A Modeling Study on the Early Electrical Development of Tropical Convection: Continental and Oceanic (Monsoon) Storms

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Manuscript received 18 May 1993; in final form 20 December 1993

ABSTRACT

Numerical modeling studies of continental tropical and maritime tropical convection were conducted using the two-dimensional, nonhydrostatic, cloud electrification model developed at the South Dakota School of Mines and Technology. The model contains six classes of water (water vapor, cloud water, cloud ice, rain, snow, and graupel) and a full set of ion equations. All hydrometeors are permitted to exchange charge. Charge transfer between microphysical species is accomplished through a noninductive charging parameterization following Takahashi.

The goal of the numerical experiments was to examine the kinematic and microphysical differences that lead to marked differences in observed electrification between the break (continental) and monsoon (oceanic) convective regimes observed near Darwin, Australia. The break regime is associated with deep, intense convection that forms in high-CAPE (convective available potential energy) environments. Normally, copious amounts of lightning accompany break period convective events. Monsoon conditions are associated with heavy rain and relatively weak convection that forms in moderate to low-CAPE environments. Very little lightning activity is normally observed in the monsoon.

Three numerical simulations ranging from high- to low-CAPE conditions are presented. The results indicate that the electrification of the simulated storms critically depends on the juxtaposition of the level of charge reversal (LCR), which is in turn dependent on temperature and liquid water contents, and the particle interaction region, which is the level where ice particle collisions occur and thus where noninductive charging can take place. In the high-CAPE (break period) case, the LCR is located several kilometers below the interaction region, and strong in-cloud electric fields develop as a consequence. In the low- to moderate-CAPE (monsoon) cases, the LCR and interaction region are closely located in the vertical. As hydrometeors move across the LCR in both directions, the charge on their surfaces continually changes sign, thus preventing the development of a significant in-cloud electric field. It is further hypothesized that in conditions of zero to extremely low CAPE, the particle interaction region would be situated below the LCR, leading to the development of an inverted dipole (positive charge underlying negative charge), such as may occur in the stratiform regions of mesoscale convective systems.

1. Introduction

Satellite observations of global lightning patterns from the Defense Meteorological Satellite Program (DMSP) (Orville and Spencer 1979; Orville and Henderson 1986) and meteorological records (Livingston and Krider 1978; Lhermitte and Kreibiel 1979; Williams 1985), have shown that lightning activity is considerably more frequent over tropical landmasses compared to the tropical oceans. Recent observations taken during the Down Under Doppler and Electricity Experiment (DUNDEE) (Rutledge et al. 1992) near Darwin, Australia, (deep Tropics; 12°S) have shown that the frequency of lightning flashes associated with precipitating convective systems in that region was strongly dependent on whether the convection developed in an air mass of continental tropical (cT) or maritime tropical (mT) characteristics. Typically, thunderstorms that formed in cT environments during the DUNDEE were prolific producers of lightning with flash rates occasionally as high as 60 min⁻¹, while mT convection seldom exceeded flash rates of 10 min⁻¹ and, more typically, often produced no lightning at all, despite heavy rainfall. To gain a better understanding of these observations, this study presents results from cloud model simulations aimed at identifying the underlying dynamical and microphysical differences that can lead to the marked differences in the observed lightning rates between tropical continental and tropical oceanic convection.

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2. Observations and background for the present study

One objective of the DUNDEE was to study the dynamical and electrical differences of maritime tropical and continental tropical convective clouds. Analysis of the Carnegie curve (Whipple 1929) reveals that one of the three major regions of global lightning activity is over the Indonesian “maritime continent” region (Ramage 1968), which includes Darwin at its southern tip. For this reason, the Darwin area is an ideal region to conduct a field experiment on the dynamics and electrification of tropical convection.

Two contrasting weather regimes occur in the vicinity of Darwin during the wet season (Australian monsoon). The weather regimes are strongly dependent on the location of the intertropical convergence zone (ITCZ) or monsoon trough, which tends to traverse Darwin several times in the course of a wet season. When the ITCZ is north of Darwin, low-level southerly flow originating from the Australian continent is deflected to easterly flow by the Coriolis force and subjects Darwin to an air mass of cT characteristics. The convection in this regime is often triggered over the elevated topography to the south and east of Darwin in the afternoon, when heating is a maximum. Consequently, the convection in this region exhibits a strong diurnal cycle. Morphologically, the convection frequently organizes into intense squall lines, but deep, isolated convective cells, reaching heights of 18 km or greater are also observed (Keenan and Carbone 1992). The convective available potential energy (CAPE) in the cT air mass is generally very high, often exceeding 2000 J kg$^{-1}$, primarily due to high values of equivalent potential temperature in the boundary layer and drier air aloft (Rutledge et al. 1992). The lightning activity from the cT storms is intense with peak lightning flash frequencies often exceeding 20 min$^{-1}$. Storms with flash rates as large as 60 min$^{-1}$ have been observed. The time period in which Darwin is under the influence of a cT airmass is often referred to as the “break period” because it represents a break from persistent, low-level westerly winds and widespread heavy rains that characterize monsoon conditions.

The monsoon regime results when the monsoon trough is located over or to the south of Darwin. In this case, low-level northerly or northwesterly flow originating from over the ocean places Darwin in an air mass of mT characteristics. Extensive stratiform cloudiness with embedded convection and widespread heavy rains are the dominant characteristic of the mT convection (Keenan and Carbone 1992; Keenan and Rutledge 1993). Because of the low values of wet-bulb potential temperature at the surface (Williams and Renno 1993) and reduced solar heating during the periods of northerly and westerly flow, the CAPE in the mT air mass is typically much less than 1000 J kg$^{-1}$ (Rutledge et al. 1992). Lightning flash rates during “monsoon” periods (low CAPE) were observed to be considerably lower than those found during “break” periods (higher CAPE). Williams et al. (1992) also discussed this finding and furthermore found that the precipitation yield per lightning flash was one to two orders of magnitude larger in monsoon systems compared to break period storms.

A positive correlation between the flash rate$^1$ and CAPE was discussed by Rutledge et al. (1992). The positive correlation indirectly implies a relation between the convective vigor and dynamics of a storm as inferred through CAPE and the frequency of lightning. The relationships between CAPE, convective vigor, and electrification are further investigated in this study with the aid of a numerical model.

The observational network for the DUNDEE consisted of two Doppler radars, a surface mesonet, upper-air sounding stations, and electrical instrumentation. [See Rutledge et al. (1992) for details and locations of the various sensors.] The sensors used to study cloud electrification in the first season of the DUNDEE (1988/89) consisted of a single Lightning Location and Protection, Inc. (LLP) lightning detector (Krider et al. 1976; Mach et al. 1986), two flat plat antennae used to detect electric field changes within a range of about 40 km, and five corona points used to record total lightning activity. In the second season of the DUNDEE (1989/90), a network of four LLP sensors was deployed. The corona point network was not operated in the second season. An electric field meter was also operated during the 1989/90 season, located at the Massachusetts Institute of Technology (MIT) radar site. The field meter recorded field changes caused by lightning flashes and provided information on the sign of electrical charge overhead. Electrical observations obtained by this network are compared with model outputs discussed herein.

3. Model description

The model used in this study is the South Dakota Storm Electrification Model (SEM) (Helsdon and Farley 1987a,b). The SEM is a two-dimensional, dynamical model that uses bulk microphysical parameterizations. The model is capable of simulating both inductive and noninductive charging processes. However, only the contribution of noninductive charging (NIC) and free ion attachment to hydrometeors were considered in this study. The NIC physics in the model was simulated using a parameterization based upon the laboratory work of Takahashi (1978), as discussed in a later section.

The SEM has evolved from a series of models originally developed by Orville (1965) for the study of...
cumulus convection. The model dynamics are represented by a vorticity equation and a density-weighted streamfunction similar to those described by Takeda (1971), Schlesinger (1973a,b), and Hane (1973). A thermodynamic energy equation as described by Orville and Kopp (1977) and Lin et al. (1983) is also used in the SEM. The microphysical framework of the cloud model includes continuity equations for water vapor, cloud water, rain, cloud ice, snow, and graupel, using standard bulk microphysical techniques (Lin et al. 1983). A flowchart of the microphysical processes that are included in the SEM can be found in Helsdon and Farley (1987a).

McCumber et al. (1991) evaluated several different ice parameterization schemes in numerical simulations of tropical convection. They showed that results more consistent with observations were obtained if the high-density precipitating ice category was characterized as graupel using the Rutledge and Hobbs (1984) scheme when compared to cases that treated the high-density precipitating ice as hail, as in Lin et al. (1983). Guided by these findings, we have modified the model microphysical parameters for graupel and hail from the original Lin et al. (1983) formulation to allow for graupel particles more characteristic of maritime tropical conditions. The bulk density of graupel particles was reduced to 0.6 g cm\(^{-3}\) compared to 0.9 g cm\(^{-3}\) in the original formulation. The intercept parameter for the exponential size distribution function used to prescribe the graupel particle size distribution was increased to \(4 \times 10^{-3}\) cm\(^{-3}\) from the original value of \(4 \times 10^{-4}\) cm\(^{-3}\) characteristic of large hail in the Great Plains. The drag coefficient was increased to 1.2 from the original value of 0.6 to give reduced fall speeds more consistent with graupel particles.

\(a.\) Electrification microphysics

Simulated graupel, rain, cloud ice, snow, and cloud water can all carry electric charge in the model. Charge is transferred from one class of hydrometeors to another through any of the microphysical processes involving collisions between particles in the above classes, as well as by NIC and interaction with free ions.

In general, the charge conservation equation for species \(i\) is

\[
\frac{\partial Q_i}{\partial t} = -V_i \cdot \nabla Q_i + \frac{1}{\rho_a} \frac{\partial}{\partial z} (V_i Q_i \rho_a)
+ \nabla \cdot K_w \nabla Q_i \pm \left( \frac{\partial Q_i}{\partial t} \right)_{\text{inter}},
\]

(1)

where \(Q_i\) is electrical charge density; \(V_i\) is the terminal fall speed; \(V\) is the horizontal wind vector; \(t\) and \(z\) represent time and height, respectively; and \(K_w\) is an eddy (mixing) coefficient. For both cloud water and cloud ice, \(V_i\) is set to zero. The terms on the right-hand side of (1) are, respectively, the horizontal and vertical advection of charge, the fallout of precipitating charged hydrometeors, subgrid-scale mixing, and a term that represents interactions between different classes of hydrometeors or ions. The last term, the hydrometeor interaction term, is the most significant in this study since the NIC parameterization is included in this term.

\(b.\) The hydrometeor interaction term

In the numerical model, hydrometeors can acquire charge through ion capture by means of microphysical processes that transfer both mass and charge from one hydrometeor class to another and by noninductive charging involving rebounding collisions. The hydrometeor interaction term in (1) can therefore be represented as

\[
\left( \frac{\Delta Q}{\Delta t} \right)_{\text{inter}} = \left( \frac{\Delta Q}{\Delta t} \right)_{\text{NIC}} + \left( \frac{\Delta Q}{\Delta t} \right)_{\text{micro}} + \left( \frac{\Delta Q}{\Delta t} \right)_{\text{ion}}.
\]

(2)

The first term is the NIC transfer per time step per grid cell and the last two terms represent charge transfer through microphysical processes such as accretion or melting (e.g., charge on snow becomes a source for charge on rain following melting), and ion attachment, respectively. A complete description of the ion physics can be found in Chiu (1978) and Helsdon (1980), while the microphysical charge transfers are addressed in Helsdon and Farley (1987b). The NIC process is described in detail in section 4.

c. Determination of electrical quantities

The primary electrical quantity that is diagnosed in the model is the electric field vector. To obtain this quantity, the space charge density is computed by summing the charge contribution from hydrometeors and ions. The electric potential can then be determined by inverting Poisson's equation:

\[
\nabla^2 \phi_E = -\frac{\rho_T}{\epsilon_0},
\]

(3)

where \(\phi_E\) is electric potential, \(\rho_T\) is the space charge density, and \(\epsilon_0\) is permittivity of free space. Finally, the electric field is computed by taking the negative gradient of the electric potential:

\[
E = -\nabla \phi_E.
\]

(4)

d. Model domain, boundary conditions, and simulation termination

The model domain is 24 km \(\times\) 24 km in the \(x\)-\(z\) plane with 200-m grid spacing in the horizontal and vertical. At the top of the domain, vertical velocity, vorticity, and hydrometeor mixing ratios are all set to zero, and entropy, water vapor, and the streamfunction
are held at their initial values. Vertical velocity, vorticity, and the streamfunction are forced to zero at the lower boundary. At the lateral boundaries, the streamfunction is specified. For inflow, the normal velocity is kept at its previous value, whereas for outflow conditions an Orlanski-type radiation boundary condition is applied to forecast the new value of the normal velocity. The streamfunction along the lateral boundaries is then built up from the new values of the normal velocities. Scalar quantities are held constant at the lateral boundaries for inflow conditions and extrapolated using first-order upstream differencing for outflow conditions.

To invert (3), the boundary conditions must be prescribed. The horizontal gradient of the electric potential at the lateral boundaries is set to zero, as is the potential at the bottom of the model domain. This is consistent with the earth’s surface being considered a perfect conductor. At the top of the model domain, the potential remains a constant, which is determined by the integration of (4) with the initial vertical profile of the fair-weather electric field specified as in Helsdon and Farley (1987b). The initial ion profiles are exponentially increasing with height in a manner consistent with the fair-weather electric field profile.

The time step in the model is variable and largely governed by the drift velocity of free ions in the simulated electric field. As the electric field increases, the ionic drift velocity (the velocity acquired by free ions accelerating in an electric field) increases and the time step must be reduced in order to avoid numerical instabilities. The initial time step is 3.75 s and it is reduced to values less than 0.5 s for electric fields on the order of 200 kV m$^{-1}$. Eventually, the time step becomes prohibitively small and the model simulation is then terminated. This termination normally occurs before the electric field approaches the dielectric strength of air, or near the time of first lightning. Therefore, the model results are applicable only to the early electrification stage of the simulated storm, describing the cloud physics and dynamical and charging processes leading up to the first lightning discharge.

4. The noninductive charging mechanism

a. The process

Laboratory studies (Reynolds et al. 1957; Takahashi 1978; Jayaratne et al. 1983; Keith and Saunders 1989; Saunders et al. 1991) have shown that when graupel undergo rebounding collisions with ice particles in the presence of supercooled liquid water, the graupel and ice can carry away opposite charges on their surfaces. This process does not require the presence of an external electric field. If this process occurs within large volumes of a storm, gravitational settling can result in the storm-scale separation of the hydrometeors carrying opposite charge, thus producing an electric dipole (Williams 1985).

NIC rests on very poorly understood physics; the underlying physical processes that govern charge transfer when graupel collide with ice crystals in the presence of supercooled liquid water is unknown. For the purposes of these numerical modeling studies, it is more important to know the variables upon which the sign and magnitude of the charge transfer depends and less, in a relative sense, on the underlying physical mechanism. Laboratory studies have provided the best insight into the dependence of the NIC mechanism on various quantities. The magnitude of the charge transferred in the NIC process has been shown to be dependent on temperature, cloud water content, impact velocity, and ice crystal size (Reynolds 1954; Reynolds et al. 1957; Takahashi 1978; Gaskell and Illingworth 1980; Jayaratne et al. 1983; Jayaratne and Saunders 1985; Baker et al. 1987; Keith and Saunders 1989).

b. NIC laboratory studies of Takahashi (1978)

Takahashi (1978) published the first extensive tabulation of the dependence of the charge separated per graupel–ice particle collision on both temperature and cloud water content. The apparatus used in the experiment consisted of a riming rod (simulating a graupel particle) that was whirled within a cloud chamber. Supercooled water and small ice crystals were simultaneously present within the cloud chamber. Peripheral instruments were used to measure ice particle concentrations and sizes, cloud liquid water content, temperature, and electrical current resulting from charge separation. The results of Takahashi’s laboratory experiments are shown in Fig. 1. There are several important points that should be gleaned from this figure:

(i) Positive charging of the riming rod (graupel) occurred when the temperature was greater than $-10^\circ$C, regardless of cloud water content.

(ii) At temperatures colder than $-10^\circ$C, positive charging occurred (at both low and high cloud water contents).

(iii) Negative charging occurred in regions where cloud water contents were moderate to high when the temperature was less than about $-10^\circ$C.

(iv) The temperature at which electrification reverses from positive charge on the rimer to negative charge on the rimer, called the “charge reversal temperature,” decreases as cloud water content decreases.

Two points of interest that are not apparent from Fig. 1 are (i) no charging occurred when the riming rod stopped rotating (no collisions), and (ii) no charging occurred when the rod rotated in an environment containing only ice crystals or only supercooled liquid water.

Takahashi (1978) measured only the dependence of charge separation on temperature and cloud water content. Parameters such as impact velocity and ice crystal size were kept reasonably constant so as to minimize
these effects. Because the Takahashi (1978) study represents the most complete dataset for NIC in terms of the range of temperature and cloud water content, the Takahashi charging data were adapted for use in the present modeling study. The Takahashi results are implemented by means of a lookup scheme; given the temperature and liquid water content, a charge transfer per collision is obtained from a table whose values have been interpolated from Fig. 1 and stored in a 2D array.

c. Previous modeling studies using NIC

Takahashi (1984) conducted a modeling study that is similar in two respects to the work reported here. Takahashi’s study focused on identifying the electrical differences between continental and maritime thunderstorms, as does this study. Furthermore, both studies used the charging parameterization developed by Takahashi (1978). Takahashi’s (1984) study is different in that continental and maritime convective regimes were differentiated only by the cloud condensation nuclei concentration, while thermodynamic profiles are used to differentiate between break period and monsoon conditions in this modeling study. Therefore, different dynamics, as a consequence of widely different CAPE, and the resultant feedback into the microphysics and electrification can be captured in this study. As suggested earlier by the positive correlation between CAPE and lightning frequency, it is vitally important to simulate the dynamical differences between the continental and maritime regimes in order to capture the resultant microphysical differences leading to observed lightning differences.

Another electrical modeling study was conducted by Helsdon and Farley (1987a,b) using the SEM. The NIC parameterization used was a simple two-value representation (one value for graupel-cloud ice interactions, and another value for graupel-snow interactions to roughly account for ice particle size dependence); the sign of the charge transfer was determined by the charge reversal temperature, and the magnitude of the charge transfer was held constant for each of the two possible interactions. Inductive charging and ion attachment were also explicitly included through the inclusion of the appropriate equations.

Helsdon and Farley used the SEM to successfully simulate the early electrical development of a High Plains convective storm (as verified by in situ aircraft data). A normal polarity electric dipole (positive charge situated above negative charge) developed in the model with electric potentials near the dielectric strength of air close to the time that the first lightning was observed. The charge centers were attributed to the gravitational separation of graupel from smaller ice particles. The NIC process was determined to be essential in the initial electrification of the storm. Inductive processes became active but surpassed only the effectiveness of the NIC mechanism once a strong electric field had developed. A model run with only inductive charging failed to produce significant electrification. The general success of the Helsdon and Farley studies was the primary reason that the SEM was chosen as the model for this study.

5. Useful definitions

It is useful at this point to define the exact meanings of three terms that will be used in following sections: the particle balance level, the particle interaction region, and the level of charge reversal (LCR). The particle balance level is the level at which the terminal fall speed of a hydrometeor is balanced by an updraft of equal magnitude. For steady-state conditions, the maximum concentration of a hydrometeor type will be found at the particle balance level. If the tendency of the updraft is not zero, a particular hydrometeor species will tend to migrate toward the particle balance level. The particle interaction region is the region where the hydrometeors associated with two different species interact through collisions. In the case of graupel and snow or graupel and cloud ice, the particle interaction region defines the portion of the cloud in which NIC can occur.

The LCR is the height at which ice particles begin acquiring charge of the opposite sign as a consequence
of the charge reversal temperature characteristic of the NIC mechanism. For example, as a graupel particle with positive charge on its surface is carried above the LCR by an updraft, it will start acquiring negative charge on its surface as it undergoes further collisions, thus neutralizing the positive charge. If the LCR is below the particle interaction region of graupel and snow, or graupel and cloud ice, then graupel will always acquire negative charge, and snow and cloud ice will acquire positive charge. Opposite signs of charging will occur if the LCR is above the particle interaction region. If the LCR occurs within the particle interaction region, then graupel (snow/cloud ice) will acquire negative (positive) charge above the LCR and positive (negative) below it. The LCR presented on subsequent figures was determined by examining the charge tendency of graupel and snow. The transition from a positive to negative charge tendency identifies the LCR.

6. Case study I: Break period (high CAPE)

a. Observations

We now turn to the discussion of the simulation of a continental thunderstorm observed near Darwin on 19 January 1990. The thermodynamic profile (Fig. 2) taken at Darwin at 0000 UTC (0930 local) 19 January indicated that the environment was in a state characteristic of the break period. The CAPE\(^2\) (1910 J kg\(^{-1}\)) was well above the average break period environment of 1194 J kg\(^{-1}\) (as determined from all break period soundings from the two DUNDEE seasons). The wind structure consisted of deep easterly flow, characteristic of the break period.

Radar reflectivity data from the MIT radar at 0612 UTC showed three regions or cores with radar reflectivities exceeding 40 dBZ (Fig. 3a). The largest and most intense core (centered near \(x = 30, y = 15\)) was located directly south of the MIT radar. A vertical cross section through this region showed that the 40-dBZ contour extended to at least 6.5 km AGL (Fig. 3b). The maximum elevation angle in the present scanning sequence was not large enough to scan above the storm top at the short ranges involved, thus preventing the full depth of the storm to be studied at this time.

A vertical cross section obtained from the MIT radar at 0620 UTC (Fig. 4) indicated continued vertical development of the convection with reflectivities of 40 dBZ extending to near 8 km, well above the melting level (located at 5.3 km). Echo tops were near 13 km AGL at this time. Rain began falling at the MIT radar site near this time. By 0636 UTC, echo tops had increased to 18 km AGL, with reflectivities up to 20 dBZ extending to 17 km (not shown). The total lightning flash rate was estimated at 20 min\(^{-1}\) at this time. The vertical component of the surface electric field measured at the MIT radar site (from field mill observations) suggested significant electrification (Fig. 5), as indicated by large perturbations from the fair-weather electric field (and negative charge overhead). The onset of significant charging (at 0609 UTC) and the first lightning event at 0624 UTC are evident from the field mill record. The cell weakened only slightly as it moved north, maintaining 30-dBZ reflectivities up to 16 km and an echo top of 17 km through 0650 UTC. After this time, the cell rapidly dissipated. Radar echo tops decreased rapidly to 14–15 km, and the 30-dBZ echo contour extended to only 6 km by 0658 UTC. Coincidentally, the flash rate decreased to only 1 or 2 min\(^{-1}\).

In summary, the chronological history of the storm (in UTC hours) is as follows.

- 0552: Convection begins. First radar echoes present.
- 0612: Convection sighted 10 km to the south of the MIT radar. Radar-indicated echoes are in excess of 40 dBZ at low levels—significant perturbation of fair-weather field.
- 0620: Reflectivities of 40 dBZ extend to 8 km AGL with overall tops to 13 km AGL.
- 0624: First lightning.

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\(^2\)In this study, the CAPE is the potential CAPE, computed by finding the intersection of the mean mixing ratio in the lowest 100 mb with the temperature trace and assuming a dry-adiabatic lapse rate below the intersection level.
• 0636: Storm matures. Radar echo tops extend to 18 km; 20-dBZ reflectivities extend up to 17 km. The flash rate is 20 min⁻¹.
• 0650: Cell changes very little through this time. Reflectivities of 30 dBZ extend to 16 km AGL. Echo tops extend to 17 km AGL.
• 0658: Convection begins to rapidly dissipate.

b. Model results

The 0000 UTC Darwin sounding served as the basis for defining the thermodynamic and dynamic base state for the 19 January 1990 simulation. The temperature and moisture profiles applied to the model were Darwin values modified to reflect features of the 1300 (local time) Koolpinyah sounding below 500 mb. Koolpinyah, the site of the MIT radar operation, is situated about 30 km east of Darwin. The subcloud region was approximated by dry-adiabatic conditions. The model wind regime was obtained by using 75% of the values from an east–west projection (perpendicular to the observed storm motion) of the Darwin winds, with 2.5 m s⁻¹ added to allow for grid translation. Mesoscale convergence of $1 \times 10^{-4}$ s⁻¹ was superimposed at lower levels (with divergence aloft) as described in Chen and Orville (1980), to approximate the effects of the sea-breeze front, considered to be important to the onset of convection on this day. In addition, random perturbations were applied to the temperature and water vapor fields in the lower levels. The maximum amplitude of these perturbations, which decreased with distance from the center of the horizontal extent of the domain and with height, were $\pm 0.5$ K and $\pm 7.5\%$, respectively.

The general dynamical and microphysical evolution of the simulated cloud is shown in Fig. 6 as a series of cloud and precipitation depiction plots at selected times. The dashed lines are the streamlines that give an indication of the 2D airflow. The solid contours sepa-
rate regions of water-saturated and water-subsaturated air (i.e., cloud edges). The microphysical and electrical development of the storm can be separated into three stages. In the first stage (from $t = 0$ to $t = 18$ min; Fig. 6a) the cloud growth is slow (based on cloud-top heights and rate of change of kinetic energy, not shown) and virtually no ice species are created. For this reason, the slow development stage of the cloud is of little interest: without ice, there is no electrification possible (based on the NIC scheme used in the model).

The second stage of development (from $t = 18$ to $t = 44$ min, Figs. 6b,c) is characterized by a rapid increase in cloud-top heights, a sharp increase in kinetic energy (or vertical motion), and the development of a deep mixed-phase region. (The formation of an electric dipole coincides with this stage, as discussed below.) Finally, in the third and last stage of the cloud development (from $t = 44$ to $t = 52$ min; Fig. 6d), the cloud

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**Fig. 6.** Cloud and precipitation depiction for 19 January 1990 at (a) 18 min, (b) 33 min, (c) 42 min, and (d) 52 min of simulated time. Asterisks and dots indicate graupel and rain mixing ratios greater than 1.0 g kg$^{-1}$, respectively. Hyphens and S's indicate cloud ice and snow mixing ratios greater than 0.5 g kg$^{-1}$, respectively. Streamlines are represented by dashed lines, and the cloud boundaries are shown by the thin solid line.
taps rise more slowly, the kinetic energy levels off, and the production of ice species reaches a maximum. As will be discussed below, the development of the anvil cloud (Fig. 6d) is accompanied by a tilted electric dipole. The break period simulation was terminated at \( t = 52 \) min because of the increasing drift velocity of the free ions.

1) **Stage II: Rapid Growth Phase**

The development of an electric dipole roughly begins with the creation of ice particles at \( t = 30 \) min. The graupel, snow, cloud ice and cloud water fields, and the charge structures associated with these fields at \( t = 33 \) and \( t = 42 \) min are shown in Figs. 7–10. The LCR (level of charge reversal; defined in section 5) determined from the temperature and cloud water fields is also shown as a thick solid line. At \( t = 33 \) min, large graupel mixing ratios are found both above and below the LCR (Fig. 7a). Snow is found both below and above the LCR but is most abundant above (Fig. 7b), while cloud ice is found almost exclusively above the LCR (Fig. 7c). The spatial distribution of the ice species results in most of the particle interaction region for graupel and snow, as well as graupel and cloud ice, to be situated above the LCR, although the distance between the center of the interaction region (subjectively determined) and the LCR is no more than about 1 km. Because a majority of the graupel–snow and graupel–cloud ice collisions occur above the LCR (and in a region containing abundant amounts of supercooled cloud water, Fig. 7d), graupel (snow/cloud ice) on average acquires negative (positive) charge (Figs. 8a–d). Gravitational settling of the ice species results in a net cloud-scale separation of charge (Fig. 8d). Graupel carrying negative charge (Fig. 8a) and snow carrying positive charge (Fig. 8b) migrate toward their respective particle balance levels. Additionally, cloud ice, carrying positive charge (Fig. 8c), is carried to the upper part of the cloud by the strong updraft. The net distribution of charge results in a normal polarity (positive charge over negative charge) dipole (Fig. 8d). The effectiveness of the cloud-scale separation is somewhat limited because of negatively charged snow from below the LCR being advected into positively charged regions of snow that exist above the LCR, and negatively charged graupel from above the LCR falling below that level into a region containing positively charged graupel.

At \( t = 42 \) min (Figs. 9a–d) the graupel and snow–cloud ice interaction region is located almost entirely above the LCR, and the separation between the particle interaction level and the LCR is now approximately 2.5 km. Similar to the situation at \( t = 33 \) min, graupel is found both below and above the LCR (Fig. 9a), and snow and cloud ice are found almost entirely above the LCR (Fig. 9b–c). This juxtaposition of ice species results in a more efficient cloud-scale separation than was found at \( t = 33 \) min, because there is significantly less charge cancellation from positively (negatively) charged graupel (snow) being carried across the LCR into regions of negative (positive) charge. The vertical electric field at \( t = 33 \) min (Fig. 10a) has a maximum magnitude of about 1 kV m\(^{-1}\) and rises to about 185 kV m\(^{-1}\) at \( t = 44 \) min (Fig. 10b). The maximum space charge density is near 2.5 C km\(^{-3}\) by \( t = 44 \) min (Fig. 10c).

2) **Stage III: Mature Phase**

During the last stage of development the first convective updraft collapses, a deep downdraft develops, and new convective cells begin to form on the upshear (left) side of the complex. A series of vertical velocity plots (Figs. 11a–c) shows the remnants of the first convective updraft at \( t = 44 \) min centered near 7 km AGL and \( x = 12 \) km, the indications of a developing downdraft downshear (right) of the decaying updraft, and a new, developing updraft in the lowest 4 km (Fig. 11a). Four minutes later (\( t = 48 \) min; Fig. 11b) the old updraft has been advected farther downstream and is now clearly disconnected from the main portion of the storm by a well-defined downdraft. The new updraft has grown rapidly and extends through the entire depth of the storm. Finally, at \( t = 54 \) min (Fig. 11c), the old updraft is virtually unidentifiable, and the dominant updraft shows indications of becoming disconnected from the boundary layer air. In its place, a new convective cell is forming at \( x = 12 \) km to begin the cycle all over.

Given the progression of the vertical velocity field, the progression of the charge density field is not surprising (Figs. 12a–c). At \( t = 44 \) min (Fig. 12a), a charge dipole resulting from positively charged snow and ice overlying negatively charged graupel is located almost entirely above the developing downdraft with only a small portion over the updraft. This is reasonable since the genesis of the downdraft is largely due to precipitation loading and evaporation. The maximum vertical electric field resulting from this configuration is \(-186\) kV m\(^{-1}\). There is no appreciable electrification associated with the newly developing cell at this time.

At \( t = 48 \) min (Fig. 12b) the main body of the dipole collocated with the downdraft begins to descend while a small upshear portion continues to ascend in the remnants of the first updraft. This trend continues through \( t = 52 \) min (Fig. 12c) and gives the appearance of a clockwise rotation or tilting of the electric dipole. Clearly, Fig. 12c shows that the positive and negative charge centers are located nearly at the same height in contrast to the vertical orientation at earlier times. In addition, the dominant updraft begins to show the development of a vertically oriented dipole although the signature in the contoured field is overwhelmed by the stronger charge densities associated with the decaying updraft. Finally, a thin negatively charged screening layer develops over the time period as negatively charged ions are attracted to the positively
charged snow and cloud ice in the upper region of the convective cloud.

c. Comparison with observation

The simulated storm can be compared with observations by considering the first radar echo predicted by the model and the first observed radar echo to be simultaneous. Consequently, 0552 UTC (time of first observed echo) will correspond to $t = 21$ min in the simulation time frame. One complication that arises when directly comparing the radar data to the model results is that the proximity of the storm to the radar prevented the radar scans from topping the storm. Therefore, direct comparison of radar cloud-top heights and predicted cloud-top heights is not possible.

With the above-mentioned problem in mind, there are two times at which the model reflectivity can be compared to radar observations. The first is at 0612 UTC ($t \approx 40$ min in model time). In general, the model
Fig. 8. Charge density on ice hydrometeors and the total charge density at \( t = 33 \) min for the 19 January 1990 case. The fields shown are (a) graupel charge, at a contour interval of 0.06 nC m\(^{-2}\); (b) snow charge, at a contour interval of 0.05 nC m\(^{-3}\); (c) cloud ice charge, at a contour interval of 0.02 nC m\(^{-3}\); and (d) total charge density, at a contour interval of 0.03 nC m\(^{-3}\). Dashed contours indicate negative values, whereas positive values are indicated by solid contours. The thick black line indicates the level of charge reversal as in Fig. 7.

has a tendency to overpredict reflectivity values. Nonetheless, the height of the deep convective core and the overall reflectivity structure compare reasonably well to the observations at this time (Fig. 13a, model; Fig. 3, observations). It is also important to keep in mind that the model essentially represents an east–west cross section, while Fig. 3b is a north–south cross section. (An east–west cross section was unattainable.)

The other time at which the model can be compared to observations is at \( t = 50 \) min (0620 UTC), shortly before the termination of the simulation. Similar to the earlier time, the simulated reflectivity (Fig. 13b) is in fairly good agreement with observations (Fig. 4), but the simulated values are generally higher than the observed values. Once again, it is important to keep in mind that the model and observations represent orthogonal cross sections.

We consider the overprediction of reflectivity values to be a problem. However, we have con-
ducted sensitivity studies (discussed in a later section) and found that the slope and intercept parameters for the microphysical species can be changed without significantly affecting the overall electrical development. The changes in the slope and intercept parameters, however, do change the values of the predicted reflectivity. Therefore, if parameters had been chosen through trial and error to give more realistic values of reflectivity, we assert that the electrical development would not be significantly altered. Availability of computer resources limited the amount of simulations that we could undertake.

In addition to the radar, measurements of the vertical electric field (Fig. 5) were recorded and can be compared to the growth of the simulated electric field (Fig. 14). Be sure to note the timescale change in Fig. 5. This is an unavoidable consequence of the manner that the data was recorded. The predicted vertical electric field begins to grow near $t = 31$ min in association with the cloud entering the rapid growth phase. The observed electrification began at 0554 UTC ($t = 23$).
Between $t = 36$ and $t = 40$ min, the simulated field leveled off, and the same tendency was observed between 0600 ($t = 29$) and 0609 UTC ($t = 38$). It is not clear what caused the leveling off in the observed storm, but it is due to the multicellular convective cycle in the model. We suspect that the electrification of the observed storm is also responding to the cycle of convective updrafts. Finally, the simulated field grows rapidly from $t = 40$ to $t = 47$ min and then begins to decrease. In the observations the field grows rapidly from 0609 ($t = 38$) to 0615 UTC ($t = 44$) and then decreases. Again, the increase and decrease in the vertical electric field is related to the convective cycle simulated in the model and likely attributable to the same process in the observations. Overall, the trends of the simulated vertical electric field seem to agree well with the observations.

7. Case study II: Monsoon case (low CAPE)

a. Observations

The ITCZ passed through Darwin from the north on 12 January 1990. Prior to the passage of the mon-
soon trough, a sounding taken at 0000 UTC 12 January 1990 (used to initialize the model) revealed a nearly saturated thermodynamic profile with easterly flow aloft (Fig. 15). The winds in the low levels were north to northeasterly. The sounding had a CAPE of approximately 300 J kg⁻¹, which is considerably less than the CAPE in the previous break period simulation.

The MIT radar observed two convective systems on this day. Both systems were in their mature stages when they were within radar range; hence, the full evolution of the systems was not captured. At 0340 UTC, radar echo tops from the first system were near 10.5 km (Fig. 16a). The largest reflectivities were near 40 dBZ, extending to a depth of only about 5 km AGL, and were oriented in a poorly organized northwest–southeast line (Fig. 16b). The system was associated with the passage of the ITCZ. The second system was a more organized squall line that passed through the DUNDEE area between 0730 and 1010 UTC. The reflectivity and Doppler velocity data have been extensively analyzed for this case by Keenan and Rutledge (1993). Radar data for this later time are shown in Figs. 16c,d. The largest reflectivities are again near 40 dBZ, with echo tops to approximately 12 km AGL (Fig. 16c). The echo pattern at z = 1.5 km is characterized by weak
convection (reflectivities greater than 30 dBZ) embedded within stratiform precipitation (Fig. 16d).

The electric field record at the MIT radar site indicated very little electrical activity (not shown). There were no lightning events recorded by the electric field mill and there were no cloud-to-ground events recorded by the LLP network within a 50-km radius of the MIT radar site during the 24 h beginning 0000 UTC 12 January 1990.

The important features to note for the monsoon case are the following:

(i) a nearly saturated thermodynamic profile, typical of the monsoon;

(ii) low CAPE, typical of an "average" monsoon event (around 300 J kg\(^{-1}\));

(iii) radar reflectivity values rarely exceeding 40 dBZ above about 5 km;

(iv) radar echo tops generally less than 10–12 km AGL; and

(v) no observed or recorded lightning activity.

These features are markedly different compared to similar parameters in the previous break period case.

b. Model results

Unlike the previous case, we are not simulating an isolated convective storm for the 12 January 1990 case.
Rather, the microphysical and electrical development for the more widespread monsoon precipitation (as represented by the observations described above) is simulated. The 0000 UTC Darwin sounding of 12 January 1990 was used to initialize the simulation of the monsoon case. The subcloud region was modified to approximate dry-adiabatic conditions. The model wind regime was perpendicular to the observed storm motion to the west-southwest. A convergence of $1.5 \times 10^{-4}$ s$^{-1}$ was applied to approximate the synoptic-scale upward motion associated with the monsoon trough. Random temperature perturbations were used to initiate convection. The general evolution of the monsoon system is shown as a series of cloud depictions similar to those shown for the break period simulation. Unlike the break period case, the monsoon simulation did not have well-defined growth stages (Fig. 17). Through the first half hour of the model run, a nearly uniform cloud deck developed and deepened with time. Warm rain processes produced limited rain, but no electrification occurred. After 40 min (Fig. 17a), more extensive regions of ice developed and produced heavier precipitation (from the melting of ice) by $t = 60$ min (Fig. 17b). An extensive ice cloud developed by $t = 80$ min (Fig. 17c).

At $t = 40$ min, convective circulations embedded in the stratiform cloud began to develop with peak vertical motions of 14 m s$^{-1}$ (which were very limited in horizontal extent; see Fig. 18a). In contrast to the break period simulation, the peak updrafts are found in the

![Electric Field Component Tendencies for High Cape Simulation 19 Jan 1990](image)

Fig. 14. A time series of the domain maximum vertical and horizontal electric field for the 19 January simulation, as predicted by the model simulation.

![Thermodynamic Sounding at 0000 UTC 12 January 1990 at Darwin](image)

Fig. 15. Thermodynamic sounding at 0000 UTC 12 January 1990 at Darwin.
lower troposphere, centered near the 0°C level. In the break period simulation, the peak updrafts were centered near 8 km. Coincident with the updrafts, maximum liquid water mixing ratios of 3.5 g kg⁻¹ were present (Fig. 18b). Graupel was created in the convective updrafts by riming (Fig. 18c) and snow existed in a stratiform layer about 2.5 km in thickness, centered near 7.5 km (Fig. 18d). With the juxtaposition of appreciable quantities of snow and graupel, the NIC mechanism began to operate near this time.

By \( t = 60 \) min (Fig. 19), a large percentage of the troposphere was saturated and cloud tops reached 10 km. The cloud water field (Fig. 19b) that was stratiform in nature at earlier times was now becoming highly convoluted because of convective motions. The vertical velocity field (Fig. 19a) shows several narrow updraft and downdraft cores with peak velocities located near 4 km AGL. Although weak upward motion extends to near 10 km (cloud top), vertical velocity values greater than 10 m s⁻¹ are confined below 8 km. A band of snow was present and extended through most of the horizontal domain between approximately 6 and 10 km AGL (Fig. 19c). The highest snow mixing ratios were concentrated near 8 km. As before, graupel (Fig. 19c) was created in the updrafts, but rarely penetrated much above 7 km. The LCR is shown as a thick black line through Fig. 19 and indicates that the particle interaction region of graupel and snow is centered roughly on the LCR (near 7 km AGL). Like the early stages of the break period simulation, when the LCR runs through the particle interaction region, cloud-scale charge separation is not favored.

At \( t = 80 \) min, shortly before the termination of the model run, little change took place in the model output.

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**Fig. 16.** Radar observation from the MIT C-band Doppler radar on 12 January 1990. (a) Vertical cross section through monsoon precipitation at 0340 UTC. (b) Horizontal map of reflectivity at \( z = 1.5 \) km. The dashed line indicates the cross section shown in (a). (c) As in (a) except for 1020 UTC. The dashed line indicates the cross section shown in (c). (d) As in (b) except for 1020 UTC. Reflectivity values are in dBZ. The location of the radar is denoted by the cross on (b) and (d).
fields compared to $t = 60$ min. The LCR was still located roughly through the center of the graupel and snow interaction region (near 7.5 km). Because the LCR was situated near the particle interaction region throughout the simulation, no pronounced cloud-scale charge separation was anticipated.

The total charge density at $t = 80$ min is shown in Fig. 20. As expected from the location of the LCR and the particle interaction region, and in contrast to the break period simulation, no organized charge structure was present (Fig. 20a). Analysis of the charge density on snow and graupel at $t = 60$, 70, and 80 min (not shown) indicated pockets of graupel (snow) with opposite charges that form and dissipate with time. No steady charge structure was simulated, and therefore, no long-lasting vertically oriented charge dipole was maintained. Figure 20b shows the vertical electric field structure at $t = 80$ min. Charge separation has occurred as shown in Fig. 20a but results in only weak electrification. Electric field values do not exceed 40 kV m$^{-1}$, well below that required to initiate lightning.

In general, this monsoon simulation did not predict the development of a persistent dipole or any other well-defined electrical structure, despite relatively large
values of cloud water, graupel, and snow. There are two reasons for this behavior. The primary reason is that the LCR was situated through the center of the graupel and snow interaction region. As the cloud water content changes in the interaction region, so does the graupel (snow) charging tendency. This prevents pockets of charge from becoming persistent features. For example, a region of graupel may charge negatively in accordance with the temperature and liquid water content at a particular time. However, because of the high temporal and spatial variability of the cloud water field, the cloud water content in the region of the negative graupel may change such that the graupel has a positive charge tendency, thus causing charge to be neutralized in the region. The second reason that inhibits the development of charge structures is the gravitational settling or convective transport of graupel (snow) across the LCR toward the graupel (snow) particle balance level. This causes the species to rapidly neutralize upon collision with oppositely charged species.

Fig. 18. Model results for the 12 January 1990 case at $t = 40$ min. The fields shown are (a) vertical velocity, contour interval of 2 m s$^{-1}$; (b) cloud water mixing ratio, contour interval of 0.4 g kg$^{-1}$; (c) graupel mixing ratio, at a contour interval of 0.7 g kg$^{-1}$; (d) snow mixing ratio, contour interval of 0.07 g kg$^{-1}$. The level of charge reversal is denoted by the thick black line as in Fig. 7.
c. **Comparison with observation**

The widespread precipitation, embedded convection, and well-defined radar bright band evident from the radar observations was rather well simulated by the model. The model-predicted radar reflectivity field is shown in Fig. 21 (at \( t = 80 \) min), which can be compared to the observed field shown in Fig. 16c. The largest reflectivities in the stratiform portion of the simulated cloud, 35–42 dBZ, are found near the 0°C level (5 km). In the convective cores, the reflectivity in the simulation is in excess of 50 dBZ, but these cores did not extend much above the 0°C level. These simulated values compared reasonably well to the observed values. The cloud-top heights were near 12 km both in the simulation and in the observations. At \( t = 80 \) min into the simulation, at which time the model-simulated storm could be considered “mature,” the model and observations seem to be in reasonable qualitative agreement.

A few comments regarding how the model simulations compare with microphysical observations of monsoon precipitation are in order here. Three studies (Churchill and Houze 1984; Houze and Churchill...
1984; Gamache 1990) have reported in situ microphysical measurements taken during flights through mesoscale convective systems similar to the monsoon convection described and modeled here. The measurements were taken aboard the National Oceanic and Atmospheric Administration (NOAA) WP-3D aircraft equipped with Particle Measurement Systems (PMS) probes during the Winter Monsoon Experiment (WMONEX). The flights were conducted between 0° and −25°C in the Gamache study and near 8 km AGL (about −17°C) in the Houze–Churchill studies. In addition to ice particle imaging, vertical velocity and cloud liquid water were also recorded.

In the convective regions of the observed systems, a peak vertical wind speed of 17 m s⁻¹ was encountered but generally updrafts were less than 10 m s⁻¹. The simulated vertical velocity is consistent with these observations. The model-predicted peak velocity was near 18 m s⁻¹ but occurred below 8 km (the flight level for the Houze–Churchill studies). Generally, simulated vertical velocities near the 8-km level were less than 10 m s⁻¹. Cloud water contents never exceeded 0.2 g m⁻³ according to the aircraft observations, even in convective regions. The values predicted by the model, at the 8-km level (flight level), are consistent with the airborne observations. Above 8 km, the simulated cloud is nearly glaciated, while abundant water is present below about 7 km.

Despite the lack of supercooled water at flight level, most of the imaged particles appeared to be rimed as reported by Churchill and Houze (1984). At the time that the aircraft recorded the 17 m s⁻¹ updraft peak, soft hail (graupel) was observed hitting the windshield. Churchill and Houze concluded that the particles had been grown at lower levels where liquid water was presumably more abundant, and then carried aloft to flight level. Although the supercooled water contents and updraft velocities were larger than anticipated in this simulation, the predicted values appear to be consistent with the limited airborne observations of monsoonal convection.

8. Additional simulations and sensitivity studies

Two cloud simulations that represent low-CAPE (12 January 1990) and high-CAPE (19 January 1990) environments observed during the DUNDEE have been
discussed. In addition to these two case studies, we have carried out a simulation for a moderate-CAPE environment. We have also conducted several microphysical and lateral boundary sensitivity studies. The results of these simulations will now be briefly discussed.

a. Moderate-CAPE simulation

The 0000 UTC 30 November 1988 Darwin sounding (not shown) was characterized by north to northwesterly flow up to near 450 mb and westerly flow above. Wind speeds were less than 15 m s\(^{-1}\) throughout the depth of the troposphere. The CAPE was 700 J kg\(^{-1}\). The model was initialized with the 0000 UTC Darwin sounding with modifications similar to those described in the two previous case studies. A narrow convective line with an extensive area of stratiform cloudiness and precipitation was observed in the DUNDEE region on this day. Radar-observed cloud tops were near 17 km AGL, but reflectivity values greater than 20 dBZ were rarely observed above 5 km. Peak lightning rates were approximately 3 min\(^{-1}\).

The simulated mixing ratios of graupel and snow, the total charge density, and the vertical electric field are shown in Figs. 22a–d at \(t = 60\) min, after which the model run was terminated. As before, the level of charge reversal is shown as a thick black line. The fields are similar to the high-CAPE simulation with a few important exceptions. The major differences between this simulation and the break period (high CAPE) simulation is that in this simulation the cloud water (not shown), snow, and graupel mixing ratio values (Figs. 22a,b) are lower, the center of the charge dipole is located lower in altitude (Fig. 22c), and the charge centers are situated farther apart. Also, the magnitude of the electric field at \(t = 60\) min is slightly in excess of 115 kV m\(^{-1}\) (Fig. 22d), which is about half of what the break period simulation produced in 52 min. A time series of the magnitude of the vertical electric field for the high-, moderate-, and low-CAPE cases is shown in Fig. 23. A discussion of the interrelation of the three different CAPE simulations will be addressed in more detail in a later section.

b. Later boundary and microphysical sensitivity studies

Additional simulations for the high- and low-CAPE cases were conducted with expanded horizontal domains to examine the effects of the close proximity of cloudy regions to the lateral boundaries. In addition, all three cases were rerun with the graupel microphysical parameters made even more "graupellike" (density of 0.4 g cm\(^{-3}\) and size distribution intercept parameter of \(4 \times 10^{-2}\) cm\(^{-4}\)). The expanded domain cases compare favorably with the standard cases, indicating that the artificial lateral boundaries are not exerting a significant influence on the model solutions. The cases with modified graupel microphysics, however, do indicate a more pronounced effect although the cases follow the trends of the original set of experiments. The electrical development is slightly weaker in the high- and moderate-CAPE cases, while the low-CAPE case indicates stronger, but still weak, electrical development.

Another aspect of the electrical development confirmed by the sensitivity tests is that the vertical electric field is dominated by the negative component except for the very earliest stages in the high-CAPE cases. The low-CAPE case displays no pronounced preference for either component of the vertical electric field, and the moderate CAPE was usually dominated by the negative component. During periods of rapid electrical buildup, negative vertical electric fields dominated in all cases.

9. Discussion and summary

Three cloud simulations that represent low-CAPE (12 January 1990), moderate-CAPE (30 November 1988), and high-CAPE (19 January 1990) environments observed during the DUNDEE have been discussed. Comparison of the three simulations indicates a correlation between CAPE and the strength and rate of electrification (Fig. 23). This is consistent with observations from the DUNDEE that also showed a correlation between CAPE and electrification, as inferred through lightning occurrences (Rutledge et al. 1992). Based on the simulations, there appear to be many critical factors that account for the relationship between CAPE and storm electrification.

a. The particle interaction region and the level of charge reversal

As a consequence of the large vertical velocity differences, with the largest vertical velocities in the break period case carrying graupel higher into the storm, the mean height of the graupel–snow interaction region was highest in the high-CAPE simulation and lowest in the low-CAPE monsoon simulation. There was also a correlation between the juxtaposition of the LCR and particle interaction region and the CAPE.

In the high-CAPE simulation, the particle interaction region was above the LCR. All the graupel in the region acquired negative charge and the snow positive charge. Based on the simulations, this is the most efficient charging scenario, with net negative charge on the graupel near the graupel balance level and net positive charge on the snow near the snow balance level. In the low-CAPE monsoon simulation, the vertical velocities were insufficient to carry a large fraction of the graupel above the LCR. As a result, the LCR was situated through the middle of the particle interaction region. This is a "worst case" charging scenario, since roughly half the graupel charged positively and half charged negatively. Charge cancellation resulted when the grau-
pel migrated toward the graupel balance level. The moderate-CAPE model run simulated a graupel–snow interaction region that was mostly (but not entirely) above the level of charge reversal. Consequently, most of the graupel charged negatively, and even after charge cancellation near the graupel balance level, a net negative charge remained on the graupel.

Three charging scenarios associated with different levels of CAPE are conceptualized in Fig. 24. The first two, high and low CAPE, were discussed above. In the third scenario, the CAPE is near zero, a situation that might be found in the stratiform region of a midlatitude mesoscale convective system (Rutledge 1991). In this case, the vertical velocities are small enough (generally less than 1 m s\(^{-1}\)) that the graupel and snow interaction region and the respective particle balance levels remain almost entirely below the level of charge reversal. Consequently, the graupel acquires charge of one sign (which would be positive according to NIC laboratory studies), leading to the development of an inverted di-
pole. This is consistent with positive charge overhead and positive cloud-to-ground lightning that has been observed in these situations (Engholm et al. 1990). The modeling study of Rutledge et al. (1990) also predicted an inverted dipole in the stratiform region of a mesoscale convective system, thus promoting positive cloud-to-ground discharges.

b. The effect of cloud water on ice mixing ratios and the level of charge reversal

The larger vertical velocities resulting from larger values of CAPE produce higher mixing ratios of graupel, snow, cloud ice, and cloud water. The higher mixing ratios result in more numerous potential charge separation interactions between the graupel and snow or cloud ice. This is consistent with the CAPE—cloud mass—radar reflectivity scaling argument that was presented in Rutledge et al. (1992).

Larger values of cloud water have two other effects. First, it provides for efficient growth of ice species through accretion or acts as a vapor source for depositional growth. Second, and perhaps more importantly, the temperature at which charge reversal occurs decreases as liquid water content decreases. This is an important effect since it appears from the simulations that the farther the graupel and snow interaction region is from the level of charge reversal, the more effective cloud-scale charge separation is. In the case of the high-CAPE simulation, not only was the interaction region situated above the level of charge reversal but, as a consequence of large liquid water contents, the level of charge reversal was essentially shifted to higher temperature, or levels lower in the cloud, and the altitude difference between the LCR and the particle interaction region increased.

c. The location and separation of storm-scale charge centers

In the high-CAPE simulation, the pockets of charge that resulted in the normal polarity dipole were found well above the freezing level as the positive charge center was at −20°C or colder, and the negative charge center was at −12°C or colder. This is reasonable since the updrafts were strong enough to keep graupel and snow from falling (relative to the ground), thereby shifting the particle balance levels (for graupel and snow) well above the LCR. The negative and positive charge centers in the moderate-CAPE monsoon occurred at slightly higher temperatures compared to the situation in the break period simulation. This had little effect on the positive charge aloft, but the negatively charged graupel was close enough to the melting level that some of the charge was transferred to rain through melting. Furthermore, the updraft was weak enough to allow the rain to fall, and the initially compact negative charge center was transformed into an elongated core of charge that extended through the depth of the downdraft by t = 60 min. The effective center of the negative charge fell toward ground with the rain, and as the distance from the positive charge center increased, the strength of the vertical electric field was compromised. This was another factor in limiting the electrification of the moderate-CAPE monsoon. Since the low-CAPE monsoon simulation produced no long-lived electrical structure, a discussion of the location and separation of storm-scale charge centers is meaningless.

d. The effect of ion attachment

In both the moderate and high CAPE cases, ion attachment decreased the effectiveness of the NIC mechanism. The favored region of ion attachment was at cloud top, above the positive charge center. A well-defined negative screening layer developed in both simulations and primarily neutralized positively charged cloud ice. Had ion attachment not been permitted, it is likely that a stronger positive charge region would have developed and subsequently produced stronger electric fields.

e. Summary

Three simulations of the electrification of tropical convection provide insight into the physical mechanisms that connect environmental CAPE and the degree of storm electrification. The most critical factor seems to be the juxtaposition of the graupel and snow interaction region and the level of charge reversal. If the level of charge reversal is not found in the interaction region, noninductive charging will proceed at a rate that is determined primarily by the number of graupel and snow interactions, temperature, and cloud water content. This situation was realized in the 19 January 1990 break period simulation.
Although the storm-scale electrification varied greatly from one simulation to another, electrification by means of the noninductive charging mechanism was active in all the simulations. That is, charge transfer at the hydrometeor scale occurred in all three simulations.

Acknowledgments. This research was supported by the National Science Foundation under Grant ATM-9015485 from the Physical Meteorology and Mesoscale Dynamics Programs and Grants ATM-8821119 and ATM-9200698 from the Physical Meteorology Program. We appreciate the careful typing of this manuscript by Lisa Berg. The reviewers are also acknowledged for their helpful comments, especially Dr. Joanne Simpson and Dr. Brad Ferrier. Prof. Earle Williams also provided helpful comments on an earlier version of the manuscript.

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