in Section 3.3.1 of the IPCC (1990) report. (An increase in λ represents an increased climate change caused by a given climate forcing.) There are clear differences among current GCMs in their response to changes in snow cover as measured by the ratio λ/λ_S (where λ_S is the sensitivity parameter for fixed snow). Here a value of $\lambda/\lambda_S > 1$ indicating a positive snow feedback is found in most models. Also shown are the corresponding feedbacks in the case of an equivalent cloudless atmosphere, found by separately averaging the models' clear-sky radiative fluxes at the top of the atmosphere (TOA) in order to isolate the effects of clouds on snow

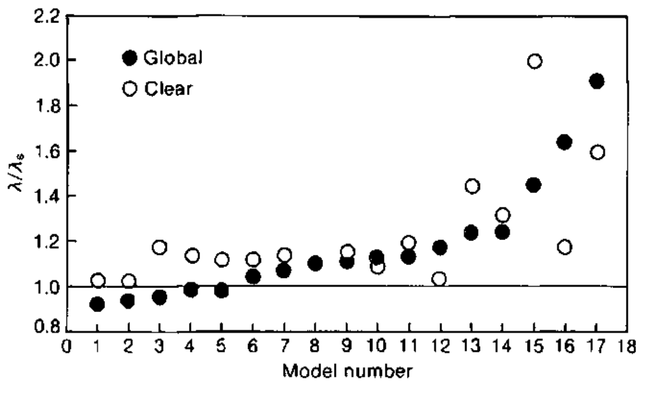

Figure B18: The snow feedback in terms of the ratio of the sensitivity parameter (λ) to that with fixed snow (λ_S), as simulated in 17 atmospheric GCMs. The full circles denote global values while the open circles are for clear-sky conditions. (From Cess *et al.*, 1991.)

feedback. The clear-sky values of λ/λ_s range from negligible snow feedback to as much as a two-fold feedback amplification.

In general, clouds are thought to reduce positive snow feedback by shielding the TOA albedo change. This mechanism, however, cannot account for the feedback sign reversals or the cloud-induced amplification of positive snow feedback exhibited by some models. As discussed by Cess *et al.* (1991), the sign reversal in some cases is caused by cloud redistribution, while in others it occurs as a consequence of cloud-induced longwave interactions; there is also a significant longwave feedback associated with snow retreat in some models. These results are complementary to those of Cohen and Rind (1991), who found that the suppression of surface air temperature changes by positive snow anomalies is considerably reduced by a negative feedback involving the non-radiative components of the surface energy budget.

Uncertainties in the size of the surface albedo feedback associated with sea-ice remain. As Ingram *et al.* (1989) noted, estimates of the strength of feedback depend not only on the model used, but also on the method used to provide the estimate. Covey *et al.* (1991) estimated an upper limit to sea-ice-albedo feedback by carrying out simulations in which the albedo of sea-ice was changed to that of the open ocean. They found an enhancement of globally and annually averaged absorption of radiation of 2 to $3Wm^{-2}$, comparable to the change on doubling CO_2. They concluded that for a warming of the magnitude expected on doubling CO_2, sea-ice-albedo feedback is likely to be smaller than the feedback from water vapour and potentially smaller than that from clouds. This is consistent with Ingram *et al.* (1989) who estimated a feedback of $0.2\text{-}0.3Wm^{-2}K^{-1}$ from sea-ice-albedo changes, whereas estimates of the strength of water vapour feedback are typically about $1.5Wm^{-2}K^{-1}$ (see for example, Mitchell, 1989). Meehl and Washington (1990) found that changing their sea-ice albedo formulation to one which was liable to induce ice melt led to both a warmer control climate (about 1K) and a greater sensitivity (about 0.5K) to doubling CO_2. Further discussion of sea-ice modelling is given in Section B5.4.

B3.5 Climate Sensitivity

There is no compelling new evidence to warrant changing the equilibrium sensitivity to doubled CO_2 from the range of 1.5 to 4.5°C as given by IPCC 1990.

The climate sensitivity is defined as the equilibrium change in global average surface air temperature due to a doubling of CO_2 (Section 3.2 of IPCC, 1990), and is a measure of the response of a climate model to a change in radiative forcing. The climate sensitivity may be thought of as partly a direct radiative effect (estimated to be of the order of 1.2°C for a doubling of CO_2) and partly the effect of feedbacks that act to enhance or suppress the radiative warming. The IPCC "best guess" for the climate sensitivity is 2.5°C, with a range of uncertainty from 1.5 to 4.5°C. Values of the climate sensitivity estimated from recent equilibrium GCM simulations for doubled CO_2 are summarized in Table B2, and generally fall within the range given in Table 3.2a of IPCC 1990. These recent simulations convey little new information as they all (with the exception of the LMD model) employ a relative humidity-based cloud scheme and fixed cloud radiative properties as in earlier models. The coupled model results (at the time of CO_2 doubling) give lower bounds, as discussed in Section B2.2.

Recent additional estimates of the climate sensitivity have been made by fitting the observed temperature record to the evolution of temperature produced by simple energy-balance climate/upwelling-diffusion ocean models, assuming that all the observed warming over the last century or so was due solely to increases in greenhouse gases (see IPCC (1990), Sections 6 and 8). Schlesinger *et al.* (1991) estimate a climate sensitivity of 2.2±0.8°C allowing for the effect of sulphate aerosols, while Raper *et al.* (1991) obtain a value of 2.3°C with a larger range of uncertainty due to their allowance for natural variability (see Wigley and Raper, 1990) and they estimate a value of 1.4°C with no aerosol effect. The effect of including sulphates (or other factors that could act to oppose the greenhouse warming) in such calculations is to increase the estimated climate sensitivity, since the observed climate warming is then ascribed to a reduced radiative forcing (see Sections A2.6 and C4.2.4).

In summary, there have been a number of further studies that have a bearing on estimates of climate sensitivity. New equilibrium GCM simulations have widened the range slightly to 1.7°C (Wang *et al.*, 1991a) and 5.4°C (Senior and Mitchell, 1992a), but no dramatically new sensitivity has been found. Energy-balance model considerations bring previous estimates of sensitivity (IPCC, 1990) more in line with the IPCC "best guess".

B4 Advances in Modelling Atmospheric Variability

1. Although studies are incomplete in many respects, the ability of models to replicate observed atmospheric behaviour on a wide range of space and time-scales is encouraging. As noted in IPCC 1990, this ability provides some evidence that models may be usefully applied to problems of climatic change associated with the greenhouse effect.

2. Model experiments with doubled CO_2 give no clear indication of a systematic change in the variability of temperature on daily to interannual time-scales, while the changes of variability for other climate features appear to be regionally (and possibly model) dependent.

B4.1 Introduction

Since the 1990 IPCC report there has been increased emphasis of the analysis of the ability of models to simulate atmospheric variability on time-scales ranging from diurnal to decadal in both control and perturbed climate simulations. Such studies are critical to the assessment of the effects of climate change, since it is increasingly recognized that changes in variability and the occurrence of extreme events may have a greater impact than changes in the mean climate itself (Katz and Brown, 1991). Notwithstanding these results, there is at present only limited confidence in the ability of climate models to infer changes in the occurrence of interannual and regional events.

The following discussion is arranged in order of increasing time-scales from diurnal to decadal. Particular attention is given to new studies that include simulated changes in variability due to doubling atmospheric CO_2, and to relating new results to the findings of IPCC 1990. Other studies are included to document the improvement in the capability of models to simulate selected aspects of atmospheric variability.

B4.2 Diurnal Cycle

Inclusion of the diurnal cycle (now a feature of many GCMs) permits the simulation of high-frequency temporal variations. For example, a recent simulation by Randall *et al.* (1991a) has demonstrated that the diurnal variation of the hydrologic cycle in the tropics can be realistically simulated, while Cao *et al.* (1992) have found that the simulated diurnal range of surface temperature is generally comparable to observations except for a slight underestimate in the mid-latitudes. On doubling CO_2, Cao *et al.* (1992) found a slight decrease in the globally-averaged diurnal range of surface temperatures, as noted in IPCC 1990, although locally they found that factors such as reduced soil moisture, receding snow lines or reduced cloud cover could lead to an increase in the diurnal temperature range with doubled CO_2. Historical observations over 25% of the global land area show a decreased diurnal temperature range, although the reasons for this change, which is largely a result of an increase in minimum temperatures, are not yet clear (see Section C3.1.5).

B4.3 Day-to-Day Variability

In several of the models used in CO_2 studies, the day-to-day variability of surface temperature is greater than that observed, particularly in higher northern latitudes (Meehl and Washington, 1990; Portman *et al.*, 1990). Cao *et al.* (1992) found that in a UKMO model the daily variability of surface air temperature is overestimated in high northern latitudes in winter, but is less than observed over mid-latitude continents in summer.

On doubling CO_2, Cao *et al.* (1992) found a general reduction in day-to-day temperature variability in winter in mid- to high latitudes over North America and over the ocean. The reduction was particularly pronounced in regions where winter sea-ice had melted. Mearns *et al.* (1990), however, found no clear pattern of change in daily temperature variability whereas P. Whetton (personal communication) found general increases in daily temperature variability throughout Australia in both winter and summer. Mearns *et al.* (1990) found a general increase in the day-to-day variability of precipitation due to increased CO_2, whereas Boer *et al.* (1991b) found an overall decline in the daily variance of sea-level pressure due to doubling CO_2, with the largest decrease occurring in the North Atlantic in winter.

B4.4 Extreme Events

The simulation of extreme events is an important aspect of a model's performance, and is closely connected with the question of natural variability. An important extreme event is the occurrence of tropical cyclones (or hurricanes) which GCMs cannot simulate in detail, though they do simulate tropical disturbances (see IPCC 1990, Section 5.3.3). Broccoli and Manabe (1990) and Haarsma *et al.* (1992) found that the spatial and temporal distributions of modelled tropical disturbances are similar to those observed. On doubling CO_2 Haarsma *et al.* (1992) found that the number of simulated tropical disturbances increased, with little change in their average structure and intensity. However, as noted in IPCC 1990, Broccoli and Manabe (1990) found an increase in the number of tropical storms if cloud cover was prescribed, but a decrease if cloud was generated within the model.

In the 1990 IPCC report a consistent increase in the frequency of convective precipitation at the expense of large-scale precipitation was noted, with the implication of more intense local rain at the expense of gentler but more persistent rainfall events. This tendency has been found in recent simulations using the CSIRO model (Gordon *et al.*, 1991; Pittock *et al.*, 1991) and a high-resolution UKMO model (J. Gregory, personal communication). Pittock *et al.* (1991) find a systematic increase in the frequency of heavy rain events with doubled CO_2 (Figure B19) and a consequent decrease in the return period of heavy rainfall (Figure B20). These changes are related to a systematic increase in the frequency of penetrating convection in the tropics and mid-latitudes on doubling CO_2 (see IPCC, 1990).

B4.5 Blocking and Storm Activity

Studies of the simulated variation of storm activity with enhanced CO_2 are difficult to generalize as they use different measures for storminess and address different regions. R. Lambert (personal communication) found a

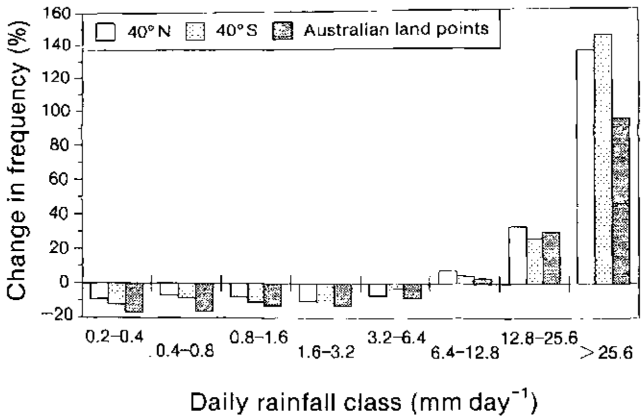

Figure B19: Changes in the frequency of occurrence of daily rainfall classes with doubled CO_2 in the CSIRO model. (From Pittock *et al.*, 1991.)

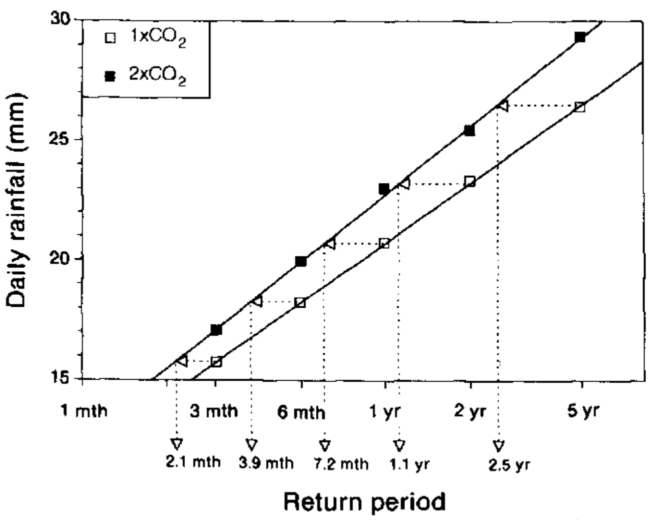

Figure B20: Daily rainfall amount vs. return period in Australia as simulated by the CSIRO model for both control (1×CO_2) and doubled CO_2. (From Pittock *et al.*, 1991.)

significant reduction in the number of cyclonic events in the CCC model between 30°N and the North Pole during winter, but a significant increase in the number of strong cyclones. (Boer *et al.* (1991b) noted a decrease in the variance of 1000 hPa height in the same experiment; see Section B4.3.) Senior and Mitchell (1992b) found a northwest shift in the filtered variance of geopotential height (a surrogate for storm tracks) in the North Atlantic in winter in the high resolution UKMO model. B. Hoskins (personal communication), on the basis of experiments using the time-mean data from Senior and Mitchell's simulations, attributed the changes in storm tracks to changes in the horizontal temperature gradient and increases in atmospheric water vapour. Mullan and Renwick (1990) conducted a detailed analysis of the results of the CSIRO model in the Australian-New Zealand region, and found a slight increase in the number of storms in this region for all seasons.

The simulation of persistent large-scale anomalies such as blocking has proved difficult to forecast with extended-range NWP models (Tibaldi and Molteni, 1990; Miyakoda and Sirutis, 1990), although GCMs can simulate some of the statistical properties of blocking in the Northern Hemisphere as recently shown by Hansen and Sutera (1990) for the NCAR CCM and by Kitoh (1989) for the MRI GCM. Tibaldi (1992) has examined the space-time variability of blocking in the MPI (ECHAM) model, using the blocking index of Tibaldi and Molteni (1990) applied to 5-day mean December to February 500 hPa fields in the Northern Hemisphere. While the variance of both the low- and high-frequency components is reasonably well simulated, the magnitudes are systematically underestimated.

B4.6 Intra-Seasonal Variability

An important intra-seasonal variability is that associated with the 30-60 day oscillation. This oscillation appears to influence onset and break phases of the monsoons, and thus has a considerable impact on the prediction of tropical rainfall (Lau and Peng, 1990). These oscillations may also play a role in the onset of El Niño events. Park *et al.* (1990) and Zeng *et al.* (1990) have shown that some GCMs simulate intra-seasonal variations that are similar in structure to those observed, but they are generally of smaller amplitude.

B4.7 Interannual Variability

As in IPCC 1990, no meaningful change in the interannual variability of temperature has been found with increased CO_2, apart from a general reduction in the vicinity of the winter sea-ice margins.

Some of the models used in climate studies exaggerate the interannual variability of surface temperature in high latitudes in winter (Mearns, 1991, reporting on simulations by Oglesby and Saltzman, 1990; Cao *et al.*, 1992). The simulated changes in the variability of temperature with doubled CO_2 vary from model to model; Rind (1991) found decreases over much of the northern continents in winter in the GISS model, as did Georgi *et al.* (1991) using a regional model driven by output from the NCAR GCM simulations of Washington and Meehl (1991). On the other hand, Mearns (1991) found both increases and decreases in temperature variability, while Cao *et al.* (1992) found that increases were widespread in the tropics and subtropics, and in summer over North America and western Europe, with significant reductions confined to regions where sea-ice receded in winter.

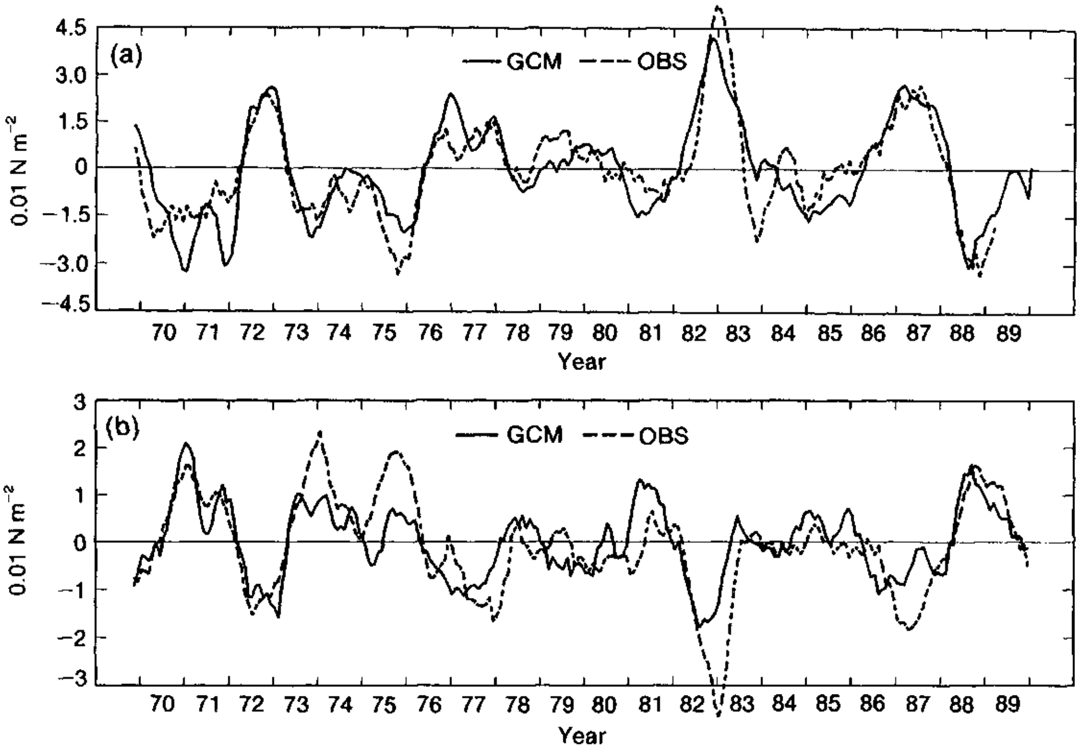

Figure B21: The 1970-1989 time-series of (a) the zonal surface wind stress anomalies averaged between 180°-140°W and 4°N-4°S, and (b) the Southern Oscillation Index. The observed variation is given by the dashed line, and that simulated by the MRI GCM is given by the solid line. (From Kitoh, 1991.)

B4.8 ENSO and Monsoons

A dominant mode of atmospheric variability in the tropics is that associated with tropical sea surface temperature anomalies in the Pacific. These tropical ocean-global atmosphere oscillations are called El Niño-Southern Oscillation, or ENSO. As noted in IPCC (1990), atmospheric models are capable of giving a realistic simulation of the seasonal tropical atmospheric anomalies at least for intense El Niño periods if they are given a satisfactory estimate of the anomalous sea surface temperature (SST) in the tropical Pacific. This conclusion is supported by more recent results (König *et al.*, 1990; Kitoh, 1991; B. Hunt, personal communication). For example, as illustrated in Figure B21, a rather good simulation has been made of the observed interannual variations of zonal surface wind stress over the central equatorial Pacific and of the observed large-scale interannual variations of sea-level pressure as represented by the Southern Oscillation Index. Using prescribed sea surface temperatures from 1987 and 1988, the broad characteristics of monsoon interannual variability have been reproduced (Palmer *et al.*, 1991; Kitoh, 1992). Using the coupled IAP model with observed initial conditions, Zeng *et al.* (1990) and Li *et al.* (1991) report a successful simulation of summer precipitation anomalies in the monsoon region of southeast Asia.

Using tropical ocean-atmosphere models, Lau *et al.* (1991), Nagai *et al.* (1991), Philander *et al.* (1991) and Latif *et al.* (1991) have successfully simulated interannual variations that resemble some aspects of ENSO phenomena, although the amplitude of the simulated SST anomalies is generally less than that observed (as noted in IPCC, 1990) and model resolution appears to have an important influence on the simulated behaviour. Meehl (1990) claims some success in simulating ENSO-like disturbances in a global low-resolution ocean-atmosphere GCM, though it should be recalled that not all tropical SST variations are associated with ENSO. As shown in Figure B22, the patterns of tropical sea-level pressure changes that are characteristic of ENSO in the present climate are also present with doubled CO_2 (Meehl *et al.*, 1991b). The patterns of the tropical precipitation and soil-moisture anomalies are similar in the control and doubled CO_2 cases, with the dry areas becoming generally drier and the wet areas becoming generally wetter with increased CO_2. These results, however, are yet to be confirmed with other models. On increasing the concentration of greenhouse gases in the MPI global coupled model, Lal *et al.* (1992) found no evidence for a significant change in the mean onset date of the Indian Monsoon or for changes in the precipitation in the monsoon region.

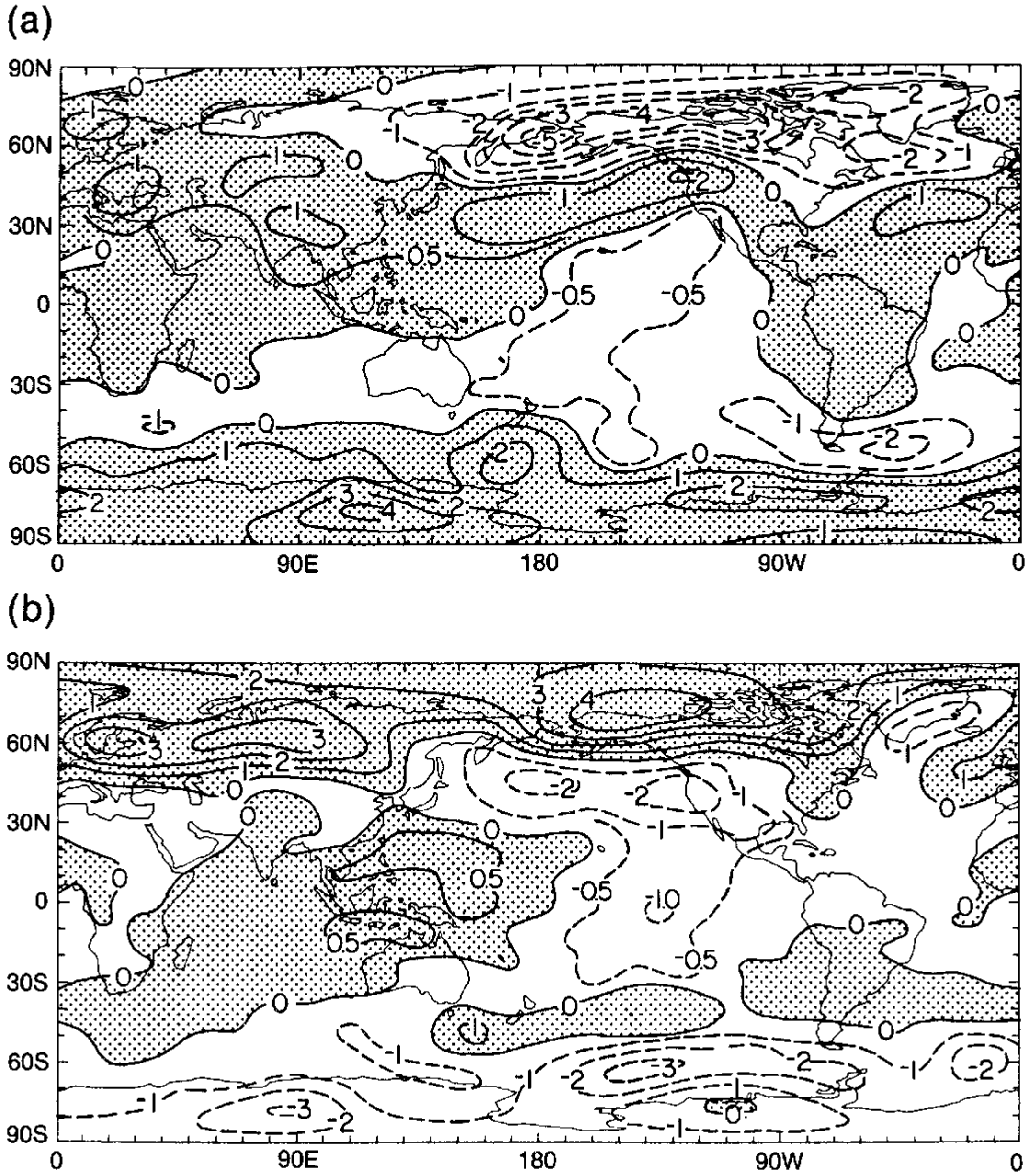

Figure B22: The change in DJF sea-level pressure (hPa) between composite warm ENSO events and the 15-year mean simulated by a coupled ocean-atmosphere model for (a) normal CO_2, and (b) doubled CO_2. (From Meehl *et al.*, 1991b.)

B4.9 Decadal Variability

There are now a sufficient number of model integrations over 50-100 year (and longer) periods to provide preliminary information on the simulation of atmospheric decadal variability. (See Section B4.3 for a discussion of simulated decadal variability in the ocean.) Whether simple models, low-resolution GCMs coupled to an oceanic mixed layer (Houghton *et al.*, 1991; B. Hunt, personal communication) or full global ocean-atmosphere GCMs (Section B2.2) are used, the presence of considerable natural variability (i.e., in the absence of changes in external forcing) on decadal time-scales is characteristically found in model control runs. Typical magnitudes of such decadal variations are several tenths °C in surface air temperature, as illustrated in Figure B2 for the control run of the GFDL coupled model that was used in the transient CO_2 simulation discussed in Section B2.2. In this integration the surface flux corrections applied at the ocean surface were evidently successful in preventing a multi-decadal drift of the mean temperature. Such internal variability may mask at least part of the changes due to increased greenhouse gases.

B5 Advances in Modelling the Oceans and Sea-Ice

B5.1 Introduction

Although there is generally less experience in ocean modelling than there is in atmospheric modelling, since the preparation of the IPCC (1990) report there have been a number of studies with high-resolution and other ocean

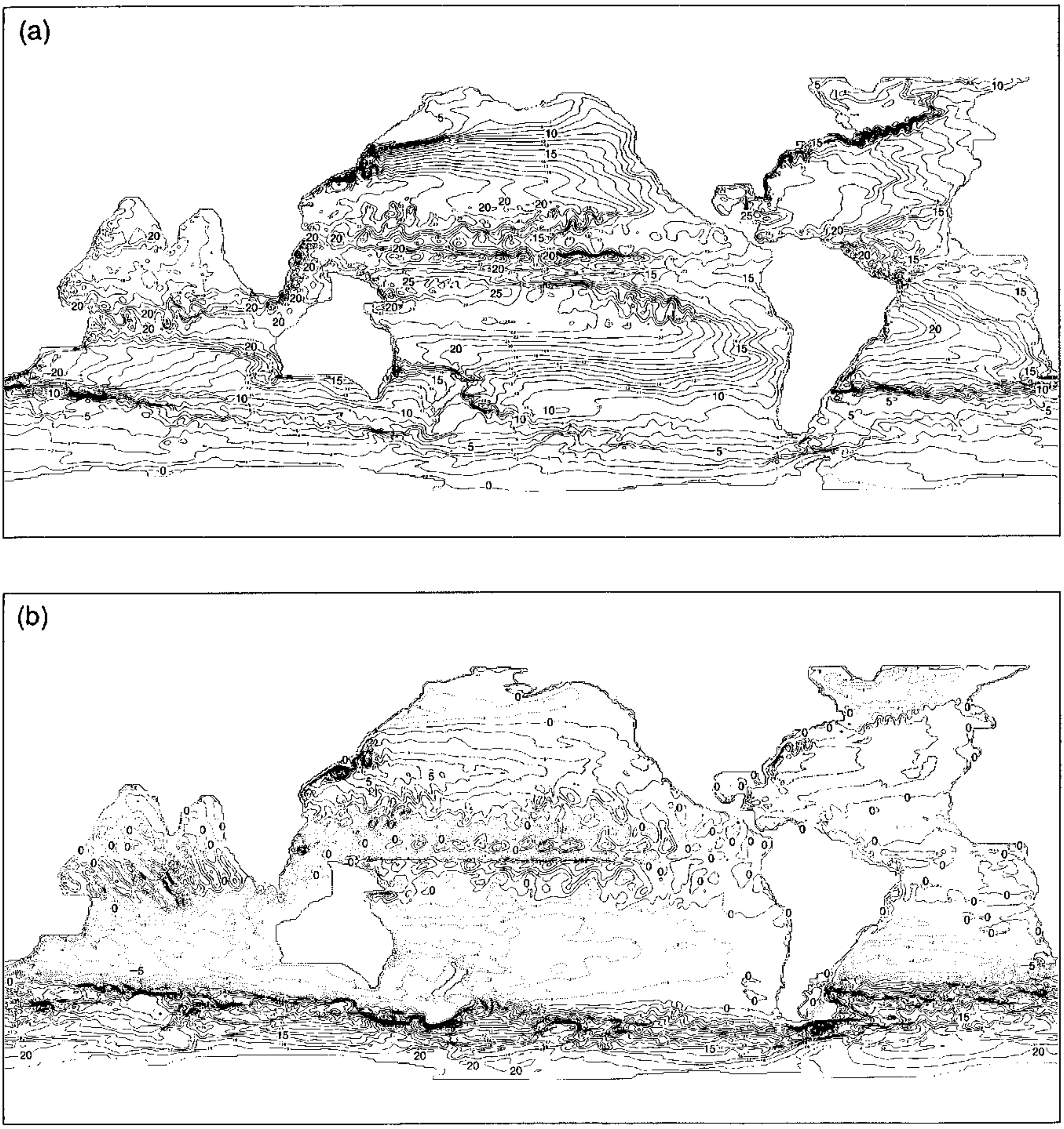

Figure B23: (a) The instantaneous 160m temperature at 1°C intervals, and (b) the volume transport streamlines at 10 Sv intervals, as simulated in an eddy-resolving global ocean GCM with 1/2° resolution. (From Semtner and Chervin, 1988.)

models that have potentially important consequences for the simulation (and identification) of climate change with coupled ocean-atmosphere GCMs (Anderson and Willebrand, 1991).

B5.2 Eddy-Resolving Models

Recent simulations with the WOCE community ocean model (Bryan and Holland, 1989) for the North Atlantic with a 1/3° resolution have yielded a distribution of eddy energy that agrees reasonably well with the GEOSAT altimetric observations, although the simulated amplitudes are generally too small. A qualitatively similar agreement of simulated and observed variability was found by Semtner and Chervin (1988) with a 1/2° model of the global ocean, examples of which are shown in Figure B23. First results from the FRAM 1/4° model for the Antarctic Circumpolar Current (Webb *et al.*, 1991) indicate that the distribution of maximum variability associated with mesoscale eddies is reasonably well simulated, although the eddy minima appear to be too low. The FRAM model also shows a strong meridional (Deacon) cell in the Antarctic Ocean, with a transport of over 20Sv. The upwelling branch of this cell, which is seen in both high- and low-resolution models, occurs at those latitudes where the coupled ocean-atmosphere models show delayed greenhouse gas warming (see Section B2.2). A recent experiment by Böning and Budich (1991) has shown that the simulated eddy energy increases substantially (along with significant variability on time-scales of several years) when the horizontal resolution is increased to 1/6°. This raises the question (of practical importance for climate modelling) whether or not mesoscale ocean eddies must be resolved in climate calculations. Mesoscale effects are so inhomogeneous and so tenuously connected to the larger-scale currents that a simple parametrization of their effects seems unlikely. However, studies using models that omit salinity variations suggest that the total meridional heat transport is only slightly influenced by eddies, since the eddy flux tends to be compensated by a modification of the mean flow induced by the eddies themselves (Bryan, 1991). Further research on the climatic (as opposed to the synoptic) role of ocean eddies is clearly required.

B5.3 Thermohaline Circulation

A great deal of interest has recently focussed on the behaviour of the thermohaline circulation of the North Atlantic, whereby North Atlantic deep water is formed. This sinking is part of a global system of ocean transports known as the Conveyor Belt Circulation (Gordon, 1986). Many aspects of the thermohaline circulation have been simulated in a 1/2° global ocean model (Semtner and Chervin, 1991), and its basic dynamical structure and sensitivity are being illuminated by both theoretical studies and simulations with idealized models. In the present climate, high-latitude winter cooling of relatively saline surface water in the North Atlantic causes it to sink and subsequently to move out of the basin by deep southward currents; this flow is compensated by the northward flow of warm surface water from the tropics, whose salinity may be increased relative to the other high-latitude oceans by enhanced evaporation (Stocker and Wright, 1991).

Coupled ocean-atmosphere models produce sinking in high latitudes in the North Atlantic, though it is not clear how realistically they do so. It may also be noted that Dixon *et al.* (1991) have recently succeeded in simulating the observed spreading of CFCs in the Southern Ocean with the same GFDL coupled model that was used in the transient CO_2 experiments; this increases confidence that the vertical mixing parametrization in the oceanic part of the model is a reasonable approximation. However, it should be noted that heat may affect the vertical stability in the ocean and hence may not behave in the same way as CFCs. Using a model with flux adjustments, Manabe *et al.* (1991) find that the effect of enhanced greenhouse warming is to reduce the rate of deep water formation. (Washington and Meehl (1989) also note a reduction in deep water formation in a model without flux adjustments.) This in turn decreases the flow of warm surface water from the south and probably contributes to a delay in the greenhouse response in the northern North Atlantic noted earlier in Figure B4.

As shown with ocean-only models, variability on decadal and longer time-scales is associated with the advection of salinity anomalies through the deep convection regions (Marotzke and Willebrand, 1991; Weaver and Sarachik, 1991; Weaver *et al.*, 1991; Mikolajewicz and Maier-Reimer, 1990). In the GFDL coupled model (R. Stouffer, personal communication) similar variability on time-scales of order 50 years is found. In an integration started from different initial conditions, a GFDL model (Manabe and Stouffer, 1988) in fact converged to a second equilibrium solution in which the North Atlantic was both fresher and colder (as shown in Figure B24) with a thermohaline circulation that was weakly reversed. The existence of multiple equilibrium states for the global ocean circulation raises the question of their stability, i.e., how easily the system may change from one state to another. Maier-Reimer and Mikolajewicz (1989) used the uncoupled MPI ocean GCM to investigate the Younger-Dryas event and found that the North Atlantic thermohaline circulation was sensitive to relatively small variations in the strength and location of the surface fresh water input, such that a breakdown of the circulation could occur within a few decades. Other recent experiments with zonally averaged ocean models have demonstrated that transitions between multiple equilibrium states can be triggered by relatively modest changes in the large-scale precipitation (Marotzke and Willebrand, 1991; Stocker and Wright, 1991).

The possibility of the natural collapse of the North Atlantic thermohaline circulation has obvious implications for our ability to predict future variations of the climate system on decadal and longer time-scales. The North Atlantic circulation, however, appears to be more robust to change than the ocean-only model results suggest. Atlantic and Pacific deep sea sediment cores show that, with the exception of a possible short interruption during the Younger-Dryas event, no major break in the thermohaline circulation has occurred since the end of the last ice age in spite of the variability that has occurred in both the atmosphere and ocean (see, for example, Keigwin *et al.*, 1991). Preliminary results from a long integration of a coupled ocean-atmosphere model (R. Stouffer, personal communication) suggest that the thermohaline circulation is more stable in a coupled model than in an ocean-only model with restored SSTs.

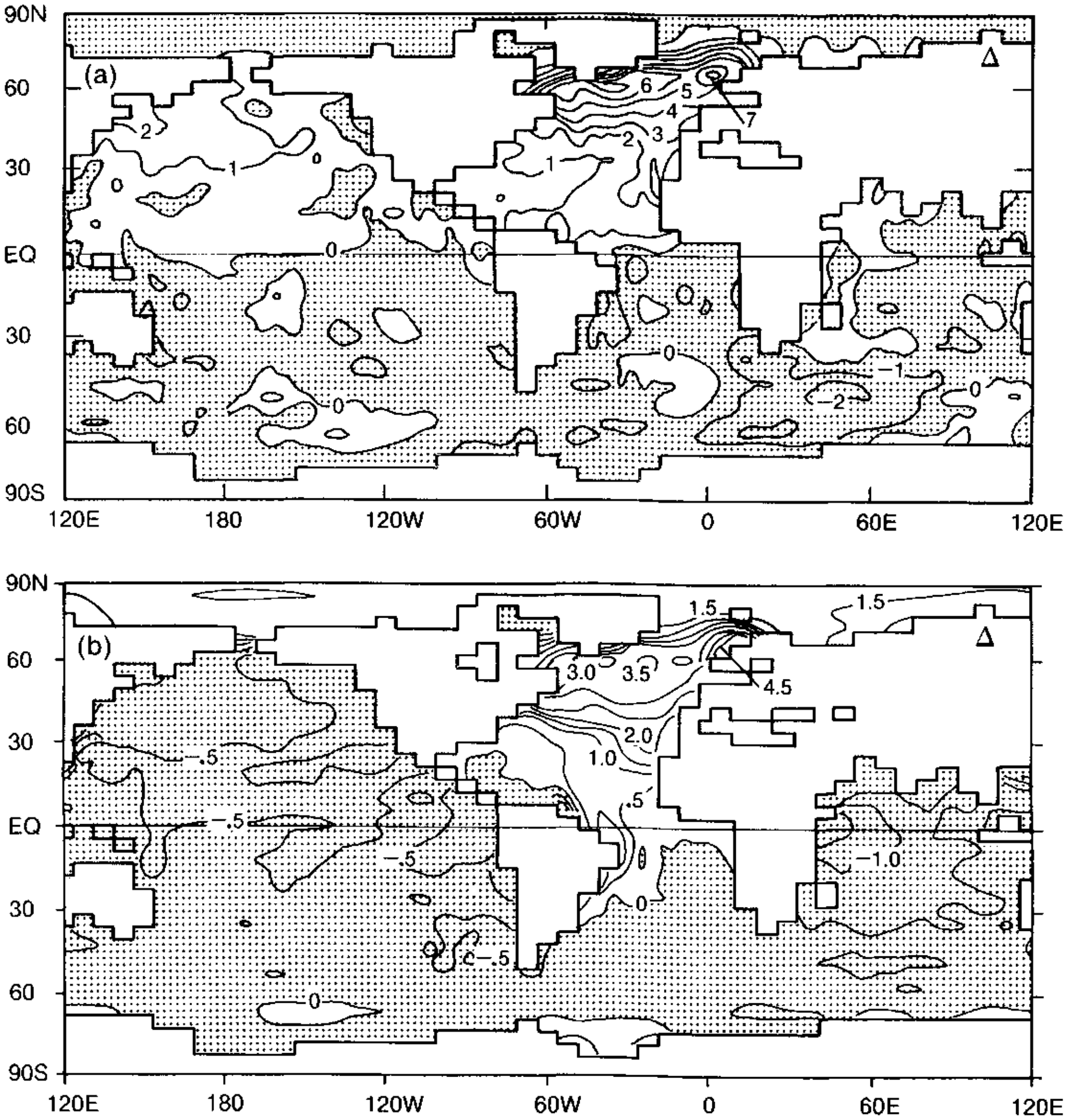

Figure B24: The difference in (a) mean surface temperature (°C), and (b) surface salinity (parts per thousand), in the North Atlantic in an equilibrium solution of the GFDL coupled ocean-atmosphere model with an active thermohaline circulation, relative to an alternative solution without thermohaline circulation. (From Manabe and Stouffer, 1988.)

B5.4 Sea-Ice Models

Ocean-only and coupled ocean-atmosphere models generally use thermodynamic sea-ice models with (at most) only simple advection schemes for the sea-ice. Experiments with sea-ice models coupled to mixed-layer pycnocline models (Owens and Lemke, 1990) have shown that the net freezing rate may depend on the continuum-mechanical properties used for the sea-ice. Inclusion of sea-ice dynamics may therefore be required to properly model the surface boundary conditions of the ocean. A model that contains most of the characteristics of sea-ice believed to be relevant for climate investigations was proposed by Hibler (1979) and has recently been reformulated in a simplified version by Flato and Hibler (1990).

Experiments with dynamic-thermodynamic sea-ice models have shown that such models are generally less sensitive to changes in the thermal forcing than are the simplified sea-ice models used so far in climate studies (Hibler, 1984; Lemke *et al.*, 1990). This reduced sensitivity is a result of a negative feedback between the sea-ice dynamics and thermodynamics. It is therefore possible that the pronounced response of the polar regions found in experiments with increased CO_2 (see Section B2.2 and IPCC (1990), Section 5) will be modified by the inclusion of ice dynamics. On the other hand, dynamic sea-ice models are more sensitive than thermodynamic models to changes in the wind forcing, and most atmospheric circulation models do not reproduce the observed wind field in polar regions very well. Further improvement of sea-ice models therefore depends upon advances in the parametrization of polar processes, including the effect of fog and low stratus clouds on the polar radiation balance. It may also be noted that models typically filter their solutions in polar latitudes for purely numerical reasons, and this may tend to degrade the simulation of sea-ice.

B6 Advances in Model Validation

B6.1 Introduction

Climate models show an increasing ability to simulate the current climate and its variations. This improvement is due to a combination of increased model resolution and improved physical parametrizations. The correctness of simulated results for the present climate at both global and regional scales is a desirable but not a sufficient condition for increased confidence in their use for simulations of future climate change.

B6.2 Systematic Errors and Model Intercomparison

Every climate modelling group strives to identify their models' systematic errors as a natural part of the continuing process of model development. Ideally, each physical parametrization in a climate model should be individually calibrated against appropriate observational data, although in many (if not most) cases this is not possible. Instead, with a particular set of parametrizations and a particular resolution, a model's performance is evaluated against the available seasonal climatological distributions of the large-scale variables such as temperature and circulation. While atmospheric and oceanic models continue to improve, the overall characterization of their performance given in Section 4 of IPCC (1990) remains valid (Gates *et al.*, 1990). The situation is somewhat different, however, for coupled atmosphere-ocean GCMs, for which only a provisional error assessment is possible due to the limited number of integrations that have been performed (see below and Section B2.2) and because of the limited ocean data currently available.

Progress in atmospheric model validation since the IPCC (1990) report has occurred both in the form of documented improvements in specific GCMs and in the form of intercomparisons of large numbers of independent GCMs. An example of an improvement in a particular model is the change in the parametrization of tropical evaporation and convection that has removed a major systematic error in the ECMWF model (see Figure B25).

Recent intercomparisons of atmospheric models have either focussed on the simulation of the present climate (as in Boer *et al.*, 1991a) or have considered the radiative forcing induced by a prescribed climate change (as in Cess *et al.*, 1991) (see Section B3.3.3). On behalf of the Working Group on Numerical Experimentation (WGNE), Boer *et al.* (1991a) have collected the mean DJF and JJA distributions of selected atmospheric variables as

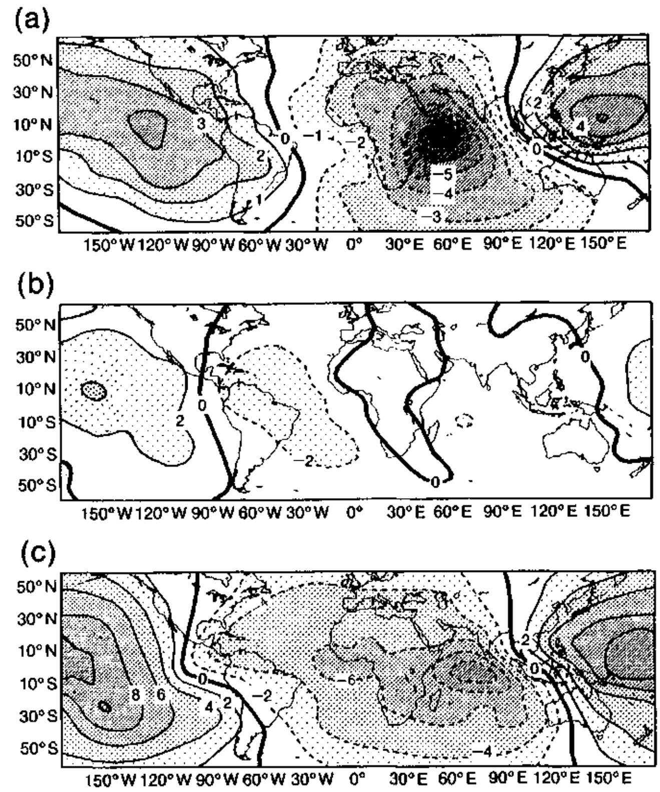

Figure B25: The 200 hPa velocity potential anomalies ($10^6 m^2 s^{-1}$) for JJA 1988: (a) as given by ECMWF analyses referenced against the mean during 1986-1990; (b) as simulated by the ECMWF model over the same interval with observed SST and standard surface flux parametrizations for heat and moisture, and (c) as simulated with revised parametrizations accounting for fluxes at low windspeeds. (From Miller *et al.*, 1991.)

simulated in the control runs of 14 atmospheric GCMs. Despite the fact that the models differed greatly in resolution and in the nature and sophistication of their parametrizations of physical processes, a number of common systematic errors where found. The simulated climate was colder than that observed on average, and all models showed a cold bias in the middle- and high-latitude upper troposphere and in the tropical lower stratosphere (see Figure B26). These temperature errors induce corresponding errors in the zonal wind distribution in accordance with the thermal wind relationship, with the result that most GCMs display too westerly a flow in the upper troposphere and in high latitudes. Taken as a whole, models appear to be more sensitive to changes in physical parametrizations than to modest changes in resolution.

The principal features of the Hadley, Ferrel and polar cells that characterize the atmospheric circulation are simulated by all models, along with appropriate seasonal shifts. There is, however, considerable variation in the strengths of the models' simulated meridional circulations.

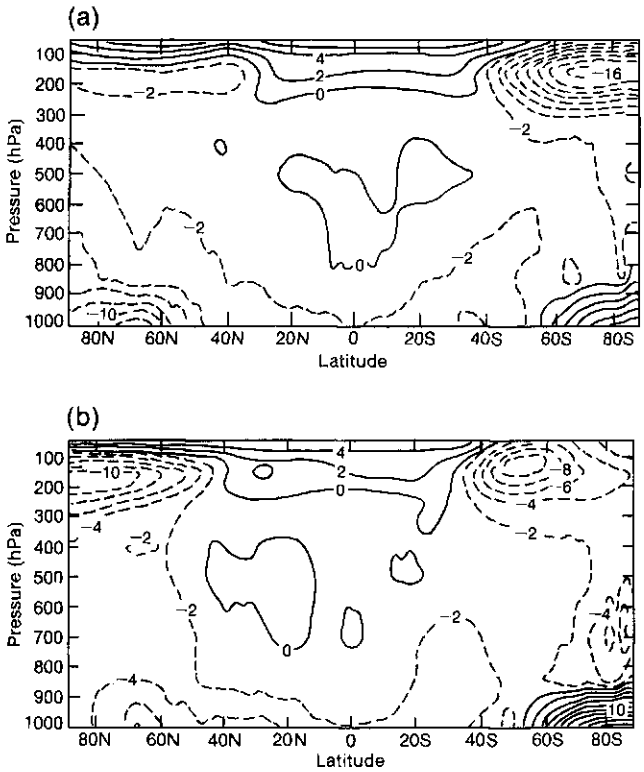

Figure B26: The systematic error (simulated minus observed) in mean temperature (°C) for (a) DJF, and (b) JJA for the UKMO atmospheric GCM. (From Boer *et al.*, 1991a.)

At the same time there has been improvement in the simulation of mean sea-level pressure in a number of models, due partly to increased resolution and partly to the introduction of the parametrization of gravity wave drag. This improvement is illustrated in Figure B27a. The largest differences among models are found near Antarctica, where the extrapolation of pressure to sea level may introduce artificial variations.

There has not been as much improvement in the simulation of precipitation by current GCMs. As shown in Figure B27b, there is a systematic over-estimation of the precipitation over much of the Northern Hemisphere (as well as an apparent under-estimation in the mid-latitudes of the Southern Hemisphere), and the models' simulated precipitation in the tropics varies considerably around that observed. There are, however, uncertainties in the observed data themselves, especially over the tropical oceans and in the Southern Hemisphere. Relatively large uncertainties in the simulation of regional precipitation are shown in the intercomparison of atmospheric GCMs being undertaken by the Monsoon Numerical Experimentation Group (MONEG) of the WCRP (1990b).

In a companion study to the model intercomparisons of Cess *et al.* (1990, 1991) (see Section B3.3), an intercomparison of the surface energy fluxes in the GCMs has recently been made (Randall *et al.*, 1991b). In this study

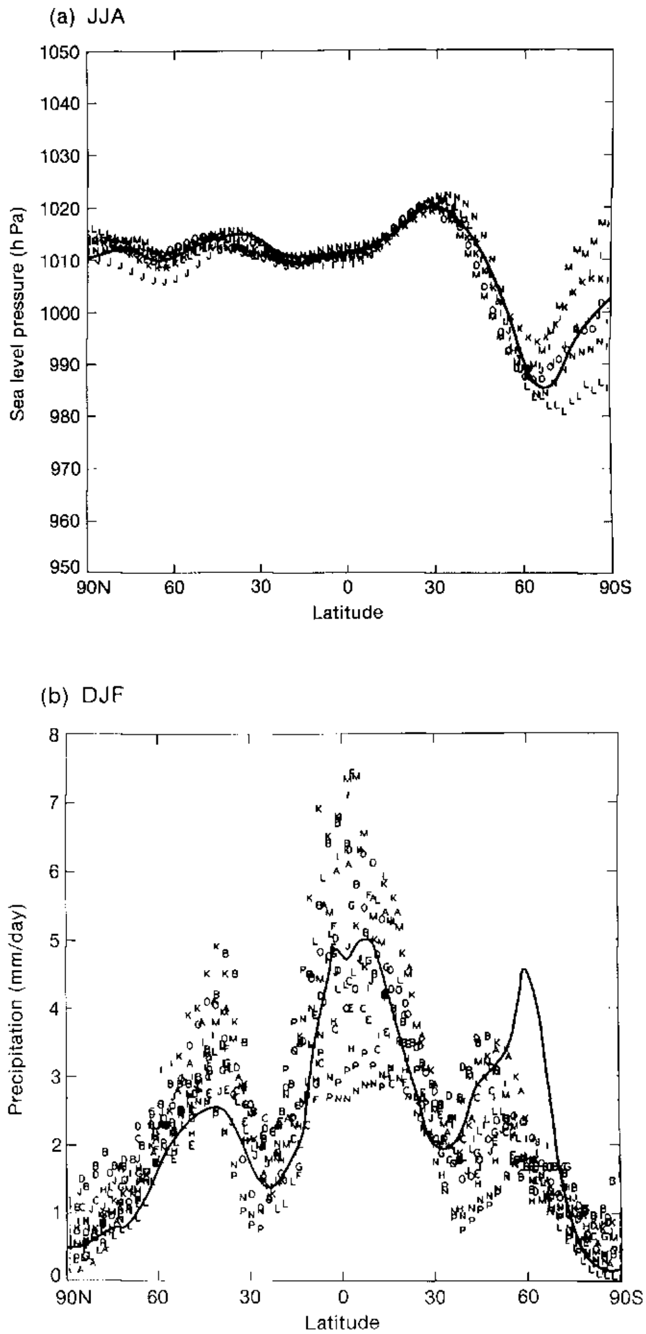

Figure B27: (a) The zonally-averaged JJA sea level pressure (mb) simulated by current high resolution GCMs; (b) the zonally-averaged DJF precipitation (mm/day) simulated by current GCMs. Different symbols indicate different models and the solid line is the observed climatological average according to Jaeger (1976). (From Boer *et al.*, 1991a.)

considerable variation is found in the models' simulated increases in precipitation, evaporation and total atmospheric water vapour in response to a uniform 4°C SST increase. As shown in Figure B28, the models also simulate an overall increase in the net infrared clear-sky flux at the surface, reflecting the positive temperature/water vapour feedback (Section B3.2) in the models.

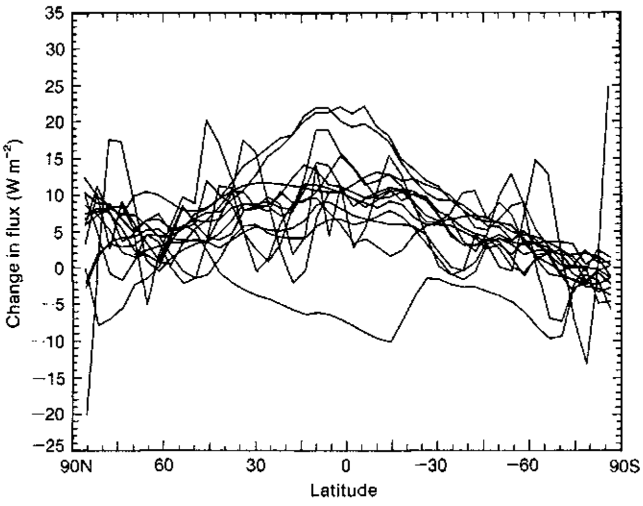

Figure B28: The change in the zonally-averaged longwave clear-sky flux at the surface as simulated by 14 atmospheric GCMs in response to a uniform 4°C increase of SST relative to July climatology. (From Randall *et al.*, 1991b.)

In the Boer *et al.* (1991a) intercomparison, as in most previous intercomparisons of climate models, results have been collected as available from the modelling community and no attempt has been made to run the models under common conditions and for common periods of time. In an effort to achieve a more systematic intercomparison of models, these attributes have recently been incorporated into the WGNE/PCMDI Atmospheric Model Intercomparison Project (AMIP), in which virtually all of the world's atmospheric GCMs will simulate the decade 1979-1988 with common values of the CO_2 concentration and solar constant and with observationally-prescribed but monthly-varying SST and sea-ice distributions (Gates, 1991). The AMIP involves many of the same modelling groups that participated in the earlier WCRP Intercomparison of Radiation Codes in Climate Models (ICRCCM), the results of which have recently been published (Ellingson *et al.*, 1991; Fouquart *et al.*, 1991).

The identification of systematic errors and the intercomparison of ocean models are at a less advanced stage than for atmospheric models. When forced with "observed" surface temperatures, salinities and wind stresses, ocean models have been moderately successful in reproducing the observed large-scale circulation and water mass distribution (see Section B5). The principal systematic errors common to most ocean models are an underestimate of the meridional heat transport as inferred from observations, and the simulation of the main thermocline to be too deep and too diffuse. No common systematic errors have as yet been identified in the deeper ocean in view of the paucity of observational data, although there are considerable differences in the simulated rate and location of deep water formation and in the strength of deep ocean circulation.

In order to examine ocean model performance in more detail, an intercomparison of tropical Pacific ocean models has recently been undertaken by TOGA-NEG (WCRP, 1990a; D. Anderson, personal communication), in which several models have been integrated with the same representation of surface forcing. In general, the SST error in all models shows a similar pattern, with the central equatorial Pacific being too cold and the equatorial western Pacific too warm throughout the year. Water along the western coast of South America is also generally too warm, and the equatorial undercurrent tends to be too weak. These discrepancies may be at least partly caused by inadequate surface forcing.

An intercomparison of the behaviour of the tropical Pacific SST in coupled ocean-atmosphere models (in which the ocean and/or the atmosphere is a GCM) has recently been made by Neelin *et al.* (1992). These models show a wide range of behaviour in their simulation of tropical interannual variability, and some are more successful in simulating ENSO-like events than others. The runs were not made under the same conditions, however, and there is a wide range of model sophistication represented.

The comparison of simulated past climates with palaeo data provides a further test of models, though the usefulness of such tests is limited by the quality of data and by our knowledge of the changes of past climates. A Palaeoclimate Model Intercomparison Project (PMIP) has been proposed (NATO Advanced Research Workshop on Palaeoclimatic Modelling, 27-31 May 1991, Saclay, France) to promote the understanding of the response of climate models to past changes in forcing.

Although model intercomparison of the sort described here does not address many important questions in climate modelling (such as how to determine which parametrizations are "best" or how best to couple the atmosphere and ocean), intercomparison has proven useful as a collective exploration of model parameter space and as a benchmark for the documentation of model improvement. It is characteristic of model intercomparisons that no single model is found to be superior to all other models in all respects. It should also be recognized that the various intercomparisons are designed for different purposes: those of Cess *et al.* (1990, 1991) are concerned with the models' response to changes in forcing (and are therefore relevant to the response to possible future changes of greenhouse gases), while those of Boer *et al.* (1991a) and those being undertaken in AMIP (Gates, 1991) are concerned with the models' ability to simulate the present climate. As model development continues (and as increased computer resources become available), there will be both an opportunity and a need to conduct further climate model intercomparisons, especially of coupled ocean-atmosphere models in both their control configuration and in experiments with increasing greenhouse gases.

B6.3 Data for Model Validation

The availability of appropriate observational data is a critical factor in the validation (and improvement) of climate models, and some progress has been made in the assembly of global data sets for selected climate variables. The recent global compilations of average monthly surface air temperature and precipitation by Legates and Willmott (1990a, b) are believed to be improvements over earlier atlases of these variables, and new assemblies of land-based precipitation (Hulme, 1992) and soil moisture (Vinnikov and Yeserkepova, 1991) are available. It should also be noted that the diagnostics made from the operational analyses of global numerical weather prediction models (Hoskins *et al.*, 1989; Trenberth and Olson, 1988) are important sources of data for the validation of selected aspects of atmospheric models. The climatological ocean atlas of Levitus (1982) has been supplemented by the compilation of surface variables from the COADS data set (Wright, 1988; Oberhuber, 1988; Michaud and Lin, 1991). These data are proving useful in the validation of both ocean and coupled ocean-atmosphere models, as are the observations of transient tracers in the ocean (Toggweiler *et al.*, 1989; Dixon *et al.*, 1991).

For other variables such as cloudiness, precipitation, evaporation, run-off, surface heat flux, surface stress, ocean currents and sea-ice, however, the observational data base remains inadequate for the purposes of model validation. These variables are relatively difficult to observe on a global basis, and are not easily inferred from compilations of conventional circulation statistics such as those of Oort (1983). The best prospect for their systematic global estimation is probably the procedure known as re-analysis, whereby modern data assimilation techniques are used to retrospectively initialize a comprehensive atmospheric GCM on a daily basis for a number of past years. Such projects are being planned by both ECMWF and NMC, and the possibility of applying the technique to ocean models is under active consideration.

Finally, it should be noted that the development of a more comprehensive global climate data base is a key element of the World Climate Research Programme (WCRP). The observational plans that have been developed in recent years for the Global Precipitation Climatology Project (WCRP, 1988a), for the World Ocean Circulation Experiment WOCE (WCRP, 1988b), for the Tropical Ocean-Global Atmosphere project TOGA (WCRP, 1990a), and for the Global Energy and Water Cycle Experiment GEWEX (WCRP, 1990c) are focussed on the acquisition of data that are necessary for the further development and validation of global atmospheric and oceanic models, and have culminated in the proposal for a Global Climate Observing System (GCOS) (WCRP, 1991).

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