The response of a general circulation model to cloud longwave radiative forcing. I: Introduction and initial experiments

By A. SLINGO and J. M. SLINGO
National Center for Atmospheric Research, Boulder, Colorado 80307, U.S.A.

(Received 8 July 1987; revised 15 January 1988)

SUMMARY

A new version of the NCAR Community Climate Model (CCM1) is used to study the effect of cloud radiative forcing on model simulations. Previous attempts to determine the role played by clouds in climate and the general circulation of the atmosphere are reviewed first. The concept of cloud radiative forcing is discussed and the forcings in the shortwave and longwave spectral regions are contrasted. At low latitudes, the cloud longwave forcing is primarily within the atmosphere, rather than at the surface. It is thus appropriate to study its effect with a model which employs fixed sea surface temperatures (s.s.t.s).

The impact of cloud longwave forcing is studied in 510-day integrations for constant January conditions. The experiments isolate the forcing by tropical and extra-tropical clouds. Tropical cloud forcing warms the tropical upper troposphere and accelerates the subtropical jets. There are subtle interactions between the forcing, the clear-sky longwave heating and the latent heating in the tropics. These additional diabatic heating terms oppose the forcing in some regions and enhance it in others.

The tropical cloud forcing strengthens the precipitation maxima at low latitudes. There are also changes in the extra-tropical flow, including the excitation of a pattern in the 200 mb geopotential height differences which is similar to those found in previous studies of the effect of s.s.t. anomalies. This confirms that the forcing may be as important as latent heat release in determining the atmospheric response to such anomalies. The model is excited with comparable strength by the extra-tropical cloud forcing. Finally, some concerns regarding the generality of these results and their applicability to other models and the real atmosphere are discussed. These concerns suggest the need for sensitivity studies, some of which are already in progress.

1. INTRODUCTION

This paper represents a first step towards quantifying the role of cloud–radiation interactions in the general circulation of the atmosphere. As is well known, these interactions may be considered separately in two spectral regions, corresponding to the incoming shortwave radiation from the sun and the outgoing longwave radiation emitted by the earth–atmosphere system. In the shortwave, clouds reflect solar radiation back to space and thus cool the system. In contrast, since cloud top temperatures are generally lower than those of the underlying atmosphere and surface, clouds reduce the outgoing longwave radiation and hence enhance the greenhouse warming of the system. A determination of the net radiative effect of clouds on climate thus requires quantification of these two opposing forcings.

Much of the interest in the possible impact of cloud–radiation interactions on climate has come from the need to reduce uncertainties in climate-change experiments. Using a simplified sector model with an idealized land–sea distribution, Wetherald and Manabe (1980) ran two versions, one with model-generated cloudiness and a second with imposed cloud distributions, and compared the sensitivity of each version to different values of the solar constant. In the first set of experiments the cloud distributions changed in response to the external forcing, but the sensitivity of the model was very similar to that in the second set of experiments. This was due to a compensation between the effect of clouds in the shortwave and longwave regions. Wetherald and Manabe concluded that cloud feedbacks were unimportant, but in subsequent investigations with a global model...
and realistic geography, they found that cloud feedbacks enhanced the sensitivity of the surface temperature to doubled CO$_2$ by about 30 per cent (Wetherald and Manabe 1986). This brought them into agreement with the results discussed by Hansen et al. (1984). In both studies, the enhanced warming was produced by increases in the amount of high level cloud in the tropics, leading to a reduction in the outgoing longwave radiation and hence amplification of the initial warming. However, both studies ignore the potentially important changes in cloud radiative properties which may be expected to accompany the cloud amount changes (Somerville and Remer 1984). The magnitude of the cloud feedback thus remains one of the greatest sources of uncertainty in current predictions of the effects of increasing CO$_2$ on climate.

Since cloud–radiation interactions strongly influence the modelled response to radiative perturbations such as changes in the solar constant or CO$_2$ amounts, they may also contribute to the unperturbed circulation. The first study of the effect of clouds on the general circulation was performed by Hunt (1978), who integrated a hemispheric model with annual-mean external forcing and studied the effect of removing zonal-mean clouds from the radiative computations. Surface and tropospheric temperatures increased, due to the dominance of the shortwave effects, but there were only small changes in the zonal-mean circulation. This result has been criticized as misleading, however, as a number of possibly important feedbacks were suppressed by assuming a flat earth with no land–sea contrast and through the use of fixed water vapour and surface albedo distributions (G. E. Hunt et al. 1980).

Satellite observations of the earth's radiation budget show that there are substantial asymmetries in the cloud distribution, so the assumption of zonal-mean cloudiness in Hunt's study is also unrealistic. To study the effect of such asymmetries, Meleshko and Wetherald (1981) compared an integration of the GFDL spectral model for northern summer conditions and imposed zonal-mean cloudiness, with one in which a geographical cloud distribution was used. In the latter experiment, the cloud cover was reduced over most land areas and increased over the oceans compared with the former integration. Over land, both the solar insolation and surface longwave cooling were thus increased, although the former was dominant. Land surface temperatures rose by 2–4 K, enhancing the continental heat lows and modifying the wind and precipitation patterns. These changes were induced by the increased asymmetry in the distribution of diabatic heating in the second experiment. The sea surface temperatures were fixed, so these results are applicable only for time-scales over which this restriction is valid, corresponding roughly to the 30 days over which the differences were computed. The potential climatic impact of the different cloud distributions was thus underestimated, since in reality the increased cloud cover over the oceans would cool the sea surface and thus enhance the land–sea contrast.

In addition to geographical variations in cloudiness, there are of course also temporal variations associated with synoptic systems. Shukla and Sud (1981) investigated the importance of such changes by comparing an integration of the GLAS model for northern summer conditions in which the cloud cover was predicted (and hence varied in time) with one in which the mean cloud distribution from this run was imposed. There were significant differences between the two integrations, indicating that temporal variations in the cloud cover influenced the model's circulation. With the imposed cloud distribution, the fixed asymmetric thermal forcing increased the generation of eddy available potential energy and its conversion to eddy kinetic energy. At 50°N, the stationary component of the 500 mb height variance was also increased.
Le Treut and Laval (1984) performed an experiment which combined features of each of the above. They compared January and July control integrations having prescribed, zonal-mean cloudiness with runs in which the cloud cover was interactively predicted by the model. As found by Meleshko and Wetherald, many of the most significant changes in the simulations were produced over the oceans, presumably being remotely forced by changes in the surface energy budget over the continents. The simulations of precipitation and evaporation patterns were generally more realistic in the integrations with interactive cloud. Le Treut and Laval found a similar sensitivity of the atmospheric energy cycle to the cloudiness as was obtained by Shukla and Sud, with an increase of the transient eddy kinetic energy when cloudiness was interactive.

In the above studies, many of the changes were caused by the effect of clouds on the incoming shortwave flux at the surface, enhancing the thermal forcing of the land surface and thus indirectly changing tropospheric temperatures and the general circulation. In contrast, Ramanathan et al. (1983) demonstrated that cloud longwave interactions within the atmosphere had a profound impact on the circulation in the NCAR Community Climate Model (CCM). Altering the emissivity of high cloud and the number of model layers in which it was allowed to form changed the zonal-mean temperature of the tropical upper troposphere by over 5 K. This change enhanced the equator-to-pole temperature gradient and so, through the thermal wind relationship, there was a corresponding acceleration of the subtropical jets by up to about 9 m s$^{-1}$. Each of these changes represents a substantial impact of clouds on the model's circulation.

Analyses of model output and satellite data to determine the effect of clouds on the radiation budget (and hence also estimate the cloud feedback) have traditionally involved the sensitivity parameter $\delta$, which is the partial derivative of the net radiation with respect to the cloud amount (e.g. Hartmann et al. 1986). While this approach seems reasonable, in practice it is fraught with problems because of the need to know the cloud amount, which is a notoriously difficult quantity to determine accurately from observations. Additionally, the techniques used to define the cloud amount and the assumed cloud radiative properties may be quite different between observational studies and models, or between different models. Comparison of values of $\delta$ from different sources can thus be very misleading. What is needed is a parameter which does not require explicit knowledge of the cloud amount. A simple analysis fulfilling this requirement was proposed by V. Ramanathan (Charlson and Ramanathan 1985; Hartmann et al. 1986; Ramanathan 1987). For a region in which there is partial cloud cover, the effect of clouds on some area-mean radiative flux is simply taken to be the difference between that flux and its value for clear skies. The sign of the difference is changed so that implied heating appears as a positive quantity. This difference is called the cloud radiative forcing. For many purposes this is a more useful parameter than $\delta$, because it gives the effect of clouds directly in energy units. It can also be derived with some confidence from satellite data because it is much easier to derive the clear-sky flux than the cloud amount.

In the next section, cloud radiative forcing is discussed in more detail. It is also argued that, in order to use a general circulation model to study the effects of cloud shortwave radiative forcing on climate, it is necessary to employ an interactive ocean surface. On the other hand, many of the effects of the longwave forcing can be studied with a non-interactive ocean surface (i.e. with imposed sea surface temperatures and sea-ice extents). It thus seems logical that the longwave forcing should be studied first. In subsequent sections, results from a series of experiments with a new version of the CCM ('CCM1') are used to study the effect of cloud longwave radiative forcing on the atmospheric general circulation.
2. CLOUD RADIATIVE FORCING

Following Ramanathan (1987), let $F$ be the area-mean radiative flux (in W m$^{-2}$) at the top of the atmosphere reflected by, or emitted from, a region with partial cloud cover and let $F_c$ be the flux for clear skies. Then the cloud radiative forcing $C_t$ is

$$C_t = F_c - F. \quad (1)$$

It is convenient to consider the cloud radiative forcing separately in the shortwave and longwave spectral regions. Diagnostics are shown from the control integration of CCM1 (for perpetual January conditions), described in the next section.

(a) Cloud shortwave forcing

The cloud shortwave forcing, $C_t(S)$, is

$$C_t(S) = S(\alpha_c - \alpha) \quad (2)$$

where $S$ is the incoming solar flux, $\alpha$ is the albedo and $\alpha_c$ is the albedo for clear skies. Maps of $\alpha$, $\alpha_c$ and $C_t(S)$ from the control experiment are shown in Fig. 1. The albedo $\alpha$ may be compared with the Nimbus 7 Narrow Field Of View data for December 1979 to February 1980, shown by Hartmann et al. (1986, Fig. 4(c)). The modelled distribution is in reasonable agreement with the satellite data, although there are some deficiencies which were also present in an earlier version of the model (Charlock and Ramanathan 1985, Fig. 7). In the tropics, the albedo maxima over the regions of deep convection are weaker than in the Nimbus 7 data. The albedo minima over the subtropical oceans are too close to the western coasts of South Africa and South America. In reality, there are extensive sheets of stratus and stratocumulus over the cool waters adjacent to the coasts, with clearer skies to the west, but in the model this pattern is reversed. Problems with the representation of such cloud sheets are a common feature of current models (ECMWF 1985). Results to be presented in a future paper demonstrate that in the CCM this error may be corrected by replacing the cloud prediction scheme by that developed for the ECMWF model (J. M. Slingo 1987). Finally, the albedos are also too low in the southern hemisphere mid-latitudes. This discrepancy is due to cloud amounts which are too low, a possible consequence of the serious underestimate of the strength of lows in the southern hemisphere depression belt, which is a feature of the model when run with low horizontal resolution (e.g. Pitcher et al. 1983). The global-average albedo is 29.8 per cent, which compares favourably with satellite data (Hartmann et al. 1986, Table 1).

The clear-sky albedo $\alpha_c$ (Fig. 1(b)) is generally much smaller than $\alpha$ at low latitudes, because the albedo of the underlying surface is much lower than that of clouds. One obvious exception is the relatively high reflectivity of the North African desert. At high latitudes, however, the surface albedo is generally large due to snow cover and sea-ice, so the contrast between the two maps is less evident.

The cloud shortwave forcing $C_t(S)$ (Fig. 1(c)) is everywhere negative, corresponding to the cooling of the system by the albedo effect. The global-average value is $-51.1$ W m$^{-2}$. The forcing is a maximum over the southern hemisphere mid-latitudes, which is where both the insolation and the cloud cover are large. As noted above, the model underestimates the cloud cover in this region, so the values of cloud forcing should actually be larger than those shown here.

Apart from cloud reflection, the atmosphere is relatively transparent to shortwave radiation, so the cloud shortwave forcing is felt primarily at the surface rather than within the atmosphere (e.g. Ramanathan 1987). Figure 1(c) shows that, in this season, most of
Figure 1. (a) Shortwave albedo and (b) clear-sky shortwave albedo, from the control experiment. The contour interval is 10 per cent. Values greater than 30 per cent are hatched and those smaller than 20 per cent are stippled. (c) Cloud shortwave forcing of the earth–atmosphere system. The contour interval is 20 W m$^{-2}$ and values less than $-80$ W m$^{-2}$ are hatched.
the forcing is over the southern oceans. In the northern summer, the forcing is larger over the northern continents, but it is also significant over the northern and tropical oceans. In order to investigate the effect of this forcing on climate, one would therefore need a coupled ocean–atmosphere model, so that sea surface temperatures (s.s.t.s) were calculated rather than prescribed, as otherwise the response would be misleadingly small. Such an investigation would require considerable computing resources to integrate the model to equilibrium, so it was decided to concentrate on the effects of cloud longwave forcing for the present study.

(b) Cloud longwave forcing

The cloud longwave forcing, \( C_f(L) \), is

\[
C_f(L) = \text{OLR}_c - \text{OLR}
\]  

(3)

where \( \text{OLR} \) is the outgoing longwave radiation and \( \text{OLR}_c \) is the value for clear skies. Maps of \( \text{OLR} \), \( \text{OLR}_c \) and \( C_f(L) \) are shown in Fig. 2. As with the shortwave albedo, the \( \text{OLR} \) map (Fig. 2(a)) may be compared with Nimbus 7 data (Hartmann et al. 1986, Fig. 5(c)). The modelled \( \text{OLR} \) distribution agrees well with the Nimbus 7 data and the global average of 239.8 W m\(^{-2}\) also falls within the range of the satellite measurements (Hartmann et al. 1986, Table 1). The minima in the tropics over South America, South Africa and Indonesia are located correctly, although the central values are not as low in the model as in these observations. Over the tropical oceans there are some areas where the \( \text{OLR} \) exceeds 300 W m\(^{-2}\), which is probably the result of insufficient cloud cover, as these are also the regions where the shortwave albedo is too low.

Comparison between the \( \text{OLR} \) distribution and that for clear skies (Fig. 2(b)) shows differences which are similar to those seen earlier in the albedos. In the tropics, the value of \( \text{OLR}_c \) is typically about 300 W m\(^{-2}\) and is less near the poles due to the lower surface and atmospheric temperatures. At high northern latitudes, the values are lowest over the cold continents and largest over the relatively warm North Atlantic Ocean. The greatest contrast between Figs. 1(a) and (b) is therefore in those regions where persistent high (and therefore cold) cloud is found above a warm surface, that is in the areas of deep tropical convection and in the North Atlantic storm track. These are the regions which show the largest values of the cloud longwave forcing (Fig. 2(c)), as was also noted by Hartmann et al. (1986).

The longwave forcing is positive everywhere, showing that clouds warm the system by enhancing the greenhouse effect. The global-average value is 30.2 W m\(^{-2}\). Since this is lower than the average shortwave forcing, the net effect of clouds in this model is to cool the system, in broad agreement with the results from other models (Cess and Potter 1987).

Of particular importance in the distribution of longwave forcing is the coincidence of the largest values with areas of deep tropical convection. This coincidence suggests that there may be important interactions between the forcing and the convection, which is of course responsible for most of the cloudiness in these regions. Before examining those interactions, however, it is first necessary to consider the partitioning of the longwave forcing between the surface and atmosphere.

(c) Cloud longwave forcing of the surface and atmosphere

The effect of clouds on longwave heating rates may be understood by considering Fig. 3, which shows schematically the perturbation of the fluxes and heating rates produced by high and low clouds inserted into two extreme atmospheric profiles. The calculations were made with the radiation scheme incorporated into the Meteorological
Figure 2. (a) Outgoing longwave radiation and (b) clear-sky outgoing longwave radiation, from the control experiment. The contour interval is 25 W m$^{-2}$ and values smaller than 225 W m$^{-2}$ are hatched. (c) Cloud longwave forcing of the earth-atmosphere system. The contour interval is 20 W m$^{-2}$ and values greater than 40 W m$^{-2}$ are hatched.
Office model, as described by Slingo and Wilderspin (1986). The scheme was run first for clear skies, followed by runs with complete cover by low and then high cloud, the latter being of course considerably lower in the subarctic winter profile than in the tropical profile because of the shallower troposphere. The emissivity of low cloud was taken to be unity (i.e. the clouds act as black bodies), whereas that of high cloud was assumed to be 0.5. A value smaller than unity is implemented simply by multiplying the cloud amount by the emissivity.

The changes in the OLR produced by the clouds are listed at the top of Fig. 3, while the changes in the downward longwave radiation at the surface are given at the bottom of the figure. Consider first the OLR. Since temperature generally decreases with height, the temperature of a cloud top is usually less than that of the underlying atmosphere and surface, so the longwave emission from the cloud top is smaller than the upward flux in a clear atmosphere at the same level. The OLR is thus reduced by the presence of cloud, giving rise to greenhouse warming of the earth–atmosphere system. The magnitude of the OLR decrease depends on the difference in temperature between the cloud top and the atmosphere and surface below, and so is largest for the high (cold) cloud in the tropical atmosphere and is smallest for the low cloud in the subarctic winter atmosphere, where the cloud top temperature is almost the same as that of the surface.

A similar argument applies to the downward radiation at the surface. Cloud base emission is generally higher than the downward flux from a clear atmosphere at that
level, so clouds increase the downward radiation and warm both the atmosphere below cloud base and the surface. The surface warming is greatest for low clouds, because for these the differential between clear and cloudy regions is greatest. It is also larger in the subarctic winter profile than in the tropical profile, because the atmosphere in this case has a much smaller water vapour content and is thus more transparent. Hence there is a greater contrast between the downward radiation in clear and cloudy skies, as well as less damping of that contrast by the water vapour emission and absorption below cloud base.

Substantial changes are also produced in the heating rates at the level of the clouds. Cloud tops are cooled because their emission exceeds that incident from above. This cloud top cooling has only a weak dependence on the cloud height. On the other hand, cloud bases are warmed because the atmosphere and surface below are at higher temperatures. This warming is strongly dependent on the cloud height; as the cloud is raised so the temperature differential increases. For most clouds the cloud top cooling exceeds the cloud base warming, so that a net cooling is produced locally. However, for high clouds in a tropical atmosphere the cloud base warming can be several tens of kelvins per day (e.g. Webster and Stephens 1980). For cloud bases above about 10 km, the warming of the base thus exceeds the cooling of the top, so there can be a net warming of the cloud layer (e.g. Paltridge and Platt 1976, Fig. 10.9).

In summary, in the longwave spectral region clouds warm both the earth–atmosphere system and the surface. Cloud bases warm and cloud tops cool. Most clouds cool the atmosphere locally, but warm the atmosphere below cloud base. High tropical clouds are sufficiently cold that they can warm the atmosphere locally.

The above discussion clarifies the partitioning of the cloud longwave forcing between the surface and the atmosphere. In a model it is obviously a simple matter to repeat the calculation in Eq. (3) for the downward longwave radiation at the surface and so derive the cloud longwave forcing of the surface, which is shown in Fig. 4(a). The values are everywhere positive so that, as explained above, clouds warm the surface. The warming is greatest at high latitudes and smallest at low latitudes, where the substantial water vapour mixing ratios in the boundary layer damp the impact of clouds on the surface fluxes (e.g. Feigelson et al. 1982).

The difference between the cloud longwave forcing at the top of the atmosphere and that at the surface gives the cloud longwave forcing of the atmosphere itself, which is shown in Fig. 4(b). Note that over most of the globe clouds cool the atmosphere, as expected from the earlier discussion. However, the presence of extensive sheets of high cloud over regions of tropical convection produces a significant warming. Comparison of Figs. 2(c) and 4(b) shows that in these regions the majority of the cloud longwave forcing is felt within the atmosphere, rather than at the surface. This result is important, for three reasons. Firstly, it indicates that there must be a direct interaction between the convective heating and the cloud longwave forcing, because both occur within the atmosphere where there is deep convection. Secondly, the peak values of the cloud longwave forcing of the atmosphere over Indonesia exceed 80 W m$^{-2}$. Even if this heating were spread throughout the troposphere (which is unlikely), the mean diabatic heating from this term would be about 1 K/d, which is by no means negligible compared with the latent heat released by convection (Ramanathan 1987). Thirdly, it should be possible to study the effect of this forcing on atmospheric convection and dynamics with a model incorporating prescribed s.s.t.s, which is tantamount to ignoring the cloud longwave forcing of the ocean surface.
In this model, the peaks in the cloud shortwave forcing of the atmosphere corresponding to those in Fig. 4(b) are at least a factor of 5 smaller in magnitude, so that the cloud longwave forcing dominates within the atmosphere.

For these reasons, it was decided to use the CCM with prescribed s.s.t.s to study the effect of cloud longwave forcing, concentrating here on the forcing within the atmosphere. The integrations were carried out for perpetual January conditions. At this time of year the inter-tropical convergence zone in the eastern hemisphere is situated over Indonesia, which minimizes the possible importance of the cloud longwave forcing of the surface, as in this region most of the surface is ocean. In addition, this is also the season when the general circulation favours the propagation of large-scale disturbances from the tropics to high northern latitudes (Hoskins and Karoly 1981) and so it should be possible to study the contribution of the radiative forcing to this process.
(d) Discussion of cloud radiative forcing

The cloud radiative forcing diagnostic must be interpreted with care to avoid confusion, because the word ‘forcing’ has certain connotations. In conventional usage, forcing implies an external influence which induces a response in a system, but the forcing is independent of the response. Because of the closed nature of the climate system, examples of such non-interaction are few, although changes in the solar output or in the earth’s orbit clearly provide external forcings which are independent of the climatic response. Clouds, however, are a highly interactive component of the system, their presence being in part a response to, for example, large-scale upward motion and the hydrological cycle. In addition, atmospheric temperatures and humidities and the surface temperature are affected by the presence of clouds, so that even the clear-sky radiative fluxes are not independent of the cloud cover and vary in time as the cloud cover changes. The magnitude of the cloud radiative forcing is therefore influenced by internal interactions and feedbacks, so it does not constitute an external perturbation. Nevertheless, it does provide a measure of the instantaneous impact of clouds on radiative fluxes and heating rates and as such is as useful a diagnostic as, for example, convective heating and the heating by large-scale adiabatic motion. It can be determined easily in numerical models and compared with the estimates which are being made during the Earth Radiation Budget Experiment (ERBE Science Team 1986). It is also appropriate to consider the role of changes in the cloud radiative forcing on climate and the atmospheric general circulation, in the present experiments by removing the forcing altogether as a sensitivity study to determine its importance.

3. Model and Integrations

(a) Model

The original version of the NCAR Community Climate Model (‘CCM0’) and results from simulations with it were described by Pitzer et al. (1983) and Ramanathan et al. (1983). Williamson et al. (1987) describe in detail the changes which were incorporated into CCM1. The most relevant changes for the present study are outlined below.

The standard version of the new model has 12 layers. The sigma layer structure is illustrated in Fig. 5, excluding a label for the second highest layer at \( \sigma = 0.025 \).

The variable (i.e. non-black) longwave emissivity of high cloud described by Ramanathan et al. (1983) is included. The emissivity of a cloud is calculated from the liquid water content, which is given by the amount of moisture condensed in the previous hour of model time. The effective cloud amount used in the longwave radiation calculations is the product of this emissivity and the cloud amount. As in CCM0, the layer cloud amount is taken to be 0.95 when large-scale condensation occurs and is zero otherwise. The convective cloud amount is taken to be 0.30 when moist convective adjustment is taking place and is also zero otherwise. Clouds are not allowed to form in either the surface layer (\( \sigma = 0.991 \)) or the top three layers (\( \sigma = 0.060, 0.025 \) and 0.009).

Major changes were made in the radiation scheme. In the longwave, improved parametrizations are used for the absorption by water vapour (Ramanathan and Downey 1986), carbon dioxide (Kiehl and Briegleb 1987, personal communication) and ozone (Ramanathan and Dickinson 1979). In the shortwave, the Kratz and Cess (1985) treatment for water vapour absorption and the Kiehl and Yamanouchi (1985) method for oxygen absorption are used. The surface albedo parametrization of Briegleb et al. (1986) is included.
Several changes were made in other areas of the model, including revised vertical finite differencing and vertical and horizontal diffusion. Of these, the inclusion of stability-dependent vertical diffusion coefficients for heat, moisture and momentum is particularly important. The diffusion coefficients were considered to be sufficiently large in unstable conditions that dry convective instability would be removed by the vertical diffusion scheme. The separate dry convective adjustment scheme is therefore disabled in the troposphere. The moist convective adjustment scheme remains unchanged.

(b) Integrations

Table 1 lists the integrations used in this study. The control is a 1200-day perpetual January integration ('X223') at R15 spectral truncation (roughly equivalent to a resolution of 4.5°/7.5° latitude/longitude), run previously to assess the climatology of the new model. The first 510 days were later re-run by the authors in order to obtain additional diagnostics. In the first cloud radiative forcing experiment (CF1), the cloud longwave forcing of the atmosphere was removed over the whole globe. In the second (CF2), it was removed only in the tropics (±30° latitude). As noted in Table 1, these three integrations enable the relative role of tropical and extra-tropical clouds to be determined. Difference fields are presented as, for example, Control – CF1 in order to show the effect of the clouds in the control integration. The cloud radiative forcing integrations were run for 510 days. Results from all the integrations are shown for the mean of days 61 to 510. The five consecutive 90-day means in this period were used to calculate t-statistics to assess the statistical significance of the differences between the experiments. The number of degrees of freedom is 8 (5 samples from each experiment minus 2). Values of t larger than 3.36 indicate significance at better than the 1% level. In practice, it is prudent to consider a difference as significant only when t is somewhat larger than this value. A critical value of 4 is used here.

<table>
<thead>
<tr>
<th>TABLE 1. MODEL INTEGRATIONS</th>
</tr>
</thead>
<tbody>
<tr>
<td>1. Control: January control integration of CCM1 ('X223')</td>
</tr>
<tr>
<td>Cloud longwave atmospheric integration (CLAF) experiments</td>
</tr>
<tr>
<td>2. CF1: CLAF removed throughout the atmosphere</td>
</tr>
<tr>
<td>3. CF2: CLAF removed only in the tropics (±30°)</td>
</tr>
<tr>
<td>Difference fields shown</td>
</tr>
<tr>
<td>Control – CF1: Shows the effect of all clouds</td>
</tr>
<tr>
<td>Control – CF2: Shows the effect of tropical clouds</td>
</tr>
<tr>
<td>CF2 – CF1: Shows the effect of extra-tropical clouds</td>
</tr>
</tbody>
</table>

The cloud longwave forcing was removed by replacing the atmospheric longwave heating rates by their clear-sky values. The effect of clouds on the longwave flux at the surface was retained, however. These experiments thus concentrate on the direct atmospheric radiative forcing by clouds, with the effect on the surface longwave fluxes kept as close as possible to that in the control. This is not a concern over the oceans, as the model uses fixed s.s.t.s, but the model does respond to the surface energy budget over the land. Nevertheless, it is anticipated that the effect of the cloud longwave forcing over the land surface on the results shown here will be small, because over land this term is generally much smaller than that for the atmosphere, except at high latitudes (Fig. 4). A further reason for minimizing the importance of land surface processes in these experiments is that this version of the model does not include interactive soil moisture, the evaporation over land being a specified fraction of that from a saturated surface (Williamson et al. 1987).
4. Results

Many of the results in this and later sections are presented in the form of meridional cross-sections, the vertical coordinate being linear in sigma (\(\sigma\)). This is necessary because the physical tendencies can show large changes from one sigma level to the next (particularly at the top of the boundary layer) and in the interpolation to a pressure grid important detail can be misrepresented.

(a) Cloud cover and cloud forcing

Figure 5(a) shows the zonal-mean of the fractional cloud cover in the control experiment. At all latitudes there is a maximum at low levels, corresponding to boundary layer cloud. This cloud extends over two sigma layers, due to a constraint in the radiation scheme which prevents the cloud amount at \(\sigma = 0.926\) from exceeding that at \(\sigma = 0.811\). This was included in the original version of the model (CCM0) in order to prevent the prediction of unrealistically large amounts of cloud in the lower layer, due to cloud-induced longwave cooling which was not offset sufficiently by vertical turbulent fluxes. Above the boundary layer there is a minimum in the cloud cover, which is particularly well marked in the tropics. At high latitudes there are secondary maxima at about \(\sigma = 0.355\), while in the tropics this maximum is centred on \(\sigma = 0.165\).

The simulation shown in Fig. 5(a) is in broad agreement with climatologies derived from surface-based observations (e.g. London 1957; Warren et al. 1986), although the mid-tropospheric minimum in the tropics is somewhat more marked than in those data.

The derived cloud liquid water content is generally large in the lower troposphere, and decreases with height. The cloud longwave emissivity is therefore unity at low levels and decreases upwards, the lowest values being just below the tropopause (Fig. 5(b)). The effective cloud amount seen by the longwave part of the radiation scheme (Fig. 5(c)) is thus the same as that shown in Fig. 5(a) at low levels, but is smaller in the upper troposphere. The strong vertical gradient in the emissivity in the upper troposphere also acts to reduce the height of the upper cloud maxima, typically by one sigma layer.

From the discussion in section 2(c), this configuration would be expected to produce strong cooling at the top of the low cloud, with warming beneath, cooling associated with the high cloud in the extra-tropics but warming from the high cloud in the tropics. Such a pattern is indeed observed in the vertical profile of the zonal-mean cloud longwave forcing (Fig. 5(d)). This diagnostic was obtained by extending the definitions of cloud radiative forcing in section 2 (which are in terms of the fluxes at some level) to include the flux divergences (i.e. the heating rates). It was calculated by taking the difference between the longwave heating rates used by the model, which include the effects of the clouds, and the clear-sky values, which were archived in the re-run of the control experiment. Figure 5(d) thus shows the zonal means of the heating rate differences which, when vertically integrated, produce the map shown in Fig. 4(b). It is important to realize that the cloud forcing of the model atmosphere is by the heating rate differences summarized in Fig. 5(d) and that there is an infinite number of such profiles which are all consistent with Fig. 4(b). This has implications for both modelling and observational studies which are discussed further in section 7.

Note that the lower tropospheric forcing in Fig. 5(d) is very strong and that the warming induced by the high cloud in the tropics is distributed throughout the middle and upper troposphere, with a maximum at \(\sigma = 0.355\). Above the tropospheric warming the lower stratosphere is cooled, because the high clouds reduce the upward longwave fluxes which contribute to the radiative heating.
Figure 5. Zonal-mean cloud amounts and associated quantities from the control experiment. (a) Fractional cloud amount. The contour interval is 0.05 with values greater than 0.1 hatched. (b) Cloud longwave emissivity. The contour interval is 0.2. The values were set to zero in the lowest layer and the top three layers, where the model assumes that the cloud amounts are zero. (c) Effective cloud amount used in the longwave part of the radiation scheme. This is the product of the first two fields. The contour interval and hatching are as in (a). (d) Cloud longwave forcing. This is the difference between the total longwave heating rates and those for clear skies. The contour interval is 0.25 K/day, with negative values hatched.

Differences between the pattern shown in Fig. 5(d) and the actual changes to the model temperatures brought about by the forcing (given by the difference between the control and experiment CF1) will indicate where the model's physical parametrizations and large-scale dynamics have responded strongly to modify the effect of the perturbation. In the following two sub-sections, the temperature and the related wind changes are therefore examined first, followed by the various physical tendencies which contribute to the temperature changes.
Zonal means of the temperatures and zonal wind components from the control are shown in Fig. 6. The temperature distribution is similar to that from CCM0 and agrees reasonably well with observations. However, tropospheric temperatures in the tropics are still too low by several kelvins. The zonal-mean winds are of course related to the temperatures through the thermal wind equation, but some of the differences compared with CCM0 are easier to see in this field. For example, both subtropical jets are stronger and hence more realistic, particularly in the southern hemisphere, in comparison with recent climatologies (e.g. see Slingo and Pearson 1987, Fig. 5(a)).

![Figure 6. Zonal-mean (a) temperatures and (b) zonal wind components from the control experiment. The contour interval for temperature is 10 K with values lower than 200 K hatched. The contour interval for wind is 10 m s⁻¹ with negative values hatched.](image)

The differences between the zonal-mean temperatures and winds from the control and CFI are shown in Fig. 7, together with the associated t-statistics. As expected, cloud longwave forcing produces a warming of the tropical upper troposphere, which is highly significant. The peak warming is at $\sigma = 0.245$, somewhat higher than the level of the cloud-forcing maximum shown on Fig. 5(d). The lower stratosphere is cooled, as expected. The net effect is thus a significant destabilization of the upper troposphere by the cloud longwave forcing. There is a strong warming of the north polar stratosphere, but this is not statistically significant as the natural variability in this region in January is substantial. The expected cooling of the upper troposphere at high latitudes may also be seen, particularly over the south pole where the cloud cover exceeds 25%.

The zonal-mean wind differences (Fig. 7(b)) show a substantial and highly significant acceleration of both subtropical jets. It is shown in section 5 that in the northern hemisphere this acceleration is not uniformly distributed in longitude, but is particularly large over the Atlantic Ocean due to a strong normal mode response. As with the temperature differences, the deceleration in the north polar stratosphere is not significant.
Figure 7. Zonal-mean differences between Control and CF1 of (a) temperature and (b) zonal wind component. The corresponding $t$-statistics are shown in (c) and (d) respectively. The contour interval is 1 K in (a), 2 m s$^{-1}$ in (b), 10-0 in (c) (but also including the ±4-0 contours) and 4-0 in (d). Negative values are hatched. For clarity, some of the contours in the southern hemisphere maximum have been omitted from (d).

(c) Diabatic heating

Various zonal-mean diagnostics relating to the diabatic heating in the control and to the differences between the control and CF1 are shown in Figs. 8 to 10. The longwave heating rates shown in Fig. 8 are those used by the model and hence are for both clear and cloudy skies in the control, but for clear skies only in CF1. The difference field shown in Fig. 8(c) is thus the sum of the cloud longwave forcing shown in Fig. 5(d) and the difference between the clear-sky longwave heating rates, which is shown in Fig. 9(b). The latent heating (Figs. 10(a) and (b)) is the sum of that due to moist convective adjustment and stable (large-scale) condensation. The total diabatic heating (Figs. 10(c) and (d)) is the sum of the longwave and shortwave radiative heating, the latent heating, the heating by horizontal diffusion and the heating by vertical diffusion, which for the
lowest model layer includes a contribution from the surface sensible heat flux. The definitions are essentially the same as in the analysis of CCM0 by Boville (1985). However, in CCM0 the dry convective adjustment scheme contributed to the diagnosed convective heating, whereas in CCM1 this scheme is disabled in the troposphere and the corresponding physical process contributes to the vertical diffusion (see section 3(a)). The convective cooling of the lowest layer is hence much smaller in Fig. 10(a) than shown by Boville (his Fig. 6).

One notable feature of Fig. 7(a) is that the expected temperature decrease at the top of the boundary layer is weak. This is due to a strong compensation between radiative and convective processes. In the control, the total longwave cooling (Fig. 8(a)) is a maximum at $\sigma = 0.811$, at the top of the low cloud layer (Fig. 5). There is a subsidiary maximum in the tropics in the lowest layer, which is due to cooling by the water vapour continuum. In CF1 (Fig. 8(b)), removal of the cloud maximum enhances the cooling by the continuum, because of increased cooling to space in the effectively cloudless atmosphere. The cloud longwave forcing thus acts to generate an unstable lapse rate in the lowest model layers (Fig. 8(c)), and hence to the enhanced latent heating shown in

---

Figure 8. Zonal-mean longwave heating from (a) Control, (b) CF1 and (c) Control – CF1. The contour interval is 0.5 K/day, with negative values hatched in (c).
Fig. 10(b) (with comparable contributions from both convective heating and large-scale condensation). This heating generates temperature increments which are similar in magnitude to those from the cloud longwave forcing, but of the opposite sign. The net temperature change shown in Fig. 7(a) is therefore small.

The model's response to the cloud longwave forcing is complicated by the fact that there are changes in both the atmospheric temperatures and the humidity mixing ratios, and both of these alter the clear-sky longwave heating. The forcing produces an increase in the humidity mixing ratios in the middle troposphere in the tropics (Fig. 9(a)), a result of enhanced vertical transport of moisture by the convective adjustment scheme. Above $\sigma = 0.811$ the atmosphere is both warmed and moistened, increasing the clear-sky longwave cooling (Fig. 9(b)). In this region, the changes in the clear-sky cooling therefore oppose the cloud longwave forcing, the net effect of which is to lower the peak of the forcing (compare Fig. 5(d) with Fig. 8(c)).

In contrast, at $\sigma = 0.926$ there is a substantial increase in the clear-sky longwave heating, corresponding to a reduction of the cooling by water vapour in the control compared with CF1. This increases the convective destabilization of this layer compared with that above. The reduction in the cooling is mainly due to the increased humidity mixing ratios above this layer, which increase the downward longwave flux and hence decrease the cooling both in the lower layers and at the surface. This effect is greatest in the southern subtropics, where it is enhanced by smaller humidity mixing ratios in this layer (Fig. 9(a)).

A final example of the interaction between the cloud longwave forcing and the clear-sky longwave heating may be seen in the tropical stratosphere. In this region there are only small differences between the total longwave heating (Fig. 8(c)) and diabatic heating (Fig. 10(d)) in the two experiments. Comparison of Figs. 5(d) and 9(b) shows that the cooling induced by the cloud longwave forcing is almost completely balanced by a relative increase in the clear-sky heating. In the control, the lower temperatures induced by the cloud forcing lead to smaller clear-sky cooling rates, which appear on Fig. 9(b) as a relative warming. These compensate for the cloud longwave forcing and maintain the equilibrium at the lower temperature.

Figure 9. Zonal-mean differences between Control and CF1 of (a) humidity mixing ratio and (b) clear-sky longwave heating. The contour interval is $2 \times 10^{-4}$ in (a) and 0.2 K/d in (b), with negative values hatched.
For such long-term means, the total temperature tendency is close to zero, so the distributions of diabatic heating and of adiabatic (i.e. dynamical) heating are inverses of each other (e.g. Boville 1985, Fig. 2). In the control (Fig. 10(c)), for example, the strong diabatic forcing of the Hadley circulation may be seen as heating in the rising branch at about 10°S and cooling in the descending branch at about 15°N. These tendencies are balanced by adiabatic cooling and warming, respectively. The primary contributions to the diabatic forcing come from the latent heating (Fig. 10(a)) and the longwave cooling (Fig. 8(a)). Mid-latitude cells associated with the storm tracks may also be seen.

The difference between the diabatic heating in the control and CF1 is shown in Fig. 10(d). The pattern is very similar to that in the control with, for example, positive values in the rising branch of the Hadley cell and negative values to the north and south. The maxima of several tenths of a degree per day are significant compared with the values in the control. This indicates that the cloud longwave forcing acts to enhance the diabatic forcing of the meridional circulation, and in particular that it strengthens the Hadley circulation at low latitudes. The effects on tropical rainfall are examined in the following sub-section.

![Figure 10](image-url)  
Figure 10. Zonal-mean latent heating from (a) Control and (b) Control – CF1. Zonal-mean diabatic heating from (c) Control and (d) Control – CF1. The contour interval is 0.5 K/d in (a), (b) and (c) and 0.2 K/d in (d). Negative values are hatched.
(d) Convection and vertical velocity

The distribution of total precipitation (convective plus large-scale condensation) from the control experiment (Fig. 11(a)) is in reasonable agreement with observed estimates (e.g. Slingo and Pearson 1987, Fig. 7(a)). Note that the peaks in the cloud longwave forcing of the atmosphere (Fig. 4(b)) coincide with the precipitation maxima over South America, southern Africa and Indonesia. The difference precipitation field (Fig. 11(b)) shows that over both South America and Indonesia the cloud longwave forcing enhances the precipitation maxima and suppresses the precipitation in surrounding areas. The pattern is more complicated over southern Africa, although even here there is a net increase in the precipitation. The precipitation changes are large and significant; greater than 6 mm/d over both South America and Indonesia, with associated t values (not shown) of around 4 and larger negative values of t in the surrounding areas. An increase of 6 mm/d corresponds to a heating of about 180 W m\(^{-2}\), which is double the peak cloud longwave forcing of about 90 W m\(^{-2}\) over Indonesia. The precipitation differences thus make a substantial contribution to the change in the vertically-integrated diabatic heating (Fig. 11(c)). This demonstrates that the model’s response is not due simply to the cloud forcing acting in isolation, but to the combined effects of the forcing and the changes induced in the latent heating.

The relationship between the cloud forcing, the latent heat release and the dynamical response may be seen more clearly by concentrating on a particular region. The distributions of total precipitation and 500 mb vertical velocity for the Indonesian region are shown in Fig. 12. In the control, there are two maxima in the precipitation field which are associated with upward motion. In the upper troposphere there are significant amounts of high cloud, which give rise to the maxima in the distribution of the cloud longwave forcing of the atmosphere (Fig. 4(b)). In CF1, the distributions are very similar to those in the control, although the values of the maxima are smaller. The difference fields thus show positive precipitation maxima and negative vertical velocity maxima which are coincident with the peaks in the control experiment. The effect of the cloud longwave forcing is therefore to enhance the total precipitation and vertical velocity in areas where these are already large. This result is consistent with the diagnostics shown in the previous sub-section and demonstrates the strong influence of the forcing on tropical convection and dynamics.

5. Extra-tropical response

The simulation of the 200 mb wind field in the control experiment (Fig. 13(a)) compares favourably with observations (e.g. the 250 mb data presented in Fig. 4(a) of Slingo and Pearson (1987)). In the southern hemisphere the jet shows only weak longitudinal asymmetry, whereas this is very marked in the northern hemisphere, with jet maxima correctly positioned over the western boundaries of the Atlantic and Pacific Oceans. Figure 13(b) shows that removal of the cloud longwave forcing in CF1 weakens the jets, particularly in the region of the Atlantic maximum. The difference field in Fig. 13(c) shows that the acceleration of the jets by the cloud forcing shown in Fig. 7(b) is fairly uniformly distributed in longitude in the southern hemisphere. However, in the northern hemisphere the acceleration is located preferentially in the vicinity of the jet maxima. The effect is particularly large for the Atlantic jet, the magnitude of the perturbation representing a significant fraction (roughly one third) of the jet strength in the control. An anomalous circulation is produced, with easterlies both to the north and south (note the pattern of the arrows on Fig. 13(c)). The accelerations of the jet maxima are highly significant; values of t for the wind speed increases are 7.3 over the Atlantic and 8.7 over the Pacific.
Figure 11. Total precipitation (convective plus large scale) from (a) Control and (b) Control – CF1. The contours in (a) are at 1, 2, 5 and every 5 mm/d thereafter, with values greater than 5 mm/d hatched. The contour interval in (b) is 2 mm/d with negative values hatched. (c) Vertical integral of the diabatic heating for Control – CF1. The contour interval is 0-5 K/d with negative values hatched.
Figure 12. Total precipitation and 500 mb vertical velocity from Control, CF1 and Control – CF1, for the Indonesian region. The contour intervals are 3 mm/d in (a) and (c), 2 mm/d in (e), $30 \times 10^{-3}$ Pa s$^{-1}$ in (b) and (d) and $20 \times 10^{-3}$ Pa s$^{-1}$ in (f). Negative values are hatched.
d) CF1: 500mb vertical velocity in $10^{-3}$ Pa/s

e) Control-CF1: Total precipitation in mm/day

f) Control-CF1: 500mb vertical velocity in $10^{-3}$ Pa/s

Figure 12 (Continued).
Figure 13. 200 mb wind vectors and isotachs for (a) Control, (b) CF1 and (c) Control – CF1. The contour interval for the isotachs is 10 m s^{-1} in (a) and (b) and 5 m s^{-1} in (c). The length of the vectors is proportional to the wind speed. A vector whose length is 15^o of longitude (the horizontal spacing between vectors) represents 50 m s^{-1} in (a) and (b) and 20 m s^{-1} in (c).
The preferential acceleration of the Atlantic jet compared with that over the Pacific is associated with a strong and statistically significant anomaly pattern in the 200 mb geopotential height (Fig. 14). In the northern hemisphere, cloud forcing results in a deep negative anomaly centred over Newfoundland, with a smaller anomaly near the dateline. In the tropics, the warming of the troposphere shown in Fig. 7(a) increases the heights by on average about 100 m. This change is large compared with the variability of tropical temperatures on the 90-day timescale, hence the large values of \( t \) in Fig. 14(b). The juxtaposition of a height maximum over the Atlantic at about 25°N and the negative anomaly to the north gives rise to a strong height gradient at 40°N, which is another manifestation of the acceleration of the jet. The height changes are much smaller in the high latitudes of the southern hemisphere and are not statistically significant.

The height anomaly pattern in the northern hemisphere shows a marked barotropic structure. At both 500 mb and 850 mb the pattern in Fig. 14(a) is preserved, the major anomaly centres showing no significant changes in position. Such barotropic patterns have been generated in modelling studies of the atmospheric response to s.s.t. anomalies (e.g. Blackmon et al. 1983; Geisler et al. 1985; Palmer and Mansfield 1986a, b). In those

---

**Figure 14.** (a) 200 mb geopotential height difference between Control and CF1. The contour interval is 40 m. (b) The corresponding \( t \)-statistic. The contour interval is 10. Negative values are hatched in both plots.
studies, s.s.t. anomalies (equal to twice the values obtained by Rasmusson and Carpenter (1982) from compositing observations during the mature phases of several El Niño events) generated upper tropospheric height anomalies in the northern hemisphere with amplitudes up to about 200 m. The simulated patterns of maxima and minima were similar to those observed during the periods of s.s.t. anomalies. In particular, the Pacific/ North America (PNA) pattern identified by Wallace and Gutzler (1981) was reproduced. The driving mechanism for these patterns is believed to be changes in the distribution of diabatic heating over the tropical Pacific, associated with changes in the distribution of precipitation forced by the s.s.t. anomalies. It is generally assumed that the anomalous diabatic heating comes from variations in the latent heat release (i.e. the precipitation changes themselves). However, the results shown here demonstrate that cloud longwave forcing is equally capable of generating height anomalies of this magnitude, although as discussed earlier some of the change in the diabatic heating comes from changes induced in the precipitation field. Nevertheless, precipitation changes must be associated with changes in the distribution of cloudiness, so they must also produce changes in the cloud forcing. One may conclude that changes in cloud forcing are an inevitable and potentially important component of the changes in diabatic heating produced by s.s.t. anomalies. This possibility does not seem to have been widely recognized in previous studies.

In the experiments with the composite El Niño s.s.t. anomaly, height minima are generated over the North Pacific just east of the dateline and over the eastern seaboard of North America. The largest feature is a maximum over north-western Canada. Geisler et al. (1985) found that moving the s.s.t. anomaly towards the east produced little change in the positions of the extra-tropical features, although their amplitudes decreased. They interpreted this result as evidence for a preferred mode which was excited less strongly as the region of anomalous precipitation moved eastwards. In contrast, when Palmer and Mansfield (1986b) used an idealized s.s.t. anomaly over the extreme western Pacific, their extra-tropical height anomaly pattern shifted to the west. In both studies it was concluded that Indonesia is the most efficient region for the excitation of the extra-tropical response.

Comparison of the extra-tropical height anomaly pattern in Fig. 14(a) with those generated in the experiments referred to above reveals some interesting differences. The pattern in Fig. 14(a) is shifted to the west with respect to the results with the composite El Niño s.s.t. anomaly. If the pattern is forced primarily by the increased precipitation over Indonesia shown in Figs. 11 and 12, then this result is consistent with the shift found by Palmer and Mansfield when the extreme western Pacific s.s.t. anomaly was used. On the other hand, the strengths of the features in Fig. 14(a) differ markedly from those in any of the s.s.t. anomaly experiments. These differences may reflect the fact that the cloud longwave forcing shows a much more complicated pattern (Fig. 4(b)) than the s.s.t. anomaly fields, with maxima not only over Indonesia but also over South America and southern Africa. The complexity of the cloud forcing pattern is possibly also responsible for the lack of clear evidence in the tropics of pairs of geopotential height anomalies to the north and south of the centres of anomalous diabatic heating, which are very clear in the s.s.t. anomaly experiments (e.g. Palmer and Mansfield 1986a, Fig. 14). It is possible that the maxima on Fig. 14(a) over the North and South Atlantic represent such a pair forced by the anomalous diabatic heating over South America (Fig. 11(c)), so that this region may be responsible for reinforcing the acceleration of the North Atlantic jet. In future work it would therefore be interesting to examine the response to each of the cloud-forcing maxima separately, to determine the extent to which the patterns in the height fields may be explained by the superposition or interference of the responses to each of the maxima.
6. RELATIVE IMPORTANCE OF TROPICAL AND EXTRA-TROPICAL CLOUD LONGWAVE FORCING

(a) Zonal-mean temperatures and zonal wind components

In the previous section it was assumed that much of the model's response was due to the tropical cloud forcing. This hypothesis may be tested by comparing the results from the control and experiment CF2. As noted in Table 1, Control—CF2 gives the effect of just the tropical cloud forcing (i.e. that within ±30° of the equator). Figure 15 shows such difference fields for the zonal-mean temperatures and zonal wind components. Comparison with Fig. 7 confirms that most of the tropical response is indeed due to the local cloud forcing. The tropical temperature differences are practically identical to those in Fig. 7(a), except in the stratosphere (see below). The zonal wind changes are also similar, the acceleration of the subtropical jets being within about 1 m s⁻¹ in Figs. 7(b) and 15(b).

![Figure 15. As Fig. 7, but for Control – CF2 and omitting the t-statistics.](image)

Outside the tropics, there are differences between Figs. 7 and 15. These may be quantified by taking the difference between the two plots, which is equivalent to taking the difference between the results from experiments CF2 and CF1. Such differences show the effect of the extra-tropical cloud longwave forcing (Table 1) and are presented in Fig. 16, together with the t-statistics. Within the tropics, the only statistically significant temperature differences are those in the stratosphere, as noted above. In this region, the low temperatures lead to radiative relaxation times which are extremely large (e.g. Kiehl and Solomon (1986, Fig. 5) give values of about 80 days). As a result, the temperatures are extremely sensitive to small perturbations, so that even weak forcing from the extra-tropics is capable of inducing a significant temperature change.

The zonal wind differences are of marginal significance, the largest t value in this region on Fig. 16(d) being 3.91. In the extra-tropics, however, there are differences which appear to be significant. The cooling of the upper troposphere at high latitudes in Fig. 7(a) is primarily a result of the extra-tropical forcing by high cloud, as is part of the warming in the mid-troposphere at latitudes just poleward of 30°. There are also significant contributions to the cooling at σ = 0.811 by low cloud. While these temperature differences are associated with quite large values of t, those for the wind changes are smaller (Fig. 16(d)) and show marginal significance, except in the southern hemisphere.
Figure 16. As Fig. 7, but for CF2 – CF1. The contour interval is 5 in (c) and 4 in (d).

(b) 200 mb winds and heights

The 200 mb wind and geopotential height differences between the control and CF2 are shown in Fig. 17. All the features described earlier for the wind differences on Fig. 13(c) also appear on Fig. 17(a), although with somewhat reduced intensities. The preferential acceleration of the Atlantic jet is preserved and Fig. 17(b) shows that this is still the result of a marked height gradient between a maximum to the south and a minimum to the north. The extra-tropical mode evident in Fig. 14(a) may also be seen in Fig. 17(b), although the intensity is much reduced. The height changes in the tropics are similar in these figures, as would be expected from the zonal-mean temperature differences.
RESPONSE OF A GCM TO RADIATIVE FORCING. 1

a) Control–CF2 : 200mb wind and isotachs (m s\(^{-1}\))

![Image of wind and isotachs](image)

b) Control–CF2 : 200mb geopotential height (m)

![Image of geopotential height](image)

Figure 17. Differences between Control and CF2 at 200 mb of (a) winds and (b) geopotential heights. The plot characteristics are as in Figs. 13(c) and 14(a), respectively.

The 200 mb heights and \(t\)-statistics for the difference between experiments CF2 and CF1 (Fig. 18) show small values in the tropics, although there are significant changes over Indonesia and Australia. In the northern high latitudes, comparison of Fig. 18(a) with Fig. 17(b) shows that the mode in Fig. 14(a) is excited at least as strongly by extratropical cloud forcing as by tropical forcing. This is somewhat surprising, in view of the fact that similar modes appearing in s.s.t. anomaly experiments have been interpreted as being due to forcing from anomalies in the tropical diabatic heating. It suggests that the extra-tropical dynamics can be influenced by \textit{in situ} diabatic forcing, which if true would be an important conclusion. However, care must be taken in the interpretation of this result, as discussed below.

(c) \textit{General comments}

The results shown above indicate that extra-tropical cloud forcing makes a contribution to the local simulation which can be detected above the model’s natural variability. This conclusion is probably reliable where the differences can be simply understood in terms of the local forcing, for example the zonal-mean temperature
changes. The values of $r$ are obviously not as large as those obtained when the tropical forcing is included. One should therefore be cautious about the reliability of these conclusions, as these integrations are not long enough to give entirely stable statistics, particularly at high latitudes.

Unfortunately, it is not possible to determine unequivocally how much of the response is solely due to the extra-tropical cloud forcing, particularly for a quantity such as the 200 mb height change. The reason is that there are differences between the tropical precipitation distributions in CF1 and CF2 (Fig. 18(c)). These differences might be forced by the extra-tropics but they may also be due to the model’s natural variability in the tropics. It is thus not possible to determine how much of the height change shown in Fig. 18(a) is due to extra-tropical cloud forcing compared with that excited by the tropical precipitation differences.

7. Discussion

The results shown in this paper demonstrate that the CCM responds strongly to cloud longwave radiative forcing, confirming the sensitivity of the simulated general circulation to the longwave radiative properties of clouds, found in an earlier version of the model by Ramanathan et al. (1983). This forcing warms the upper troposphere in the tropics by about 4 K and also cools the lower stratosphere by over 6 K. It accelerates the subtropical jets, that in the northern hemisphere by over 7 m s$^{-1}$ in the zonal mean. The forcing also increases the precipitation maxima at low latitudes and strengthens the Hadley circulation. A strong barotropic height perturbation is excited in the northern hemisphere, with similarities to those generated by s.s.t. anomalies in the tropical west Pacific. The perturbation is strong over the Atlantic Ocean, where the jet speeds are accelerated by up to 19 m s$^{-1}$ by the forcing.

The experiments reported here are idealized, in that the cloud longwave forcing of the atmosphere was removed completely in order to determine its gross effect. However, this does not mean that the perturbations applied are extreme compared with those to be expected in reality. For example, at the peak of the 1982/83 El Niño/Southern Oscillation (ENSO) event in January 1983, the displacement of the tropical convection maximum from the west Pacific towards the east gave rise to OLR anomalies of +65 W m$^{-2}$ over northern Indonesia and −88 W m$^{-2}$ on the equator at about 155°E (Ardanuy and Kyle 1986, their Fig. 3(d)). These anomalies were of course primarily due to clouds and the fact that they are comparable with the cloud longwave forcing of the system shown in Fig. 2(c) demonstrates that perturbations similar in magnitude to those considered here can arise during ENSO events. Clearly, more work is needed to assess the role of cloud radiative forcing in the ENSO phenomenon, not only by the longwave forcing in the atmosphere but also by the shortwave forcing at the ocean surface.

It appears that cloud longwave radiative forcing is an important component of the atmospheric response to an applied perturbation, such as that provided by a sea surface temperature anomaly. Much of the response appears to be due to the forcing by high cloud in the tropics, as also found by Ramanathan et al. (1983). The response is complicated by strong interactions between the forcing and the latent heat released by convection and large-scale condensation. The cloud radiative forcing and the forcing by latent heat release are thus intimately related.

There is much evidence from both observational and theoretical studies that the radiative forcing by high cloud has an important influence on tropical convection and dynamics. Data for the Indonesian region obtained during the Winter Monsoon Experiment (WMONEX) show that extended sheets of upper-tropospheric cloud are a common
Figure 18.  (a) 200 mb geopotential height difference between CF2 and CF1. The contour interval is 40 m.  
(b) The corresponding t-statistic. The contour interval is 2.  (c) The precipitation difference. The contour interval is 2 mm/d. Negative values are hatched on all the plots.
feature (Webster and Stephens 1980). These cloud layers produce substantial perturbations to the local radiative heating profile, particularly in the longwave spectral region. Such perturbations are significant when compared with the diabatic heating associated with latent heat release. Houze (1982) showed that the extensive upper cloud layers are associated with the mature and later stages in the development of a tropical cloud cluster. Radiative and condensational heating within the cloud and evaporative cooling of precipitation at lower levels combine to modify the heating profile from the main updraught regions, with increased heating at upper levels. As a result, the vertical motion field on the scale of the cloud cluster, which in the mean balances the heating, is also modified. Hartmann et al. (1984) showed that when such a modified heating profile was used to force a linear model, it led to a marked improvement in the simulation of the east–west Walker circulation, compared with when a profile more typical of just the deep convective elements was used. The present results also demonstrate that the radiative forcing is capable of influencing the circulation on the largest scales, with appreciable effects on the flow fields at high latitudes.

Despite the supporting evidence from the studies cited above, there remains a concern that the cloud forcing in the model may still be unrealistic. In particular, the vertical profile of the cloud forcing is strongly influenced by the assumptions which are made regarding the thickness of the cloud. As an illustration, the radiation scheme in the Meteorological Office 11-layer model (Slingo and Wilderspin 1986) was used to compute the vertical profile of the cloud longwave forcing for three different cloud layers (Fig. 19). The clouds were placed in the tropical atmospheric profile used by Slingo and Wilderspin. The vertical profile of the cloud forcing is given by the difference between the atmospheric heating rates for complete cloud cover and those for clear skies. The

![Figure 19. Vertical profile of the cloud longwave forcing of the atmosphere (in K/d) for complete overcast by a cloud with its top at 125 mb and its base at three different levels.](image-url)
cloud top pressure is 125 mb, so that the cloud longwave forcing of the system is 192 W m\(^{-2}\) in each case. The clouds are six, two and one sigma layers thick, with bases at 790, 270 and 195 mb respectively. These are referred to as the thick, intermediate and thin clouds. The cloud longwave forcing of the surface is 32, 8 and 5 W m\(^{-2}\) respectively. For the intermediate cloud, there is strong cloud top cooling and cloud base warming, the values being similar to those obtained by Webster and Stephens (1980). There is a marked radiative destabilization of the cloud layer which has a pronounced influence on convective motions within the cloud, as noted by Webster and Stephens. Some fraction of the radiative divergence may thus be compensated by vertical motion and condensation within the cloud layer, so that the effect on the large-scale fields is reduced. Beneath the cloud there is a deep warming. The lower base of the thick cloud leads to a much reduced cloud base warming, but enhanced warming in the mid-troposphere compared with the other clouds. In contrast, the thin cloud does not resolve either the warming of the base or the cooling of the top, showing only the net warming of the cloud layer. This case appears unrealistic, but it should be remembered that most general circulation models (including the CCM) make the assumption that layer clouds are only one sigma layer thick, so this is the forcing profile injected into such models. The total atmospheric cloud forcing is in each case within 10 per cent of 175 W m\(^{-2}\), but the vertical distribution is radically different, with the maximum for the thick cloud at 125–195 mb being practically a mirror image of the forcing for the thin cloud. Since the induced vertical motion field may be expected to depend on the cloud forcing profile, this difference suggests that the model's response could be sensitive to the cloud thickness.

As noted in section 4(a) and confirmed by the results discussed above, there is an infinite number of vertical profiles of the cloud radiative forcing which are consistent with a given value for the system (at the top of the atmosphere) or integrated over the atmosphere itself. If, in reality, the atmospheric response is sensitive to the vertical profile of the forcing, it follows that agreement between the modelled distributions, such as those shown in Figs. 2(c) and 4(b), and the equivalent fields derived from satellite data is a necessary but not a sufficient constraint on the realism of a model. Estimates are therefore needed from observations of the vertical profile of atmospheric radiative heating in cloudy conditions, particularly in regions of deep tropical convection, although this may prove an extremely difficult requirement to achieve. For the present, further modelling studies are needed to determine the sensitivity to the cloud radiative forcing profile.

The response of the atmosphere to cloud longwave forcing is the result of interactions between the radiative heating by clouds, the heating by convection and the large-scale dynamics. In order to discover how robust the results obtained in this study are, it is thus necessary to determine their sensitivity to each of these components. The control and experiment CF1 were therefore repeated with a different cloud prediction scheme which, unlike that in CCM1, gives the fractional coverage by both layer and convective clouds as continuous functions of the model variables. The scheme was that implemented in the ECMWF operational medium range forecast model (J. M. Slingo 1987). Results from these experiments will be presented in a future paper. In addition, integrations are planned in which the standard convective adjustment scheme is to be replaced with a modified version of the penetrative convection scheme of Donner et al. (1982). Some of these experiments have also been repeated with the Meteorological Office model, which should provide further information on the generality of the results obtained with the CCM.
ACKNOWLEDGMENTS

We gratefully acknowledge R. Anthes, J. Coakley, R. Dickinson, V. Ramanathan and W. Washington for making our visit to NCAR possible and for their support of this work. We also thank Gloria Williamson for help in learning how to modify and run the CCM, Rick Wolski for much advice and practical help on the CCM Modular Processor and Bruce Briegleb for introducing us to the Interactive Data Analysis Processor (IDAP), which was used to generate most of the plots in this paper. Valuable comments on the manuscript were received from J. Coakley, R. Dickinson, J. Kiehl, P. R. Rowntree and W. Washington.

REFERENCES

Ardanuy, P. E and Kyle, H. L.
Blackmon, M. L., Geisler, J. E. and Pitcher, E. J.
Boville, B. A.
Briegleb, B. P., Minnis, P., Ramanathan, V. and Harrison, E.
Cess, R. D. and Potter, G. L.
Charlock, T. P. and Ramanathan, V.
Donner, L. J., Kuo, H.-L. and Pitcher, E. J.
ECMWF
ERBE Science Team
Feigelson, E. M., Kondratyev, K. Ya. and Prokofyev, M. A.
Geisler, J. E., Blackmon, M. L., Butes, G. T. and Muñoz, S.
Hartmann, D. L., Hendon, H. H. and Houze, R. A.
Hoskins, B. J. and Karoly, D. J.
Houze, R. A.
Hunt, B. G.
Hunt, G. E., Ramanathan, V. and Chervin, R. M.
Kiehl, J. T. and Solomon, S.

1985 The thermal balance of the NCAR Community Climate Model. ibid., 42, 695–709
1987 Exploratory studies of cloud radiative forcing with a general circulation model. Tellus, 39A, 460–473
1985 The albedo field and cloud radiative forcing produced by a general circulation model with internally generated cloud optics. J. Atmos. Sci., 42, 1408–1429
1982 The significance of thermodynamic forcing by cumulus convection in a general circulation model. ibid., 39, 2159–2181
1985 ‘Workshop on cloud cover parameterization in numerical models’. 26–28 November 1984. ECMWF, Shinfield Park, Reading
1984 Some implications of the mesoscale circulations in tropical cloud clusters for large-scale dynamics and climate. J. Atmos. Sci., 41, 113–121


Slingo, J. M. 1987 The development and verification of a cloud prediction scheme for the ECMWF model. *ibid.*, 113, 899–927


Wetherald, R. T. and Manabe, S. 1980 Cloud cover and climate sensitivity. *ibid.*, 37, 1485–1510