The response of a general circulation model to cloud longwave radiative forcing. II: Further studies

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**SUMMARY**

Following on from the first part of this study, the impact of cloud longwave forcing on the general circulation has been studied further using a series of 510-day, constant January integrations with the NCAR Community Climate Model. The sensitivity of the global response to the vertical profile of the forcing has been assessed by replacing the cloud prediction scheme by that used in the ECMWF model. The results confirm those reported in Part I and emphasize the influence of tropical cirrus clouds on the local thermal structure and on the strength of the subtropical jets. The impact on the model's hydrological cycle of the cloud longwave forcing associated with boundary-layer clouds is also shown to be important.

Several integrations are described which assess the relative importance of the three major tropical forcing maxima over Indonesia, South Africa and South America in determining both the local and remote responses. For each region, the perturbation to the upper tropospheric diabatic heating by the radiative effects of the cirrus clouds excites an anticyclonic vorticity pair, located near the longitude of the forcing and almost symmetric about the equator. The influence of these anticyclones on the upper tropospheric circulation is substantial. The results also indicate that a perturbation to the diabatic heating over South America, provided in this case by the cloud longwave forcing, may have an important effect on the Walker circulation and on the extra-tropical flow. The implications of these results for the problem of deforestation are discussed.

1. **INTRODUCTION**

Cloud radiative forcing (CRF) is defined as the effect of clouds on the radiation balance at the top of the atmosphere. It is obtained by differencing the total shortwave and longwave fluxes from estimates of them for clear skies. Clouds reflect much more solar radiation back to space than the clear atmosphere and thus cool the system in the shortwave (with the possible exception of clouds overlying a brighter snow surface). In contrast, since cloud-top temperatures are generally lower than those of the underlying atmosphere and surface, clouds reduce the outgoing longwave radiation and thereby enhance the greenhouse warming of the system. The Earth Radiation Budget Experiment (ERBE) data averaged over the globe (Ramanathan et al. 1989, Table 3) show that, in the annual mean, the shortwave cooling of about 48 W m\(^{-2}\) exceeds the longwave warming of about 31 W m\(^{-2}\), so that the net radiative effect of clouds on the present climate is a 17 W m\(^{-2}\) cooling.

As discussed in Part I (Slingo and Slingo 1988, hereafter referred to as SS), the relative contributions by clouds to the atmospheric and surface radiation budgets differ for the shortwave and longwave components of CRF. Through most of the shortwave spectrum, atmospheric and cloud absorption is weak so that the shortwave CRF is felt mainly as surface cooling. However, cloud absorption can be significant, particularly in the near-infrared portion of the spectrum, and probably drives the diurnal cycle often observed in marine stratocumulus (Turton and Nicholls 1987). In a recent study of the shortwave absorption by cirrus clouds, Ramaswamy and Ramanathan (1989) showed that this process could lead to a substantial warming of the upper troposphere of the NCAR Community Climate Model (CCM), dependent on the ice-water path of the clouds.

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The partitioning of the longwave CRF between the atmosphere and the surface is subtle, depending on the temperature of the cloud and thus on its height. Clouds warm the surface by enhancing the downward flux compared with clear skies, the effect being greatest for low clouds. In comparison, the time-mean atmospheric component may vary from a cooling of around 20 W m\(^{-2}\) in areas where low clouds dominate, to a warming in excess of 80 W m\(^{-2}\) for high tropical clouds. The distribution of this warming (or cooling) within the atmosphere depends on the vertical profile of the cloudiness. All clouds experience cloud-top cooling because their emission exceeds that incident from above, whilst cloud bases are heated by the warmer underlying atmosphere and surface. For high tropical clouds the warming at the base can be large enough to produce a net longwave heating of the cloud. The combination of cooling at the top and warming at the base generates convective instability and hence turbulence and entrainment. The degree to which a model can resolve the vertical radiative heating profile associated with the cloudiness may thus have important consequences for other aspects of the hydrological cycle (Randall et al. 1989).

SS used the CCM to study the impact of the cloud longwave atmospheric forcing (hereafter referred to as CLAF) on the simulated general circulation for fixed January conditions. They showed that CLAF interacted with the latent-heat release, enhancing the convective activity and influencing both the tropical and extra-tropical circulations. The importance of the forcing has since been supported by results from the UCLA/GLA general circulation model (Randall et al. 1989). The vertical profile of CLAF in their model was substantially different from that in the CCM. Nevertheless, Randall et al. (1989) concluded that their results were generally consistent with SS and that the effects of CLAF were primarily due to the anvil cirrus clouds.

Whilst the results of SS show that CLAF is important, the paper also raises questions about the generality of the results and the sensitivity to the vertical profile of CLAF. The present paper describes further experiments with the CCM in which the model's cloud-prediction scheme is replaced. The change in the cloud parametrization gives a different vertical cloud distribution, thus modifying the profile of CLAF, particularly in the upper troposphere and in the vicinity of the boundary layer. In addition to the longwave forcing studied here, the shortwave cloud absorption may also be important in some circumstances, particularly that by cirrus as demonstrated by Ramaswamy and Ramanathan (1989). The present paper may thus be seen as a parallel investigation to that reported by Ramaswamy and Ramanathan.

The main areas of CLAF occur in the tropics and are associated with minima in the outgoing longwave radiation (Fig. 1). These are coincident with the three major convection zones over South America, South Africa and Indonesia in January. This complexity in the cloud-forcing pattern may account for differences in the extra-tropical response found by SS, when compared with that in sea-surface-temperature (SST) anomaly experiments. In this paper, a series of experiments will be described in which the roles of the individual cloud-forcing maxima are studied separately, to determine how much of the extra-tropical behaviour can be explained by superposition or interference of the response to an individual forcing maximum.

In the tropical atmosphere, the east–west circulations are closely related to the diabatic heating pattern (e.g. Webster 1983). For example, the rising part of the Walker cell occurs over the convectively active region of Indonesia and the west Pacific. The descending branch is observed over the east Pacific and helps to suppress any convection in that area. Its sensitivity to the distribution of diabatic heating is evident in an El Niño/Southern Oscillation (ENSO) event, when there is disruption of the Walker circulation. In this paper the remote impact of the individual CLAF maxima on the diabatic heating
Figure 1. Cloud longwave forcing for January. (a) ERBE observations 1986, (b) CCM1 result with the new cloud prediction scheme, and (c) atmospheric component of the forcing from CCM1 with the new cloud prediction scheme. The contour interval is 20 W m$^{-2}$. Negative values are stippled and values greater than 40 W m$^{-2}$ are hatched.
of other regions of the tropics, through their impact on the east–west circulation patterns, will also be considered.

Observational and modelling studies have identified Indonesia and the west Pacific as important regions for exciting an extra-tropical response, typically a Pacific/North American (PNA) pattern in the height anomalies (e.g. Horel and Wallace 1981; Blackmon et al. 1983). An SST anomaly in this region drives an atmospheric diabatic heating anomaly, associated with changes in both the latent and radiative heating. These studies have concentrated on the impact of perturbations over the oceans, whilst the present study may also be seen as an investigation into the sensitivity of the general circulation to anomalies in the diabatic heating over the major tropical continents. Such perturbations may arise through changes in surface characteristics as a result of man’s activities—the massive deforestation of the Amazon Basin being an example. The resulting changes in surface albedo and the hydrological cycle may force perturbations in the atmospheric diabatic heating, part of which will be the cloud longwave forcing. It is possible that such perturbations may also influence the extra-tropical circulation. Whilst changes in CLAF are only a small part of the complete perturbation, the experiments in the present paper should also identify the importance of the atmospheric component of the CRF in any study of tropical deforestation.

2. DESCRIPTION OF THE EXPERIMENTS

All experiments were run with version 1 of the CCM (CCM1) (Williamson et al. 1987; Hack et al. 1989). The integrations described in SS used the cloud scheme developed by Ramanathan et al. (1983), in which the cloud amount was set to 0.95 whenever largescale condensation occurred and to 0.3 whenever moist adjustment took place; otherwise the cloud cover was set to zero. In addition, an important constraint was imposed on boundary-layer clouds such that these clouds must always fill two model layers. This was intended to prevent an unrealistic, positive feedback between cloud amount and radiative cooling in the absence of adequate vertical turbulent fluxes.

In the experiments described in this paper, the cloud-prediction scheme was replaced by a continuous, fractional cloud-cover parametrization, developed for the ECMWF model (Slango 1987). This method allows for a basic cloud geometry of three layer clouds and one convective tower. The cloud cover for the convective tower is derived from the precipitation rate calculated by the model’s convection scheme, in this case the moist adjustment method. The fractional cloudiness for the layer clouds is based on the model’s relative humidity. Upper tropospheric clouds have an additional dependence on deep convective activity, whilst boundary-layer clouds are also related to the model’s thermal structure and vertical velocity. The scheme was tuned for CCM1 to give a radiation budget in reasonable agreement with ERBE data. This involved modifying the threshold relative humidities for the formation of layer cloud and adjusting the dependence of convective cloudiness on the model’s precipitation rate. Further details of the implementation of the scheme in the CCM are given in Slango and Slango (1991).

In addition to changing the cloud parametrization, the calculation of the cloud longwave emissivity was also modified. The standard version of CCM1 uses a variable (i.e. non-black) cirrus emissivity which depends on the cloud liquid-water content, calculated from the condensed moisture in the previous hour of model time (Ramanathan et al. 1983). This parametrization tends to produce rather low values of emissivity for the cirrus clouds associated with deep tropical convection (Kiehl and Ramanathan 1990). The results of various studies have indicated that these clouds have emissivities closer to
unity (Ackerman et al. 1988). Whilst a dependence of emissivity on the predicted cloud liquid-water or ice-water path is clearly desirable, it can only be achieved if, firstly, the model’s moisture distribution is realistic and, secondly, the formation and dissipation of the clouds are reasonably represented. Kiehl and Ramanathan (1990) have concluded that the moisture distribution is deficient in CCM1, leading to the low values of emissivity for the tropical cirrus clouds. The realistic prediction of cloud liquid-water and ice-water paths also depends on sophisticated cloud microphysics, which has been incorporated in other models (Smith 1990) but not as yet in the CCM. Similarly, the dependence of cloud radiative properties on liquid-water and ice-water paths and on droplet and crystal sizes (and shapes) is still uncertain, although this should improve with the results from field experiments such as FIRE (Starr 1987; Albrecht et al. 1988).

Bearing in mind the above difficulties, the treatment of cloud longwave emissivity was simplified with the new scheme by assuming that all clouds including and above model level $\sigma = 0.355$ ($\sigma$ is the ratio of the pressure at this level to its surface value) have an emissivity of 0.75; below that level it was taken as unity. This value was chosen to give reasonable agreement between the model’s simulation of the outgoing longwave radiation and that observed by ERBE. Figure 1(a) shows the cloud longwave forcing for January 1986 observed by ERBE. (Note the areas of missing data, particularly over the cloudy regions of the tropics and the southern hemisphere depression belt. These arise because ERBE was unable to find any clear pixels during the month.) The model’s cloud longwave forcing (Fig. 1(b)) is in very good agreement with ERBE, confirming that the choice of cirrus emissivity was well-founded. For the model, the portion of the forcing which is felt by the atmosphere, i.e. CLAF, is shown in Fig. 1(c) and can be compared with Fig. 4(b) in SS. The improved simulation of the outgoing longwave radiation has given slightly greater warming by CLAF in the tropical convection regions and in the storm tracks of the northern hemisphere extra-tropics.

All integrations were run with rhombooidal truncation at 15 wave numbers (roughly equivalent to a resolution of 4.5°/7.5° latitude/longitude). The model was integrated for 510 days using perpetual January conditions and the results from the last 450 days were averaged. In the ‘control’ integration (named thus because differences from it show the longwave radiative effect of clouds) the atmospheric longwave heating rates were replaced by their clear-sky values. In the perturbation experiments, CLAF was applied in various locations by using the cloudy longwave heating rates in the atmosphere. As in SS, the surface radiative fluxes use the cloudy values in all the experiments. Table 1 lists the experiments; the extent of the applied forcing in each case is shown in Fig. 2 and can be related to the maxima in CLAF evident in Fig. 1(c).

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<th>TABLE 1. MODEL INTEGRATIONS</th>
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<td>1. ‘Control’: No CLAF globally</td>
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<td>2. CLAF applied everywhere</td>
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<td>3. CLAF applied only over Indonesia</td>
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Differencing experiment 2 from the ‘control’ demonstrates the global impact of CLAF. The results are directly comparable with those reported in SS and show the impact of a different vertical profile of CLAF. The results from experiments 3, 4 and 5 will be used to assess the relative importance of the three major maxima in CLAF in the tropics. Their influence on the tropical and extra-tropical circulation will be studied using a stream-function analysis. The results will be used to consider how much of the total impact of CLAF can be explained by a linear combination of the individual responses.
Experiment 6 was run to elucidate the results of experiment 5 and to examine further the role played by South America in the global circulation. The remote influence of the individual forcing maxima on the other major tropical convection zones, through their effect on the east–west circulations, will be studied using the velocity potential field.

3. Results

(a) Impact of CLAF globally

Figures 3(a) and 3(b) show the zonal mean of the fractional cloud amount for the standard version of CCM1 (as used in SS) and for the integration with the new cloud scheme (experiment 2), respectively. With the new scheme, the constraint that the boundary-layer clouds should occupy two model layers was removed. The effect of this can be seen in the vertical distribution of the clouds in the lower troposphere. In the tropics, there is evidence of the deep clouds associated with the ITCZ and the major convection zones over Indonesia, South America and South Africa. In consequence, the mid tropospheric minimum in cloud amount obtained with the standard method in SS is not so marked.

The vertical profile of cloud longwave forcing (Figs. 3(c) and 3(d)), implied by these cloud distributions, was obtained by taking the difference between the longwave heating rates used by the model, which include the effects of clouds, and the clear-sky values. The strong cooling at the top of the low cloud, with warming beneath, can be seen. Both versions of the model show the cooling associated with the high clouds of the extratropics and polar regions. However, the upper tropospheric warming by the tropical cirrus clouds is much more marked with the new cloud parametrization, as is, to a lesser extent, the implied stratospheric cooling above these clouds. This result is primarily due to the change in the emissivity of these clouds, rather than the cloud amount. As noted by Ramaswamy and Ramanathan (1989) and Slingo and Slingo (1991), the standard version of CCM1 is too cold by at least 6 K in the upper troposphere. Changes to the radiative heating of the cirrus clouds, whether it be shortwave or longwave, substantially improves the cold bias near the tropopause. A similar effect is reported in Slingo and Slingo (1991) with a different emissivity calculation.
Figure 3. Zonal-mean cloud amounts for (a) standard CCM1 and (b) new cloud scheme. The contour interval is 0.05. Values greater than 0.1 are hatched. Zonal-mean cloud longwave forcing for (c) standard CCM1 and (d) new clouds. The contour interval is 0.25 K day⁻¹. Negative values are hatched.
RESPONSE OF A GCM TO RADIATIVE FORCING. II

Despite differences in the cloud distributions, both horizontally and vertically, the impact of CLAF globally was very similar to that obtained with the original version of CCM1. The main results can be summarized as follows.

(1) The upper troposphere is warmed in the tropics and cooled at high latitudes.
(2) The subtropical jets show substantial acceleration consistent with the increased equator–pole temperature gradient.
(3) Tropical precipitation maxima are increased coincident with the areas of maximum CLAF, particularly over the continents.
(4) Extra-tropical height anomalies are excited, resembling the wave trains, such as the PNA pattern, commonly seen with tropical SST anomalies.

The differences between the zonal-mean temperatures and winds from the control and experiment 2 are shown in Fig. 4 and can be compared with Fig. 7 in SS for the standard CCM1. There is a substantial warming of the upper tropical troposphere by CLAF, consistent with the more radiatively active cirrus clouds in the new cloud scheme. For the same reason, the associated stratospheric cooling is more marked, suggesting that the radiative properties of high-level clouds may have important implications for stratospheric thermal structure. Figure 4 shows that the enhanced equator-to-pole tropospheric temperature gradient leads to a substantial acceleration of the subtropical jets. Unlike the results of SS, CLAF by the new cloud scheme gives rise to a marked increase in the upper tropospheric tropical easterlies. Indeed, the simulation of these easterlies is much improved with this version of the model.

As in SS, the main areas of warming due to CLAF produce an in situ increase in diabatic heating, which approaches 2 K day\(^{-1}\) when integrated over the depth of the atmosphere (Fig. 5(a)). The partitioning of the diabatic heating anomaly into its latent and radiative components (Figs. 5(b) and 5(c)) shows that, whilst the latent heating clearly determines the pattern of the anomaly field, more than one third of the diabatic heating anomaly can be attributed to the radiative warming by the upper-level clouds associated with the convective systems.

Despite the stabilization of the upper tropical troposphere by CLAF (Figs. 3(d) and 4(a)), the results given in SS, and those reported here, show a positive response by the precipitation. SS suggested that the deep radiative warming by CLAF, throughout the troposphere, was compensated by adiabatic cooling from additional ascent, this ascent giving enhanced convergence of moisture in the lower troposphere and thus increased convective activity. The change in vertical velocity around the tropical belt, averaged between 10\(^\circ\)N and 30\(^\circ\)S, shows that this is also true in this study (Fig. 6).

An interesting feature of these results is the different response by the convection to CLAF over land and ocean. Figure 5(b) shows that the increase in latent heating occurs mainly over the continents, whereas over the west Pacific where CLAF is strongly positive, there is no significant increase, in contrast to the results given in SS (their Fig. 11). As already discussed, an enhancement of the latent heating can be explained by increased ascent and hence additional moisture convergence. However, the moisture source for convection is also derived from the surface evaporation, which can be substantial over the warm waters of the tropical oceans. Figure 7 shows the change in the surface evaporation when CLAF is applied globally for both the standard CCM1, as used in SS, and for the version with the new clouds. For the standard CCM1, the change in surface evaporation is generally small (Fig. 7(a)), apart from some coastal areas which are affected by on- or off-shore winds. However, in this study the effect of CLAF on the surface evaporation is substantial (Fig. 7(b)), particularly over the west Pacific, and probably accounts for the small response by the precipitation in that region.
Figure 4. Zonal-mean differences generated by introducing CLAF globally. (a) temperature and (b) zonal wind component. The contour interval is 1 K in (a) and 2 m s⁻¹ in (b). Negative values are hatched.
Figure 5. Differences generated by introducing CLAF globally for the vertically integrated (a) diabatic heating, (b) latent heating and (c) radiative heating. The contour interval is 0.5 K day$^{-1}$ in (a) and (b) and 0.25 K day$^{-1}$ in (c). Negative values are stippled. Note that 1 K day$^{-1}$ averaged over the depth of the atmosphere is equivalent to a flux convergence of approximately 119 W m$^{-2}$, or a precipitation rate of about 4.1 mm day$^{-1}$. 
The impact of CLAF on the surface evaporation in this study is an interesting result which merits an explanation. In Fig. 7(b) two regimes are clearly evident, with decreases over much of the tropical and subtropical oceans, but substantial increases over the cold waters of the eastern subtropics where stratocumulus occurs and is well predicted by the new cloud scheme (Slingo and Slingo 1991). In the model, the surface evaporation \( (E) \) is computed using:

\[
E = D_{w} \rho_{h} C_{H} |V|(q_{s}(T_{s}) - q_{h})
\]

(1)

where the subscript \( h \) denotes the first model level above the surface, \( q_{s}(T_{s}) \) is the saturation specific humidity at surface temperature, \( T_{s} \), and \( |V| \) is the magnitude of the near-surface wind velocity. \( D_{w} \) is a wetness factor and \( C_{H} \) is a stability-dependent drag coefficient. In addition, the vertical diffusion uses stability-dependent diffusion coefficients.

There is little indication that the surface winds have changed substantially to account for the changes in surface evaporation seen in Fig. 7(b). However, the structure of the boundary layer has been modified by the CLAF associated with the low clouds, thus influencing both the near-surface humidities and the vertical diffusion of moisture. Except over the cold waters of the eastern subtropics, the boundary layer has a well-mixed structure, typical of a convective regime. When low clouds are present, the radiative cooling generates and/or strengthens the boundary-layer inversion so inhibiting the transport of moisture out of the boundary layer into the free atmosphere. This leads to a moistening of the boundary layer, an increase in the near-surface humidity and thus to a reduction in the surface evaporation. This process is marked in this version of the model because there is no parametrization of the vertical transport by shallow cumulus, the effects of which can be substantial (Tiedtke et al. 1988).

Over the cold sea surface near the eastern boundaries of the subtropical oceans, the effect of CLAF associated with low clouds is substantially different. The clear-sky
boundary layer is stable over the cold waters so that when low clouds are introduced, the radiative cooling associated with them destabilizes the boundary layer which tends more towards a well-mixed structure. This increases the vertical transfers of heat and moisture, in part associated with the stability dependence of the vertical diffusion coefficients. The lowest layer thus becomes cooler and dryer, leading to enhanced evaporation. In the model, the radiative cooling of the cloud layer is balanced by condensational heating, rather than being partially offset by cloud-top entrainment (Chen and Cotton 1987). This process, not currently included in the CCM, would also act to dry the boundary layer and thus enhance the evaporation. Further discussion on the impact of stratocumulus on the surface evaporation can be found in Slingo and Slingo (1991).

In the standard CCM1, used in SS, the impact of the low clouds was quite different, as is evident from the change in surface evaporation (Fig. 7(a)). As noted in section 2, the vertical structure of the low clouds was constrained so that the clouds filled two
model layers, with the consequence that the cloud-top cooling occurred within the free atmosphere above the boundary layer, rather than within the boundary layer itself. In addition, because the cloud filled two model layers, the cloud-base warming was resolved (compare Fig. 3(c) with 3(d)). This warming occurred in the boundary layer and thus the low clouds had a substantially different effect on the boundary-layer structure than in the case with the new cloud scheme.

The position of the cloud-top cooling in relation to the boundary layer has important ramifications for the response by other aspects of the model’s physics, in addition to the impact on the surface evaporation already described. In the standard CCM1, the cloud radiative cooling was compensated by moist convective adjustment, which has the property of transferring moisture vertically as well as heat. The radiative effects of the low clouds thus served to move moisture out of the boundary layer into the free atmosphere, making more moisture available for the major convective regions. Thus the radiative forcing of the low clouds in the standard CCM1 had a positive impact on the latent-heat release, accounting for some of the local response to CLAF seen in SS. In contrast, the new cloud scheme correctly places the cloud-top cooling within the boundary layer where it is balanced by the heating associated with large-scale condensation. In this case, the cloud-top cooling does not force a vertical transfer of moisture into the free atmosphere, with the consequence that there is a drying of the tropical troposphere above the boundary layer and a reduction in the convective latent-heat release with this version of the model (Slingo and Slingo 1991). The CLAF associated with the boundary-layer clouds thus acts as a negative feedback with respect to the deep convection, whilst the opposite was true with the version used in SS.

It is clear from the above discussion that the radiative effects of the low clouds can have a substantial impact on the hydrological cycle of the model. Over the tropical continents, there is a negligible change in the surface evaporation so that the response by the vertical motion, to the additional warming by CLAF, dominates and there is a positive response by the convection (Fig. 5(b)). Over Indonesia and the west Pacific, however, the reduction in evaporation, associated with the radiative effects of the low clouds, is substantial, and, despite the enhanced ascent seen in Fig. 6, there is a little evidence of an increase in latent-heat release. These results suggest a subtle balance between the effects of CLAF associated with clouds in the free troposphere and those in the boundary layer. This balance may change when further refinements are added to the model physics, such as parametrizations of shallow cumulus and cloud-top entrainment.

It seems likely that a change in the horizontal distribution of diabatic heating (Fig. 5(a)) would influence the model’s simulation of the tropical east–west circulations, particularly the Walker circulation. Figures 8(a) and 8(b) show the velocity potential field at 200 mb for the ‘control’ integration and for experiment 2, respectively, and Fig. 8(c) the difference produced by global CLAF. (The velocity potential is a scalar field whose gradient equals the velocity vector of the irrotational flow; thus convergent flow has a maximum, positive value at its centre, and divergent flow a minimum value. The divergence field, implied by the velocity potential, applies only to the largest, planetary scales.) Figure 8 demonstrates that CLAF has increased the divergence aloft for Indonesia and South America; this implies enhanced ascent in these regions, consistent with the results shown in Fig. 6. However, the compensating subsidence, indicated by the regions of convergent flow in Fig. 8, is shifted from predominantly over North Africa and the Caribbean in the ‘control’ with no CLAF (Fig. 8(a)), to being shared with the east Pacific in experiment 2 with global CLAF (Fig. 8(b)). This shows that the strength of the Walker circulation, implied by the maximum in the velocity potential difference, shown in Fig. 8(c), is sensitive to the effects of CLAF.
Figure 8. Velocity potential field at 200 mb for integrations with (a) no CLAF and (b) global CLAF. (c) is the change in velocity potential produced by global CLAF, i.e. (b) minus (a). The contour interval is $1 \times 10^5 \text{m}^2\text{s}^{-1}$. Negative values are stippled.
It might be expected that the weaker perturbation to the diabatic heating over Indonesia, compared with South America (Fig. 5(a)), would produce a lesser change in the velocity potential in that region than is suggested by Fig. 8(c). In this respect, the use of single-level fields is somewhat misleading, since the height of the minima in velocity potential perturbation varies geographically. Detailed examination of the model-level fields revealed that the enhanced divergence occurred predominantly at $\sigma = 0.245$ over Indonesia. Over South America, however, the perturbation to the velocity potential field reached a minimum of $-4.1 \times 10^6 \text{m}^2\text{s}^{-1}$ at $\sigma = 0.165$ implying maximum additional divergence at that level. Such variation in the height of the main outflow is consistent with the varying strength of the response by the convection, evident in Fig. 5(b).

An important result of SS was the excitation of a height anomaly pattern in the extra-tropics of the northern hemisphere. Whilst the response had some of the features of the well-known PNA pattern (Wallace and Blackmon 1983), there were differences which might be related to the complex forcing field provided by global CLAF, compared with the single diabatic heating anomaly typical of El Niño. However, there were similarities in the magnitude of the response to that obtained with a modest west Pacific SST anomaly (Palmer and Mansfield 1986b), which led SS to conclude that CLAF may be as important as latent-heat release in determining the atmospheric response to such SST anomalies. In the analysis presented here, the use of height anomaly fields is replaced with a stream-function analysis. There are two distinct advantages to this approach. Firstly, the stream-function anomalies describe differences in the wind field, which in the extra-tropics will closely follow the height anomalies because of geostrophic balance. Secondly, in the tropics where geostrophy is weak, substantial wind perturbations can be associated with relatively small height perturbations.

Figure 9(a) shows the difference in the stream function at 200 mb for the experiments with and without CLAF reported in SS. As expected, the stream-function anomalies closely resemble the height anomalies in the extra-tropics (see Fig. 14(a) in SS). However, in the tropics there is much more structure to the stream-function anomalies which can be related to changes in the wind field evident in Fig. 13(c) of SS.

The difference in the stream function generated by CLAF with the new cloud scheme is shown in Fig. 9(b). The Student’s $t$-test values for the equivalent height anomalies are shown in Fig. 9(c). (The $t$-test was performed on height because standard deviations were only kept for that field. Bearing in mind the above comments on the use of stream function, there is likely to be good correspondence in the extra-tropics.) There is evidence of a wave-train response in the northern hemisphere, with a ridge over Europe and a trough upstream over the eastern seaboard of North America. The trough over the north Pacific, just west of the dateline, is produced, although the ridge over north-western Canada is not as well defined as in SS. The differences in the detail of the response are not surprising, given the change in the climatology of the model with the new cloud distribution. There is preferential acceleration of the Atlantic and Pacific jets, the core speed of the Atlantic jet increasing by $17 \text{m s}^{-1}$ compared with $19 \text{m s}^{-1}$ in SS. Both the ridge over Europe and the trough over the north-west Pacific are significant, based on the $t$-test results.

Figure 9(b) shows clear evidence of twin anticyclones to the north and south of the major tropical diabatic heating anomalies, possibly associated with the excitation of an internal Rossby mode as demonstrated in the theoretical studies of Gill (1980). However, his results, using a linear model, only show a symmetric response when the anomalous heating is placed on the equator. Also the model results are not indicative of equatorial trapping, in contrast to Gill’s theoretical model, since the anomalous subtropical westerlies are as strong as the equatorial easterlies. Similar anticyclonic dipoles have been
Figure 9. Stream-function anomalies at 200 mb produced by global CLAF with (a) standard CCM1 and (b) new clouds. The contour interval is $5 \times 10^5$ m$^2$ s$^{-1}$. Negative values are stippled. (c) is the $t$-statistic for the equivalent height anomalies at 200 mb, produced by global CLAF in the model with new clouds. The contours are placed at 0, ±2, 4, 10, 20 and every 20 thereafter. Values ≥4 and ≤−4 are stippled. Absolute values of $t$ ≥ 4 are considered significant at the 95% level.
Figure 10. Perturbation to the diabatic heating around the tropical belt produced by global CLAF, averaged between 10°N and 30°S. The latent heating is shown in (a) for standard CCM1 and (b) for new clouds. The longwave radiative heating is shown in (c) for standard CCM1 and (d) for new clouds. The contour interval is 0.5 K day⁻¹ in (a) and (b) and 0.25 K day⁻¹ in (c) and (d). Negative values are stippled.

observed during the 1982/83 El Niño (Sardeshmukh and Hoskins 1985). As in this study, they formed symmetrically about the equator despite the asymmetry of the anomalous heating. This, and other aspects of the observed response to the El Niño, led Sardeshmukh and Hoskins (1985) to stress the importance of nonlinearity for understanding the local upper tropospheric response.

Figure 9(b) suggests that the strength of the anticyclones is directly related to the strength of the diabatic heating perturbations (Fig. 5(a)). It is interesting to note,
however, that despite similar diabatic heating anomalies in SS (their Fig. 11), the development of these anticyclones is much less marked (Fig. 9(a)). This suggests that they are sensitive to the vertical profile of the diabatic heating anomaly and that the magnitude of the upper tropospheric heating may be crucially important. Vertical cross-sections around the tropical belt (averaged between 10°N and 30°S) of the heating anomalies due to latent-heat release (both convection and large-scale condensation) and longwave radiation, associated with the effects of CLAF, are shown in Fig. 10 for both
the standard CCM1 and for the version with the new clouds. The impact of the differing vertical structure of the low clouds, already discussed, is immediately apparent. The perturbation to the latent heating is dominated by the low clouds in the standard CCM1 (Fig. 10(a)), with relatively small changes in the middle and upper troposphere. With the new clouds (Fig. 10(b)), the additional latent heating over the major convection areas occurs at, or below, 500 mb, consistent with the behaviour of moist adjustment schemes which tend to place the heating too low in the troposphere (Donner et al. 1982; Tiedtke 1984). There is no indication from the results shown in Figs. 10(a) and 10(b) that the perturbation to the latent heating could be responsible for the differing upper tropospheric circulation anomalies.

The longwave radiative heating, however, does display some important differences (Figs. 10(c) and 10(d)). Apart from the lower tropospheric changes, which are largely compensated by condensation processes (Figs. 10(a) and 10(b)), the more radiatively active cirrus clouds of the new cloud scheme have produced substantial perturbations to the heating in the upper troposphere (Fig. 10(d)). It would seem reasonable to conclude that the anticyclonic dipoles are excited, primarily, by the upper-level heating associated with the cirrus clouds, again emphasizing the importance of a realistic treatment of these clouds. However, the role of the latent heating is clearly not negligible since the strongest anticyclones are not related to the strongest cirrus warming. As Fig. 9(b) shows, South America produces the greatest circulation anomaly associated with the largest enhancement of precipitation (Figs. 5(b) and 10(b)). The result, that the vertical profile of the diabatic heating may be crucially important for the development of these circulation anomalies, is consistent with the conclusion reached by Sardeshmukh and Hoskins (1985).

(b) Relative roles of the major tropical CLAF maxima

This section describes the results of experiments 3–5 in which the CLAF associated with the major diabatic heating maxima over Indonesia, South Africa and South America is applied independently for each region. It should be emphasised that CLAF is only a perturbation on the already substantial diabatic heating of these regions, associated with latent-heat release and the clear-sky radiative cooling. The impact in the tropics of the individual areas of CLAF will be discussed in the context of their local and remote influences on the east–west large-scale circulation, as depicted by the perturbation to the upper-level velocity potential (Fig. 11), and hence on the precipitation distribution. Figure 12(a) shows the precipitation distribution between 40°N and 40°S from the ‘control’ integration, and Figs. 12(b)–(d) the difference in the precipitation generated by applying CLAF over the three major diabatic heating maxima. So that these changes can be interpreted in the context of the total impact of CLAF everywhere, Fig. 12(e) shows the precipitation difference produced by global CLAF.

When CLAF is applied over Indonesia there is a strengthening of the velocity potential minimum compared with the ‘control’ integration (Fig. 11(a)), implying enhanced ascent in that region. There is no apparent strengthening of the Walker circulation, the compensating subsidence occurring predominantly over the northern Indian Ocean. Palmer and Mansfield (1986b) noted a similar response in their study of the impact of a warm west Pacific SST anomaly. Their results showed a marked increase in upper-level divergence above the anomaly, but with the compensating convergence occurring over the Indian Ocean rather than over the central and east Pacific. Figure 12(b) shows little evidence of a local enhancement of the precipitation over Indonesia, but rather a southwards shift in the area of maximum convection, consistent with the results shown in Fig. 12(e) for global CLAF. A southwards displacement in the precipitation is also evident over South America and to some extent over the Indian Ocean.
Figure 11. Perturbation to the velocity potential at 200 mb produced by CLAF only over (a) Indonesia, (b) South Africa and (c) South America. The contour interval is $1 \times 10^6$ m$^2$s$^{-1}$. Negative values are stippled.
Figure 12. Total precipitation distribution and anomalies between 40°N and 40°S. (a) Precipitation distribution for the integration with no CLAF. The contours are at 1, 2, 5 and every 5 mm day$^{-1}$ thereafter. Values greater than 5 mm day$^{-1}$ are stippled. Precipitation anomalies produced by CLAF over (b) Indonesia, (c) South Africa, (d) South America and (e) the whole globe. The contour interval is 1 mm day$^{-1}$. Negative values are stippled.
Indeed, the suppression of convection over the northern Indian Ocean, evident in Fig. 12(e), seems to be primarily forced by the introduction of CLAF over Indonesia. Over South Africa, however, there is a moderate increase in precipitation (in excess of 3 mm day$^{-1}$) which may be related to circulation changes over the adjacent Indian Ocean, implied by the shift in the convection in that region already noted.

The South African monsoon provides the smallest of the three major heating maxima and consequently its impact on the velocity potential field is relatively minor (Fig. 11(b)). The addition of CLAF again leads to a southwards displacement of the precipitation maximum although there is also a slight increase (Fig. 12(c)). This result suggests that the positive anomaly in the precipitation over South Africa, seen in Fig. 12(e) with global CLAF, is associated with an in situ response to the local CLAF as well as the remote effects of the CLAF over Indonesia.

There is a marked response to the imposition of CLAF over South America (Fig. 11(c)). There is a dramatic increase in divergence, and hence ascent, associated not only with the additional radiative warming by CLAF, but also with a substantial increase in convective precipitation which exceeds 7 mm day$^{-1}$ (Fig. 12(d)). This rainfall enhancement is somewhat greater than the 5 mm day$^{-1}$ seen in Fig. 12(e) with global CLAF, and suggests some influence from the remote effects of the other main areas of CLAF. The compensating subsidence, implied by areas of upper-level convergence, is increased over the east Pacific. This suggests that the strength of the subsiding branch of the Walker circulation may be sensitive to the diabatic heating over adjacent South America. This point will be discussed further in section 3(c), where the results of the additional experiment 6 will be used to substantiate the specific role played by South America.

The results shown here support the expectation that a perturbation of the diabatic heating in one location of the tropics will influence the diabatic heating in another remote location, through an induced change in the longitudinal circulation. The main impact is on the latent heating by convection, which will be sensitive to the environmental vertical motion, although a change in convection naturally implies a change in the CLAF as well.

The impact of the three major tropical heating maxima on the extra-tropical circulation is analysed using the stream function at 200 mb, as before. Figure 13 shows the difference in the stream function when experiments 3–5 are compared with the 'control' integration with no CLAF. The stream-function anomalies produced by CLAF over Indonesia (Fig. 13(a)) demonstrate the excitation of the anticyclonic dipoles to the north and south of the area of forcing. This result is consistent with that obtained by Palmer and Mansfield (1986b; their Fig. 16(c)) for a warm west Pacific SST anomaly of moderate strength. However, unlike their results, there is no evidence of a substantial extra-tropical response and certainly no acceleration of the North Atlantic jet. One possible explanation is the smaller (by about a factor of two) implied diabatic heating anomaly in this study compared with the precipitation anomaly described in their paper. Similarly, the earlier study reported in SS shows a greater enhancement of the precipitation over Indonesia, with a greater diabatic heating anomaly, hence their conclusion that Indonesia may have been instrumental in exciting the extra-tropical response. While it is clear from Fig. 12(b) that there is little enhancement of the precipitation over Indonesia, the excitation of the anticyclonic stream-function anomalies can again be attributed to the upper tropospheric radiative heating anomaly associated with the cirrus clouds (see section 3(a)).

When CLAF is applied only over South Africa, the stream-function anomalies also show anticyclonic dipoles to the north and south of the forcing (Fig. 13(b)). Otherwise, the impact of South African CLAF on the extra-tropical circulation appears to be quite small. Again, the diabatic heating anomaly is relatively minor, compared with that
Figure 13. Stream-function anomalies at 200 mb produced by CLAF only over (a) Indonesia, (b) South Africa and (c) South America. The contour interval is $2.5 \times 10^4 \text{m}^2\text{s}^{-1}$. Negative values are stippled.
commonly seen in SST anomaly studies, particularly since there is no substantial in situ enhancement of the precipitation, as discussed above.

In experiment 5, with CLAF imposed over South America, the combination of the increased radiative warming (Fig. 1(c)) and the enhancement in the local precipitation (Fig. 12(d)) leads to a substantial positive diabatic heating anomaly over South America. Consequently, the stream-function anomaly field (Fig. 13(c)) shows pronounced anticyclonic flow to the north and south of the forcing. There is an increase in the easterly component of the zonal flow over South America, reducing the equatorward extent of the subtropical westerlies in that region. There is also a weak wave train propagating north-eastwards into the extra-tropics with its origins over Indonesia.

The above results show a clear association between the individual areas of CLAF and the upper tropospheric response in the tropics and subtropics, characterized by an anticyclonic vorticity pair that is located near the longitude of the forcing, and is almost symmetric about the equator. Although the strength of the subtropical anticyclonic dipoles has a substantial impact on the mid-latitude flow, there is little evidence from Figs. 13(a)–(c) of a preferred location for exciting an extra-tropical anomaly pattern. Indeed, none of the perturbations to the extra-tropical flow, shown in Fig. 13, were statistically significant in terms of Student's t-test.

Figure 14 shows the anomaly pattern obtained when the individual anomalies from the three forcing maxima are added. This can be compared with the same field for global CLAF (Fig. 9(b)). In many respects, for the tropics and subtropics, the stream-function anomalies generated by global CLAF (Fig. 9(b)) are the result of a linear combination

![Stream-function anomalies at 200 mb produced by a linear combination of the anomalies shown in Figs. 13 (a)–(c). The contour interval is 5 x 10^6 m^2 s^{-1}. Negative values are stippled.](image)

of the anticyclonic dipoles produced by the three main areas of CLAF. Figure 14 shows a similar extra-tropical response to that seen in Fig. 9(b), due to the juxtaposition of the weak areas of anomalous cyclonic flow over western Canada, the North Atlantic and the north-west Pacific, evident in Figs. 13(a)–(c). Thus some part of the extra-tropical response can be directly attributable to the linear combination of the three main areas of tropical CLAF. It is interesting to note that, whilst the extra-tropical responses to the individual areas of forcing were not statistically significant, when the three areas are imposed together the resultant circulation anomalies are significant (Figs. 9(b) and 9(c)).
Specific role of South America

It is clear from the previous section that a perturbation to the diabatic heating over South America can have important consequences for the tropical and subtropical circulation. The results of the additional experiment 6 are used to investigate further the importance of South America by considering the impact of CLAF against a more realistic basic state with CLAF included at all other locations. Figure 15 shows the velocity potential at 200 mb for experiment 6; it can be compared with Fig. 8(b) from experiment 2 with global CLAF. The specific role played by South America in determining the strength of the Walker circulation is evident, with an increase in the subsidence over north Africa at the expense of the east Pacific, confirming the earlier result that changes in the diabatic heating over South America can have a substantial impact on the subsiding branch of the Walker circulation. The change in precipitation when CLAF is included over South America (experiment 2 minus experiment 6) shows an increase in the local precipitation rate by 5 mm day\(^{-1}\) (Fig. 16), smaller than the corresponding increase of 7 mm day\(^{-1}\) obtained in experiment 5 when CLAF was introduced only over South America (Fig. 12(d)), but identical to the enhancement obtained with global CLAF (Fig. 12(e)).
If the stream function for experiment 6, with no CLAF over South America, is considered as the basic state, then the change in the stream function produced by experiment 2, with global CLAF, will show the impact of introducing CLAF over South America. This should thus be comparable with the stream-function perturbation produced when South American CLAF is applied to a basic state with no CLAF (Fig. 13(c)). Figure 17(a) shows the stream-function anomalies at 200 mb produced by experiment 2 versus experiment 6, and again the development of the twin anticyclones to the north and south of the forcing are clearly identifiable. However, there are substantial differences in the extra-tropical response, compared with Fig. 13(c), particularly over the North Atlantic and Newfoundland. This suggests that the basic state of the model, to which the perturbation is applied, is important in determining the response of the general circulation, particularly in the extra-tropics. Such a conclusion was also reached by Palmer and Mansfield (1986a) when they considered the response to SST anomalies in models which differed in their treatment of orography. The significance of the extra-tropical
response can be seen in Fig. 17(b) which shows the \( t \)-test values for the equivalent height anomaly field. In contrast to the results from experiment 5, where the extra-tropical response was barely significant (values of \( t \) less than 3), experiment 6 shows the development of significant barotropic cyclonic circulations in the mid latitudes and anticyclonic behaviour in the polar regions of the northern hemisphere. The effects of these on the North Atlantic jet are considerable (Fig. 18), with a strengthening of the jet core by 7\( \text{m s}^{-1} \) and a movement of the main jet southwards towards north Africa.

![Diagram](image)

Figure 18. 200 mb wind vectors and isotachs for (a) CLAF applied everywhere except South America and (b) CLAF applied everywhere. The contour interval for the isotachs is 10\( \text{m s}^{-1} \). The length of the vectors is proportional to the wind speed. A vector whose length is 7.5° of longitude represents 50\( \text{m s}^{-1} \).

4. DISCUSSION

Both versions of the model show that inclusion of global CLAF excites a response in the northern hemisphere extra-tropics (Fig. 9). However, the main areas of CLAF lie in the equatorial latitudes (Fig. 1(c)), predominantly in the region of upper tropospheric easterlies. Thus, from theoretical considerations, it is unlikely that they could excite a mid-latitude response. However, as Hoskins (1986) notes, whilst the diabatic heating anomaly itself may not lie in a regime favourable to the excitation of a stationary wave response in middle latitudes, there may be a substantial perturbation to the divergence field away from the forcing which could lie in a region of westerly upper tropospheric
flow. That this may be so is endorsed in the study by Sardeshmukh and Hoskins (1988) which shows that the advection of vorticity by the divergent component of the flow is an important term in the large-scale vorticity balance, and may lead to a Rossby wave source in the subtropical westerlies remote from the heating anomaly which lies in the tropical easterlies. It has not been possible to compute the distribution of the vorticity source terms from the integrations described in this paper, but it would certainly be of interest to do so if the role of the cirrus radiative heating is to be understood more completely.

Whilst anomalies in diabatic heating over the tropical Pacific Ocean have received a good deal of study, particularly because of the global impact of El Niño, the importance of the diabatic heating over South America has received relatively little attention. Yet deforestation provides perturbations to the diabatic heating, through a reduction in precipitation and associated CLAF, which may be similar in magnitude to those forced by tropical SST anomalies. Over the past few years there has been increasing concern that the deforestation of the Amazon Basin may have serious consequences for the world’s climate. Apart from acting as sinks for carbon dioxide, the tropical rain forests represent important components of the hydrological cycle. Indeed, a substantial part of the local rainfall is recycled by canopy interception and re-evaporation. A recent modelling study (Lean and Warrilow 1989) has shown that deforestation leads to a reduction in the local precipitation rate of 2 mm day\(^{-1}\) averaged over South America (north of 30°S), with peak values of 5 mm day\(^{-1}\) which are equivalent to a diabatic heating anomaly of the order of 150 W m\(^{-2}\), neglecting any changes associated with cloud longwave forcing (Rowntree, private communication). A similar model response to deforestation has been reported by Nobre et al. (1989).

The results of the integrations described in this paper have implications for the effects of deforestation. Firstly because a change in the cloud longwave forcing would be an inherent part of the induced change in precipitation, and secondly because the diabatic heating anomaly of about 220 W m\(^{-2}\), generated in this study, is similar in magnitude to that obtained in the deforestation experiments noted above. In addition, there appears to be a positive feedback between CLAF and precipitation such that a decrease in the cloud forcing would lead to a further reduction in the rainfall.

The sensitivity of the subsiding branch of the Walker circulation to the diabatic heating over South America suggests that the east–west circulation is not entirely dominated by heating over Indonesia. The reduction of diabatic heating as a result of deforestation may thus have implications for the climate of the tropical east Pacific, particularly the suppression of convection in that region. The extra-tropical response obtained when experiment 2 was compared with experiment 6 (Figs. 17 and 18) also emphasizes the importance of South America in the general circulation. The possibility that diabatic heating anomalies over this region might influence mid-latitude weather has been raised by Hoskins and Sardeshmukh (1987) in their study of the 1985/86 northern hemisphere winter. They concluded that unusual diabatic forcing over the South American/Caribbean region may have been the catalyst for the development of a European/Atlantic block in the first half of February 1986. A recent study by Branković et al. (1990) of the same blocking event also demonstrated that it might be sensitive to the diabatic forcing over South America.

5. CONCLUSIONS

The results described in this paper endorse the conclusion drawn by SS that CLAF plays an important role both locally and remotely in the general circulation. Both studies
emphasize the substantial impact of the cirrus clouds. In determining the structure of the Walker circulation, the importance of the upper tropospheric radiative warming of the anvil cirrus clouds of a mature tropical cloud cluster has also been demonstrated by Hartmann et al. (1984), using a linear steady-state model. In addition, they concluded that the elevated heating of the mature cluster might force different mid-latitude circulation anomalies. As in SS, this study has only considered the longwave cloud forcing within the atmosphere, since the shortwave cloud forcing is felt primarily at the earth’s surface and therefore can only be studied meaningfully with a model with an interactive ocean. Nevertheless, the results of Ramaswamy and Ramanathan (1989) have shown that the shortwave effects of cirrus may give an upper tropospheric radiative heating of as much as 0.5 K day\(^{-1}\). This would serve to accentuate the warming described in this study and would be likely to affect the circulation in a similar manner.

The role of the CLAF associated with the low clouds is an important result of this study. The sensitivity of the surface evaporation to the treatment of these clouds is substantial. It is clear that the interaction between these clouds and the hydrological cycle is uncertain and warrants further investigation. There can be little doubt that the impact of these clouds would vary considerably if the effects of shallow cumulus and cloud-top entrainment were included in the model.

The sensitivity of the model to both the upper and lower tropospheric cloudiness provides ample demonstration that the model’s general circulation is sensitive to the magnitude and vertical distribution of CLAF. Thus it would seem imperative that any model used in studies related to climate change should be capable of producing the global distribution of cloud forcing and its temporal variations with reasonable fidelity. There is also a clear need for more information on the vertical structure of the cloud forcing and on the radiative properties of ice clouds, since the generation of the upper tropospheric anticyclonic anomalies seems primarily related to the upper tropospheric diabatic heating.

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