On CO₂ climate sensitivity and model dependence of results

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(Received 6 September 1985; revised 18 July 1986)

SUMMARY

The regional response of climate models to small perturbations is shown to be highly dependent on the unperturbed simulation. An experiment in which CO₂ concentrations are doubled and sea surface temperatures are enhanced by 2 K has been carried out with two general circulation models which differ considerably in their control climates. The resulting changes in tropical precipitation in each model simulation are related to the increase in atmospheric water vapour which leads to enhanced precipitation in the main regions of low-level atmospheric convergence. Since these regions of convergence occur in slightly different locations in the unperturbed simulations, the distribution of changes is also different.

Differences in control simulations must be taken into account when comparing results from different models (for example, on doubling atmospheric CO₂); otherwise unduly pessimistic conclusions may be reached concerning the consistency of model results. One may be able to make subjective allowance for the effect of known deficiencies in the unperturbed simulation on the model's response before using the simulated changes in, for example, impact studies.

A detailed examination of one of the experiments reveals that the change in precipitation is limited by the heat balance of the atmosphere, and indicates the importance of treating accurately the radiative perturbation due to changes in water vapour. The magnitude of the model's response is shown to be consistent with that found in three-dimensional climate models which include a simple representation of the ocean.

1. INTRODUCTION

Over recent years, increasing use has been made of numerical models in climate research. Three-dimensional general circulation models have been used extensively to study the atmospheric circulation, and, not surprisingly, it is found that the simulated climate varies with the model used. Attempts have been made to understand the reasons for differences in simulations (for example, GARP 1979), but the results are generally speculative because the models compared differ in more than one way. In principle, discrepancies between two models could be resolved by removing the differences one at a time (for example, Hansen et al. 1983); in practice this would involve extensive cooperation between the institutions involved, and may be impossible, for example, where the physical processes of a particular model are interdependent. Thus, we are left with the unsatisfactory position in which model simulations differ from each other (and observed climate) for reasons which are not well understood.

Many climate models have been used to study the effect of increased atmospheric CO₂ concentrations on climate (Dickinson 1985; United States Department of Energy 1985). It is estimated that a doubling of atmospheric CO₂ concentrations would increase the net radiative heating below the tropopause by 4 W m⁻², which is small compared with the 240 W m⁻² outward long-wave flux at the top of the atmosphere. Thus a doubling or quadrupling of CO₂ amounts can be regarded as a perturbation on present day climate. This is confirmed by numerical studies where it has been found that changes in circulation are small (for example, Manabe et al. 1981; Washington and Meehl 1984). It follows that the results of such sensitivity studies may be highly dependent on the simulation of the unperturbed climate, which in turn depends on the model used. Given the wide variety of models used to date, it is not surprising that the results obtained are often contradictory.

The dependence of a model's sensitivity on its unperturbed climate should not be confused with dependence on the physical processes included in the model. For example, it is possible to construct two versions of a model, one with prescribed cloud, another with model-generated cloud, which would produce indistinguishable simulations of present day
mean climate, but would respond quite differently to a given perturbation. Thus, an accurate representation of present day mean climate is not sufficient to guarantee that a model's sensitivity is correct. Here we are more concerned with the differences in sensitivity between models which include the same physical processes but produce different errors in the control simulation. For small perturbations, the simulated changes in climate will inevitably be strongly related to the control simulations. Since it will be some time (if ever) before present climate can be simulated without appreciable error, the likely effect of such errors on the simulated response to a given perturbation must be assessed.

Some progress towards this end has already been made. Hansen et al. (1984) and Spelman and Manabe (1984) have demonstrated the dependence of the sensitivity of global mean surface temperature on the areal extent of sea-ice, or sea-ice and snow cover. A warming of the surface will remove some of the highly reflective snow or ice cover, producing further warming through enhanced absorption of solar radiation, often referred to as 'temperature-albedo feedback'. If the temperature in the control simulation is reduced, the latitude of the snow or ice boundary is reduced, the length of the boundary is increased and the mean solar elevation at the boundary becomes higher. Temperature-albedo feedback becomes stronger, and occurs over a larger area, and the sensitivity of the surface temperature is increased.

There is also evidence that the nature of the control simulation can affect changes on a continental scale. Several studies of the effect of increasing atmospheric CO₂ (for example, Manabe et al. 1981; Mitchell and Lupton 1984) show a drying of the surface of the mid-latitude continents in the summer, despite the accumulation of extra soil moisture during winter. Much of this excess moisture gained in winter is lost as run-off in spring, when the model surface reaches field capacity over much of the area. A longer drying season and reduced precipitation then produce a drying of the surface in summer. In contrast, Washington and Meehl (1984) find increases in zonal mean soil moisture throughout the year although some areas dry out in summer, and the increase is least pronounced in winter. In their model, the soil does not become saturated in spring in either the control or enhanced CO₂ integration (Meehl, private communication) so one may speculate that all the excess moisture accumulated in the winter and spring in the enhanced CO₂ case is retained in the soil and must be lost over the summer by evaporation. The presence of enhanced soil moisture at the beginning of summer may lead to increased precipitation (Rowntree and Bolton 1983; Yeh et al. 1984). Indeed, Washington and Meehl find an increase in precipitation in northern mid-latitudes in summer, which further counteracts the tendency for enhanced evaporation to dry out the surface.

Little can be said about the smaller-scale changes found in experiments with increased CO₂. Many simulations carried out to date have used low horizontal resolution and so cannot resolve regional detail. Furthermore, the comparison of the control and anomaly integrations has been limited to a few years, which may be insufficient to distinguish regional changes from the year to year variability on that scale. This is unfortunate, since predictions on a regional scale are required before the potential social and economic impact of climate change due to increased CO₂ can be assessed.

Mitchell (1983a), using a model with relatively high horizontal resolution, did find changes in regional precipitation which were statistically significant and physically reasonable. On increasing sea surface temperatures and doubling CO₂ amounts, significant changes in the precipitation were produced over the eastern subtropical continents in summer and in other regions. The consistency of these regional changes has encouraged us to repeat the experiment using a different model, in order to assess the model dependence of the results, and whether or not any discrepancies between the responses
could be related to differences in simulations of the unperturbed climate. We concentrate largely on the model dependence of changes in precipitation, although other fields are considered.

In this paper, it is not our intention to give a detailed explanation of the differences in the control simulations in terms of the model formulations. It would require additional experiments in which the differences in model formulation are removed one at a time to do so convincingly, a project which is at present beyond our resources.

Here, we:

(1) Describe briefly the climatology of the Meteorological Office 11-layer general circulation model (11LM) and compare it with that of the Meteorological Office 5-layer general circulation model (5LM).

(2) Describe the changes in the 11LM due to doubling CO₂ and enhancing sea surface temperatures by 2 K, and compare them with the changes found using the 5LM, relating the differences in the response to the differences in control simulations, where appropriate.

(3) Identify some of the physical mechanisms associated with changes in the hydrological cycle.

(4) Discuss briefly the implications of the above work for numerical studies of climate change.

2. THE MODELS, EXPERIMENTAL DESIGN AND MODEL CLIMATOLOGIES

(a) The models

The 11-layer model used is a global finite difference model with a regular 2.5°×3.75° latitude/longitude grid, and 11 sigma (σ = pressure/surface pressure) layers which are irregularly spaced, being concentrated near the boundary layer and the tropopause. Earlier versions of the model have been used by Slingo (1985a), Cunniongton and Rowntree (1986) and Palmer and Mansfield (1986), and were based on the 5-layer model described by Corby et al. (1977).

In this version, computational stability and noise are controlled by a multipoint filter near the poles, and by nonlinear diffusion. The seasonal and diurnal variation of solar radiation are represented, and the radiative fluxes are a function of temperature, water vapour, ozone and carbon dioxide concentrations, and prescribed zonally averaged cloudiness. Long-wave fluxes are calculated using an emissivity approximation (the 'old' scheme described by Slingo and Wilderspin (1986)). Low, medium and convective cloud are assigned a solar albedo of 0.7 and a long-wave emissivity of unity, whereas high cloud is given a reflectivity of 0.2, and by halving the amount, a long-wave emissivity of 0.5. Sea surface temperatures and sea-ice extents are prescribed from climatology, and updated every five days. The heat flux through sea-ice is included.

The treatment of the boundary layer, which occupies the lowest three model layers, is based on method I of Clarke (1970). The calculation of the fluxes of heat, moisture and momentum from the surface is based on Monin–Obukhov similarity theory (see Mitchell et al. 1985). Evaporation is limited, where appropriate, by soil moisture, following Manabe (1969). Convection is modelled using a penetrative scheme in which both entrainment of environmental air and detrainment from the convective parcel are allowed at each level (Lyne and Rowntree 1976). Excess moisture due to condensation
following convection or large-scale ascent is removed as precipitation. Further details of the model are given by Slingo (1985b).

The 5-layer model uses a similar finite difference scheme on a quasi-uniform 330 km horizontal grid, with five equally spaced sigma layers in the vertical (Corby et al. 1977). The treatment of radiation, sea surface temperature, sea-ice extent, cloud amount, and soil moisture, described by Slingo (1982), is similar to that in the 11-layer model. The penetrative convection scheme is less intricate than that used in the 11-layer model. The boundary layer height is modelled explicitly and used to control surface fluxes and, where appropriate, convection.

The main differences in the formulation of the two models are listed in Table 1. A comparison of a version of the 11 LM on a 220 km quasi-uniform grid with the 5 LM was made by Mitchell and Bolton (1983).

(b) Experimental design

The control integration in the 11-layer model experiment was run for eight years starting from real initial data from the First GARP Global Experiment (25 July 1979). The anomaly integration (in which CO₂ amounts were doubled, and sea surface temperatures were enhanced by 2 K) was started from 1 March of the second year of the control, and run for three years. Seasonal differences calculated from averages of the eight years of the control and the three years of the anomaly simulation are presented; annual mean quantities are computed from the period corresponding to the last two years of the anomaly integration. The 5-layer model control integration was run for three years.

<table>
<thead>
<tr>
<th>TABLE 1. MAIN DIFFERENCES IN MODEL FORMULATIONS</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>GRID</strong></td>
</tr>
<tr>
<td>5 layers, equally spaced 300 km in horizontal</td>
</tr>
<tr>
<td>11 layers, concentrated near surface and tropopause 2.5°×3.75°</td>
</tr>
<tr>
<td><strong>BOUNDARY LAYER</strong></td>
</tr>
<tr>
<td>1 layer (up to σ = 0.8)</td>
</tr>
<tr>
<td>Explicit boundary layer height</td>
</tr>
<tr>
<td>Bulk aerodynamic formula</td>
</tr>
<tr>
<td>Stable/Unstable, land/sea drag coefficient</td>
</tr>
<tr>
<td>Full evaporation when soil moisture = 10 cm</td>
</tr>
<tr>
<td>Run off when soil moisture = 20 cm</td>
</tr>
<tr>
<td>3 layers (up to σ = 0.79)</td>
</tr>
<tr>
<td>Vertical diffusion, 'Clarke' scheme</td>
</tr>
<tr>
<td>Drag coefficient continuous function of stability and roughness length</td>
</tr>
<tr>
<td>Full evaporation when soil moisture = 5 cm</td>
</tr>
<tr>
<td>Run off when soil moisture = 15 cm</td>
</tr>
<tr>
<td><strong>RADIATION</strong></td>
</tr>
<tr>
<td>Temperature and humidity interpolated to 10 equally spaced layers for radiation</td>
</tr>
<tr>
<td>No absorption of reflected solar beam</td>
</tr>
<tr>
<td>Snowfree albedo a function of latitude</td>
</tr>
<tr>
<td>Albedo over snow a function of snow depth</td>
</tr>
<tr>
<td>Fluxes calculated on model layer boundaries</td>
</tr>
<tr>
<td>Reflected solar beam absorbed</td>
</tr>
<tr>
<td>Snow-free albedo constant (=0.2)</td>
</tr>
<tr>
<td>Albedo over snow constant (=0.5)</td>
</tr>
<tr>
<td><strong>CLOUD AMOUNTS, ALBEDOS SIMILAR</strong></td>
</tr>
</tbody>
</table>

| **PENETRATIVE CONVECTION**                    |
| Detrains only at upper levels                 |
| May affect a given layer more than once per timestep |
| May entrain and detrain at any level          |
| Only affects a given layer once per timestep  |

| **DIFFUSION**                                 |
| Nonlinear diffusion of potential temperature θ |
| Nonlinear diffusion of humidity               |
| Nonlinear diffusion of aT + bθ                |
| Linear diffusion of humidity                  |
and the anomaly for $2\frac{1}{2}$ years (Mitchell 1983a). Differences are computed over the final two years of the anomaly integration.

(c) The 11 LM climatology, and comparison with that of the 5 LM

(i) Temperature. The zonally averaged temperatures simulated by the 11 LM (Fig. 1) are generally lower than climatological estimates (Newell et al. 1972; Oort and Rasmusson 1971) except in the stratosphere in the tropics. The largest discrepancies occur in high latitudes where temperatures are some 15 to 25 K lower than observed in the stratosphere throughout the year, and over 10 K lower than observed in the troposphere in winter. Otherwise zonally averaged tropospheric temperatures are mostly within 2 to 3 K of observed values. The 5 LM does not resolve the tropopause in the tropics (Fig. 2). Temperatures in high latitudes are generally much lower than observed in the stratosphere, as in the 11 LM, and in the troposphere over the Antarctic in winter. The 5 LM simulates the low-level surface inversion found in polar latitudes in winter, though the model exaggerates this feature in the Antarctic. On the other hand, the 11 LM underestimates the strength of these surface inversions due to excessive vertical mixing in the lowest three layers in the boundary layer formulation.

(ii) Wind. Both models reproduce qualitatively the seasonal variation of the strength and latitude of the subtropical westerly jet, and secondary westerly maximum near 55°S in winter (Figs. 1, 2). In each case, the strength of the subtropical jet is overestimated by 5 to 10 ms$^{-1}$ in the summer hemisphere. In the winter hemisphere, the 11 LM partially resolves the stratospheric polar night jet; in both models the flow in the topmost

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Figure 1. Zonally averaged temperatures (dashed lines, contours every 10 K) and zonal component of wind (solid lines, contours 5 ms$^{-1}$, areas of easterly flow stippled) from the 11 LM, averaged over eight years of the control simulation. (a) June to August. (b) December to February.
levels in mid-latitudes is excessive, consistent with the stronger than observed enhanced meridional temperature gradient at upper levels. In the 11LM, there is a secondary maximum in westerly flow near 55°N. This error, which appears to be a feature of high resolution models (for example, Manabe et al. 1979), is perhaps due to the neglect of enhanced drag induced over rough or high terrain (Palmer et al. 1986). The secondary maximum in the simulated flow is not unlike that observed in winter in the southern hemisphere where orographic effects are likely to be much less important. In the 5LM, the polar surface easterlies extend further equatorward than observed, particularly in the northern hemisphere in winter. Despite this there is, as observed, a maximum in low-level flow in December, January, February near 45°N in the 5LM, whereas it is near 55°N in the 11LM.

(iii) Pressure at mean sea level. During June to August (JJA, Fig. 3(a)) the 11LM captures some of the more subtle features of the observed circulation such as the weak westerly flow in northern mid-latitudes (Schutz and Gates 1972). The subtropical anticyclones in both hemispheres are correctly placed. The simulation of the Asian monsoon is marred by a spurious low centre in the China Sea. The depth and latitude of the antarctic circumpolar trough are close to the climatological values.

The pattern during December to February (DJF, Fig. 3(b)) is less satisfactory with pressure 20 mb lower than observed over the arctic and up to 12 mb higher than observed over the Mediterranean (Schutz and Gates 1971), producing the band of excessive westerly flow near 55°N noted earlier. The northward extents of the Siberian and North
American high pressure cells are severely curtailed. The low latitude and southern hemisphere patterns are similar to those observed, though the southern hemisphere anticyclones are displaced slightly poleward.

In the 5LM simulation (Mitchell 1983a), the simulated pressure over the poles is higher than observed, and the antarctic circumpolar trough is shallower and further north than observed throughout the year. Otherwise, the JJA pattern is in close agreement with climatological data. During DJF, the Aleutian and Icelandic lows are deeper than observed, but any tendency to excessive westerly flow (particularly over the northern mid-latitude continents) is much less pronounced than in the 11LM.

(iv) Precipitation. The simulated distribution of precipitation in the 11LM is generally similar to climatological estimates (Moller 1951; Jaeger 1976). The main errors occur over the tropical oceans where the simulated precipitation is excessive. For example in JJA (Fig. 4(a)), peaks of 20 mm d\(^{-1}\) are found in the Bay of Bengal and the west Pacific, whereas precipitation over the tropical continents (northern India, Africa near 10°N) is less than observed. In DJF (Fig. 4(b)), there is still a tendency towards excessive precipitation over the oceans (for example, north of Madagascar and west of Mexico).
Figure 4. Precipitation from control integration of the 11 LM (8-year mean). Contours at 1, 2, 5, 10, 20 and 40 mm d\(^{-1}\); stippled where greater than 5 mm d\(^{-1}\). (a) June to August; (b) December to February.

In the 5 LM, precipitation is excessive over land (Fig. 5), as can be seen over the Indonesian Islands and South Africa throughout the year, and in parts of South America in JJA. Indeed, the annual mean precipitation over land is almost 60% greater than in the 11 LM (Table 2). A more detailed account of the hydrological cycle in the 5 LM is given by Mitchell (1983b).

<table>
<thead>
<tr>
<th>TABLE 2. GLOBAL ANNUAL MEAN PRECIPITATION IN THE 11 LM AND 5 LM MODELS, AND CHANGES DUE TO DOUBLING CO(_2) AND ENHANCING SEA SURFACE TEMPERATURES (mm d(^{-1}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>Precipitation</td>
</tr>
<tr>
<td>----------------</td>
</tr>
<tr>
<td>11 LM</td>
</tr>
<tr>
<td>Globe</td>
</tr>
<tr>
<td>Land</td>
</tr>
<tr>
<td>Sea</td>
</tr>
<tr>
<td>Globe</td>
</tr>
<tr>
<td>Land</td>
</tr>
<tr>
<td>Sea</td>
</tr>
</tbody>
</table>
3. **The models' responses to doubling CO₂ and enhancing sea surface temperatures by 2 K**

**(a) Global annual mean**

The globally averaged response of the 11-layer model is similar to that of the 5-layer model (Table 2). The increase in sea surface temperature raises the saturation vapour pressure at the ocean surface and evaporation increases by 5.6% (6.6% in the 5-layer model). The total atmospheric moisture content increases by 20% (18%) and precipitation by 5.6% (4.9%) (Table 3). The troposphere warms by 3.1 (3.0) K and the land surface

<table>
<thead>
<tr>
<th></th>
<th>Tropospheric temperature (K)</th>
<th>Atmospheric moisture content (%)</th>
<th>Surface temperature</th>
<th>Precipitation (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>11 LM</td>
<td>3.08</td>
<td>20</td>
<td>2.26</td>
<td>5.6</td>
</tr>
<tr>
<td>5 LM</td>
<td>3.02</td>
<td>18</td>
<td>2.21</td>
<td>4.9</td>
</tr>
</tbody>
</table>
TABLE 4. VERTICAL DISTRIBUTION OF THE CHANGES IN TEMPERATURE AND HUMIDITY IN THE 11LM EXPERIMENT (GLOBAL ANNUAL AVERAGE)

<table>
<thead>
<tr>
<th>o level</th>
<th>Change in temperature (K)</th>
<th>Change in humidity (g kg⁻¹)</th>
<th>% change in humidity mixing ratio</th>
<th>% change in relative humidity (% of saturation)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>0·987</td>
<td>0·937</td>
<td>0·844</td>
<td>0·718</td>
</tr>
<tr>
<td></td>
<td>2·33</td>
<td>2·33</td>
<td>2·41</td>
<td>2·70</td>
</tr>
<tr>
<td></td>
<td>1·29</td>
<td>1·17</td>
<td>1·07</td>
<td>0·62</td>
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<tr>
<td></td>
<td>15</td>
<td>16</td>
<td>18</td>
<td>21</td>
</tr>
<tr>
<td></td>
<td>0·1</td>
<td>0·3</td>
<td>0·2</td>
<td>−0·2</td>
</tr>
</tbody>
</table>

Note. Relative humidities are calculated from seasonal averages of temperature, humidity and pressure.

warms by 3·1 (2·9) K as a result of the increased downward flux of infrared radiation from the warmer, moister, CO₂-enriched atmosphere. In the evaluation of climatic impacts, the regional variations in supply of moisture to the surface (precipitation minus evaporation, or \( P - E \)) is of more interest than the global mean change in precipitation. The mean magnitude of \( P - E \) increases from 1·15 (1·72) mm d⁻¹ by 0·15 (0·17) mm d⁻¹ or 13 (10)%, substantially more than the percentage change in global mean precipitation.

The percentage increase in humidity mixing ratio in the 11LM reaches a maximum in the upper troposphere where temperature increases are largest (Table 4). Note also that the fractional change in saturation vapour pressure for a given change in temperature is larger at lower temperatures. The changes in globally averaged humidity at each level are broadly consistent with the assumption that relative humidity remains constant.

(b) Zonal means

(i) Temperature and zonal wind. The top of the model cools, following increased radiation loss to space from the enhanced CO₂, whereas the troposphere warms (Fig. 6). The model’s moist convective processes tend to adjust the vertical profile towards the moist adiabatic lapse rate, which becomes smaller with increasing temperature, so that temperature changes are amplified with height. In the 11LM the 5 K increase at 300 mb in the tropics is consistent with increasing the surface temperature by 2 K and maintaining a moist adiabatic temperature profile. Note that in studies carried out with models employing a moist convective adjustment scheme (for example, Manabe and Stouffer 1980; Washington and Meehl 1984) the amplification of the temperature increase with height is much less pronounced than in studies where a penetrative convection scheme has been used as here and in Hansen et al. (1984).

[Figure 6. Changes in zonally and annually averaged atmospheric temperature in the 11LM due to doubling CO₂ and enhancing sea surface temperatures. Contours every 1K, shaded where negative.]
In a moist convective adjustment scheme, adjacent pairs of layers are considered in succession for instability (or potential instability). If the profile between two layers is unstable, the air in the two layers is mixed until the instability is removed. Thus, if the scheme is presented with a saturated unstable profile, illustrated schematically in Fig. 7, it will produce a uniform profile of wet-bulb potential temperature extending to level a.

In a penetrative scheme, air from a given level can rise to the limit given by convective parcel theory. In principle, air parcels adjacent to the surface in our example can penetrate to level b, though in practice, this limit would not be reached as most penetrative schemes allow for entrainment of environmental air which reduces the buoyancy of convective parcels. Even so, the penetrative scheme will warm to a higher level and (because both schemes conserve moist static energy) produce a slightly cooler profile at lower levels. In summary, the penetrative scheme will produce a larger response at upper levels to a given surface warming as it moves heat and moisture to higher levels than the convective adjustment scheme.

The height and magnitude of the maximum warming decrease towards the poles, increasing the meridional temperature gradient in the upper troposphere, particularly near 40° of latitude. In the subtropics a tongue of warmer air extends downwards from the upper troposphere, reducing the meridional temperature gradient at mid-levels in the tropics. One may speculate that air near the top of the ascending branch of the Hadley circulation, where the warming is largest, is carried poleward and subsides in the subtropics, further warming the middle troposphere. There is a secondary maximum warming in high latitudes which, in the winter hemisphere (not shown), varies considerably from year to year.

Figure 7. Schematic illustration of the effects of moist convective adjustment and penetrative convection schemes on an unstable (saturated) profile of wet-bulb potential temperature against height (solid line). The moist convective adjustment scheme will leave a uniform $\theta_w$ profile (dashed line), with warming extending to a. The penetrative scheme considers individual convective parcels, so in principle the warming can penetrate to b. The final profile (not shown) will be warmer at upper levels and cooler at lower levels than in the case of convective adjustment.
Figure 8. Changes in zonally averaged west to east wind in the 11 LM due to doubling CO$_2$ and enhancing sea surface temperatures. Contours every 1 m s$^{-1}$, shaded where negative. Dash-dot lines indicate the latitudes of maximum west to east wind in the control integration. (a) June to August; (b) December to February.

The changes in zonally averaged west to east wind (Fig. 8) are largely those expected assuming geostrophic balance with the changes in geopotential gradients implied by the temperature changes in Fig. 6. For example, near 40° of latitude, there is enhanced westerly flow particularly at upper levels associated with the increased meridional temperature gradient, whereas at some latitudes further poleward, the westerly flow is weakened. Outside the tropics, the sign of the changes in zonal wind is constant with height. The increased flow near 75°N in December to February (Fig. 8(b)) and near 50°S

Figure 9. As Fig. 6, but for the 5 LM.
in June to August is associated with a tendency for the winter surface pressure trough to be deeper or displaced poleward in the anomaly integration.

The annual mean changes in temperature in the 5-layer model were remarkably similar (Fig. 9) to those in the 11 LM, except in the upper troposphere near 40° of latitude, where in summer a spurious warming occurs due to a deficiency in interpolation of the humidity profile for use in calculating the radiative fluxes (see Mitchell 1983a). As a result of this error, the associated increase in meridional gradient is further poleward than in the 11-layer model, and the enhanced westerly flow in summer is found poleward of the maximum in the control (Fig. 10) rather than equatorward as in the 11-layer model.
The zonally averaged surface pressure trough is found near 60°N and 60°S in local winter in the 5-layer model, and further poleward at 70°N and 65°S in the 11-layer model. In both models, the winter depression belts and associated westerlies intensify, with the enhanced westerly flow occurring further poleward in the 11 LM. However, in the 11 LM in the northern hemisphere in winter (Fig. 8(b)) the enhancement increases with height, and there is a further region of increased flow at upper levels near 30°N. In the 5 LM (Fig. 10(b)) there is a single band of enhanced westerly flow near 65°N in winter which

Figure 12. Zonally averaged precipitation from the control (solid line) and anomaly integrations (dashed line), and differences (dotted line) in the 11-layer model (mm d⁻¹). (a) June to August; (b) December to February.
ON CO₂ CLIMATE SENSITIVITY

weakens with height. This discrepancy between the responses in the models is due in part to an erroneous low-level maximum in westerly flow near 55°N in the 11 LM.

(ii) Relative humidity. In the 11 LM, the annual mean relative humidity calculated from seasonal geographical means of temperature, humidity and pressure (Fig. 11) is reduced in the upper troposphere, with maximum reductions of over 10% in the equatorial region. The main increases occur in the lower stratosphere except in the subtropics and in the

Figure 13. As Fig. 12, but for the 5 LM. (Note that in Mitchell (1983a), the seasons are incorrectly labelled in Fig. 9.)
tropics in the middle troposphere; small increases are found in the lower troposphere at most latitudes. These changes are similar to those found in subsequent experiments using versions of the 11 LM model with interactive cloud (Mitchell et al. 1986) and by Wetherald and Manabe (1986) in CO$_2$ experiments using prescribed and model-generated cloud.

(iii) Precipitation. As noted earlier, the moisture content of the atmosphere increases in the anomaly integration. Mitchell (1983a) argued that in the absence of changes in circulation, this should lead to an increase in precipitation (or more accurately, precipitation minus evaporation, $P - E$) in regions of low-level mass convergence due to increased moisture convergence, and conversely a decrease in precipitation in regions of low-level mass divergence. If one assumes that storage in the atmosphere and diffusion of moisture can be ignored, the equation for moisture (see Corby et al. 1977) can be integrated in the vertical to give

$$\frac{1}{p_*} \int (\nabla \cdot (\nabla p_* g)) \, d\sigma = E - P$$

(1)

where $p_*$ is surface pressure, $\nabla$ is the velocity on $\sigma$ levels, $g$ is the specific humidity and $\sigma$ is the vertical coordinate. Noting that the main contribution to the vertical integral comes from the lower troposphere, and assuming that the fractional increase in specific humidity in the lower troposphere is fairly uniform then $P - E$ will increase in regions of convergence and decrease in regions of divergence. In the 11LM, zonally averaged precipitation increases in the main regions of convergence in the tropics and in middle to high latitudes (Fig. 12). The enhancement in precipitation in winter mid-latitudes is further increased by stronger low-level mass convergence accompanying the deepening of the depression tracks noted above. In the subtropics, precipitation changes little although evaporation increases markedly, so precipitation minus evaporation and hence moisture convergence are reduced.

In general, the peaks in precipitation become more intense (for example at 60°N, 10°S, 50°S in December to February (Fig. 12(b))), though the mid-latitude peak shifts poleward in winter (from 35°N in December to February and 40°S in June to August). The broad pattern of the response is similar in the 5-layer model experiment (enhancement of the equatorial and mid-latitude peaks, poleward shift of the winter mid-latitude peak and little change in the subtropics (Fig. 13)), though the distribution of the changes differs in detail. This to a large extent reflects the differences in the two control simulations; for example, in the 11-layer model the simulated precipitation shows two peaks in the equatorial regions (Fig. 12), associated with more than one zone of increase, whereas in the 5-layer model both the equatorial peak in precipitation and associated regions of increase are more coherent (Fig. 13).

In summary, the zonally averaged responses of the models are similar in many respects, and most of the discrepancies that do occur can be explained directly in terms of the differences in model formulation, or related to differences in the unperturbed simulation.

(c) Geographical distribution of the changes in precipitation

Although simulated precipitation patterns are subject to considerable temporal variability, many of the regional changes in both the 11LM (see the following subsection) and the 5 LM (Mitchell 1983a) are statistically significant. The most intense changes occur in the tropics (Fig. 14), with large increases in regions of strong low-level convergence. In the subtropics and middle latitudes, the response is less pronounced, though over land precipitation is generally reduced, especially during June to August (Fig. 14(a)). Precipitation is enhanced in high latitudes, particularly in winter.
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In the tropics the detailed changes are not random, being closely related to the unperturbed pattern. Regions of increase generally coincide with areas where precipitation is already intense in the control simulation, as one would expect if the changes are dominated by the increase in atmospheric moisture as opposed to changes in circulation. Thus in the 11 LM north of the equator during June to August, rainfall is enhanced over the central Pacific and western Atlantic oceans, off the south-eastern United States, in the Bay of Bengal and to the east and west of the Philippines (Fig. 14(a)), corresponding to maxima in the control integrations (Fig. 4(a)). Increases occur south of the equator in the Indian and central Pacific Oceans where there is strong low-level convergence in the control integration.

This pattern is repeated in December to February (Fig. 14(b)) north of the equator over the tropical east Pacific Ocean and east of the Philippines, and south of the equator in the central Pacific and western Indian Oceans, east of South America and Australia and over Indonesia, but not north of the equator over the west Atlantic and Indian Oceans, and over Indonesia and South America near 20°S where rainfall is decreased over wet regions.

In mid-latitudes the relationship between the control simulation and the response is less obvious. For example, in December to February (Fig. 14(b)) precipitation is enhanced over southern Alaska and west of Scandinavia as expected, but also in the north Atlantic well to the east of the rainband off eastern north America. Precipitation decreases in many regions where it is already meagre, for example over the south-west of North America and from the Mediterranean eastwards in June to August. Over land, changes in $E - P$ may be limited by the availability of soil moisture. In regions of prevailing moisture divergence (for example, over much of the mid-latitude continents in summer) an increase in moisture divergence ($E - P$) will enhance the drying of the surface, and may ultimately produce a reduction in both evaporation and precipitation, with little change in $P - E$.

A similar correlation between the patterns of precipitation in the unperturbed simulation and patterns in the response is evident in the 5-layer model (Figs. 5, 15). As noted in section 2, in the 5-layer model the heaviest precipitation occurs over land (around the Caribbean, over east Africa and the Indonesian islands, along the southern and eastern coasts of North America and Asia in June to August, (Fig. 15(a)) and over northern and eastern South America, east Africa and the Indonesian islands during December to February, (Fig. 15(b)), whereas in the 11-layer model, the heaviest precipitation is found in the same general regions, but occurs predominantly over the ocean (Fig. 14). Although in both models the increase in atmospheric moisture appears to be the dominant mechanism producing the changes in precipitation, the distribution of the response is different because the unperturbed simulations are different. Thus, in the 5-layer model there is a much larger response over land than in the 11-layer model (Table 3). This is reflected in the geographical distribution of the response, as around the southern coasts of the Caribbean and south-east Asia during June to August (Figs. 14(a), 15(a)). In short, the pattern of the model's response is strongly dependent on the unperturbed simulation.

In order to quantify the relationship between the model's response and the unperturbed simulation, we have calculated the spatial correlation between $P - E$ and changes in $P - E$ (see appendix). If the atmospheric circulation is unchanged and the fractional change in humidity is uniform, the correlation coefficient should be unity (see Eq.(1)). The spatial correlation is -36 (-28 in the 5 LM) for the annual mean if the whole globe is considered. The correlation is -39 if only the tropical oceans (30°N to 30°S) are considered, and is considerably higher in the extratropics (poleward of 30°, see appendix). Our
estimate of the correlation coefficient is biased downwards due to the temporal variability of the model. If a simple correction is made to allow for this (see appendix), the above figures are increased to 0.59 (0.35 in the 5LM) and 0.63 respectively. Thus for example, our hypothesis accounts for approximately 40% of the spatial variance in the changes in $P - E$ over the tropical oceans in the 11LM if the contribution from temporal variations is removed.

\textit{(d) Statistical significance of results}

Simulations made with a general circulation model exhibit apparently random year to year fluctuations which are a manifestation of the model's inherent natural variability. In order to distinguish those differences between the anomaly and control simulations which are unlikely to have occurred by chance, and are therefore probably due to the enhancement of sea surface temperatures and CO$_2$ amounts, we have performed a simple significance test.

A 2-tailed Student's $t$-test was carried out using seasonal means (three from the anomaly, eight from the control) of precipitation in the 11LM experiment (Fig. 16). Results of a similar test on the 5LM results were reported by Mitchell (1983a). Here,
there are systematic patterns of change significant at the 90% level of confidence or higher in high latitudes in December to February, and in the tropics during June to August. The patterns are less well organized in the tropics during December to February, although the increases in regions of existing heavy precipitation are generally significant.

The limitations of the above test must be borne in mind. First, as meteorological patterns are correlated in space, coherent areas of apparently significant changes will occur by chance. Hannoschock and Frankignoul (1985) have tried an approach proposed by Hassellmann (1979) using a reduced basis set which may overcome this shortcoming, but is unsuitable for use with fields such as precipitation which vary on a small spatial scale. Second, for random samples from the same population, a certain fraction of points (on average $A$) will appear significant at the $(1 - A)$ level of confidence. One can test if the field from the anomaly experiment is significantly different from the field in the control (Preisendorfer and Barnett 1983; Livezey and Chen 1983), but such tests do not indicate where the fields are significantly different. Third, since the distribution of precipitation deviates from the normal distribution (Reed 1986) assumed in the test, the significance levels should be regarded as relative rather than absolute. Most emphasis should be placed on those aspects of the model's response which have been explained on a physical basis.

Figure 15. As Fig. 14, but for the 5LM.
TABLE 5. PERCENTAGE AREA COVERED BY GRID SQUARES WHERE THE CHANGES IN PRECIPITATION ARE SIGNIFICANT AT THE 90% LEVEL OF CONFIDENCE

<table>
<thead>
<tr>
<th>Season</th>
<th>Spring</th>
<th>Summer</th>
<th>Autumn</th>
<th>Winter</th>
</tr>
</thead>
<tbody>
<tr>
<td>% area significant at 90% level of confidence</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Anomaly v. control</td>
<td>22.4</td>
<td>27.4</td>
<td>24.0</td>
<td>20.8</td>
</tr>
<tr>
<td>Control v. control</td>
<td>12.3</td>
<td>6.3</td>
<td>10.6</td>
<td>9.4</td>
</tr>
</tbody>
</table>

Upper row, the difference between the anomaly and control integration; lower row, the difference between two different periods from the control integration (see appendix). The seasons refer to the northern hemisphere: winter comprises December, January and February, and so on.

From Table 5, it can be seen that the percentage area with precipitation changes significant at the 90% level of confidence is consistently higher than 20% throughout the year. According to Livezey and Chen (1983, Fig. 6) this change in the field of precipitation is significant at the 90% level of confidence provided the number of spatial degrees of freedom is greater than 25. There is no generally accepted method for establishing the number of degrees of freedom, but given the small spatial scales on which precipitation

(a)

(b)

Figure 16. Student's t-tests on differences in precipitation in the 11 LM experiment (three years' data from the anomaly integration; eight years' from the control). A two-sided test has been used, allowing nine degrees of freedom, and grid boxes with changes which would occur by chance on less than ten per cent of occasions are stippled. Areas of significant increases are stippled heavily, areas of decrease are stippled lightly. Also shown are the standard deviations of the seasonal mean for the control integration. Solid contours at 1, 5 mm d⁻¹, chain dotted contour at 2 mm d⁻¹. (a) June to August; (b) December to February.
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varies in the model, it is exceedingly unlikely that it is less than 25. Also given in Table 5 are the percentage areas with precipitation changes significant at the 90% level, when two sets of three years’ data (four for spring) from the control integration are compared. The average value (9-7%) is close to the expected value (10%) for a large number of tests from a population of normally distributed variables.

4. DISCUSSION

The main thrust of this paper has been the dependence of simulated climate change on the distribution of climate in the control simulation. Here we consider some additional aspects of the 11LM experiment.

(a) Assessment of the changes in the hydrological cycle

The similarity of the global mean response in the two models indicates that the magnitude of the simulated changes is governed by the same factors. In this sub-section we consider the nature of the simulated response and compare it with that expected on the basis of certain simple assumptions.

(i) Precipitation and the atmospheric heat balance. Increasing the surface temperature from 15°C (the global mean) by 2 K increases the saturation vapour pressure at the surface by 13%. If we make the naive hypothesis that there are no marked changes in the static stability, wind speed and relative humidity in the bottom layer then evaporation and hence global mean precipitation would increase by approximately the same amount. The simulated increase in evaporation and precipitation is only 5-6% (4-9% in the 5LM), or less than half that expected on the basis of the above assumptions.

Kandel (1981) examined the surface energy budgets reported in several studies on the effect of increased atmospheric CO₂ and concluded that the sensitivity of surface temperature (when free to respond to the change in CO₂) depended critically on the constraints on evaporation and atmospheric humidity. In the present work, the change in surface temperature over the oceans is prescribed and is therefore independent of the surface energy budget. However, the changes in latent heat release to the atmosphere must balance other changes in the atmospheric heat budget. This we now consider in detail to identify the factors which influence changes in latent heat release, and hence precipitation in the 11LM experiment.

The 4-9 W m⁻² increase in latent heat release (Table 6) is balanced by a small decrease in the flux of sensible heat, and a 4-2 W m⁻² increase in radiative cooling. The latter comprises a 2-6 W m⁻² warming due to the increased absorption of solar radiation, and a cooling of 6-8 W m⁻² due to the enhanced divergence of long-wave radiation (note

<table>
<thead>
<tr>
<th>TABLE 6. GLOBAL ANNUAL MEAN CHANGES IN ATMOSPHERIC HEAT BALANCE (11LM, W m⁻²) DUE TO DOUBLING CO₂ AND ENHANCING SEA SURFACE TEMPERATURES BY 2 K</th>
</tr>
</thead>
</table>
| Radiative | -4-2 | [solar
| Latent | 4-9 | long-wave
| Sensible | -0-7 | -6-8 |

Note. These figures are for the atmosphere in isolation, not the combined atmosphere–surface system. The relative contributions from solar and long-wave radiation are given in the second column.
TABLE 7.  ESTIMATED CHANGES IN RADIATIVE FLUXES AND ATMOSPHERIC HEATING RATES DUE TO CHANGES IN CO₂, WATER VAPOUR AND TEMPERATURE (W m⁻²)

<table>
<thead>
<tr>
<th>Radiative component</th>
<th>Contribution to changes</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>CO₂</td>
</tr>
<tr>
<td>Net downward long-wave flux</td>
<td>Top</td>
</tr>
<tr>
<td></td>
<td>Surface</td>
</tr>
<tr>
<td></td>
<td>Heating</td>
</tr>
<tr>
<td>Solar</td>
<td>Heating</td>
</tr>
<tr>
<td>Net</td>
<td>Heating</td>
</tr>
</tbody>
</table>

The changes diagnosed during the model simulation are given in the final column. The contributions to the long-wave components were estimated by taking the annual mean profile at each grid point, running the radiation code with the changes applied one at a time, and globally averaging the results. The contributions to solar heating were estimated using a globally averaged single-column model.

that here we are considering the heat balance of the atmosphere in isolation, not the atmosphere and surface). We first consider the radiative changes in more detail.

The contribution of the increases in CO₂, water vapour and temperature to the enhanced long-wave cooling have been estimated by taking the annual mean profile at each grid point, running the radiation code with the changes applied one at a time, and globally averaging the results (Table 7). The effect of applying all three changes simultaneously was also estimated (Table 7, penultimate column), and found to be in close agreement with that found in the integration of the full model shown in Table 7 (last column).

Doubling CO₂ amounts reduces the outgoing long-wave flux at the top of the atmosphere by 2.1 W m⁻² (Table 7). The increase in the downward flux at the surface is smaller, as the lower atmosphere is optically thick at the wavelengths at which CO₂ is active, due largely to the overlap between the absorption bands of CO₂ and water vapour. As a result, increasing CO₂ amounts produces a direct warming of the atmosphere (see also Ramanathan 1981).

The importance of the overlap with water vapour is demonstrated by the results depicted in Fig. 17(a), obtained by running the radiation code for a single column of the atmosphere using globally averaged annual mean profiles with and without water vapour. The downward flux at the surface due to doubling CO₂ increases by 3.3 W m⁻² in the absence of water vapour (dashed line, Fig. 17(a)), but this reduces to 1.2 W m⁻² if the unperturbed water vapour profile is included (solid line). The effect of including water vapour is largely confined to the lower atmosphere, and changes the difference in the upward flux at the top of the atmosphere by less than 0.6 W m⁻².

The change in the downward flux at the surface due to the increase in water vapour is greater than at the top of the model (Table 7) for the following reasons. Water vapour absorbs long-wave radiation not only in the rotational/vibrational bands which overlap strongly with the absorption bands of CO₂, but also in the atmospheric window (8-12 μm) due to the water vapour continuum. The slab emissivity, \( \varepsilon \), for water vapour continuum absorption is parametrized according to the exponential relation

\[
\varepsilon = 1 - \exp(-k_1 up - k_2 ue)
\]  

(2)

where \( u \) is the absorber amount, \( p \) is pressure, \( e \) is the water vapour partial pressure and \( k_1, k_2 \) are temperature-dependent coefficients (Slingo and Wilderspin 1986). The second
term is important where the water vapour pathlengths and partial pressures are large (the lower troposphere, especially in the tropics). Hence, unlike the case of enhanced CO₂, the change in downward flux due to enhanced water vapour continues to increase from the middle troposphere to the surface due to increased emission in the water vapour continuum. As a result, enhancing the concentration of water vapour, in contrast to enhancing CO₂, produces a slight cooling of the atmosphere at infrared wavelengths.

The effect of the water vapour continuum is demonstrated in Fig. 17(b), again using results from the single-column model. The change in the downward radiative flux due to enhanced water vapour, neglecting continuum absorption, reaches a maximum near \( \sigma = 0.5 \) and decreases towards the surface (dashed line; this is qualitatively similar to the effect of increased CO₂ (solid line, Fig. 17(a)). The inclusion of continuum absorption enhances the change in downward flux at the surface from 1.6 W m⁻² to 5.4 W m⁻². (There is little change in the upward flux at the top of the atmosphere.) If the water vapour rotational bands are neglected, then the increase in water vapour produces an enhancement of the downward long-wave flux which increases monotonically with depth (dotted line, Fig. 17(b)). This effect is largely unchanged when the overlap with the other absorption bands is included, as much of the continuum absorption occurs across the atmospheric window where the atmosphere is optically thin.

The increase in surface and atmospheric temperatures produces a net cooling of over 6 W m⁻², largely due to increased cooling to space. We estimate, on the basis of single-column experiments, that about 0.6 W m⁻² of the 2.6 W m⁻² increase in solar absorption is due to doubling CO₂ and the remaining 2.0 W m⁻² is due to increases in
water vapour. Thus we conclude that doubling CO₂ reduces the atmospheric radiative cooling by about 1.5 W m⁻²; that water vapour has little effect on atmospheric heating rates, since the enhanced long-wave cooling is largely compensated by solar heating, and that the temperature increases enhance radiative cooling by about 6 W m⁻² (Table 7). With these findings in mind, we consider further the atmospheric heat balance and the implications of our naive hypothesis.

As noted earlier, the 4 W m⁻² net increase in radiative cooling is balanced by increased latent heat release (and a small reduction in sensible heating). The 13% increase in evaporation and precipitation deduced from the simple assumptions outlined at the beginning of this sub-section would produce an increase in latent heat release of 12 W m⁻². It would require a substantial alteration in the vertical distribution of the changes in temperature and humidity to increase the 4 W m⁻² to the 12 W m⁻² postulated above through enhanced radiative cooling. For example, further temperature increases in the boundary layer would increase the static stability at the surface reducing evaporation; increases at upper levels would increase the static stability of the troposphere and suppress convective instability. Similarly, further increases in boundary layer humidities would reduce evaporation. Further increases in moisture at upper levels would be limited by saturation, and would not contribute greatly to the net radiative cooling of the atmosphere. Thus it seems that constraints on changes in static stability and radiative cooling reduce the changes in evaporation and precipitation below that expected on the basis of constant relative humidity.

Given that the atmosphere cannot support a 13% increase in the intensity of the hydrological cycle, what mechanisms limit the increases in precipitation and evaporation? The decrease in relative humidity at upper levels (Fig. 11) is consistent with a reduction in the changes in condensation and precipitation below the ‘expected’ 13%; similarly, the increase in relative humidity near the surface is consistent with a ‘shortfall’ in the increase in evaporation. We now consider in more detail the ‘reduction’ in evaporation, assessing first the changes over the oceans, since they form the greater part of the earth’s surface, and the evaporation rate there is much greater than over land (Table 8).

The surface humidity (the saturation specific humidity at the surface temperature) increases by 12.9% over the oceans, whereas evaporation is enhanced by only 6.5% (Table 8). The lapse of humidity between the surface and the lowest layer increases by only 8.9%, due to a slight increase in the relative humidity at the lowest level, and the temperature increase in the lowest level being 0.33 K larger than that at the surface. The mean change in the surface exchange coefficient is small, so the remaining discrepancy between the hypothetical and simulated changes in evaporation must be attributed to a reduction in the magnitude of the surface wind.

Over land, the annual mean evaporation is unchanged (Table 8) although the lapse of humidity between the saturation value at the surface and the actual value in the lowest layer increases by 27.4%. There is a general decrease in soil moisture (not shown) which,

| TABLE 8. Global annual means over land or sea of evaporation, and changes in selected parameters due to doubling CO₂ and enhancing sea surface temperatures by 2 K |
|-----------------|----------|----------------|-----------------|
|                 | Evaporation (mm d⁻¹) | Change (%) | Change in surface humidity (%) | Change in humidity lapse at surface (%) |
| Land            | 1.51     | 0.0           | 20.8            | 27.4                        |
| Sea             | 3.64     | 6.5           | 12.9            | 8.9                         |
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in areas where evaporation is already limited by surface conditions, will counteract the increase in potential evaporation accompanying the rise in surface temperature.

(ii) Precipitation minus evaporation ($P - E$). From Eq. (1), if it is assumed that changes in circulation (represented by $V$) are small and that the fractional increase in humidity is uniformly distributed, then the changes in $|P - E|$ will be proportional to $q$, and the 20% rise in humidity should be accompanied by a similar increase in the magnitude of $P - E$. However, $|P - E|$ increases by considerably less (13%), as the fractional increase in specific humidity is not distributed uniformly, but increases with height (Table 4) and the largest contribution to the vertical integral in Eq. (1) comes from the lower troposphere, since humidity decreases rapidly with height, and the mass convergence in the middle troposphere is small. Even so, the fractional change in $|P - E|$ is smaller than the change in humidity in the boundary layer, indicating that changes in circulation, or horizontal variations in the fractional change in humidity, also contribute to changes in $P - E$.

(b) Effects of increased CO₂

In this sub-section we estimate the changes in global mean surface temperature that would be in equilibrium on doubling atmospheric CO₂, and the accompanying change in global mean precipitation. We follow the approach taken by Mitchell (1983a) to assess the 5LM response, assuming that the response of the model to small changes in sea surface temperature is linear, and that the effects of CO₂ and sea surface temperature are additive. In addition, we will assume that doubling CO₂ alone produces an increase of 3·5 W m⁻² in net surface heating, as found using the 5LM (Mitchell 1983a). Surface heating is reduced by 1·2 W m⁻² due to doubling CO₂ and increasing sea temperatures by 2 K in the 11LM. A 2K increase in sea surface temperatures alone might therefore be expected to reduce the net surface heating by 4·7 (i.e. 3·5 + 1·2) W m⁻². For the changes in CO₂ and sea surface temperatures to be in balance, the change in net surface heating should be zero, which would require a sea surface temperature increase of 1·5 (i.e. 2 × 3·5/4·7) K, or a reduction of 25% from the present experiment, to balance the 3·5 W m⁻² increase in surface heating due to doubling CO₂. The corresponding changes in global mean surface temperature (allowing for land) and precipitation would also be 25% less than the values in Table 3 at 1·7 K and 4·2% respectively, in broad agreement with the 2·0 K and 3·4% obtained by Manabe and Stouffer (1980) using an atmospheric model coupled to a rudimentary model of the upper ocean (if one assumes that a quadrupling of CO₂ amounts has twice the effect of doubling).

In view of the relationship between precipitation and the atmospheric heat balance discussed in sub-section 4(a)(i), a better estimate of the precipitation change at equilibrium can be given. Assuming the tropospheric temperature change, and hence the associated changes in sensible heat flux and net long-wave cooling of the atmosphere are also reduced by 25%, the resultant cooling of the atmosphere due to changes in temperature alone would be 0·75 × (0·7 + 6·3) = 5·25 W m⁻² (see Tables 6 and 7). This must be balanced by contributions due to changes in water vapour (-0·3 W m⁻²), CO₂ (1·5 W m⁻²) and precipitation. This implies an enhancement of latent heat release of 4·05 W m⁻², corresponding to a 4·7% increase in precipitation.

The surface warming is smaller than in other studies of the effect of doubling CO₂ amounts (Hansen et al. 1984; Washington and Meehl 1984; Wetherald and Manabe 1986). In those studies the response to increased CO₂ was amplified by reductions in sea-ice extents and changes in cloud cover; here the distributions of cloud and sea-ice were prescribed and unchanged.
5. Summary and Concluding Remarks

The responses of two general circulation models to doubling CO₂ concentrations and prescribing a 2 K increase in sea surface temperatures have been compared. The main findings are:

1. The large-scale responses are similar in two models. The following features were identified as being common to both experiments.
   a. The models produce a vertical profile of temperature changes consistent with maintaining a moist adiabatic lapse rate, giving temperature increases which increase with height from about 2 K near the surface up to 5 K in the tropical upper troposphere.
   b. To a first approximation, the atmospheric moisture content increases so that relative humidity is unchanged. The main deviations from this generalization in the 11 LM occur in the upper troposphere, where relative humidity is reduced, and in the lower stratosphere and near the surface, where it is increased.
   c. Precipitation is increased where it is already heavy due to the increase in specific humidity leading directly to increased moisture convergence. Conversely, precipitation changes little or decreases where it is meagre in the control simulation.
   d. The magnitude of the changes in the supply of moisture to the ground \((P - E)\) is roughly proportional to the change in atmospheric moisture content in the lower atmosphere, which in turn is related to the change in atmospheric temperature through the Clausius-Clapeyron relation. The change in global mean precipitation is smaller, being limited by the atmospheric heat balance. The evidence presented in this study indicates that it is important not only to represent the radiative perturbation due to increased CO₂ accurately, but also that due to changes in water vapour, including the effects of the water vapour continuum.

2. On a regional scale, although there is some agreement, important differences do occur, as shown for precipitation. Many of these discrepancies can be related to differences in the control simulation. This is to be expected since the changes in the model boundary conditions are small enough to be regarded as a perturbation.

3. In comparing the results from different models used to investigate the effect of perturbations on climate, for example due to increased atmospheric CO₂, one needs to assess the influence of the different model formulations. This may be extremely difficult because the models are likely to differ in more than one respect. Alternatively, one can identify the physical mechanisms which dominate the simulated response in each model. If the same mechanisms operate, one can then assess the likely effect of differences in unperturbed simulations on each model's response. Failure to do so will lead to unnecessarily pessimistic conclusions concerning the consistency of numerical studies of climate and climate change.

4. The close relationship between unperturbed climate and climate change indicates that it is necessary (though not sufficient) to use a model which simulates the present day climate accurately if it is to produce accurate estimates of climate change. It may be possible to allow for some of the discrepancies between observed and simulated climate provided the physical mechanisms which dominate the modelled response have been identified (and are realistic). This is analogous to the interpretation of forecasts from early numerical prediction models, allowance being made for known biases in the model.

5. The 11-layer model results indicate that the equilibrium changes in global mean surface temperature and precipitation due to doubling CO₂ (in the absence of changes in sea-ice extents or cloud cover) would be 1.7 K and 4.2% respectively. This is based on the
assumption that globally averaged surface temperatures, precipitation and surface heating vary linearly with small changes in sea surface temperature. A more detailed analysis of the atmospheric heat budgets suggests a slightly larger (4.7%) increase in precipitation.

ACKNOWLEDGMENTS

We thank Andrew Gilchrist, Peter Rowntree and Howard Cattle for useful comments and discussions and Bruce Ingleby for estimating the spatial correlation coefficients in the appendix. Suggestions from R. Dickinson and an anonymous referee have been incorporated. This work was supported in part by EEC Contract No. CLI-030-81-UK(H).

APPENDIX

Estimates of the spatial correlation coefficient

The spatial correlation coefficient \( \rho_{p,q} \) between fields \( p \) and \( q \) is defined by

\[
\rho_{p,q} = \frac{\text{covar}(p, q)}{\sqrt{\text{var}(p) \cdot \text{var}(q)}}
\]  

(A1)

where, for example, the spatial variance \( \text{var}(p) \) is given by \( (p_i - P)^2 \). \( p_i \) is the value of \( p \) at position \( i \), \( P = \bar{P}_i \), and overbar denotes the area-weighted mean over the region of interest. In practice, the grid point values \( p_i \) and \( q_i \) fluctuate in time due to the model's inherent variability, and so must be estimated from the available samples. With \( N \) independent samples \( p_{ij} \) (\( i = 1 \) to \( N \)), the most obvious estimate of \( \rho_{p,q} \) is obtained by replacing \( p_i \) by the temporal average \( (1/N) \sum_{j=1}^{N} p_{ij} \), and so on. However, the spatial variability tends to inflate the temporal estimate of the spatial variance. For example, if the temporal variances of \( p_i \) and \( P \) are \( \sigma_i^2 \) and \( \Delta^2 \) respectively, the expected value of the spatial variance of \( p \) estimated from a sample of size \( N \), \( E[\text{var}(p)] \), is given by

\[
\text{var}(p) + (1/N)(\sigma^2 - \Delta^2)
\]  

(A2)

where \( \sigma^2 = \bar{\sigma}_i^2 \). Now \( \sigma^2 - \Delta^2 \geq 0 \), and in practice, \( \sigma^2 \gg \Delta^2 \), so the expected value of the variance exceeds the actual value, especially for small samples. It can also be shown that

\[
E[\text{covar}(p, q)] = \text{covar}(p, q)
\]

provided temporal variations of \( p \) and \( q \) are not correlated in time.

We can estimate \( \sigma_i^2 \) and \( \Delta^2 \) from the samples \( p_{ij} \) in the normal manner by \( s_i^2 \) and \( d^2 \) respectively, where

\[
s_i^2 = \frac{1}{(N-1)} \sum_j \left( p_{ij} - \frac{1}{N} \sum_j p_{ij} \right)^2
\]

\[
d^2 = \frac{1}{(N-1)} \sum_j \left( P_i - \frac{1}{N} \sum_j P_{ij} \right)^2
\]  

(A3)

An unbiased estimate for \( \text{var}(p) \) is the sample variance reduced by \( (1/N)(\sigma^2 - \Delta^2) \), which can then be used in (A1).

The correlation of \( p \) with \( q - p \) is given by

\[
\rho_{p,q-p} = \frac{\text{covar}(p, q) - \text{var}(p)}{\sqrt{\text{var}(p)\text{var}(p) + \text{var}(q) - 2\text{covar}(p, q)}}
\]

Here \( p \) and \( q \) represent \( P-E \) sampled from the control and anomaly integration respectively, \( \rho_{p,q-p} \) represents the correlation of changes in \( P-E \) with \( P-E \). The global spatial
TABLE A1. SPATIAL CORRELATION BETWEEN P-E AND CHANGES IN P-E (USING ANNUAL MEAN DATA)

<table>
<thead>
<tr>
<th></th>
<th>Correlation coefficient</th>
</tr>
</thead>
<tbody>
<tr>
<td>11 LM</td>
<td></td>
</tr>
<tr>
<td>Uncorrected</td>
<td>-36</td>
</tr>
<tr>
<td>Corrected</td>
<td>-59</td>
</tr>
<tr>
<td>5 LM</td>
<td></td>
</tr>
<tr>
<td>Uncorrected</td>
<td>-28</td>
</tr>
<tr>
<td>Corrected</td>
<td>-35</td>
</tr>
</tbody>
</table>

Correlations and corrected values are given in Table A1. It has been assumed that $\sigma^2_i$ and $\Delta^2$ are the same in both the anomaly and control integrations, and $\sigma^2_i$ and $\Delta^2$ were derived from the eight-year control integration from the 11 LM, and from the three-year control integration of the 5 LM.

REFERENCES


GARP 1979. ‘Report of the JOC Conference on Climate Models; Performance, intercomparison and sensitivity studies’. GARP Publication Series No. 22. ICSU/WMO.


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Manabe, S. and Stouffer, R. J.

Manabe, S., Hahn, D. G. and Holloway, J. R. Jr.

Manabe, S., Wetherald, R. T. and Stouffer, R. T.
1981 Summer dryness due to an increase of atmospheric CO₂ concentration. Climatic Change 3, 347–385

Mitchell, J. F. B.

Mitchell, J. F. B. and Bolton, J. A.

Mitchell, J. F. B. and Lupton, G.
1984 A 4×CO₂ experiment with prescribed sea temperatures. Progress in Biometeorology. 3, 353–374


Mitchell, J. F. B., Wilson, C. A., Ingram, W. I. and Cunninston, W. M.

Moller, F.
1951 Quarterly charts of rainfall for the whole earth, (in German). Petermanns Geogr. Mitt., 95, 1–7

Newell, R. E., Kidson, J. W., Vincent, D. G. and Boer, G. J.

Oort, A. H. and Rasmusson, E. M.

Palmer, T. N., Schuttis, G. J. and Swinbank, R.

Palmer, T. N. and Mansfield, D. A.

Preisendorfer, R. W. and Barnett, T. P.

Ramanathan, V.
1981 The role of ocean atmosphere interactions in the CO₂ climate problems. ibid., 38, 918–930

Reed, D. N.

Rowntree, P. R. and Bolton, J. A.

Schutz, C. and Gates, W. L.

1972 ‘Global climatic data for surface, 800 mb, 400 mb: July’. ibid., R-1029-ARPA

Slingo, A.
1985a Simulation of the earth’s radiation budget with an 11-layer general circulation model. Meteorol. Mag., 114, 121–141


Slingo, A. and Wilderspin, R. C.


Wetherald, R. T. and Manabe, S. 1986 An investigation of cloud cover change in response to thermal forcing. *Climate Change*, 8, 5-24