Sensitivity of a simulated tropical squall line to long-wave radiation

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SUMMARY

A two-dimensional cloud-resolving model is used to simulate a tropical squall line and to examine its sensitivity to long-wave radiation. Model results show that the magnitude of the sensitivity to long-wave radiation depend on the organization and structure of the squall line. For one simulation, the inclusion of long-wave radiation increased surface precipitation by nearly a third. Long-wave radiation is also found to enhance the mid-level rear inflow and upper-level outflow of the squall line. Sensitivity experiments are performed to examine the role of cloud-top cooling and cloud-base warming, differential cooling between clear and cloudy regions, and domain-wide cooling.

KEYWORDS: Cloud-resolving model Deep convection Precipitation enhancement Radiative cooling

1. INTRODUCTION

Observations indicate the importance of radiative effects on convective systems. Gray and Jacobson (1977), for example, using island station data, found an early morning maximum in deep convection in the tropics. Using satellite data, Janowick et al. (1994) found that very cold cloud tops over the tropical oceans occurred most frequently between 0300 and 0600 local time.

Historically, modellers of convective clouds have ignored radiative effects. However, recent investigations indicate that radiative effects can significantly modulate the behaviour of convective systems. A number of modelling studies show that the inclusion of radiative effects significantly enhances precipitation. Dudia (1989) used a two-dimensional mesoscale model to simulate the diurnal convection over the South China Sea. He found that, when radiative heating was excluded, the convection failed to grow to the same extent and shallower weaker clouds developed. The net rainfall was only 36% of that in a simulation which included radiative heating. Tao et al. (1993) used a two-dimensional cloud-resolving model to simulate squall lines. They found that long-wave (LW) radiative cooling increased surface precipitation by 31% for a tropical squall line and by 14% for a mid-latitude squall line. Miller and Frank (1993) used a mesoscale model to study the radiative forcing of a simulated tropical cloud cluster. They found that in simulations which included radiative heating, rainfall rates peaked near midnight and that a diurnal cycle was established. This diurnal cycle was the result of the day/night difference in radiative heating. It is important that numerical weather prediction models and general circulation models represent this significant daily oscillation in tropical rainfall correctly.

Attention has focused on three mechanisms for the enhancement of precipitation by LW radiation. Gray and Jacobson (1977) proposed that the differential cooling between clear and cloudy regions generates a secondary circulation which enhances convective activity. Within the cloud interior there is little LW cooling; thus, the cold air surrounding the cloud experiences significantly greater cooling than the cloudy air. This leads to subsidence in the clear air which then leads to convergence into the cloud region. Webster and Stephens (1980) suggested that cloud-top (base) cooling (warming) destabilizes stratiform clouds and leads to convective overturning, which then enhances precipitation. The third mechanism ignores cloud–radiation interactions and suggests that the domain-wide

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cooling by LW radiation is important. Although this cooling extends to the surface, it enhances the surface fluxes which means that the equivalent potential temperature $\theta_e$ reduces more slowly in the boundary layer than in mid-troposphere. Therefore, this tends to destabilize the domain and increase the convective activity. Previous modelling studies (Dudhia 1989; Miller and Frank 1993) suggested that it is the domain-wide cooling mechanism which is of primary importance. However, these studies did not examine all three mechanisms simultaneously. In addition, these studies used mesoscale models which had a coarser spatial resolution than our cloud-resolving model and therefore may not have resolved the cloud–radiation interactions effectively. For example, Miller and Frank (1993) found that the radiative-heating rates calculated at cloud base and cloud top in their model were less than half those suggested by Webster and Stephens (1980). This was probably because of the lower vertical resolution of the mesoscale model. The importance of the vertical resolution of the model can be understood from the following argument. For clouds with high emissivity, at cloud top the long-wave radiative cooling is localized near the cloud surface and decays rapidly in the bulk of the cloud. However, the cooling rate calculated by the radiation scheme is an average for a model grid-box. Thus, this average includes the large cooling rates near the cloud-top surface and the much smaller cooling rates in the bulk of the cloud. Therefore, the deeper a model layer the greater the contribution by the bulk of the cloud to the average cooling rate, and hence the smaller the calculated cooling rate. A similar argument applies to cloud-base warming.

Modelling studies show that radiative effects can also modify the structure of convective systems. Chin (1994) used a two-dimensional cloud-model to simulate a mid-latitude squall line. He found that including LW radiative cooling enhanced the upper-level outflow and mid-level rear inflow. The simulation including LW radiation also produced a more realistic transition zone between the convective and stratiform regions. Following Tripoli and Cotton (1989), Chin (1994) suggested that the greatest impact of long-wave radiation was in the stratiform region of his simulated squall-line.

The present paper describes a simulation of a tropical squall-line using a cloud-resolving model and the results of experiments to determine the system’s sensitivity to long-wave radiation. Observations from EMEX9 (Equatorial Mesoscale Experiment; aircraft mission 9) are used to initialize the model. The EMEX9 system developed over the Arafura Sea between Australia and New Guinea and is chosen for the simulation because the synoptic conditions and observations of this system have already been described by Webster and Houze (1991). In addition, this system has been simulated by Tao et al. (1993) and Wong et al. (1993) which permits some comparison of results. This is particularly relevant given the advocacy by the Global Energy and Water Cycle (GEWEX) Cloud System Study (GCSS) for cloud-resolving models to be used to develop parametrizations for large-scale models (Browning 1994). The approach suggested by GCSS includes the use of cloud-resolving models as sources of synthetic data for use in developing the parametrizations used within large-scale models; it also includes intercomparisons of cloud-resolving models (and validation against observations) to improve the parametrizations used within the cloud-resolving models themselves. A detailed intercomparison of our cloud-resolving model and the Goddard Cumulus Ensemble (GCE) model used by Tao et al. (1993) is beyond the scope of the present paper. However, our results show that the squall-line simulations are sensitive to some aspects of model formulation, and that not all the results of Tao et al. (1993) are reproducible with our cloud-resolving model.

Section 2 briefly describes the cloud model, and section 3 describes the long-wave radiation scheme used within it. Section 4 describes the simulation, including initial conditions. Section 5 describes the results of sensitivity experiments and section 6 discusses the results and offers some conclusions.
2. Model

The model used for this study is the two-dimensional version of the Meteorological Office Cartesian large-eddy-simulation model, described by Shutts and Gray (1994) and in more detail by Derbyshire et al. (1994). This model is formulated using the deep anelastic quasi-Boussinesq equation set which is a generalization of the incompressible Boussinesq equation set. The model is thus non-hydrostatic and excludes sound waves. The model has a positivity-preserving advection scheme (Leonard 1991), a first order, stability-dependent, Smagorinsky–Lilly turbulence parametrization (Mason 1994), and parametrized microphysics. The turbulence parametrization uses a modified gradient Richardson number to account for moist processes, following MacVean and Mason (1990). The cloud microphysics scheme includes a Kessler-type parametrization for liquid water and a three-category ice-phase scheme (cloud ice, snow, graupel). The conversion rates between categories are based on the bulk-water parametrization of Lin et al. (1983). The size spectrum for each precipitation type is assumed to be an inverse exponential. Details of this scheme are given by Swann (1993).

The model uses a stretched vertical grid with 31 levels and the model top at 20 km. Height increments vary from about 400 m near the surface to 1500 m at the top. The model has 1000 grid points in the horizontal with a constant resolution of 750 m. Periodic boundary-conditions are used for the lateral boundaries. The lower surface uses a free-slip boundary-condition. Surface fluxes of heat and moisture are included using bulk aerodynamic equations. A 5 km deep absorbing layer is used at the top of the model. The intention was to repeat the simulation of Tao et al. (1993) using resolution, surface boundary-conditions and forcing which mirrored theirs as closely as possible. The most obvious difference between the two models is that Tao et al. (1993) used open lateral boundary-conditions.

3. Long-wave radiation parametrization

The long-wave parametrization scheme implemented within the cloud model is based on the scheme used by the Meteorological Office Unified Model as described by Ingrams (1993) and Slingo and Wilderspin (1986). This parametrization scheme solves the radiative transfer problem for each model column in terms of a vertically upward and a vertically downward long-wave flux. Such an approach is, strictly, only valid for a plane parallel atmosphere. For gaseous absorption, variations in the vertical are much larger than variations in the horizontal and thus the 'plane parallel' approximation will give good results. However, clouds can show significant horizontal variation and it is not clear how valid the plane parallel approximation remains in this case. Even stratiform clouds such as stratocumulus are not truly horizontally homogeneous. For example, the tops of stratocumulus clouds are undulating and this is likely to affect their radiative properties. Unfortunately radiative schemes which take into account the horizontally finite extent of clouds are computationally very expensive.

The long-wave radiation scheme includes absorption by water vapour, carbon dioxide, ozone, cloud water, cloud ice, rain, snow and graupel. Only the effects of absorption and emission are included and scattering is ignored. For gaseous absorption, a look-up table is used to obtain transmissivities from absorber path-length. This path length is scaled to include the effects of collision- and Doppler-broadening of the absorption lines. A model grid-box is assumed either to be completely filled by cloud or to contain no cloud (partial cloud cover within a grid box is not allowed). This approximation is reasonable given the relatively high resolution of the model. For cloud water, the emissivity
is calculated from an equation derived by Stephens (1984). Only the cloud-water path (cloud-water content \( \times \) layer width) is required. Thus the emissivity is given by

\[
\varepsilon_c = 1 - \exp\left(-\kappa_c \, CW \, P\right),
\]

where \( CW \, P \) is the cloud-water path and \( \kappa_c = 130 \, m^2/\text{kg}^{-1} \). The above equation is derived by assuming that the absorption efficiency \( Q_{\text{abs}} \) varies linearly with cloud-droplet radius, which is valid when the droplet radius is smaller than the wavelength of the radiation \( \lambda \). This approximation is reasonable since clouds have their greatest impact on LW radiative transfer in the window region (\( \lambda \approx 10 \, \mu m \)).

It is inappropriate to treat the radiative properties of rain water in the same way as cloud water since rain drops are much bigger than 10 \( \mu m \). Using (1) for rain would significantly overestimate the effect of rain on LW radiation. For rain, it is more appropriate to use the large-particle approximation that \( Q_{\text{abs}} = 1 \). In this case, the emissivity of rain is given by

\[
\varepsilon_r = 1 - \exp\left(-\kappa_r \, RW \, P\right),
\]

where

\[
\kappa_r = \frac{3}{4} \frac{D}{\rho_w r_{e,r}}.
\]

\( D (= 1.66) \) is the diffusivity factor, since we are dealing with broad-band irradiance, \( RW \, P \) is the rain-water path, \( \rho_w \) is the density of water and \( r_{e,r} \) is the effective radius of the raindrops. The effective radius \( r_e \) is defined to be proportional to the ratio of average volume (\( < V > \)) to average cross-sectional area (\( < A > \)):

\[
r_e = \frac{3}{4} \frac{< V >}{< A >}.
\]

The \( 3/4 \) in (4) ensures that, for a population of monodispersed spheres, the effective radius is equal to the radius of the spheres. For rain drops which are assumed to be spherical

\[
r_{e,r} = \frac{\int_0^\infty n_r(r)r^3 \, dr}{\int_0^\infty n_r(r)r^2 \, dr}
\]

where \( n_r(r) \) is the size spectra of rain drops defined such that \( n_r(r) \, dr \) is the number of rain drops per unit volume in the radius interval \( (r, r + \, dr) \). The \( 1/r_{e,r} \) dependence in (2) and (3) arises from the importance of the cross-sectional area of the rain drops in radiative transfer.

The effective radius of the drops in rain can be derived by using the empirical relationship of Marshall and Palmer (1948)

\[
n_r(r) = N_{0,r} \exp\left[-2\Lambda_r r\right].
\]

\( N_{0,r} \) is assumed to have a fixed value of \( 8 \times 10^6 \, m^{-4} \) and \( \Lambda_r \) is a function of rain water content \( w_r \):

\[
\Lambda_r = \left(\frac{\pi \rho_w N_{0,r}}{w_r}\right)^{1/4}.
\]

Thus, putting (6) and (7) into (5) and integrating by parts

\[
r_{e,r} = \frac{3}{2\Lambda_r}.
\]
In our simulations, we find that values for \( r_{e,1} \) range from about 100 \( \mu \)m to 1 mm.

The ice-microphysics scheme includes three different species. Ice crystals are created by nucleation and are assumed to be monodispersed hexagonal plates or columns. Snow particles are larger and made up of clusters or aggregates of ice crystals. These are assumed to have an inverse exponential size-distribution like (6) but with different parameters \( (N_0, \lambda) \). Graupel particles are ice particles that have become heavily rimed; again they are assumed to have an inverse exponential size-distribution. Often, for radiative transfer, the considerable size-differences between the three ice species is ignored and only the total ice-water (ice crystals+snow+graupel) is used. However, because of their small size and hence large surface-to-volume ratio, ice crystals have a disproportionately large effect on radiative transfer. By considering only the total ice-water, useful information is wasted.

Since ice crystals are generally larger than 10 \( \mu \)m, equations similar to (2) and (3) should be used to calculate the emissivity. Because of the lack of data on ice-crystal size distribution in the anvils of tropical cumulonimbus clouds, we set \( k_1 = 65 \) m\(^2\)kg\(^{-1}\). This value is used by the Unified Model and corresponds to an effective radius for ice crystals of about 20 \( \mu \)m. It is also consistent with the work of Zender and Kiehl (1994) who used a one-dimensional model which explicitly modelled the size distribution. Stephens et al. (1990) found that an ice-crystal effective radius of 16 \( \mu \)m gave good agreement with lidar-radiometer measurements. Observations from FIRE* 1986 give values for ice-crystal effective radii ranging from 36 \( \mu \)m to 153 \( \mu \)m (Stackhouse and Stephens 1991). However, because of instrument limitations, particles smaller than 44 \( \mu \)m were not detected and so information about small crystals has to be extrapolated.

The radiative effects of snow and graupel are treated in exactly the same way as those of rain, except that different values are used for the parameters for the size spectra in (6). For snow \( N_{0,s} = 3 \times 10^6 \) m\(^{-4}\) and of graupel \( N_{0,g} = 4 \times 10^4 \) m\(^{-4}\). The density of snow is assumed to be \( \rho_s = 100 \) kg m\(^{-3}\) and of graupel \( \rho_g = 300 \) kg m\(^{-3}\). In our simulations, values for the effective radius of snow particles range from about 200 \( \mu \)m to 1.5 mm and of graupel from about 200 \( \mu \)m to 4 mm.

4. MODEL SIMULATION

This section describes some of the features of the modelled squall-line and compares these with previous simulation studies by Tao et al. (1993) and Wong et al. (1993). Simulations including long-wave radiation are described, although the results from simulations with no long-wave radiation are qualitatively similar.

(a) Initial conditions

Tao et al. (1993) and Wong et al. (1993) have simulated the EMEX9 system. However, their initial conditions appear quite different. Tao et al. (1993) used a combination of observations from EMEX9. Below 400 mb, measurements from an aircraft* ahead of the system were used; above 400 mb, data interpolated from radiosonde profiles and objective analyses were used. Wong et al. (1993) used primarily the radiosonde profile for Darwin, Australia. Probably of most importance are the differences in the wind profile used. Both wind profiles have a low-level jet with maximum speed of approximately 15 m s\(^{-1}\), but for Tao et al. (1993) the jet is at about 850 mb while Wong et al. (1993) have the jet at about 550 mb. Above 700 mb the wind profile used by Tao et al. (1993) has very little

* First ISCCP* Regional Experiment.
† International Satellite Cloud Climatology Project.
‡ The Electra aircraft of the United States' National Center for Atmospheric Research.
shear and the wind speed is close to 7.5 mi s\(^{-1}\). By contrast, the wind profile used by Wong et al. (1993) has strong reverse shear between 500 mb and 100 mb, with a wind speed of \(-18\) m s\(^{-1}\) at 100 mb. Figure 1 shows the thermodynamic sounding used by Tao et al. (1993) and our simulations. Figure 2 shows the wind profiles used by Tao et al. (1993) and a wind profile similar to that used by Wong et al. (1993).

Following Tao et al. (1993), the cooling and moistening effect of a forced mesoscale upward motion on the basic state is applied. This upward motion has a maximum vertical velocity of 7 cm s\(^{-1}\) at 1300 m and decreases linearly with height to zero at 4400 m. A cold pool (5 K colder than its surroundings) is used to initiate the convection. More details are given by Tao et al. (1991) and Tao et al. (1993).
We run two sets of simulations: the first set uses the same initial conditions as Tao et al. (1993); the second set also uses the same initial conditions as Tao et al. (1993) but uses a wind profile like that of Wong et al. (1993).

(b) Model results

This section describes the results of two simulations; both include long-wave radiation. Simulation 1A uses the initial wind profile used by Tao et al. (1993) while simulation 2A uses an initial wind profile like that used by Wong et al. (1993). The evolution of the two systems is shown by the time series of maximum and minimum vertical velocities, minimum potential temperature anomaly and horizontally averaged surface-rainfall rates illustrated in Figs. 3 and 4. The life cycles of individual convective cells can be observed from the oscillations in the curves of maximum vertical velocity. New cells are generated about every 30 minutes.

Both systems develop quickly and reach the mature stage at about 4 hours. Model results show that the systems propagate by the discrete growth of new convective cells at the gust front, with decaying convection merging into the stratiform region. At the mature stage, the maximum vertical velocities are greater than 10 m s\(^{-1}\) in both simulations. In simulation 1A, the modelled squall-line moves with a speed of 12.5 m s\(^{-1}\), while for simulation 2A the propagation speed is 13.5 m s\(^{-1}\).

Simulation 2A produces a system with intense convection at the leading edge and a long trailing stratiform region. The convective cores have an upshear slope, and a narrow downdraught separates the convective updraught from the stratiform region. A mid-level rear inflow, with a double jet-core structure (two maxima) is also simulated; (see Figs. 7 and 8). Simulation 1A produces a system with a smaller stratiform region and the convective cores are nearly upright (see Figs. 5 and 6). The large separation between the convective and stratiform region is also produced in the simulation by Tao et al. (1993); (see their Fig. 2(d)). Radar observations show that the updraught has an upshear tilt in some regions of the squall line, whereas in other regions it is nearly vertical (Webster and Houze 1991).

![Figure 3](image-url)  
**Figure 3.** Time series of maximum and minimum vertical velocities, minimum potential-temperature anomaly and horizontally averaged surface-rainfall rate for simulation 1A.
Figure 4. Time series of maximum and minimum vertical velocities, minimum potential-temperature anomaly and horizontally averaged surface-rainfall rate for simulation 2A.

Figure 5. Model cloud and precipitation 4 hours into simulation 1A. The solid line encloses a region where the sum of cloud water, cloud ice, graupel and snow is greater than 0.001 g kg\(^{-1}\). The shaded regions show precipitation as described in the key. Darker shaded regions show amounts greater than 0.5 g kg\(^{-1}\) while lighter shaded regions show amounts greater than 0.01 g kg\(^{-1}\). Only part of the model domain is shown.

Figure 6. 3–6 hour time-averaged, system-relative horizontal wind field for simulation 1A. The dotted contour shows values of 0 m s\(^{-1}\), while the dashed contours are for 2, 4 and 6 m s\(^{-1}\). Solid contours are for −2, −6 and −10 m s\(^{-1}\).
The lifetime of our simulated squall lines is about 9 hours. However, the simulation by Tao et al. (1993) appears to last indefinitely. (Their simulation is run up to 16 hours and the squall line shows no signs of decaying.) This difference is found to arise from the difference in lateral boundary conditions used, periodic by our simulations and open by Tao et al. (1993). When Tao (personal communication) runs his simulation with periodic lateral boundaries the squall line also decays. Klemp and Wilhelmson (1978) showed that open lateral boundaries can be a significant source of moisture and we believe this is important. In our simulations, the precipitation causes the model domain to become drier and the subsequent lack of moisture causes the squall line to decay. In a separate simulation (results not shown), the moisture convergence resulting from the imposed mesoscale upward motion is increased by an order of magnitude. The simulated squall line persists throughout the simulation (16 hours), demonstrating the importance of the moisture inflow on squall-line longevity.

A separation technique based on rainfall rates (as described by Tao et al. 1993) is used to partition the rain into that originating in convective clouds ('convective rain') and that originating in stratiform clouds ('stratiform rain'). For simulation 1A, 58% of the total surface precipitation is classified as stratiform; this compares with a value of 42% for the EMEX9 simulation by Tao et al. (1993). For simulation 2A, 37% is classified as stratiform.

Figure 7. Model cloud and precipitation 4 hours into simulation 2A. The solid line encloses a region where the sum of cloud water, cloud ice, graupel and snow is greater than 0.001 g kg\(^{-1}\). The shaded regions show precipitation as described in the key. Darker shaded regions show amounts greater than 0.5 g kg\(^{-1}\) while lighter shaded regions show amounts greater than 0.01 g kg\(^{-1}\). Only part of the model domain is shown.

Figure 8. 3–6 hour time-averaged, system-relative horizontal wind field for simulation 2A. The dotted contour shows values of 0 m s\(^{-1}\), while the dashed contours are for 2, 4 and 6 m s\(^{-1}\). Solid contours are for \(-2, -6, -10, -15\) and \(-25\) m s\(^{-1}\).
(c) *Long-wave radiative heating*

The calculated long-wave heating rates are presented in Figs. 9 and 10 and show substantial horizontal variation through the cloud. This is related to differences within the cloud of its ice- and liquid-water-content and the species and size-distribution of its particles. Maximum cooling rates are found at cloud tops and have magnitudes of up to 30 K d$^{-1}$, which is in good agreement with the results of Tao *et al.* (1993) and Wong *et al.* (1993). Cooling in the anvil extends over a substantial depth. This is because cloud-ice amounts within the anvil are small so that absorption is relatively weak. If the anvil behaved as a black body then all the cooling would be localized at the surface (skin). Maximum cloud-base warming rates are found to be about 10 K d$^{-1}$, whereas the clear air has an average cooling rate of about 2 K d$^{-1}$. 
5. **Sensitivity to long-wave radiative effects**

Simulations are run in which LW radiative effects are switched off. (These simulations are labelled 1B and 2B.) Comparisons of horizontally averaged surface rainfall rates are shown in Figs. 11 and 12. Removing LW radiative cooling results in a slower development of the convective system and smaller rainfall-rates. Even after only two hours of model time, the difference in rainfall rates is noticeable and, at peak intensity, simulation 1A has a rainfall rate 27% greater than simulation 1B whereas simulation 2A has a rainfall rate 47% greater than simulation 2B.

![Figure 11](image1.png)

*Figure 11.* Horizontally averaged surface-rainfall rates. The solid line is for simulation 1A which includes long-wave radiation. The dashed line is for simulation 1B which ignores radiation.

![Figure 12](image2.png)

*Figure 12.* Horizontally averaged surface-rainfall rates. The solid line is for simulation 2A which includes long-wave radiation. The dashed line is for simulation 2B which ignores radiation.

Figures 13 and 14 show the effect of including LW radiation on horizontally averaged total rainfall. After 9 hours, simulation 1A produces 12% more total rainfall than 1B, whereas simulation 2A produces 29% more rainfall than 2B. The two simulations without LW radiation (1B and 2B) produce nearly the same amount of total rainfall, but simulation 2A produces significantly more total rainfall than 1A. The only initial difference between simulation sets 1 and 2 is the wind profile used. However, as described in subsection 4(b), there are several differences between the squall lines produced by simulation sets 1 and 2. Sensitivity to LW radiation can also vary between different cloud models. In simulation set 1, we use the same initial conditions as Tao *et al.* (1993). However, when they included LW radiation in their model, surface precipitation increased by 31% whereas, when we do
so, it increases by only 12%. A discussion of the radiative-forcing mechanisms bringing about the enhancement of precipitation is given in subsection 5(a).

Figures 15 and 16 show the horizontally averaged total stratiform rain for simulation sets 1 and 2. These figures show that, in our simulations, stratiform rain is not significantly enhanced by LW radiative effects and that it is the convective rain which is enhanced. A threshold rainfall-rate value of 25 mm h\(^{-1}\) is used to partition the convective and stratiform rain. However, using smaller threshold values still gives the same result: LW radiative effects do not significantly enhance stratiform precipitation. This result is different from that of Tao et al. (1993) who found that both stratiform and convective precipitation were enhanced by LW radiation.

Comparison of the 3-6 hour time-averaged, system-relative, horizontal wind fields shows that that LW radiation enhances the mid-level rear inflow. For simulation set 1, the maximum, system-relative, value of the mid-level rear inflow is increased from 4.7 m s\(^{-1}\) to 6.1 m s\(^{-1}\). For simulation set 2, the maximum, system-relative, value of the mid-level rear inflow is increased from 4.4 m s\(^{-1}\) to 5.1 m s\(^{-1}\). For simulation set 2, there is also evidence of the enhancement of the upper-level outflow by LW radiation; this is shown by Fig. 17. The rearward tilt of the updraught is also increased by LW radiation and this is also shown in Fig. 17.

(a) Radiative-forcing mechanism

Additional simulations are performed to examine the mechanism by which LW radiation enhances the surface precipitation; these experiments are described in Table 1. Only
Figure 15. Horizontally averaged cumulative stratiform surface-rainfall. The solid line is for simulation 1A which includes long-wave radiation. The dashed line is for simulation 1B which ignores radiation.

Figure 16. Horizontally averaged cumulative stratiform surface-rainfall. The solid line is for simulation 2A which includes long-wave radiation. The dashed line is for simulation 2B which ignores radiation.

Figure 17. Difference in the 3–6 hour time-averaged, system-relative horizontal wind fields from simulations 2A and 2B. Solid contours are for values of $-1$ m s$^{-1}$ and dashed contours for 1 m s$^{-1}$. 
TABLE 1. SIMULATIONS TO EXAMINE THE RELATIVE IMPORTANCE OF RADIATIVE-FORCING MECHANISMS

<table>
<thead>
<tr>
<th>Simulation</th>
<th>Cloud–radiation interaction</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>Long-wave radiation</td>
</tr>
<tr>
<td>B</td>
<td>No radiation</td>
</tr>
<tr>
<td>C</td>
<td>Reduced cloud-top cooling and reduced cloud-base warming</td>
</tr>
<tr>
<td>D</td>
<td>Reduced cloud-top cooling</td>
</tr>
<tr>
<td>E</td>
<td>Reduced cloud-base warming</td>
</tr>
<tr>
<td>F</td>
<td>Cloud and precipitation ignored by long-wave radiation parametrization</td>
</tr>
<tr>
<td>G</td>
<td>Spatially averaged long-wave cooling</td>
</tr>
</tbody>
</table>

results for simulation set 2 are shown since they show a larger sensitivity to LW radiation. Also, the most important radiative-forcing mechanism is found to be the same for both simulation sets. To examine the importance of cloud-top cooling and cloud-base warming, simulations are performed with reduced cloud-top cooling and/or cloud-base warming. Cloud-top cooling is reduced by locating grid boxes where the LW radiative cooling rate is greater than 3.5 K d⁻¹ and reducing the excess cooling rate by a factor of 10. The threshold value of 3.5 K d⁻¹ is chosen since this is the maximum cooling rate in clear air. Similarly, grid boxes which show LW radiative warming have the warming rate reduced by a factor of 10. To examine the importance of the domain-wide LW radiative cooling, two simulations are performed where both cloud-top cooling/cloud-base warming and the differential cooling between clear and cloudy regions are removed. In simulation 2F, this is achieved by ignoring cloud and precipitation in the LW radiation parametrization; in simulation 2G, the spatial variation of the LW radiative heating is removed. A full LW radiative-transfer calculation (including clouds and precipitation) is performed from which a mass-weighted domain-wide average cooling rate is derived. This is then applied at all grid points. Cloud and precipitation reduce the outgoing LW radiation at the top of the atmosphere and consequently the domain-average cooling rate. This effect is included in simulation 2G but not in 2F. The significance of the differential cooling between the clear and cloudy regions can be deduced by comparing the simulation in which cloud-top cooling/cloud-base warming has been removed (simulation 2C) with the simulations where only the domain-wide radiative cooling is included (simulations 2F and 2G). Differences in the LW radiative cooling rates between the simulations performed are shown in Fig. 18. The large cooling rate at cloud top and warming rate at cloud base are clearly visible. The differential cooling between columns containing cloud and columns containing no cloud is also apparent. Table 2 shows the total surface precipitation after 9 hours for the simulations performed.

Webster and Stephens's (1980) hypothesis can be tested by comparing simulation 2C (in which both cloud-top cooling and cloud-base warming are reduced) with simulation 2A. By reducing cloud-top cooling and cloud-base warming, the total surface-precipitation at 9 hours is reduced by 6% (comparing simulation 2C with 2A). This reduction in precipitation is the result of the reduction in cloud-base warming. When removing only cloud-top cooling (simulation 2D) the total surface-precipitation at 9 hours is changed by less than 1% (comparing simulation 2D with 2A). When only cloud-base warming is reduced the total surface-precipitation is reduced by 7% (comparing simulation 2E with 2A).

For completeness, we also examine the effect of cloud-top cooling and cloud-base warming on precipitation reaching the surface from convective and stratiform clouds ('convective surface-rainfall' (CSR) and 'stratiform surface-rainfall' (SSR)). Figure 19 shows the effect of cloud-top cooling and cloud-base warming on the CSR. The simplest interpre-
TABLE 2. TOTAL SURFACE RAINFALL

<table>
<thead>
<tr>
<th>Simulation</th>
<th>Total rain (normalized)</th>
</tr>
</thead>
<tbody>
<tr>
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<td>G</td>
<td>0.99</td>
</tr>
</tbody>
</table>

'Total rain (normalized)' denotes horizontally averaged total surface rainfall after 9 hours, the numbers are normalized with respect to simulation 2A.

Figure 18. Long-wave radiative-heating rates. The solid line shows the LW radiative-heating rate from simulation 2A for one particular cloudy column. The dashed line shows the effect of reducing cloud-top cooling and cloud-base warming using the procedure described in the text and used by simulation 2C. The dot-dashed line shows the LW radiative-heating rate in simulation 2A for a column containing no cloud. The double dot-dashed line shows the domain-wide spatially-averaged LW-heating rate, 4 hours into simulation 2G.

...tation is that cloud-base warming increases CSR whereas the effect of cloud-top cooling is negligible: simulations 2A and 2D include more cloud-base warming than simulations 2C and 2E and give more CSR; by contrast CSR in simulations 2C (reduced cloud-top cooling) and 2A (full cloud-top cooling) are almost identical. Similarly, CSR in simulations 2D (reduced cloud-top cooling) and 2A (full cloud-top cooling) are almost identical for the first eight hours of simulated time; the small difference during the last hour of the simulation is probably not significant.

Figure 20 shows the effect of cloud-top cooling and cloud-base warming on SSR. These results, however, are difficult to interpret. Differences in SSR between the simulations amount to only about 0.02 mm h⁻¹ and the simplest interpretation is that they are not significant. This view gains support from Fig. 16 which shows that LW radiative heating has a negligible effect on SSR.

Separating cases into convective and stratiform rainfall involves an arbitrary choice of threshold value of rainfall rate, and some errors in partitioning the cases may have caused the differences in Fig. 20. Because stratiform rain is less intense, errors in it are proportionately larger than for convective rain. Figure 20 shows that differences in SSR
between simulations are sensitive to the choice of threshold. In Fig. 20(a), for a threshold rate of 25 mm h\(^{-1}\) for example, simulation 2E (full cloud-top cooling and reduced cloud-base warming) produces more SSR than simulation 2A (full cloud-top cooling and full cloud-base warming). In Fig. 20(b), by contrast, for a threshold rainfall rate of 10 mm h\(^{-1}\), simulations 2A and 2E produce similar values of SSR.

Comparing simulation 2A with simulations 2F and 2G shows that it is the large-scale LW cooling rather than the cloud–radiation interaction which is primarily responsible for enhancing the surface precipitation. At 9 hours, simulation 2F produces 2% more total surface precipitation than simulation 2A, whereas simulation 2G produces 1% less than 2A. The differences in the total surface precipitation between 2A, 2F and 2G are not very significant. However, it is interesting to note that the domain-average cooling rate is greater in simulation 2F than in 2G (as discussed previously) and this can explain the greater surface precipitation from simulation 2F.

Dudhia (1989) states that convective activity is modulated by the cooling of the troposphere by long-wave radiation and forced vertical motion. Following Cotton and Anthes (1989) we note that the rate of precipitation reaching the surface \( P \) is related to the apparent heat-source due to all physical processes \( Q_1 \), radiative heating \( Q_R \) and the
surface flux of sensible heat $H_s$ by

$$P = \frac{C_p}{Lg} \int_{p_i}^{p_t} (Q_1 - Q_R) \, dp - \frac{H_s}{L}$$  \hspace{1cm} (9)

where

$$Q_1 = \frac{1}{C_p} \left[ \frac{\partial \tilde{s}}{\partial t} + \nabla \cdot (\tilde{V} \tilde{s}) + \frac{\partial \tilde{w} \tilde{s}}{\partial p} \right],$$  \hspace{1cm} (10)

$s = C_p T + g z$ is the dry static energy and other quantities have their conventional notations. Cooling by LW radiation will be partially balanced by greater latent heating due to increased convective activity. If the balance were exact then $Q_1$ would remain unchanged between the simulations including or excluding LW radiation and the increase in rainfall as a result of long-wave radiative cooling could be calculated from (9). Putting $Q_R = -2$ K d$^{-1}$, the cloud-top pressure $p_t = 200$ mb and the surface pressure $p_s = 1000$ mb gives an increase in the surface-rainfall rate of 0.27 mm h$^{-1}$. Comparing simulations 2A and 2B, the actual, average, surface-rainfall rate is enhanced by 0.15 mm h$^{-1}$ when LW radiation is
Figure 21. Effect of differential cooling between clear and cloudy regions on horizontally averaged cumulative surface-rainfall: simulations 2C, 2F and 2G.

included, since the apparent heat-source is smaller when LW radiation is included. Therefore, it may be more appropriate to treat the value of 0.27 mm h\(^{-1}\) derived above as an upper bound. However, given the simplicity of this argument, the agreement with the model results seems reasonable.

This argument also indicates that the enhancement of precipitation by radiation is approximately linearly related to \( Q_R \) and that the effect of radiation is to enhance the precipitation by some constant amount, regardless of the strength of the convective system. This is in good agreement with the results of Miller and Frank (1993) who found that radiation increased the domain-averaged rainfall by approximately 2 mm over a 24 hour period, regardless of the strength of the background forcing. Note that Miller and Frank (1993) included both solar and LW radiation, so that a value of \( Q_R = -1 \) K d\(^{-1}\) is more appropriate for their results.

To deduce the role of differential cooling between the clear and cloudy regions, simulation 2C is compared with 2F and 2G. Simulation 2C includes both large-scale LW radiative cooling and differential cooling between clear and cloudy regions, whereas simulation 2F and 2G include only the large-scale LW radiative cooling. Figure 18 shows the difference in LW radiative cooling between clear and cloudy columns. If differential cooling is important, then simulation 2C should produce significantly more precipitation than simulations 2F or 2G. However, Fig. 21 shows that the total surface-precipitation in simulation 2C is only slightly greater than that in simulations 2F and 2G between 2–6 hours and is less after six hours. Thus, differential cooling between clear and cloudy regions can be of only secondary importance in enhancing total surface-precipitation.

Chin (1994) claims that cloud-top cooling by LW radiation creates a gravity-wave duct which enhances the upper-level outflow. However, our results do not support this hypothesis. As shown in Fig. 17, LW radiation enhances the upper-level outflow but simulations 2A and 2C show that this enhancement is not due to cloud-top cooling. When the difference in the time-averaged horizontal wind fields from simulations 2A and 2C is plotted (not shown) only random noise is visible and there is no coherent structure. Our results indicate that the enhancement of the upper-level outflow and mid-level rear inflow are both the result of increased convective activity.
6. SUMMARY AND CONCLUSIONS

The two-dimensional version of the Meteorological Office Cartesian large-eddy-simulation model is used to model a tropical squall-line and reproduces many of the features observed during EMEX9. Model results show that convective systems can have significant sensitivity to LW radiation and that this sensitivity depends on the organization and structure of the convective system as well as on the cloud-resolving model used.

Our results confirm that the inclusion of LW radiation within cloud-resolving models is quantitatively important. They also show that the EMEX9 squall-line simulations are sensitive to some aspects of model formulation. In particular, the use of different lateral boundary conditions in the Goddard model and in our model gives rise to significant differences in the evolution of the squall line. The sensitivity of the simulated system to LW radiation also differs significantly between the two cloud-resolving models. Further investigation of these differences is left to a future study.

A number of sensitivity experiments are performed to identify the dominant cloud-radiative-forcing mechanism. Three mechanisms have been proposed for the enhancement of precipitation by LW radiation: cloud-top cooling/cloud-base warming; differential cooling between clear and cloudy regions; and domain-wide cooling. Results show that both the cloud-top cooling/cloud-base warming mechanism and the differential cooling between clear and cloudy regions mechanism are of minor importance and that it is the domain-wide cooling mechanism which is of primary importance. This result is in agreement with those of previous modelling studies (Dudhia 1989; Miller and Frank 1993).

The most important weakness of the long-wave radiation scheme is the treatment of absorption by cloud ice. The value of \( \kappa_{\text{ice}} \) used in the long-wave radiation scheme is equivalent to assuming that the ice-crystal effective radius is always 20 \( \mu \text{m} \). However, the actual ice-crystal effective radius will depend on quantities such as temperature and ice-water-content. A possible way to improve the long-wave radiation scheme is to use empirical data to parametrize the ice-crystal effective radius; see, for example, Sun and Shine (1993). However, experimental data are limited and mostly for mid-latitude cirrus. Existing data may not be representative for the anvils of tropical storms.

The simulations described in this paper do not include short-wave radiative heating. The simulation by Tao et al. (1993) does not include short-wave radiative heating either. Wong et al. (1993) include short-wave radiative heating and find maximum heating rates of 33 K d\(^{-1}\) at cloud tops. Therefore, short-wave radiative heating is likely to have an important effect on convective systems and will need to be included in future studies.

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