Electrification of New Mexico Thunderstorms

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ABSTRACT

The authors use a numerical model of early electrification in thunderstorms, together with observations of a series of summer thunderstorms in New Mexico, to understand the roles of certain environmental factors in determining thunderstorm electrification. The results suggest that development of lightning depends sensitively on τ_{cz}, the duration of significant updrafts in the charging zone between $-10°$ and $-25°C$. Model tests suggest that τ_{cz} is maximized for moderate cloud-base forcing and that τ_{cz} depends on environmental parameters in predictable ways. These results are used to investigate the relationships between lightning evolution and several functions of environmental quantities that have been suggested as lightning predictors.

1. Introduction

There have been a number of suggestions in the recent literature linking thunderstorm electrification to measurable environmental parameters (Goodman 1990; Williams and Renno 1991; Price and Rind 1992). These relationships are largely empirical; they have been proposed on the basis of observed climatological correlations. In this paper we examine these proposed lightning predictors for a very restricted group of storms, to investigate the physical mechanism(s) responsible for the observed correlations between predictor function and ensuing electrical behavior. We also examine the role of cloud-base forcing, which is not, in general, predictable from soundings alone, in modifying thunderstorm electrification.

We begin by summarizing the salient features of early thunderstorm evolution as they are commonly understood. In the most commonly observed thunderstorms, early electrification (that which precedes the first occurrence of lightning) can be observed in as little as some 5–10 min after cloud growth begins (Krehbiel 1986). A negative charge center develops in a region we shall henceforth refer to as the "charging zone" (CZ), at about $-10°$ to $-25°C$. There is a more diffuse positive charge zone above this CZ and sometimes a smaller positively charged region at lower altitudes.

We assume, based on field and model results (Gardiner et al. 1985; Dye et al. 1986; Norville et al. 1991; Ziegler et al. 1991), that the dominant charge transfer mechanism is that involving collisions between ice crystals and graupel, or soft hail pellets (Takahashi 1978; Jayaratne et al. 1983; Baker et al. 1987; Saunders et al. 1991) in the CZ. The instantaneous charge separation rate depends therefore on the liquid water content (LWC), the numbers and sizes of ice crystals and hail particles in the CZ. The LWC is determined in part by the strength and duration of the updraft below the charging zone. The crystal concentrations depend on the glaciation mechanism(s) and the efficiency of circulation of ice from cold temperatures to the CZ, while the numbers of the graupel pellets depend on the numbers and sizes of supercooled drops, again a function of updraft strength just below the CZ. The sizes of the graupel pellets depend in part on w_{cz}, the updraft velocity in the CZ (Ziegler et al. 1986). The crystal sizes depend on the time they spend in the CZ and thus also on w_{cz}.

The total charge separated (and hence the electric field strength) at any time t^* depends on the charge separation rate and the duration of charging prior to t^*. Thus, we expect the duration of significant updrafts in the CZ to be an important factor in the electrical development and activity of thunderstorms.

2. Environmental parameters as predictors of lightning

A number of semiempirical studies (Goodman 1990; Williams and Renno 1991; Price and Rind 1992) have established correlations linking environmental parameters (referred to hereafter as "predictors") to the development of lightning. The predictors we consider here are the following.

1) The convectively available potential energy (CAPE) of a parcel raised from cloud base z_{cb} to cloud top z_{ct}:

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TABLE 1. Statistics on storms and soundings: Socorro (1984).

Date (1984)	Cloud-base pressure (mb)	CAPE (J kg^{-1})	θ_w (K)	LI (°C)	Ri	$Z_{ct,max}$ (km)	Lightning events (min^{-1})	Comments
23 July	655	234	294.5	0.0		~9	None	Very weak electrification
27 July	644	43	294.3	−1.2		~10.5	>3	
29 July	660 (?)	208	294.8	−1.6		9	None	Very weak electrification
31 July	655	924	294.2 (293.5)	−1.6	12	14	>100 (~1.6)	
1 August	638	854	294.5	−3.0		9.5	None	Storm did electrify
2 August	631	441	294.7	−5.8	19	13.5	>10 (~0.4)	
3 August	614	412	294.9	−3.8	8.2	12	6 (~0.4)	
6 August	660 (?)	322	295.2	−2.7		~8	None	
7 August	670	1463	296.0	−2.9		11.5	?	
13 August	693	321	294.8	−2.0	4.4	12	18	
14 August	676	935	295.8	−2.3		9.5	None	Moderate (?) electrification observed
15 August	693	2038	295.3 (294.2)	−1.1	9.2	11	6 (~0.2)	Inverted dipole (initially)

$$\mathrm{CAPE} \equiv \int_{z_{cb}}^{z_{ct}} g \frac{(T_{v,ad} - T_{v,env})}{T_{v,env}} dz,$$

where $T_{v,env}$ and $T_{v,ad}$ are the virtual temperatures of the environment as computed from soundings and along a reversible moist adiabat going through cloud base at z_{cb}. The ice phase is not included in this calculation.

2) The wet-bulb potential temperature at the surface, θ_w $(z = 0)$ (Williams and Renno 1991).

3) The maximum observed cloud-top height, z_{ct}(max). Price and Rind (1992) make use of an empirical relationship between maximum total flash rate F and observed maximum cloud-top height:

$$F = 3.44 \times 10^{-5} Z_{ct,max}^{4.9}.$$

Noting this, we examine the use of cloud-top height as a predictor for electrification.

4) The shear Richardson's number Ri for the cloud layer, defined as

$$\mathrm{Ri} \equiv \frac{\mathrm{CAPE}}{(u_{ct} - u_{cb})^2/2},$$

where u_{cb} and u_{ct} are the horizontal wind velocities at cloud base and cloud top.

5) The lifted index LI (Goodman 1990) defined as the difference at 500 mb between the temperatures of the environment and of a parcel of air taken from the lowest 100 mb of the sounding (near ground) and raised adiabatically to 500 mb.

We examine here the physical mechanisms that result in the observed correlations of these five predictors, and the evolution of breakdown electric fields.

In addition to these environmental factors, the dynamic and thermodynamic properties of the air entering cloud base determine the ensuing cloud development. Cloud model studies consistently show that cloud dynamic and thermodynamic properties depend sensitively on the imposed forcing (thermal or mechanical) at cloud base. Presumably, the electrical development is also sensitive to this forcing. Since we have no data on the relationships linking lightning to the cloud-base forcing, we examine these relationships by means of a numerical model with which we have successfully simulated the development of electrification in several midlatitude continental storms (Norville et al. 1991). While the details of model-predicted relationships between forcing and ensuing cloud physical and/electrical properties depend on the model used, we expect that the general nature of these relationships is model independent. We apply the results of our model tests to the interpretation of the observations discussed herein.

3. The observations

We confine our attention to a set of 12 thunderstorms that occurred over a 3-week period in July and August of 1984 over the Magdalena Mountains in New Mexico. Some of the storms we discuss here have been discussed in detail in the literature (see, for example, Dye et al. 1989; Raymond and Blyth 1989; Blyth and Latham 1993). Because the circumstances surrounding cloud formation are somewhat similar for all the storms in the study, we hope to learn which small environmental changes play the biggest roles in determining the very large differences in observed electrical development. Our goals are to determine (a) which environmental conditions are most important in determining cloud electrification; (b) why these are important; (c) which of the aforementioned lightning predictors is most successful, and (d) why.

In Table 1 we summarize the observations of cloud structure and lightning development in the 1984 storms. We computed values for the predictors mentioned above in all cases for which we had sufficient data. The rest of this paper is devoted to an examination of these storms and the correlations shown in this table. As part of our study, we have simulated some aspects of storm development in the New Mexico environment using our cloud model, described in the next section. We use the results of these simulations to attempt to understand the relationships between predictors and lightning development shown in the table.

4. Model studies

Our model is based on that of Taylor (1989), modified by Norville et al. (1991). The model domain is axisymmetric. It consists of three communicating coaxial cylinders: inner and outer cloudy regions and an unsaturated environment. Entrainment and mixing between inner and outer cloud regions are parameterized as functions of turbulent kinetic energy (TKE). The model water and ice microphysics are explicitly calculated with 40 (80) water (ice) mass bins from 6.5×10^{-14} g to 2.2×10^{-4} (230) g. The microphysical processes included are growth by deposition, riming, and collection. All particles are assumed to be spherical. The ice-crystal density is set to 0.1 g cm^{-3}. The density of the riming particles is taken from Macklin (1962). Fall velocities for drops and solid ice particles are taken from Berry and Pranger (1974), while ice-crystal and graupel fall velocities are taken from Heymsfield (1978). The concentrations of ice nuclei are computed from the results of Fletcher (1962) with homogeneous freezing of all liquid at -40°C.

Noninductive charge transfer accompanying collisions of ice crystals and soft hail as documented in a range of laboratory conditions (Takahashi 1978; Jayaratne et al. 1983; Baker et al. 1987; Saunders et al. 1991) is assumed to be the sole source of charging within the model. There is no representation of lightning in the model, whose use is therefore limited to the study of early electrification.

We have compared the model-predicted microphysical and dynamical characteristics against observations from the New Mexico storms. The data consist of radar reflectivities, soundings, electric field measurements (ground and air based), and cloud physics measurements made during sailplane flights (e.g. Dye et al. 1989; Raymond and Blyth 1989; Blyth and Latham 1993). Although we have no in situ measurements of cloud microphysics from the 1984 field experiments, we use studies from other New Mexico storms (for example, Blyth and Latham 1993) in which at temperatures around $-10°$ to -15°C LWC was 1.75 g m^{-3} in updraft regions and approximately 0.75 g m^{-3}, with ice, in downdraft regions. Ice particles were first detected when the cloud-top temperature fell below approximately -10°C; small ice particles were observed in the updrafts with sizes indicating origins from about the -12°C level, and ice particles were observed in clouds whose tops never became colder than approximately -12°C. Model results are consistent with these observations as well as with observed radar reflectivities, cloud-top heights, and updraft velocities.

We use the model first to examine the possible chain of physical processes linking cloud-base parameters to the development of lightning. The air entering the base of a convective cloud does not always have the same thermodynamic and dynamic properties as the surrounding air. The differences between the temperature and/or velocity of the cloud-base air and those of the environment at the level of cloud base are generally referred to as "cloud-base forcing" parameters. For our purposes, cloud-base forcing can be of two types, defined in terms of the excess energy per mass of cloudy parcels at cloud base over the energy per mass of the environment at that level. There are two types of forcing.

1) "Kinetic" or momentum forcing:

$$E_{\mathrm{kin}}(\mathrm{J\ kg^{-1}}) \equiv \frac{w_{\mathrm{cb}}^2}{2},$$

where w_{cb} is the updraft velocity of air entering cloud base, assumed to have the thermodynamic properties of the environmental air at that level.

2) "Thermal" forcing:

$$E_{\mathrm{th}}(\mathrm{J\ kg^{-1}}) \equiv \int_{z_{\mathrm{cb}}}^{z_{\mathrm{ct}}} g\, \frac{(T_{v,\mathrm{ad}} - T_{v,\mathrm{env}})}{T_{v,\mathrm{env}}}\, dz,$$

where $T_{v,\mathrm{env}}$ and $T_{v,\mathrm{ad}}$ are the virtual temperatures in the environment and along a moist adiabat going through cloud base (where $T_{v,\mathrm{cb}} = T_{v,\mathrm{env}} + \delta T_v$) to z_{ct}. (That is, E_{th} is the CAPE of a parcel with the virtual temperature $T_{v,\mathrm{cb}}$ at cloud base.)

In the "real" world, of course, cloud forcing is some combination of these effects. If it appears that the two produce similar results (insofar as electrification is concerned), then, for our purposes, it is not so important to know how nature splits energy between them.

We ran our cloud model for a range of values of E_{th} and E_{kin}, leaving all other input quantities unchanged, to determine the effects of cloud-base forcing on subsequent electrification. The 2 August 1984 Socorro sounding (Fig. 1) was used for all model runs. The sounding had an inversion near 400 mb and was relatively dry with a moist layer between 650 and 450 mb. The CAPE of a parcel of air starting near cloud base, 625 mb, following a moist adiabat until the parcel was negatively buoyant, is 441 J kg^{-1}. We integrated the model forward 1900 s (by which time the cloud had either passed through a maximum in cloud-top height or reached a quasi-steady state); the imposed cloud-base anomaly was maintained for 1500 s. While the forcing was on, the environmental lapse rate below cloud base was constrained to be adiabatic and the sub-

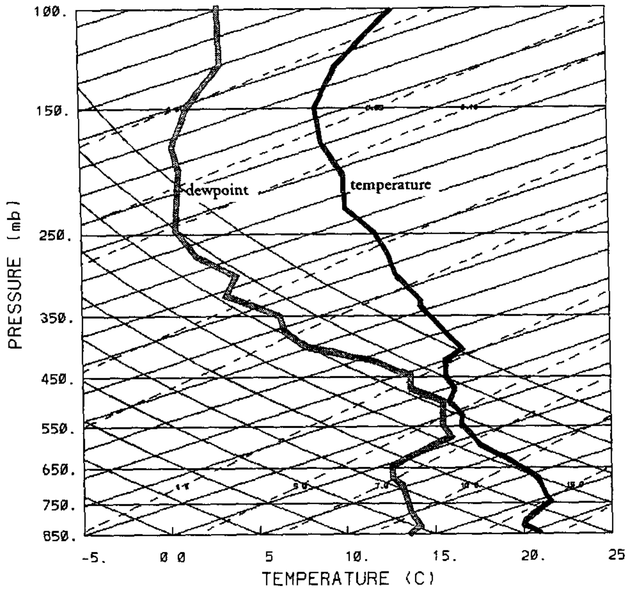

FIG. 1. Sounding taken from Socorro on 2 August 1984. Temperature and dewpoint temperature shown on a skew T–log$p$ plot.

cloud water vapor mixing ratio was held constant at the cloud-base value.

We examine the thermodynamic, dynamic, and electrical development of the model-produced storms for a range of the kinetic and thermal forcings. For model runs involving only kinetic forcing, the values for cloud-base velocity varied from 0 to 3.0 m s^{-1} ($0 \leqslant E_{\rm kin} \leqslant 4.5$ J kg^{-1}). Negative values for the vertical velocity gave the expected result—that is, no cloud formation. For model tests of $E_{\rm th}$, the cloud-base temperature was increased by up to 3.0°C and the cloud-base velocity was set to zero. This corresponds to a range $441 \leqslant E_{\rm th} \leqslant 2210$ J kg^{-1}.

a. *Effects of forcing on cloud dynamics and microphysics*

Figure 2 shows the maximum cloud-top height as a function of cloud-base forcing magnitude. For small values of either $E_{\rm th}$ or $E_{\rm kin}$, the developing cloud stops at the small inversion at roughly 400 mb. At somewhat higher forcing the cloud is able to "punch" through the inversion and continues to grow, at times up to the tropopause. With still greater forcing, the cloud initially grows well past the inversion but does not sustain growth, apparently because of increased entrainment (parameterized as an increasing function of the TKE), which causes evaporative cooling and hinders further development. Sometimes the cloud-top region continues to develop, probably driven by latent heat release during freezing of the remaining liquid drops, while there is sufficient evaporation in the cloud interior that downdrafts develop and the cloud decays. Maximum cloud-top height occurs at moderate forcing: $E_{\rm kin} = 1.1$ J kg^{-1} or $E_{\rm th} = 800$ J kg^{-1}.

Near the inversion level at about 400 mb, the buoyancy goes through a minimum for that forcing that just pushes cloud top through the inversion, but then the buoyancy rises to near neutral or positive with increased forcing. This increase is due to the fact that the inversion is near the level where glaciation becomes significant; the in-cloud temperatures just above this inversion range from $-10°$ to -15°C. Thus, for these New Mexico soundings the inversion plays an important role not only dynamically but electrically, as shown in the next paragraphs.

Note that the effects of kinetic and thermal forcing are similar in the upper regions of the cloud, including the charging zone. We can make a rough correspondence between them by examining the magnitudes of forcings that produce roughly similar cloud-top heights and maximum electric fields. We find that the effects of a change in thermal forcing $\delta E_{\rm th}$ are roughly the same as the effects of a change in kinetic forcing for $\delta E_{\rm kin} \approx 0.2\%\ \delta E_{\rm th}$ for the New Mexico storms.

b. *Effects of forcing on charging*

Since our charging parameterization is based on noninductive ice–ice collisions, it is a strong function of LWC and temperature (Saunders et al. 1991). Model runs produce negatively charged hail and positively charged ice crystals and snow for most of these test cases. It is possible, however, to cause the reverse by changing the microphysical parameters (discussed below). Strong charging takes place only for temperatures less than approximately -10°C. Thus, little, if any, charge generation occurs for clouds starting with

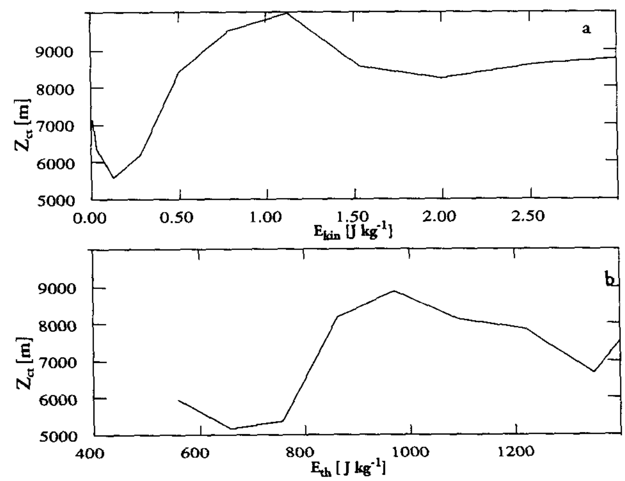

FIG. 2. Model-predicted maximum cloud-top height as a function of (a) $E_{\rm kin}$ and (b) $E_{\rm th}$ (0 km is ground level in Socorro, approximately 1800 m above sea level).

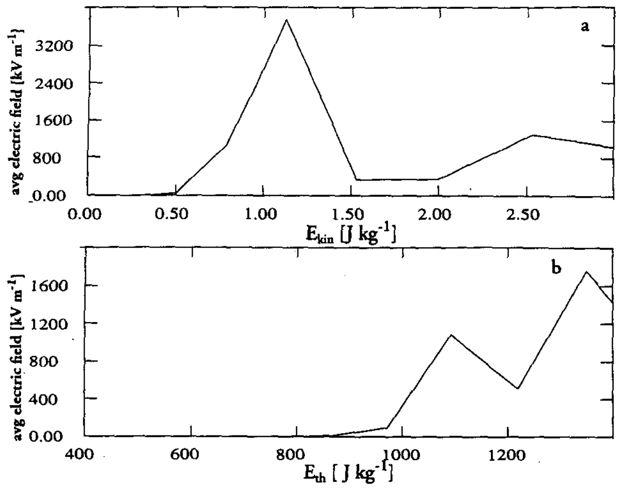

FIG. 3. Model-predicted magnitude of the average electric field between 8 and 9 km as a function of (a) E_{kin} and (b) E_{th}.

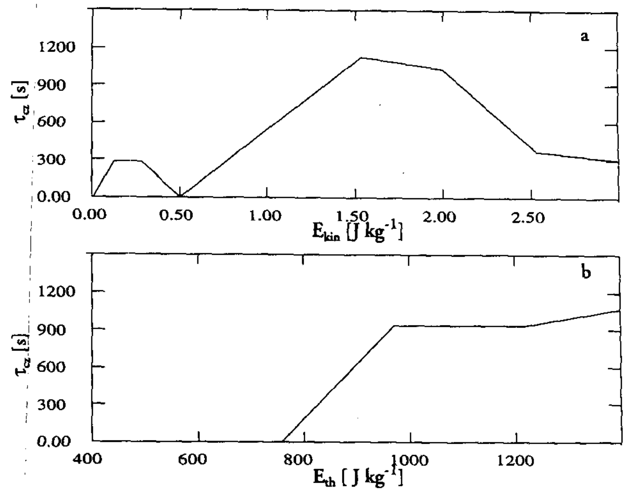

FIG. 4. Approximate amount of time that the updraft exceeds 2.5 m s^{-1} in the region of charging τ_{cz} as a function of (a) E_{kin} and (b) E_{th}.

small initial forcings, kinetic or thermal, simply because cloud top never reaches a point where appreciable amounts of ice form. A rapid transition occurs as the cloud forcing is increased and growth continues beyond the inversion to cooler temperatures.

As the magnitude of the forcing continues to increase, the cloud is unable to sustain development, as stated above. This prevents clouds driven by very high forcing from generating as much charge as those driven by more moderate forcing. The necessary conditions for charging are not maintained for high values of forcing (either kinetic or thermal); graupel does not remain suspended in the CZ long enough to separate substantial amounts of charge. Also, higher forcing places the region of maximum charge separation higher in the cloud. This is due to the effects on liquid water and graupel–ice distribution and the higher updraft velocities associated with more forcing.

Figure 3 shows the variation of the average electric field magnitude along the axis at 8–9 km as a function of forcing magnitude. The average electric field magnitude goes through a well-defined maximum for moderate forcing, both for kinetic and thermal forcing. [Note: our model has no mechanism for the removal of charge when the electric field exceeds the breakdown threshold of approximately 300 kV m^{-1}. Therefore, in the model the charge simply accumulates. We show the calculated maximum magnitude of the electric field (which can exceed the breakdown value by an order of magnitude) only as an indication of the dependence of the charging efficiency on forcing.]

The electric field magnitude seems to be related to the duration of appreciable updrafts in the CZ. We define τ_{cz} as the time for which $w_{cz} \geqslant 2.5$ m s^{-1}. Figure 4 shows the variation of τ_{cz} with forcing magnitude. In the model breakdown strength electric fields occur for $\tau_{cz} > 600$–900 s, and never for τ_{cz} under 600 s.

Figure 5 shows the magnitude of the average electric field in the CZ as a function of τ_{cz} for a model run (picked to best represent the 2 August 1984 storm). The abrupt rise of the electric field is consistent with that observed in thunderstorms (Krehbiel 1986).

Our model results suggest that for the most efficient charge separation the forcing must be strong enough to push the cloud to temperatures cold enough to generate sufficient ice for charge separation. On the other hand, the forcing must not be too high since this seems to hinder cloud development in the CZ and decrease τ_{cz}

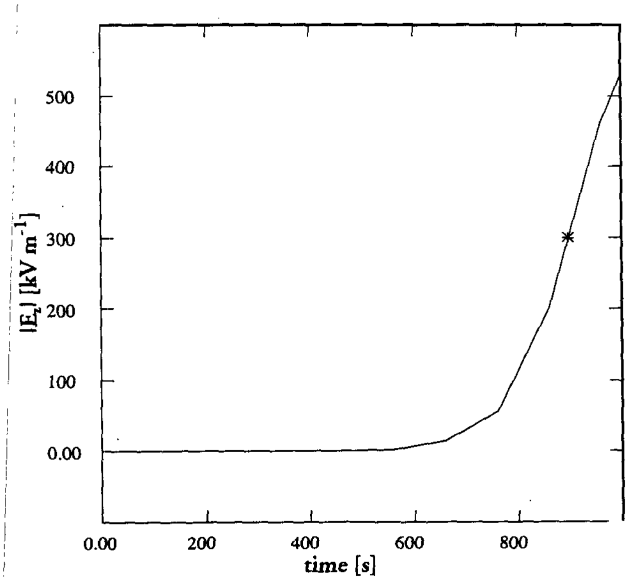

FIG. 5. Modeled magnitude of the electric field in the charging zone versus τ_{cz}. The single point represents the approximate point where breakdown would be expected to occur.

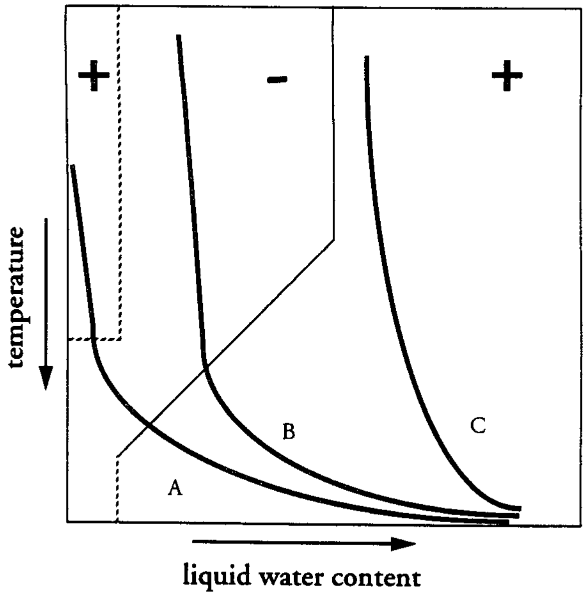

FIG. 6. Changes in charging regimes as a function of height and liquid water for changes in ice nuclei concentrations: (A) high concentration of ice nuclei, (B) intermediate values, and (C) low concentration of ice nuclei. The plus (+) and minus (−) are the signs of charge received by graupel (Saunders et al. 1991).

and the electric field (though the electric field remains above the threshold for breakdown).

c. Sensitivity of electrification to variations in ice microphysics

The preceding sections relate model-predicted electrical responses to forcings of various types and magnitudes with other microphysical quantities held constant. Changes in microphysical quantities, especially those that affect the ice particle distribution and LWC, also affect electrification in the model, since the parameterization for charge separation is quite sensitive to these.

Unfortunately, in situ observations of glaciation within clouds are limited. Therefore, we use the model simply to estimate the sensitivity of electrification to variations in some of the most uncertain ice phase procedures. Primary ice nucleation within the model is assumed to occur on ice nuclei whose concentration IN is

$$\mathrm{IN}(T) = \alpha e^{\beta(T_0 - T)},$$

where α and β are empirical constants found from laboratory experiments on freezing droplets and T_0 = 273.16 K (Fletcher 1962).

Increasing ice nucleus concentrations (by increasing α) lowers the altitude at which freezing becomes important. The early glaciation and release of latent heat of freezing allows the cloud to obtain maximum cloud-top height for smaller forcings; that is, the maxima shown in Fig. 2 move to smaller values of E_{kin} and E_{th}. If the ice nucleation concentrations are too high, all liquid water in the CZ is depleted and charging stops at relatively warm temperatures, decreasing the total charge separated.

Lowering the ice nucleus concentration slows glaciation, permitting liquid water to exist for longer periods of time and to reach colder temperatures. This reduction in latent heat release hinders cloud development, and the maxima shown in Fig. 2 move to larger values of E_{kin} and E_{th}.

Figure 6 is a qualitative illustration of how changes in ice nucleus concentrations can give rise to differing conditions within the CZ. The thick curves represent the liquid water as a function of temperature (or height) for (A) high ice nuclei concentrations, (B) intermediate values, and (C) low ice nuclei concentrations. Superimposed (light lines) are the charging regimes defined by Saunders et al. (1991). It is important to note that for the extremes represented by curves A and C, hail receives positive charge while inside the CZ. This would lead to an "inverted" dipole structure with positively charged hail residing below negatively charged ice crystals. The existence of convective storms with inverted dipoles is still open to debate. Indeed, the charge structure of this particular day, 2 August 1984, has been questioned by some (Moore et al. 1986; Winn et al. 1988).

5. Predicting lightning

A cursory look at Table 2 suggests that a CAPE threshold might be a good lightning predictor for these soundings. In Table 2 we show the results of attempts to predict the occurrence of lightning by assuming it is triggered for CAPE > 400 J kg^{-1}. This CAPE threshold is quite successful as a predictor, although there were a few days when it failed: the predicted CAPE on 27 July and 13 August fell below the threshold though those days produced electrified storms, and the 14 Au-

TABLE 2. CAPE threshold predictions.

Date (1984)	CAPE (J kg^{-1})	Is CAPE over threshold value?	Correct guess?
23 July	234	No	Yes
27 July	43	No	No
29 July	208	No	Yes
31 July	924	Yes	Yes
1 August	854	Yes	Yes (?)
2 August	471	Yes	Yes
3 August	412	Yes	Yes
6 August	322	No	Yes
7 August	1463	Yes	Yes
13 August	321	No	No
14 August	935	Yes	No (?)
15 August	2038	Yes	Yes

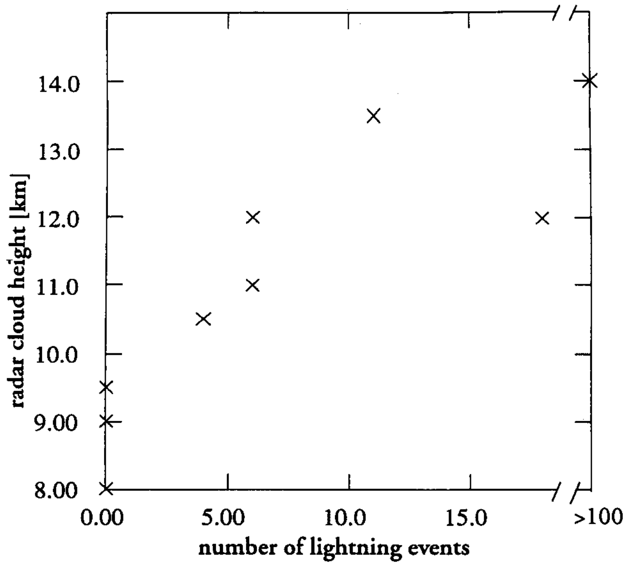

FIG. 7. Maximum radar cloud height versus number of lightning events for various New Mexico storms from 1984.

gust CAPE was above the threshold but did not produce lightning (though moderate electrification was noted). Stronger forcing could have caused the 27 July and 13 August storms to produce lightning. In fact, Sabreliner measurements on 13 August show air at cloud base was warmer than in the environment by about 1°C. On the other hand, there may have been insufficient forcing on 14 August.

The maximum observed cloud height z_{ct}(max) is quite well correlated with lightning, which occurred in all clouds whose tops rose above approximately 9.5 km and in no smaller clouds (see Fig. 7). In lower clouds, presumably, there is not enough (or no) ice present to separate charge simply because cloud top is not sufficiently cold. The use of z_{ct}(max) as a predictor requires that it be estimated in advance of cloud development. The cloud top estimated on the assumption of moist-adiabatic ascent from cloud base, $z_{ct,ad}$, is approximately as effective a predictor as CAPE. Neither θ_w nor LI were useful lightning predictors (see Fig. 8), and we do not have enough information on Ri to be able to evaluate it as a lightning predictor for the New Mexico storms, although we may expect low-Ri environments to better circulate ice to the charging region. The Price and Rind (1992) relationship between maximum flash rate and z_{ct}(max) fits the trend for the average flash rates in these storms.

To understand the success of the CAPE threshold and cloud-top height as lightning predictors, we performed a series of model tests in which the 2 August 1984 sounding was modified in order to increase the CAPE of a parcel starting from cloud base. This was done in three ways: (i) by decreasing the temperature in the 600–500-mb region, (ii) by decreasing the temperature in the 450–350-mb region (and thereby decreasing the strength of the inversion), and (iii) by a combination of the two above. For all of these, the cloud-base temperature was unchanged.

Our modifications to the sounding increased CAPE by as much as 200 J kg^{-1}. The model was run with a forcing that produced some charge separation but not enough to produce fields above the breakdown threshold when initialized with the unmodified sounding. Figure 9 shows that in each case an increase in CAPE increased the total charge separated and pushed the magnitude of the electric field to a value that would cause breakdown, although, as in the forcing study, "too much" CAPE decreased the lightning production

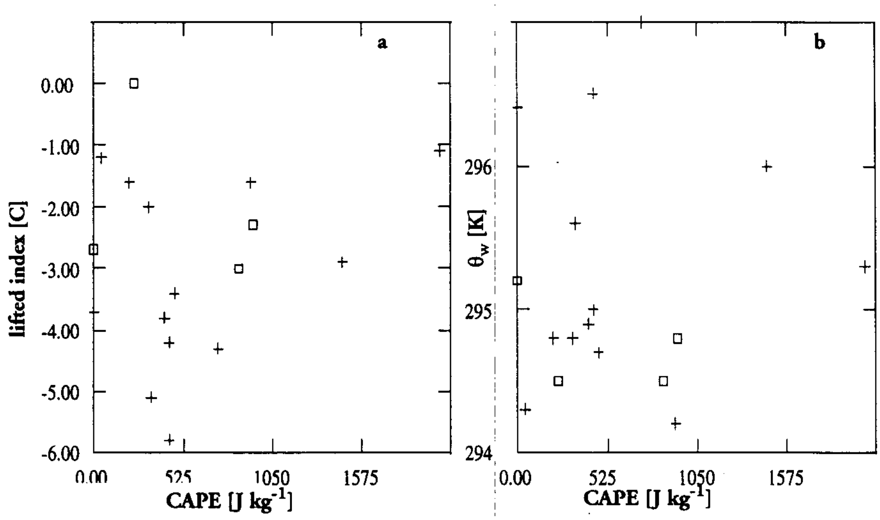

FIG. 8. (a) Lifted index and (b) θ_w versus CAPE for the New Mexico storms. The crosses represent days that produced lightning and the squares represent days that did not.

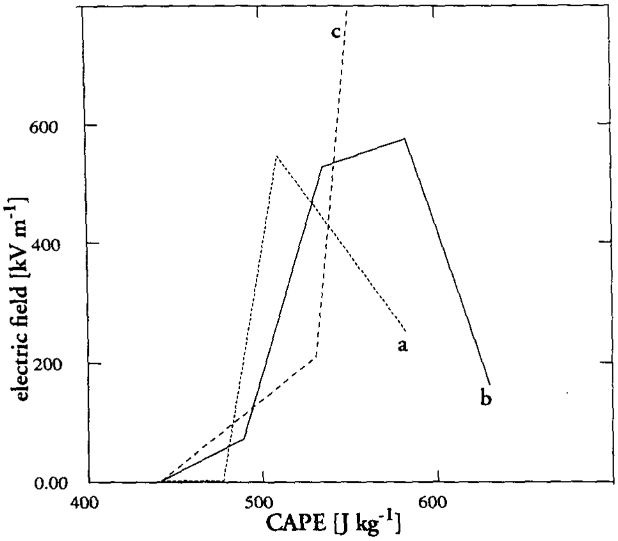

FIG. 9. Model-predicted maximum magnitude of the electric field as a function of CAPE for the modified soundings (a) modified in the 600–500-mb region, (b) in the 450–350-mb region, and (c) in both regions.

efficiency somewhat. In cases in which the inversion strength was reduced, water was transported above $-10°C$ more easily, and more charge was separated than for the unmodified sounding. When the CAPE in the lower portion of the cloud was increased, the amount of charge separated was also increased, though not as much as the previous case. Although higher CAPE in the lower portion of the cloud generated stronger updrafts and more water was transported past the inversion, the resulting increases in TKE and entrainment in the lower levels prevented large increases in charging rate; τ_{cz} also went through a maximum and then decreased as CAPE was increased, lending support to the idea that it is important in determining the amount of charge transferred.

6. Conclusions

The attempt to predict the future electrical status of a storm from one environmental quantity seems quite optimistic, given the large amount of potential variability within nature. However, the combination of observed cloud behavior and modeled results suggest that CAPE is a good lightning predictor for the New Mexico storms and it appears that this is because CAPE is highly correlated with τ_{cz}. (CAPE does work well as a predictor for eventual storm electrification; however, it does not work well as a predictor for the number of lightning events; see Fig. 10. We do not have enough information to comment on its correlation with flash rate, the quantity commonly referred to in the literature.) In contrast with some previous studies (Goodman 1990; Williams and Renno 1991), we did not find good correlations between θ_w and LI and subsequent lightning production. Both LI and θ_w are evaluated from surface and boundary layer conditions and our results would suggest that the air at the surface (at least for these storms) is not the air that enters cloud base; rather, cloud-base air typically has a mixing ratio equal to the average mixing ratio in the layer between 700 and 800 mb and a potential temperature of order 1°C warmer than that in this layer (Raymond et al. 1991). Observed maximum cloud-top height was strongly correlated with electrical development, and the predicted cloud-top height was as effective a lightning predictor as CAPE. Lightning occurs only if, either because of CAPE or through forcing, the storm is able to sustain high updraft speeds in the CZ long enough for substantial charge separation. Our model studies at high forcing magnitudes must be viewed as suggestive only, because we have no representation of lightning. They suggest, however, that for clouds forced too strongly, entrainment decreases τ_{cz} and lifts the charging zone to colder temperatures.

This study suggests that in order for thunderstorms to produce lightning, sustained updrafts within the CZ must exceed some threshold, w_{min}, for a time τ_{cz}. For the New Mexico storms, $w_{min} \approx 2.5$ m s^{-1} and $\tau_{cz} \approx 600$ s. This may be diagnosed in part from z_{ct} and CAPE as follows:

1) Predicted cloud-top height must exceed a threshold (9.5 km for the New Mexico storms) (although the model suggests that it is possible for z_{ct} to rise above this but for τ_{cz} to be too short to produce lightning).

2) The CAPE of the environment must be greater than some threshold amount; CAPE > 400 J kg^{-1} for the New Mexico storms.

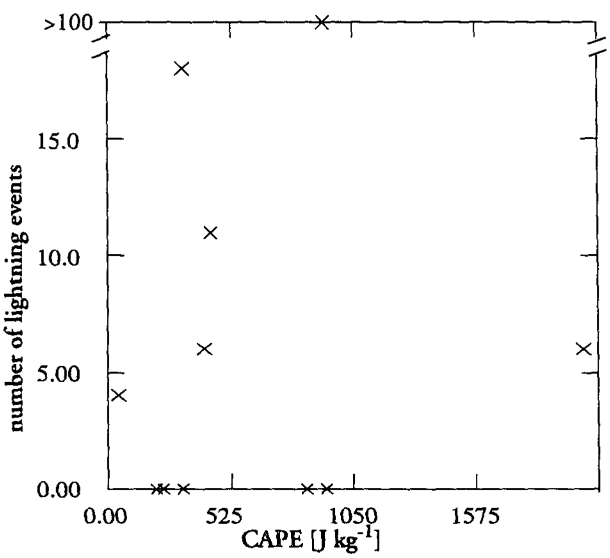

FIG. 10. Observed number of lightning events versus CAPE for the New Mexico storms.

The extent to which these quantitative statements are valid in other environments is open to question, although we predict the general statements presented here are valid for other convective thunderstorms.

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