LIGHTNING-RAINFALL RELATIONSHIPS IN AN ISOLATED THUNDERSTORM OVER THE MID-ATLANTIC STATES

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Abstract

Temporal and spatial relationships between cloud-to-ground (CG) lightning and precipitation were examined for an isolated nocturnal thunderstorm over the mid-Atlantic states. The lightning flash density field was compared to the rainfall pattern. Additionally, the volumetric and spatial distribution of rainfall were related to the concentration of CG lightning strikes. Also, the peak occurrence of CG lightning strikes within 10 km, 20 km, and 30 km of the National Weather Service forecast office at Sterling, Virginia was compared to the amount and time of the greatest rainfall and rainfall rate.

The maximum rainfall coincided well with those areas that received the highest concentration of CG lightning strikes. The greatest concentration of strikes (57% of the total storm CG strikes) produced just over half of the total volumetric precipitation over only 16% of the area that received rainfall. The heaviest rainfall on station began just after the 5-min CG lightning peaked within the 10 km, 20 km, and 30 km radii of the station. The greatest rainfall rate was recorded in the 5 to 40 min period following the peak in the 5-min CG lightning on station.

1. Introduction

There is an intrinsic relationship between lightning and precipitation which emanates from charge separation mechanisms. When water exists below freezing in multiple phases, vertically developing clouds tend to become very thermoelectrically active (Stow 1969). In all probability, several different charge mechanisms act simultaneously to achieve cloud electrification (Latham 1981; Pierce 1986). Workman and Reynolds (1949) found that precipitation was an important part of the electrification process. They observed that cloud-to-ground (CG) lightning strikes were generated from a developing cloud a few minutes after precipitation appeared from the base of the cloud. Additionally, Reynolds and Brook (1956) observed that precipitation was a necessary, but not sufficient, condition in thunderstorm electrification. The presence of a detectable radar echo alone did not automatically result in electrification unless there was rapid vertical development. Likewise, Lhermitte and Williams (1984) found that convective development and the subsequent growth of the radar echo were correlated with bursts of electrical activity.

Moore et al. (1962) examined the enigmatic relationship that exists when rain gushes (rapid discharges of rain) were observed following a lightning flash. Their findings implied that lightning may possibly cause the rain gush by greatly enhancing the coalescence of cloud droplets. Additionally, Moore et al. (1964) observed that after a CG lightning strike from an area in a weak radar echo, the area sometimes intensified rapidly. This was suddenly followed by a rapid discharge (gush) of rain or hail. Szymanski et al. (1980) observed an intra-cloud (IC) discharge that passed through a relatively weak and dissipating radar echo. Afterwards, the echo promptly began to regenerate and intensify in the same volume through which the lightning passed, apparently the result of rapidly growing precipitation particles.

Kinzer (1974) used sferics (electromagnetic signals from lightning) to correlate CG strikes with radar reflectivity for thunderstorms in Oklahoma. His results suggest that the areas of greater reflectivity were likely regions of greater CG lightning frequency and that, on the average, the lightning increased rapidly with an increase in the radial depth of reflectivity. Furthermore, there was a disproportionate increase in CG strikes as the amount of radar-estimated rainfall increased. Battan (1965) examined the relationship between rainfall and CG lightning frequency for thunderstorms in Arizona. On days when it rained heavily (mean rainfall per day per rain gage exceeded 0.1 in.) there was an average of 56 times more lightning strikes than on those days with light rain. That is, the greater the rainfall from convection over the area of study, the greater the CG strikes. Furthermore, results indicated that the rainfall at the ground and the frequency of CG strikes increased as the number, size and duration of the convective clouds increased.

Reap and MacGorman (1989) examined a large number of CG lightning strikes over two warm seasons in Oklahoma. They found a high correlation between the peak CG activity and echo intensity, especially for negative discharges to ground. Similarly, Rutledge and MacGorman (1988) collected data for the entire life of a midwestern Mesoscale Convective System (MCS). The peak in negative CG activity was well correlated with the maximum in convective rainfall, while the peak in positive CG lightning was in agreement with the maximum in the trailing stratiform rainfall.

Nielsen et al. (1990) observed a substantial lightning-rainfall correlation during the life (11 hrs) of an MCS over the southern Great Plains. The total (positive and negative) CG lightning rates closely followed the trends in convective rain flux during the entire event. Similarly, Goodman (1990) observed a high correlation between the rain flux and CG flash rate for an MCS in the Tennessee Valley. In addition, Goodman (1990) presented a comparison of the CG lightning and rainfall for the life-cycles of 10 Mesoscale Convective Complexes (MCCs; Maddox 1980). A substantial correlation existed between the highest CG lightning rates and rainfall approximately 2 hrs prior to MCC maturation. On the average, MCCs produce their heaviest rainfall just prior to MCC maturation (Kane et al. 1987).

Grosh (1978) examined the relationship between the average radar echo volume, echo height, rain flux, rainfall rate, and lightning flash rate. For a single storm, he found that there was a strong relationship between the peak rain flux, echo growth, and the lightning frequency. However, the maximum rainfall rate lagged the peak in flash rate by several minutes. Likewise, Goodman et al. (1988) found a lag of a few minutes between the peak flash rate (total IC and CG lightning) and the maximum rain flux and rain rate in a microburst-producing storm over the southeastern United States. However, the peak flash rate was well correlated with the maximum storm mass, vertically integrated liquid (VIL), echo volume, and cloud height. Ellison (1992) found the lag time between the peak CG lightning rate and the maximum rainfall rate for a convective system in New Mexico to be as much as 45 min. Buechler et al. (1990) investigated 21 small thunderstorms over the southeastern United States and found a high correlation between the peak 10-min CG lightning rate and the maximum rain flux. However, in one storm there was a lag in the maximum rain flux by 10 min. This temporal lag in precipitation is similar to findings of Piepgrass et al. (1982) in which the peak rainfall followed the peak in total flashes (IC and CG) by 5 to 10 min. Similarly, Shackford (1960) suggested a temporal lag of about 3 to 4 min between heavy rain arrival at a particular station and the first lightning strikes within a one-mile radius of the station.

On 3 June 1991, precipitation echoes first appeared (estimated from radar reports) about 0245 UTC over north-central Maryland. The convective system then intensified as it propagated southward into Virginia passing over the National Weather Service Forecast Office (NWSFO) at Sterling, Virginia (WBC) and Dulles Airport (IAD). This investigation examines the lightning-rainfall relationships of this isolated slow-moving nocturnal thunderstorm. It was characterized by a slow southward propagation which resulted in a localized precipitation maximum of 11 cm (over 4 in.). The temporal and spatial characteristics of the attendant CG lightning were examined in relation to the precipitation field and the amount of rainfall received on station. Unlike previous studies, the CG lightning flash density field was compared to the precipitation field. The synoptic environment is presented as well as several convective and kinematic variables.

2. Data

The lightning data were obtained from the National Lightning Detection Network (Orville et al. 1983; Orville 1991). The lightning detection system uses a series of magnetic direction finders spaced throughout the conterminous United States to sense CG lightning strikes. The detection efficiency is approximately 80%. The polarity, location, time, number of return strokes, signal strength, peak current and number of strikes can be discerned.

The isohyetal pattern presented was constructed from various rainfall reports, which included National Weather Service (NWS) observers, first order reporting stations (i.e., IAD [Dulles Airport], DCA [National Airport]), cooperative observers, NWS/National Meteorological Center high density precipitation reports, and even reports furnished by several fire departments. The rainfall reports were subjectively analyzed to obtain the precipitation field.

3. The Synoptic Situation

At 0000 UTC 3 June 1991, a weak surface pressure trough was positioned from coastal Delaware and Virginia northwest across southwest Pennsylvania and into central Ohio (thick dashed line in Fig. 1). By 0600 UTC, the western portion of the surface trough had moved south and extended through southern Ohio and northern West Virginia, while the eastern section moved southwest and became oriented southeast to northwest across Virginia and Maryland (dotted line in Fig. 1). Meanwhile, at 0000 UTC an equivalent potential temperature (Theta-E) axis at 850 mb extended west to east just south of the surface trough and intersected it across the Delmarva peninsula (Fig.1). Also at 850 mb, the mixing ratio was at a maximum (10 g kg^{-1}) across northern Virginia and central Maryland (Fig. 2). Similarly, an axis of 850 mb moisture convergence extended across the Mid-Atlantic with a maximum over southwest Virginia and southern West Virginia (Fig. 2). At 500 mb (Fig. 3), the mean ridge was over the Mississippi Valley and extended from Minnesota southward into Arkansas. Meanwhile, a short wave trough was propagating across the northeastern United States with 30-m height falls extending from Albany, New York south to Atlantic City, New Jersey. At 300 mb, (Fig. 4) a 26 m s^{-1} to 36 m s^{-1} (50 to 70 kt) jet stretched from the Great Lakes through northern Virginia and into New England. Assuming little change in the upper-level wind field from 0000 UTC to 0300 UTC, the convection developed just on the south edge of the strong height gradient (jet) and propagated southward in a diffluent upper-level flow. The 850-500 mb thickness pattern (not shown) extended from northwest to southeast and was also diffluent over Virginia.

Surface dewpoints ranged from 18.3°C (65°F) to 21.1°C (70°F) across Virginia and Maryland. Atmospheric stability parameters[1] based on the upper-air sounding at IAD (not shown) were characteristic of relatively slow moving single or multi-cell thunderstorms. Convective Available Potential Energy (CAPE) was 1218 J kg^{-1}. This amount of buoyant energy was just below the lower threshold for moderate instability (1500 J kg^{-1} to 2500 J kg^{-1}) defined by Weisman and Klemp (1986). The atmosphere was conditionally unstable with a Lifted Index (LI) of -4, K-Index 31, and Total Totals of 45. The precipitable water was 4.17 cm (1.64 in.), indicative of significant mean atmospheric moisture.

The thunderstorm originated over north-central Maryland around 0245 UTC. Assuming little change in the 850-mb Theta-E field, this was approximately near the intersection of the 850-mb Theta-E axis and the mean position (between the 0000 UTC and 0600 UTC) of the surface pressure trough shown in Fig. 1. The lowest 2 km of the atmosphere were characterized by an average dewpoint of 14.8°C and an average Theta-E of 341°K. Storm motion was from 347° at 4.1 m s^{-1}. The mean wind through the lowest 2 km of the atmosphere was from 163° at less than 1 m s^{-1}. Therefore, the mean storm inflow through this same layer was 166° at 5.1 m s^{-1}. The weak low-level inflow was directly opposite in direction to the storm's propagation, which should have provided for optimum moisture influx into the storm.

4. CG Lightning-Rainfall Relationship

The CG lightning strikes first started in north-central Maryland at 0350 UTC 3 June 1991 (Fig. 5). The system then propagated southward into Virginia, passing over the NWSFO at Sterling, Virginia and Dulles Airport (IAD). The convective system continued southward and crossed the Potomac River north of Quantico, Virginia (NYG) and finally dissipated over southern Maryland. There were a total of 391 CG strikes recorded from the convective system between 0350 and 0831 UTC. All strikes were negative (lowered negative charge to ground) except two. Typically during the warm season, negative CG strikes greatly outnumber positive

[1]Convective parameters were calculated by using the Skew-T/Hodograph Analysis and Research Program (SHARP; Hart and Korotky 1991).

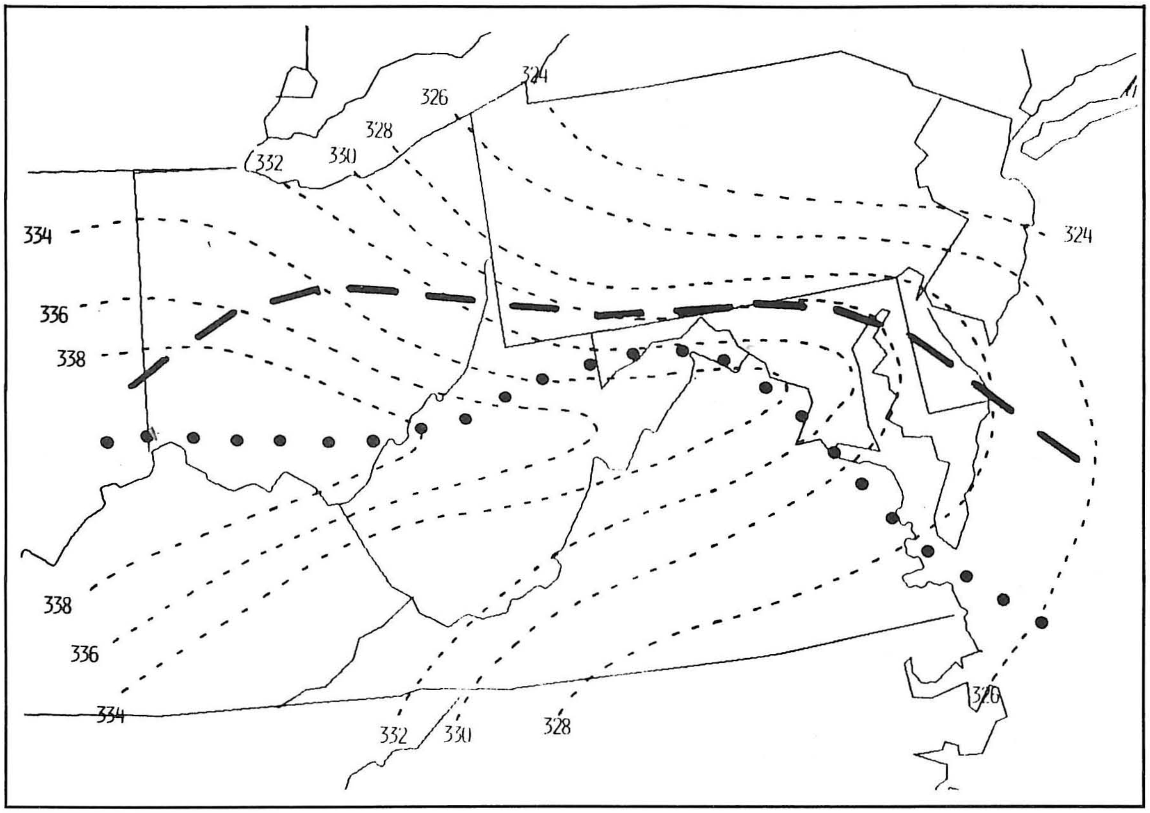

Fig. 1. Surface trough position at 0000 UTC 3 June 1991 (thick dashed line) and at 0600 UTC 3 June 1991 (dotted line) with the 850-mb Theta-E (deg K; thin dashed lines) at 0000 UTC superimposed.

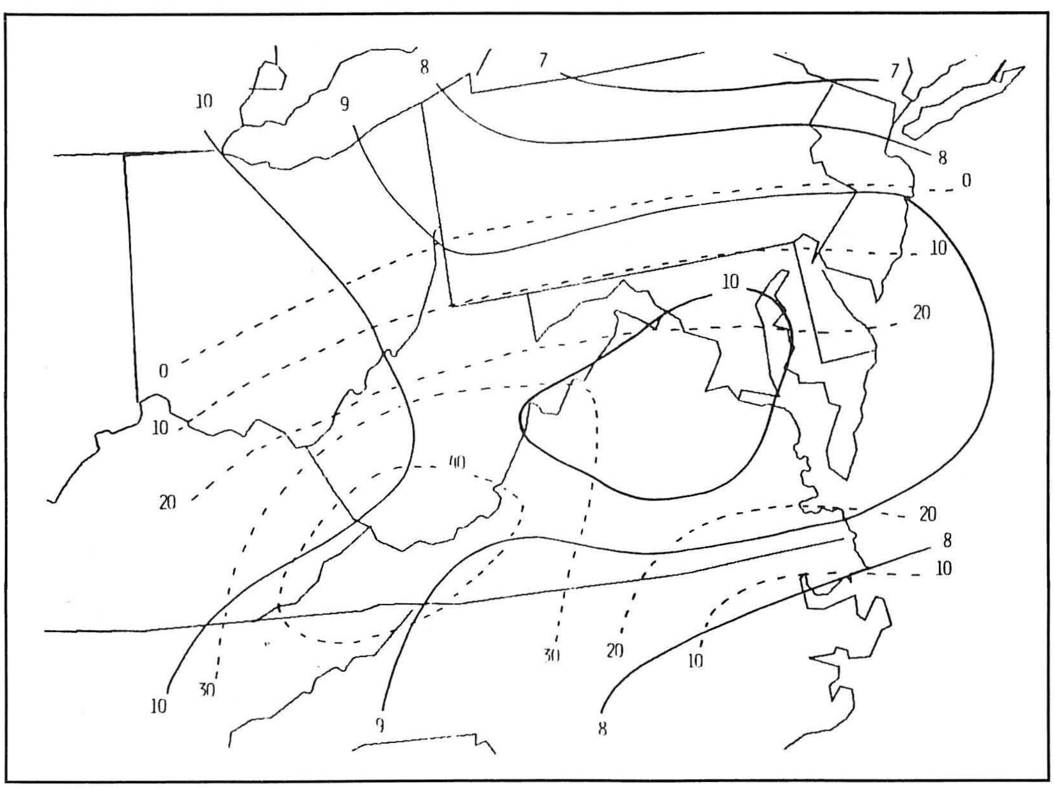

Fig. 2. 850-mb mixing ratios (g kg^{-1}; solid lines) and 850-mb moisture convergence (g kg^{-1} hr^{-1} $\times$ 10; dashed lines) for 0000 UTC 3 June 1991. (For details on the acquisition of the 850-mb moisture fields, see Foster, 1988.)

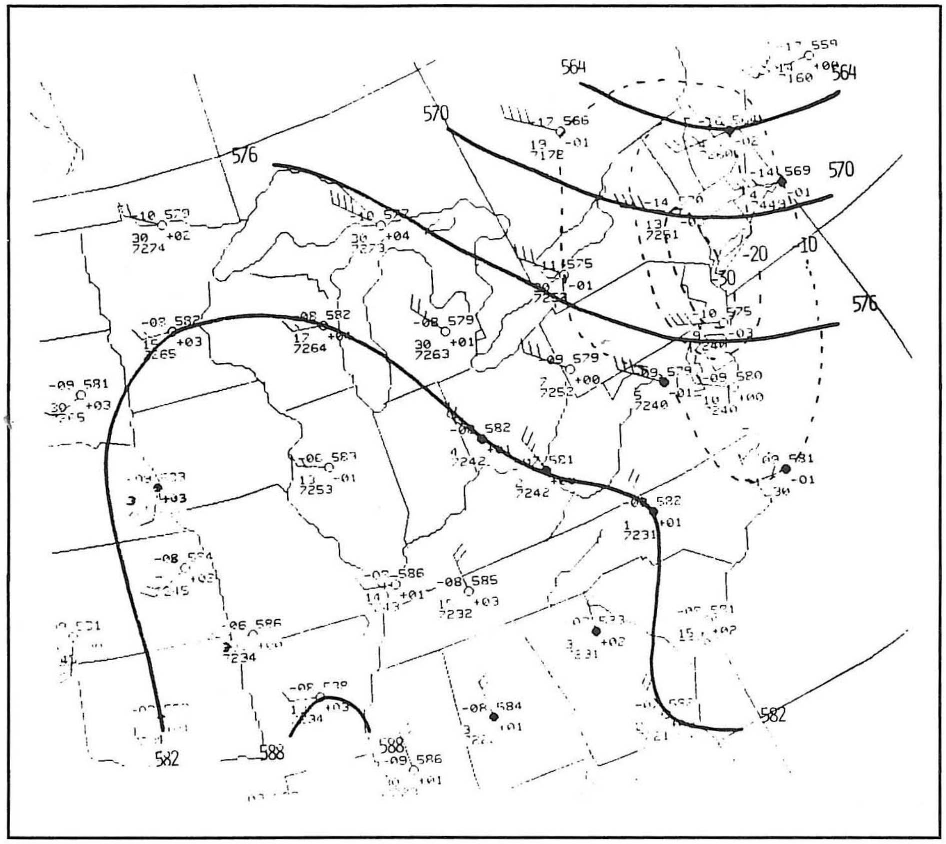

Fig. 3. Standard 500 mb station plot with wind in knots for 0000 UTC 3 June 1991. Solid lines represent the height field (dm); dashed lines represent 12-hr height falls (m).

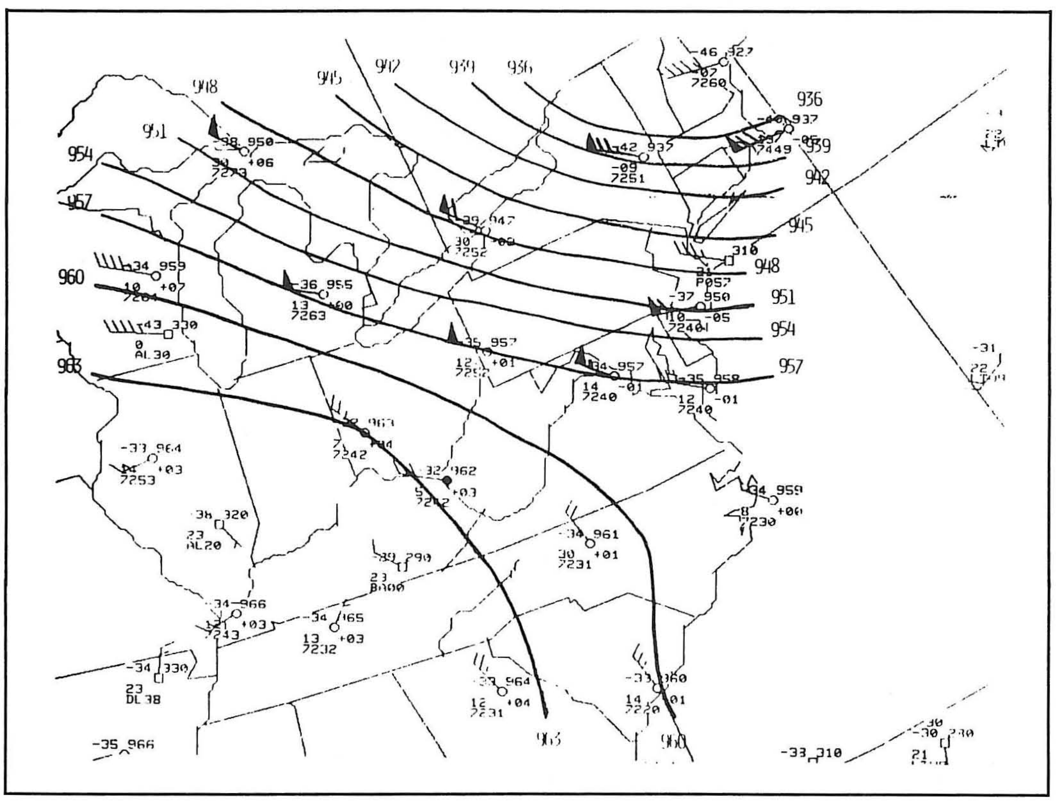

Fig. 4. As in Figure 3, except for 300 mb. Winds ≥ 50 kts are highlighted.

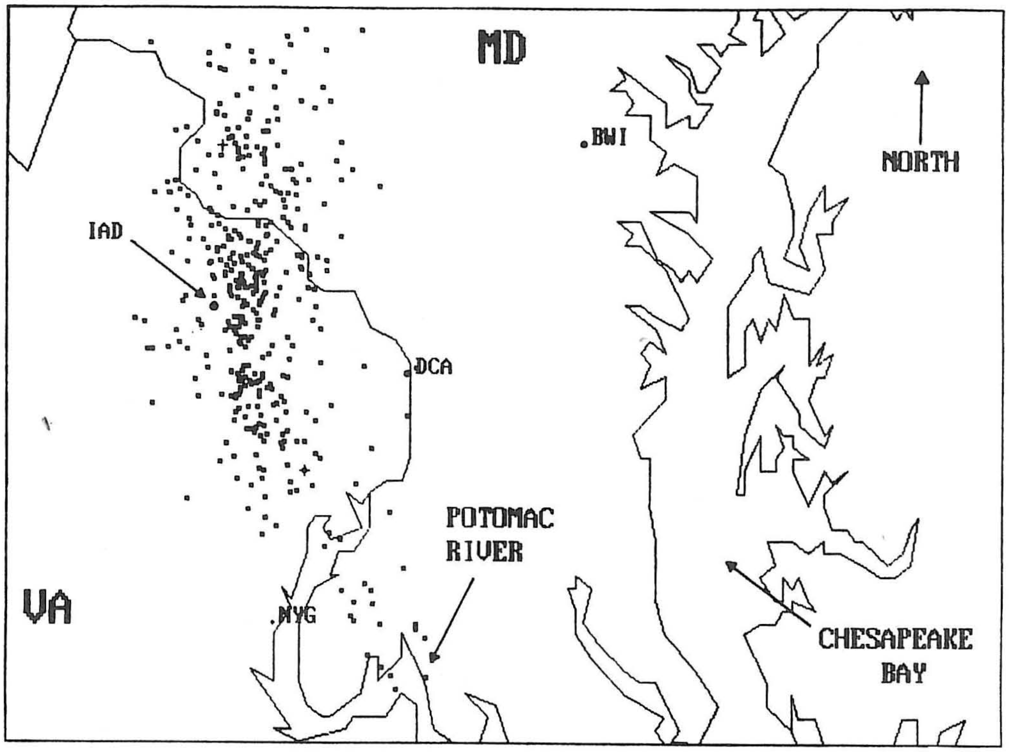

Fig. 5. Cloud-to-ground lightning strikes from 0350 UTC to 0831 UTC 3 June 1991 for an isolated thunderstorm across central Maryland and northern Virginia. Pluses (+) indicate positive strikes.

strikes (Beasley 1985; Fuquay 1982; Brook et al. 1989; Rust et al. 1981; Orville et al. 1987; Reap and MacGorman 1989; Scott 1988).

Thunderstorms commonly exhibit fluctuations in lightning frequency during their life cycles (Maier and Krider 1982). The 5-min CG lightning rate had several peaks through the storm's life time; however, the maximum 5-min CG lightning rate occurred from 0542 to 0547 UTC (Fig 6). Forty CG strikes occurred in this 5-min period. At the same time, a special radar observation was taken (0547 UTC) from the NWS radar site in Patuxent River, Maryland. The observation reported a maximum intensity VIP 6 (>57 dBZ) radar echo with a large VIP 5 (50 to 57 dBZ) echo extending to 6.4 km (21,000 ft) AGL. Figure 7 shows the outline of the VIP 5 echo core superimposed on the CG strikes for the 5-min period from 0542 to 0547 UTC. The center of the radar echo core was approximately 117 km (63 nm) to the northwest of the Patuxent River radar site. The radar was scanning at 0.5° elevation. As a result, the radar beam centerline intersected the echo (Fig. 7) at an altitude of about 1.8 km (6,000 ft), accounting for the effect of the earth's curvature on the radar beam[2]. The beam diameter at this range was approximately 3.7 km (2 nm). It can be seen from Fig. 7 that a large number of the CG strikes were associated with the VIP 5 echo. In fact, 65% of the strikes that occurred between 0542 and 0547 UTC were within the VIP 5 core (assuming a slow storm propagation within the 5-min period).

The associated rainfall pattern (Fig. 8) shows the 10 cm (4 in.) isohyet encircling IAD and vicinity (actual rainfall total at IAD was 11 cm [4.33 in.]). The overall pattern is elongated along the axis of storm propagation with the tightest precipitation gradient on the west and southwest side. Figure 9 represents the CG lightning flash density field superimposed on the rainfall pattern. The flash density contours are in flashes km^{-2} and the rainfall is described by the 2 cm, 4 cm, 6 cm, 8 cm, 10 cm isohyets. The maximum precipitation area coincides well with the maximum in flash density, although the center of the maximum rainfall (over 10 cm) is displaced over the western portion of the tightest flash density gradient. There are secondary maxima of lightning flash density both to the north (0.3 flashes km^{-2}) and south (0.5 flashes km^{-2}) of the main rainfall maximum. Because of the resolution of the precipitation reports, it was impossible to discern whether there were rainfall maxima at these locations. However, both flash density maxima lie very near or on the maximum precipitation axis.

To better examine the relationship between the precipitation field and the CG flash density, a grid was constructed. Each individual grid block was approximately 25 km^2. The grid was superimposed on the CG lightning field and the number of CG strikes in each grid box was calculated. The same grid was then superimposed on the precipitation field and a rainfall value for each grid point was ascertained. The average rainfall for each grid block was then calculated by averaging the four grid points of each grid block. The total volumetric rainfall was computed, as well as the total area that received rain. If rain fell anywhere within a particular grid block, it was assumed that rain occurred throughout the entire grid block. Based on these calculations, the number of CG strikes were compared to the average rainfall in each

[2]Height of the beam centerline given by the equation: $H = ([r^2 \cos^2 x / 9168.66] + [r \sin x]) * (6076.115)$ where r is in nautical miles, H in feet and x the elevation angle. (U.S. Department of Commerce and U.S. Department of Defense 1981)

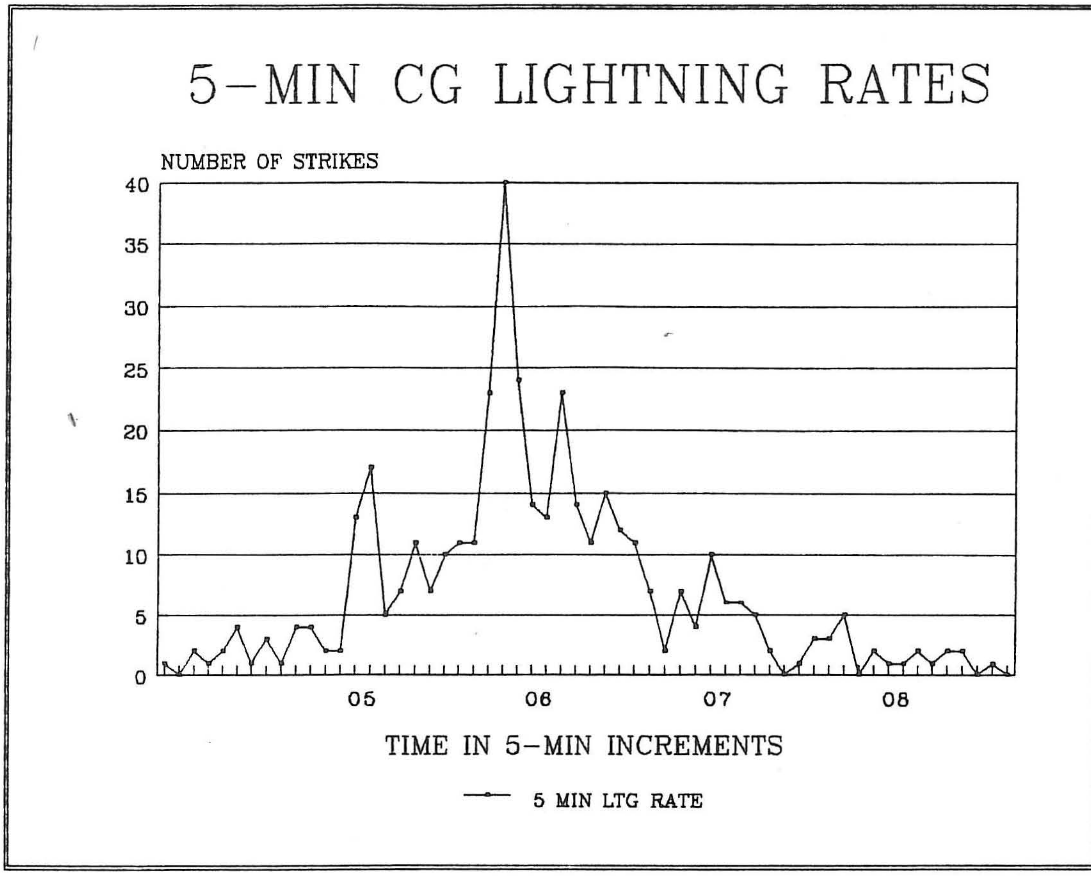

Fig. 6. 5-min CG lightning rates for thunderstorm on 3 June 1991.

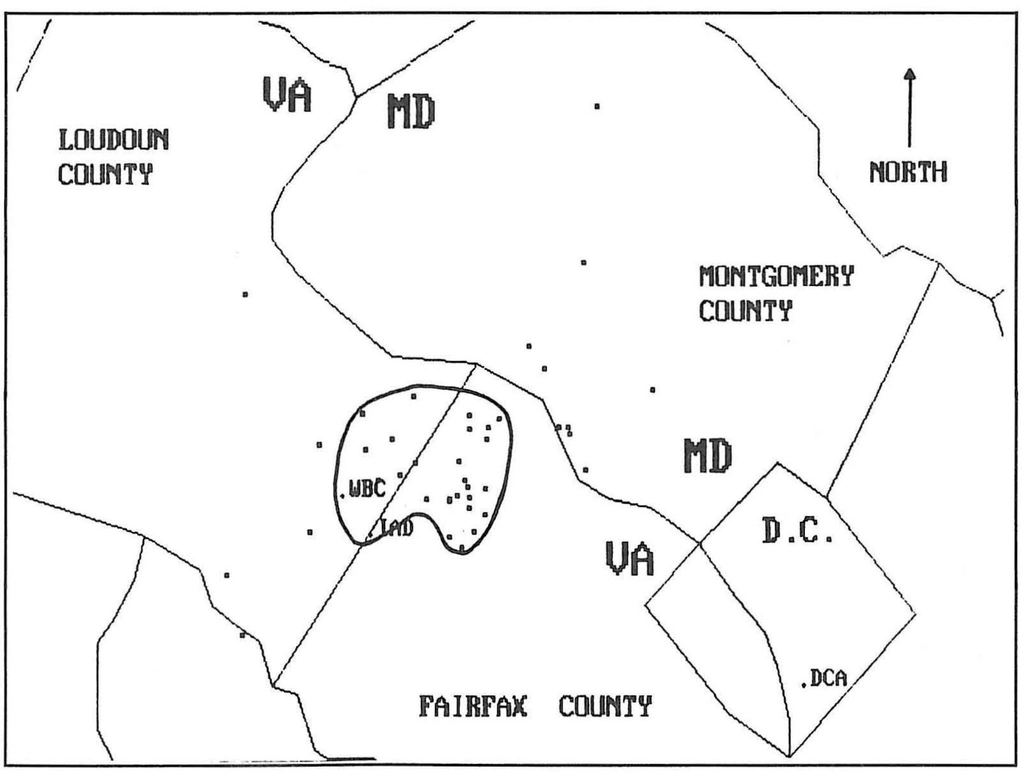

Fig. 7. Cloud-to-ground lightning strikes from 0542 UTC to 0547 UTC 3 June 1991 during maximum 5-min lightning rate. The outline of the VIP 5 core radar echo from Patuxent River, Maryland is superimposed.

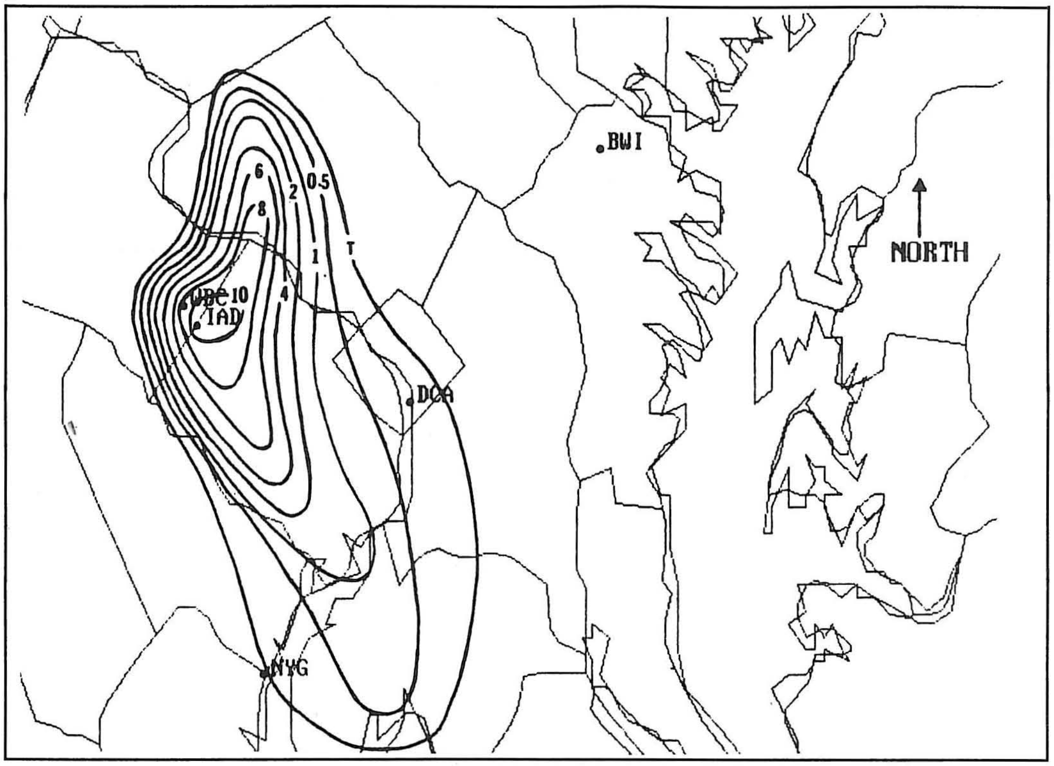

Fig. 8. Analyzed precipitation field (cm) produced by isolated thunderstorm of 3 June 1991.

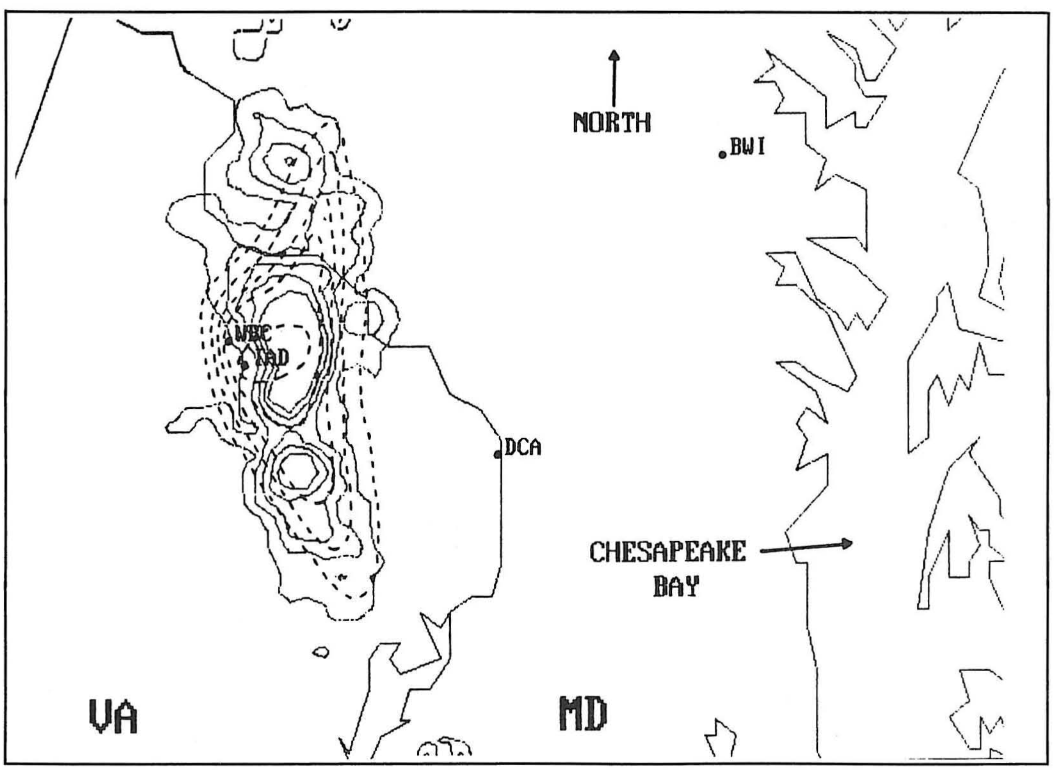

Fig. 9. Analyzed lightning flash density field (solid lines) from 0350 UTC to 0830 UTC 3 June 1991. The lightning data include both positive and negative strikes. Contours are in increments of 0.1 flashes km^{-2} (ranging from 0.1 to 0.5 flashes km^{-2}). The analyzed precipitation field (dashed lines) is superimposed with isohyets of 2 cm, 4 cm, 6 cm, 8 cm, and 10 cm.

grid block, the total area covered by rain, and the total volumetric rain.

There were 160 grid blocks that received rain (over an area about 4,000 km^2) and 114 grid blocks with CG lightning strikes. There were 68 grid blocks that received rain without detected CG strikes. There were 22 grid blocks with CG strikes either without rain or only a trace of rain recorded. The majority of grid blocks that received either very little rain or none at all, occurred primarily during the early stages of the storm system. In fact, approximately 80% of the blocks appeared in the first third of the storm's existence. The remaining blocks were located along the western periphery of the precipitation area were the rainfall gradient was the largest and they occurred about mid-way through the storm's life.

The greatest number of strikes to occur in a grid block either without rainfall or with only a trace of rain was 4, while the largest amount of average rain over a grid block without CG strikes was 3.55 cm (1.4 in.). The largest average rainfall for a grid block was 9.88 cm which occurred in two separate grid blocks and the highest number of CG strikes in any one grid block was 17.

Those grid blocks which had 5 or more strikes accounted for just over half (50.1%) of the total volumetric rainfall over only 15.6% of the area that received rain (Fig. 10). Additionally, 57% of the total CG strikes produced by the storm were concentrated in this same area. The grid blocks containing from 0 to 2 strikes were most numerous (135). They accounted for 33.6% of the total volumetric rainfall which covered about 72% of the area that received rain. However, when the average amount of rainfall in each grid block is categorized and compared (Fig. 11), those grid blocks which received 12 or more strikes, each averaged 8.7 cm of rainfall. This was almost twice as much as the next lower category (9 to 11 strikes) and well over 12 times as much as those grid blocks containing from 0 to 2 CG strikes.

To investigate the temporal relationship between the lightning and rainfall for this thunderstorm, the CG strikes within specific radii of the NWSFO at Sterling, Virginia (WBC) were compared with the rainfall recorded on station. The amount and time of rainfall for each 5 min (using the weighing rain gage) were compared to CG lightning strikes every 5 min within a radius of 10 km, 20 km, and 30 km of WBC. Figure 12 shows that the amount of CG strikes within 30 km rose sharply about 0500 UTC and then started to fluctuate. The strikes within 20 km had an initial peak at 0515 UTC, while the first strike within 10 km also occurred at this same time.

The strikes within all 3 radii started to increase sharply about the time rain began on station (0535 UTC). At 0545 UTC, the strikes within all 3 radii peaked. After an initial lull in rainfall immediately following the lightning peak (ending at 0550 UTC), the rainfall increased rapidly with 2.11 cm (0.83 in.) occurring in the next 15 min (by 0605 UTC). The heavy rain continued until 0625 UTC. In this 40-min period following the peak in all three CG strike radii, 4.22 cm (1.66 in.) of rain fell. The rainfall then diminished in the next 5-min period. Accordingly, the rainfall rate also increased rapidly about the time that all three radii peaked. The greatest rainfall rate (calculated from the weighing rain gage chart) occurred approximately between 0550 and 0625 UTC. During this period, rain fell at the rate of 7.3 cm hr^{-1} (nearly 3 in. hr^{-1}).

A secondary peak in the 20 km and 30 km CG lightning occurred at 0605 UTC, about mid-way through the highest

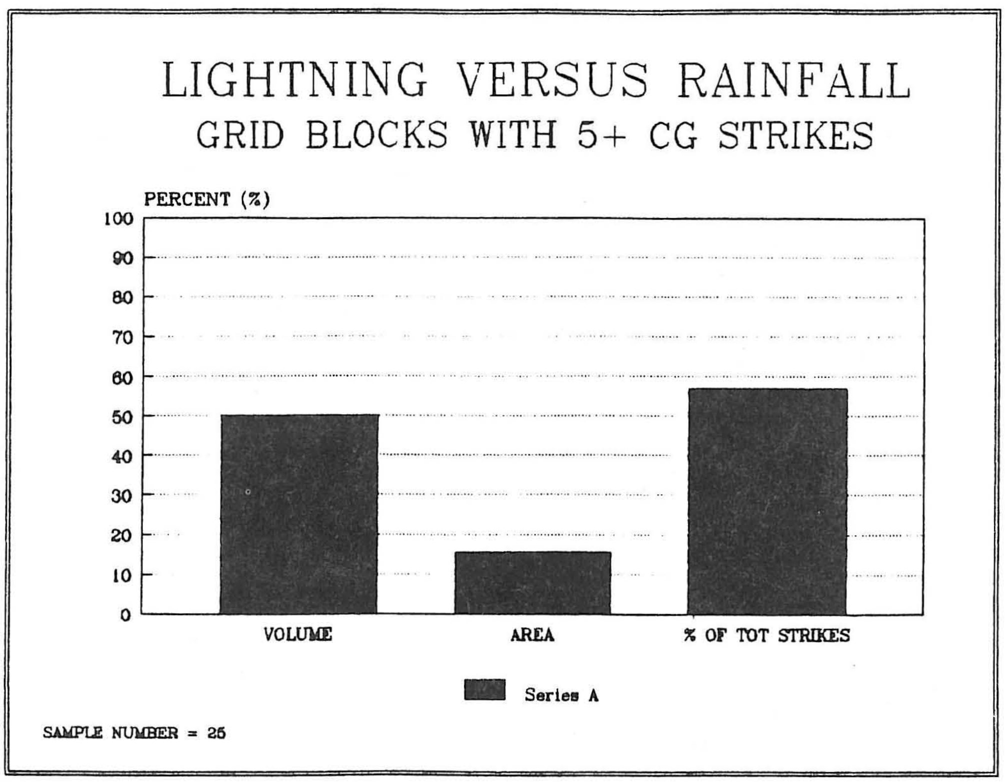

Fig. 10. Percentages of the volume of rainfall, area over which rainfall occurred, and amount of total CG strikes for grid blocks containing 5 or more strikes.

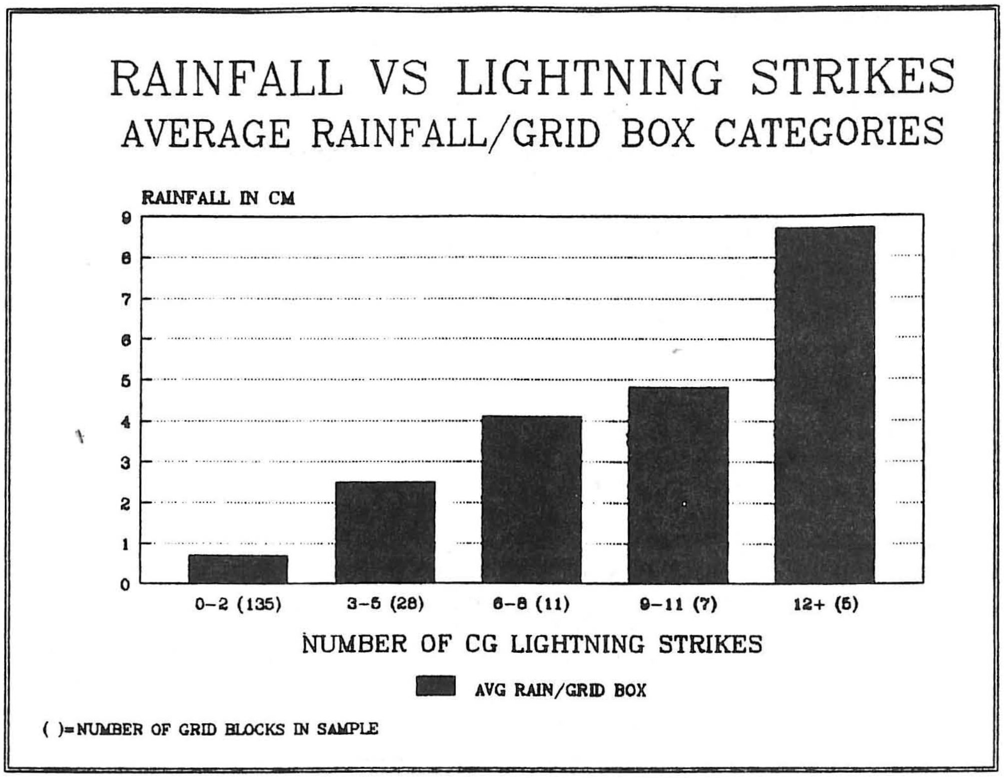

Fig. 11. Average rainfall versus CG lightning strikes per grid box and the number of grid boxes comprising the sample.

rainfall rate; however, the CG lightning within the 3 radii gradually diminished, with the CG strikes within 10 km ending at 0630 UTC, 20 km at 0655 UTC, and 30 km at 0705 UTC. There were four additional peaks in rainfall, all of which were much smaller than the initial downpour (Fig. 12). They occurred at 0705, 0720, 0815, and 0830 UTC. It must be noted that although all CG strikes declined and eventually ceased, IC flashes were physically observed on station for some time.

5. Discussion and Summary

A few investigators in the past have used rain gages for ground truth in comparing lightning characteristics to rainfall (Piepgrass et al. 1982; Grosh 1978; Battan 1965). However, the vast majority of past studies used radar and empirical reflectivity versus rainfall relationships (e.g., Z-R relationships described by Marshall and Palmer 1948, Jones 1956, and Seliga et al. 1986) in order to estimate or calculate the derived rainfall characteristics. This investigation directly relates the CG lightning flash density field to an observed precipitation field. Clearly, a correspondence existed between the areas that received the highest concentration of CG strikes and those areas that experienced the greatest rainfall. A similar relationship is apparent when the average rainfall in each grid block is compared to the number of CG strikes (Fig. 11). As the number of CG strikes in each grid block increased, the average rainfall over the grid block increased. It is also clear from Fig. 10 that just over half the storm-total volumetric rainfall fell in a relatively small concentrated area, which was described by the majority of total storm CG strikes. In other words, nearly 60% of the total number of CG strikes generated by the entire system were concentrated in only about 16% of the area that received rain. Furthermore, this same small area received over half the total volumetric water produced by the convective system.

The temporal relationship between CG lightning and heavy rain on station (WBC) was also significant. The heaviest rain started within about 5 min of the peak in CG lightning within all three radii (10 km, 20 km, and 30 km) of the station. Meanwhile, the greatest rainfall rate occurred during the 5 to 40 min period after the peak CG lightning within all three radii. Of interest are the four additional peaks in rainfall following the initial downpour. The trailing stratiform-like precipitation that existed behind the main convective core was very similar to that observed by Hunter et al. (1990). They observed an MCS in Oklahoma in which the trailing precipitation shield contained multiple embedded convective elements. These elements were pervasive and unsteady, so much so, that the term "stratiform" may have been inappropriate. Similarly, in the 3 June 1991 case, a general VIP 2 (30 to 41 dBZ) region developed behind the main convective core with multiple embedded VIP 3 (41 to 46 dBZ) echoes and even a VIP 4 (46 to 50 dBZ) echo for a time. It is possible that these convective entities were responsible for the continued IC lightning, as well as the four rainfall peaks shown in Fig. 12.

It was also shown, at least at the time of the peak 5-min CG rate (0547 UTC), that most of the strikes were collocated with the high reflectivity core. If reflectivity-rainfall relationships were used instead of measured precipitation, this would also suggest that the highest concentration of strikes was associated with the heaviest rainfall. It was also fortuitous that the peak in the storm's overall CG flash rate occurred

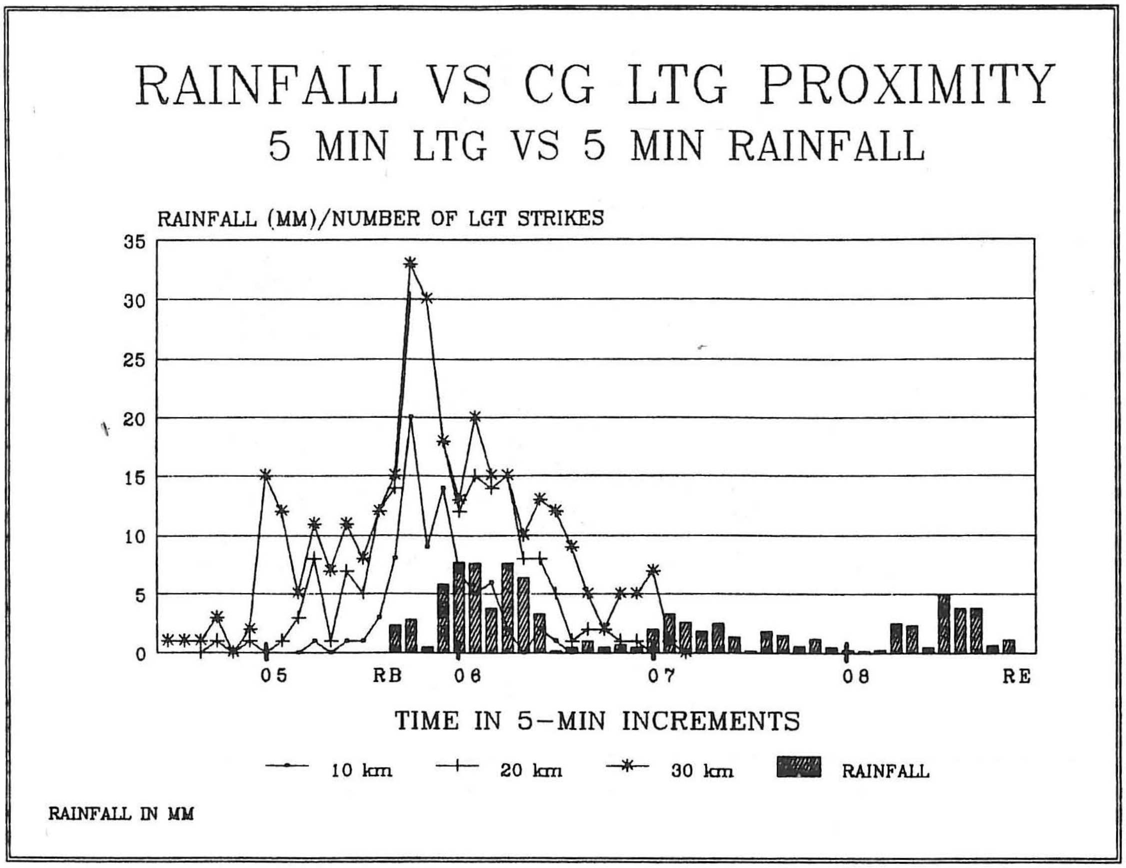

Fig. 12. 5-min CG lightning rates for strikes which occurred within 10 km, 20 km, and 30 km of the NWSFO at Sterling, Virginia compared to the amount of rain (mm) every 5 min received on station. The beginning (RB) and ending (RE) of rain are also noted.

approximately as it moved over the NWSFO at Sterling, Virginia (WBC) and Dulles airport (IAD). It can be inferred that the peak in the CG flash rate was followed by the heaviest rainfall both temporally and spatially. Thus, this implies that the most intense convective towers and associated echoes were associated with the best charge generation and the heaviest precipitation.

Other investigations have found similarities to this case, as well as differences, in the clustering of CG strikes with respect to high reflectivity cores. Engholm et al. (1990) noted that the predominant negative CG strikes were clustered near deep convection in both a squall line and a cluster of thunderstorms. Ray et al. (1987) reported that lightning tended to appear downshear of the main updraft (indicated by a weak echo region) and reflectivity core in a supercell storm. On the other hand, lightning was concentrated in the updraft and reflectivity core in a multicell storm. LaPenta et al. (1990) found that the strongest echoes in a high precipitation supercell were associated with the maximum in negative CG flash density. Cherna et al. (1984) found that lightning was clustered in the immediate area of intense precipitation cores within a squall line but found at least one cell was characterized by a high flash density around its perimeter. Mazur and Rust (1983) suggested that the maximum in lightning activity was not always coincident with the highest reflectivity cores in squall lines and there was a tendency for the maximum in lightning density to remain near the leading edge of the precipitation core. Likewise, Geotis and Orville (1983), and Orville et al. (1982) have found CG strikes to proliferate outside high reflectivity cores along the forward periphery of storms. Lopez et al. (1990) also documented an abundance of negative CG strikes on the fringe of the maximum radar echo in the high reflectivity gradient.

The differences in the displacement of the CG clusters with respect to the most intense cores is not clear. Although, it must be noted here that even in this example, the majority of strikes that were located within the VIP 5 core appeared in the forward half (with respect to propagation) of the reflectivity core. It may be that there is a difference among convective system types as distinguished by differences in low-level moisture availability, boundary forcing, or upper-level dynamics. However, as suggested by Mazur and Rust (1983), the answer may be that the relative positions of the two maxima may change during the storm's life.

The synoptic pattern presented here is a typical setting for nocturnal convection over the mid-Atlantic region of the United States during the warm season (i.e., the mean upper ridge position upstream, upper level northwest-flow, a very weak baroclinic zone and associated thickness pattern, extending northwest to southeast, and upper level diffluence). The large long-lived MCCs and MCSs which are responsible for a majority of the nocturnal convective precipitation in the Midwest (Fritsch et al. 1981, 1986) are noticeably absent in the mid-Atlantic region. However, Fleming et al. (1984) found many excessive rain producing convective systems in the eastern United States to be warm-topped, slow moving or regenerative, small, single or multi-clustered thunderstorms very similar to the convective system presented in this study.

The fact that the heaviest rain was associated with the highest concentration oi strikes and greatest CG flash rate may have considerable potential in its applicability to short-term excessive rainfall prediction. Furthermore, just over half of the volumetric rain fell in a very small percentage of the area that received rain. This area, with the highest potential for flooding, was well described by the highest concentration of CG strikes. Lightning data has already proven to be a significant diagnostic tool in an excessive rain event (Kane 1990). The relationships presented here, and in past studies, indicate the potential for both diagnosing and forecasting short-term convective events. This is especially true now that real-time lightning data are available on a nation-wide scale.

At least in this isolated thunderstorm, there was a very good correspondence between CG lightning and rainfall. For the operational forecaster that has access to real-time lightning data, the improved diagnosis and prediction of heavy rainfall will unquestionably lead to more accurate and timely flash flood warnings. Lightning data can greatly aid the forecaster in the decision making process by detecting intensity, location, and movement of storms (i.e., speed, persistent cells, training, backward propagation, etc.). In addition, lightning data does not suffer attenuation, is available in real-time, and has no gaps in coverage. Operationally, the combination of real-time lightning data, radar, and satellite imagery provides an accurate and efficient way to predict and warn for smaller-scale excessive rainfall events. As the WSR-88D radars are deployed and the Precipitation Processing (Ahnert et al. 1984) and Flash Flood Processing (Walton et al. 1986) systems are implemented, lightning data will undoubtedly play an important role in supplementing this radar information. Even more importantly, real-time lightning data will play a vital role in complementing new technologies such as the Automated Surface Observing System (ASOS) (Short and McNitt, 1991).

Acknowledgments

The author wishes to thank Patrick Moore of the National Weather Service Forecast Office in Sterling, Virginia for his thorough review. Thanks are also extended to Gary Carter, Jeff Waldstreicher, and especially Steve Hrebenach of the NWS Eastern Region Headquarters, Scientific Services Division for their reviews and perspectives. Additionally, the author wishes to thank Dr. Wayne McGovern of the NWS Techniques Development Laboratory for his ideas and suggestions. Appreciation is also extended to the National Weather Service Meteorological Observatory in Patuxent River, Maryland and to the Albany Mesoscale Climatology Group for help in obtaining data. Finally, a special thanks to Rhonda, Krystle, and Richard Kane III for their understanding, patience, and support during this project.

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