Electrification in Mesoscale Updrafts of Deep Stratiform and Anvil Clouds in Florida

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Abstract

Airborne observations in deep stratiform and anvil clouds showed extensive layers of 10- to 40-kV/m electric fields collocated with highly stratified uniform radar reflectivity of 20 to 25 dBZ from 5- to 9-km altitude. Size distributions with numerous small- and intermediate-sized ice crystals (mostly aggregates) and large aggregates were observed in these regions. We infer that the uniform electric field, radar reflectivity, and broad particle size distributions were the result of mesoscale updrafts, confirmed by high-resolution images of particles showing diffusional growth and no riming. No measurable supercooled liquid water was found in these regions from −10 to −45 °C. Calculated particle collision rates from observed distributions were >5 × 10^5 collisions per cubic meter per second in this volume. Laboratory results show that weak charge separation occurs when ice particles collide and separate even in the absence of supercooled water. We infer that charge separation occurred in the mesoscale updrafts via a noninductive mechanism in which ice particles growing by diffusion collide and transfer charge without supercooled water being present. These regions with strong, uniform fields, stratified radar reflectivity, and broad size distributions also occurred in anvils that barely reached the melting zone. Thus, we deduce that the nonriming collisional mechanism acts at middle to upper cloud levels and is not dependent upon electrification occurring near the melting zone. This mechanism should produce two oppositely charged layers of charge with the top layer residing on smaller particles often existing near the top of the cloud.

1. Introduction

Deep stratiform clouds, such as those associated with mesoscale convective systems (MCSs) are known to be highly electrified. It is commonplace to observe very extensive lightning flashes near cloud base in association with the melting layer and also positive cloud-to-ground (CG) lightning originating from the interior of the cloud. There have been numerous balloon investigations of stratiform regions associated with MCSs showing electric fields as strong as ~100 kV/m (e.g., Marshall & Rust, 1993; Stolzenburg et al., 1994; Schuur & Rutledge 2000a, 2000b) to name only a few. Stolzenburg and Marshall (2008) provide a relatively recent overview of electrical studies of thunderstorms including stratiform clouds. Stratiform clouds and precipitation falling from them have been the subject of much research in the tropics as well as at midlatitudes. Houze (1989) developed a conceptual model of typical MCSs including the trailing stratiform region that has served as a basis for modified conceptual models and simulations of MCSs. Other authors such as Schuur and Rutledge (2000b) have used this conceptual model as a basis for modeling studies of electrification within these stratiform systems. Lightning and the extensive layers of charge occurring in trailing stratiform regions also have been of considerable interest in association with transient luminous events such as SPRITES (e.g., van der Velde et al., 2006; Lang et al., 2010).

Early papers suggested that transport of charge from convective regions into the stratiform region is a main source of charge and electric field in stratiform regions and anvils. For example, Stolzenburg et al. (1994) examined five soundings taken at different times and locations in an Oklahoma squall line within a large trailing stratiform region. The soundings showed similar charge density profiles as a function of altitude, location, and time suggesting that extensive layers of charge existed in the stratiform region over extended time periods. Based on the measurements and a conceptual model, the authors argued that an extensive upper level layer of positive charge and an extensive thinner layer of negative charge at middle level could be the result of charge advection from the active convective region of the squall line. But they also thought that the lower negative charge could have been augmented by some in situ charge separation and that the charge layer near 0 °C was probably caused by in situ charge separation. In a comparison of observations with model simulations of another trailing stratiform region Schuur and Rutledge (2000a, 2000ab) concluded...
that advection of charge could not explain the observed magnitude or structure of the charge and that significant in situ charging (perhaps up to 70%) had taken place. Charge separation in their model was via the noninductive ice-graupel collision process in which supercooled liquid water (SLW) is present. (Jayaratne et al., 1983; Takahashi, 1978).

Mesoscale updrafts are a key feature of stratiform clouds that form in association with deep convection. We consider stratiform clouds as ones for which the radar reflectivity is vertically stratified and exhibits relatively uniform values over extended distances of tens of kilometers or more. Reflectivity decreases in value at upper levels without evidence of deep cellular convection or embedded convection as suggested by closed contours of reflectivity often of 30 to 35 dBZ or more. Updraft strengths in stratiform regions are often <1 m/s, while in convective regions they can be a few meters per second up to tens of meters per second in the strongest storms. This paper focuses on measurements in middle to upper levels in stratiform regions and not convective regions. The stratiform regions and anvils are a result of the decay and outflow from convective cores (Houze, 2014). He distinguishes between active and inactive stratiform clouds. Active stratiform clouds are those in which there is a weak updraft and the mean updraft speed averaged over a cloud volume is less than the terminal velocity of the larger ice particles, that is, aggregates. Inactive stratiform are those where the mean vertical air motion is 0 or descending, and Houze refers to these as debris or fallout.

Mesoscale updrafts are thought to form as a result of mesoscale forcing due to buoyant air detrained from deep convection, by radiative imbalance between top and bottom of an anvil, and/or by gravity waves. A discussion on mesoscale updrafts can be found in Cloud Dynamics by Houze (2014), chapter 5 on Cirriform Clouds, and chapter 6 on Nimbostratus Clouds. The updraft strengths are thought to range from a few centimeters per second to tens of centimeters per second. Because they are weak they are difficult to measure directly with conventional vertically pointing Doppler radar. The updraft speed is similar to fall speeds of some of the ice particles so it is not possible to separate the vertical air motion from the fall speed of the particles. A few observations have been made with Doppler wind profilers at 50 MHz, which have the capability of separating upward vertical wind motion from motion of falling precipitation in the Doppler spectra, (Ecklund et al., 1992; Williams & Ecklund, 1992) but measurements are limited to storms that pass directly over the fixed location of the profiler. Some multiple Doppler radar syntheses have deduced updrafts in well-developed stratiform regions behind large MCSs with strengths of a few tens of centimeters per second, for example, Schuur and Rutledge (2000a). Radar scans using velocity azimuth display (VAD techniques [e.g., Knupp et al., 1998]) or volume velocity processing (VVP) techniques (Boccippio, 1996) have been used to deduce the strength of mesoscale updrafts and downdrafts averaged over large domains. Mesoscale downdrafts are also common near and slightly above the melting level in deep stratiform clouds (Protat & Williams, 2011).

Mesoscale updrafts and downdrafts have a large impact on the microphysical structure of the cloud. Adiabatic expansion and cooling in upward moving air allow all ice particles, both small and large, to grow by diffusion and in combination with aggregation, mesoscale updrafts can maintain broad ice particle size distributions (PSDs). By broad size distribution we mean a size distribution containing both high concentrations of small and midsized ice crystals and irregularly shaped particles and aggregates up to several millimeters in size. In the downdraft the warming preferentially causes sublimation of the smallest particles, thereby producing a much flatter size distribution with significantly fewer small and intermediate-sized particles. Consequently, the shape of the size distribution, the nature of the ice surface, and the sign and magnitude of charge separation via ice-ice collisions (Williams et al., 1991) is different in updrafts in contrast to downdrafts. This will be discussed at greater length in sections 4 and 5 below.

There have been relatively few airborne in situ microphysical measurements in stratiform regions. Among them are the studies of Houze and Churchill (1984), Willis and Heymsfield (1989), Schuur and Rutledge (2000a), and Stith et al. (2002). Willis and Heymsfield (1989) studied the region between −8 °C and +8 °C in a detailed investigation of the melting layer. They reported small ice crystals, dendrites, and small and large aggregates. The Rosemount Icing Detector (RICE) showed “the presence of a small amount of liquid water at temperatures from −5 to −2°C.” They do not mention the presence of graupel or rimed particles. Much like the observations we report herein they show broad size
distributions containing high concentrations of small particles and small to large aggregates. Schuur and Rutledge (2000a) reported that the microphysical structure in a stratiform region of an MCS at levels warmer than $-10\,^\circ\text{C}$ was complex with small ice crystals, dendrites, aggregates of dendrites, and many irregularly shaped particles. They also reported some rimed particles with graupel up to 3 mm in diameter. While the presence of graupel is possible and others have also reported rimed particles and graupel in this temperature zone, given the resolution of the images from the Particle Measurement Systems 2-D probes, it is difficult to determine this with certainty. The focus of Stith et al. (2002) was primarily convective regions of tropical clouds but they briefly report PSDs at low, middle, and high levels in deep stratiform clouds in Brazil, Kwajalein, and Florida. The observations in Florida at middle and upper levels show broad size distributions with many small and midsized particles and aggregates of a few millimeters in size. They conclude that aggregation was a major growth mechanism in the stratiform clouds that they studied.

Mo et al. (2003) report combined electric field and some microphysical measurements made just above the 0 °C level by the South Dakota School of Mining and Technology T-28 aircraft in the same stratiform cloud examined with balloon borne measurements by Stolzenburg et al. (1994). The microphysical measurements from the Particle Measuring Systems (PMS) 2D-Precipitation probe (2D-P) with 200-μm resolution on the T-28 showed irregularly shaped particles with the larger particles being primarily aggregates of snowflakes and ice crystals which were lightly rimed if at all. Riming would have been difficult to discern with the 2D-P. More recently investigators in the Midlatitude Continental Convection Clouds Experiment project in Oklahoma, used the University of North Dakota Citation II aircraft with a detailed microphysical instrument package similar to the one we describe in section 2 to explore convective and stratiform clouds in and near Oklahoma (Jensen et al., 2016). In their Figure 5, Jensen et al. present PSDs from 7.5 km ($-27\,^\circ\text{C}$) down to 2.8 km (9 °C) along with images of the particles from the High Volume Precipitation Spectrometer, Mod. 3. Even at 7.5 km, 15-mm aggregates were present with >20 mm observed near 0 °C. Although not seen in Figure 5 because of the linear scale, high concentrations of small to midsized particles were observed at all altitudes except in and below the melting zone. In this same project, Giangrande et al. (2016) have described the microphysical processes occurring in the zone from $-10\,^\circ\text{C}$ to below the melting zone in great detail with polarization radar and an array of modern particle imaging sensors. They reported many particles types including needles, apparently from the Hallett and Mossop ice multiplication process. The region from $-10\,^\circ\text{C}$ down through the melting zone is a very complex zone and in need of additional focused microphysical and electrical studies. Electric field mills were not flown on the University of North Dakota Citation for the Midlatitude Continental Convection Clouds Experiment project.

While there are many vertical profiles of electric field from balloon platforms, these lack information on the microphysical content. Airborne in situ simultaneous measurements of electric field and microphysical content in deep stratiform regions are sparse. The only published results of which we are aware are those of Mo et al. (2003) who report measurements in Oklahoma made near 0 °C near and in the bright band by the South Dakota School of Mining and Technology T-28 aircraft and from the NOAA P-3 aircraft at about 5.8 km ($-9\,^\circ\text{C}$). Mo et al. report that the field measurements at 4.5 km showed rapid reversals in the vertical field with magnitudes up to 50 kV/m. They concluded that the measurements were consistent with a narrow undulating layer of strong positive charge near 0 °C. They also report electric field measurements made by the NOAA P-3 aircraft at $-5.8\,^\circ\text{C}$ (about $-9\,^\circ\text{C}$) that were always positive but with variable magnitudes from 0 to 25 kV/m over a distance of more than 200 km. These airborne measurements were consistent with the magnitude and sign of the balloon-borne measurements of Stolzenburg et al. (1994) made in the same stratiform cloud.

Dye et al. (2007) presented measurements of both electric field and particles in anvils and debris clouds at middle and upper levels near Kennedy Space Center (KSC), Florida, and explored relationships between electric field, radar reflectivity, and particles concentrations and sizes in these clouds. Using measurements from this same project Dye and Willett (2007) reported enhancements of electric field and radar reflectivity in long-lived anvils and suggested that the enhancements might result from mesoscale updrafts.

In this paper we present airborne observations at middle to upper levels (at temperatures colder than $-10\,^\circ\text{C}$) in several deep stratiform clouds near Kennedy Space Center, Florida. The observations include three-dimensional electric field measurements and detailed microphysical and liquid water measurements.
from an array of particle and imaging probes. As anvils were the main focus of the project, the aircraft penetrations were most frequently made at altitudes from 8 to 10 km but clouds were also sampled at lower altitudes including near the melting zone for some penetrations and during ascent or descent to or from higher altitudes. The focus of this paper is the in situ measurements from roughly $-10$ °C ($\sim 6$ km) to roughly $-45$ °C (10 km). The airborne measurements were coordinated with radar reflectivity measurements from a 5-cm radar located at Patrick Air Force Base (PAFB) and the NEXRAD national network of 10-cm Doppler radar located at Melbourne Florida. The observations showed the presence of layers of strong electric field (10 to $>40$ kV/m) extending over distances of ten to many tens of kilometers from altitudes near $-10$ °C up to 10 km or more (about $-45$ °C). These strong fields were located in and extended somewhat above regions of highly stratified radar reflectivity of 20 to 25 dBZ from 6 to 9 km and sometimes 25 to 30 dBZ from 5 to 7 km. Values up to 40–45 dBZ were sometimes present in the bright band, but the bright band was not present in all cases.

From the existence of these relatively uniform, extensive and long-lived strong fields at middle to high altitudes (temperatures $< -10$ °C), the horizontal uniformity of radar reflectivity without obvious embedded convection, and the broad ice PSDs we infer that an extensive weak, but long-acting, in situ charge separation mechanism was occurring. Dye and Willett (2007) observed these very same conditions in long-lived anvils in Florida and also concluded that in situ charge separation had to have been taking place. The broad ice PSDs containing numerous small ice particles, and large aggregates creates conditions such that the ice–ice collision rate can be high over a broad area and throughout a considerable vertical depth of the cloud. The mesoscale updraft can sustain these broad spectra as long as the updraft persists. High-resolution images of ice particles in both small and large stratiform regions and anvils away from convective cores at temperatures below $-10$ °C definitively show evidence of recent diffusional growth of ice particles with no evidence of graupel or riming. Our observations show that, if it existed at all, SLW contents were not above the noise level of the Rosemount Icing Detector of about 0.002 g/m$^3$ (Heymsfield & Miloshevich, 1989) at temperatures below $-10$ °C. The absence of SLW and graupel and the absence of riming on the ice particles rules out the classic noninductive ice riming graupel mechanism that requires the presence of SLW as the cause of the in situ charge separation.

Based on our observations, a review of older laboratory work, and the new studies of Luque et al. (2016), we infer that in situ charging at middle to upper levels was a result of noninductive charge separation from ice–ice collisions in the absence of SLW. We will refer to this mechanism as the nonriming collisional mechanism. By way of contrast we shall refer to the classic mechanism of Takahashi (1978) and Jayaratne et al. (1983), as the noninductive riming electrification mechanism with SLW.

We suggest that the nonriming collisional mechanism we discuss herein occurs in the middle to upper levels of stratiform and anvil clouds. Although the nonriming mechanism might also be acting between $-10$ °C and the melting level, we do not have sufficient measurements to investigate that issue. Very strong electric fields have also been observed near the melting level in association with the radar bright band (e.g., Shepherd et al., 1996; Stolzenburg & Marshall, 2008), but we will not address that topic in this paper. Additionally, advection of charge from the core of convective cells is almost certainly taking place, but we do not have the necessary airflow measurements to examine that topic either.

2. The Measurements

The observations were made in Florida near Kennedy Space Center in June 2000 and May and June 2001 during the Airborne Field Mill Project (ABFM II). An overview of the project and an in-depth description of the airborne, radar, and lightning measurements and instrumentation are presented in Dye et al. (2007). The airborne measurements were made from the University of North Dakota Citation II jet aircraft that had the capability of making observations from the surface to ~11-km altitude inside of anvils, deep stratiform regions, and above the cores of small cumulonimbus clouds. Instrumentation included six electric field mills and an array of particle probes capable of measuring and imaging the hydrometeors (both water and ice) over the size range of $<100$ μm to at least 2 cm. Although we have measurements of particles $<100$ μm in size, we do not include them in the size distributions shown here because particle breakup on probe tips and limited electronic response time make the concentration and size values highly uncertain even after correction (Korolev & Isaac, 2005).
The six electric field mills similar to those described by Bateman et al. (2007) were built and calibrated by National Aeronautics and Space Administration (NASA) Marshall Space Flight Center. We used a right-handed coordinate system with $E_x$ positive upward, $E_y$ positive forward, and $E_z$ relative to the aircraft. The matrix coefficients for determining the individual electric field components was based on the approach of Mach and Koshak (2007). We estimate that the uncertainty in the measured electric field out of cloud was about $\pm 10\%$. When the aircraft penetrated an ice cloud, however, the errors increased significantly because of aircraft charging. $E_y$ and $E_z$, the field components in the vertical and along the wings, respectively, are estimated in clouds to be accurate to about 20% when the field due to charge on the aircraft, $E_{mag}$, is small compared to the magnitude of the ambient electric field. The uncertainty in $E_x$ is much larger, but the values of $E_x$ are normally small compared to the total scalar magnitude of the measured field, $E_{mag}$.

The concentrations, sizes, and type of particles were measured with five separate microphysical instruments. See Dye et al. (2007) for a short description of those instruments and the uncertainties in concentration and sizes for those instruments. The PMS Forward Scattering Spectrometer Probe (FSSP) was designed to measure cloud droplets in the range of $\sim 3$ to $50 \mu m$ but responds to ice particles as well. Concentration measurements from the FSSP in mixed phase regions and ice-only clouds have long been known not to be reliable because of ice particle shattering on the tube inlet (Gardiner & Hallett, 1985; Field et al., 2003; Jensen et al., 2009). Although we show measurements from the FSSP in some figures we use those measurements in this work only as a qualitative indicator of when water droplets might be present when there are departures between the FSSP and PMS 2D-Cloud probe (2D-C) traces as shown in some figures in section 3. At the speed of the Citation the 2D-C gives shadow images of particles from nominally $60 \mu m$ to $1 mm$ with $30 \mu m$ resolution. The Stratton Park Engineering Co. (SPEC) High Volume Particle Sensor (HVPS; Lawson et al., 1998) was important as it allowed us to image the larger particles in the nominal range of $<1 mm$ to $5 cm$ with a sample volume sufficiently large to give statistically meaningful concentrations. The HVPS also allowed us to determine the size of the largest particles (aggregates). The data from the 2D-C and the HVPS were processed using the minimum enclosing circle approach. (Wu & McFarquhar, 2016). When the HVPS was not functioning the 2D-P was flown. It has a resolution of $\sim 200 \mu m$ and covers the range from $400 \mu m$ to several millimeters but has a smaller sample volume than the HVPS.

The SPEC Cloud Particle Imager (CPI; Lawson et al., 2001) was important for this study because it gave us images of particles with $2.3 \mu m$ resolution over its effective size range of $10 \mu m \sim 2 mm$. Measurement and imaging of the larger particles is limited by the small sample volume and breakup of particles on the probe inlet. Images of particles from the CPI in the regions of interest were examined to determine if the imaged particles exhibited recent diffusional growth, an indicator of mesoscale updrafts, or riming, a clear marker of SLW or if small graupel were present.

Measurements from a RICE (Heymsfield & Miloshevich, 1989) gave us the ability to detect the presence of SLW in these clouds from temperatures of about $-2$ to $-40 ^\circ C$ or rule out the presence of SLW at values more than the detection limit of the RICE. Work by Heymsfield and Miloshevich (1989) suggests a lower detection limit of $0.002 g/m^3$ for the unit they calibrated. Detailed examination of RICE measurements showed no evidence of SLW above the detection limit in the regions of interest. See Heymsfield and Milosevich for an evaluation of the RICE, the King constant temperature hot wire sensor (which was flown for this project) and the Johnson-Williams (J-W) hot wire device in regions with many ice particles. The J-W hot wire has been used by a number of investigators such as on the NOAA P-3 aircraft in other studies of deep stratiform regions (e.g., Mo et al., 2003; Schuur & Rutledge, 2000a). Both the King probe and the J-W device falsely respond to ice particles, especially large aggregates at warm temperatures, and their baselines drift as an aircraft ascends, descends, and changes speed or temperature.

Radar measurements were obtained from a WSR-74C radar located at Patrick Air Force Base (PAFB) about 25 km south of KSC and the WSR-88D NEXRAD radar located at Melbourne, Florida, about 18 km to the southwest of the 74C radar. The location of the 74C radar was used as the origin in all of our radar plots. The 74C radar provided support for all launch operations at KSC and the Air Force Eastern Range. The 74C was a C-band (5.3 cm) horizontally polarized weather radar without Doppler capability. A complete volume scan was made every 2.5 min. The NEXRAD 88D was the S-band 10 cm Doppler weather radar
used by the National Weather Service in the national radar network. A complete volume scan took ~5 min. See Dye et al. (2007) for additional information and a comparison of reflectivity measurements from the two radars.

The universal format data from both radars were converted to a Cartesian grid with 1-km horizontal and vertical spacing over a 225 × 225-km domain using SPRINT (Mohr et al., 1986). SPRINT was configured to perform a bilinear interpolation with a maximum acceptable distance of only 0.2 km to relocate a closest point estimate with no range interpolation. More details of this process can be found in Dye et al. (2007) and a NASA Technical Report (Dye et al., 2004). Scan gaps occurred when the difference between adjacent elevation sweeps exceeded the beam width of the radar. These gaps in combination with the gridding software sometimes produce a concentric ring appearance of the bright band, cloud tops, and cloud bases in the cross-sectional displays of the reflectivity measurements shown here.

Attenuation of the 74C measured reflectivity was apparent behind regions of heavy precipitation or when the radome of the 74C was wetted because of precipitation. The 74C observations were manually checked for each flight to determine times when attenuation might have occurred. Observations from these time periods were not used. Both radars have a cone of silence directly above the radar that was not scanned because it lies at an elevation angle higher than the elevation of the highest sweep angle. These circular voids in the radar measurements can be seen in some Constant Altitude Plan Precipitation Indicator plots (CAPPIs) presented in following sections.

Two lightning detection systems were used during ABFM II to determine the occurrence, location, and frequency of lightning discharges. The Lightning Detection and Ranging system (Lennon & Maier, 1991), which is a total lightning system, used time-of-arrival techniques to locate the sources of very high frequency radiation at 63 to 69 MHz emanating from parts of the lightning discharge. The Cloud-to-Ground Lightning Surveillance System provided the locations and times of CG return strokes (Maier, 1991).

3. Examples of Electrification in Deep Stratiform Clouds

During ABFM II the UND Citation aircraft was used to investigate 10 cases of deep stratiform clouds as well as a comparable number of anvils. In this section we present measurements from four separate stratiform cases to illustrate the electric field magnitudes and sign, the radar reflectivity development and structure, and the ice PSD and characteristics of the particles themselves that were found in these four cases. Our primary goal in this section is to emphasize the similarity of extensive regions of strong electric fields (>10 kV/m), the highly stratified radar reflectivity, and the uniformity of small and large particle concentrations in these regions. Section 4 will present measurements of the broad size distributions and show CPI images of ice particle characteristics that we observed in different cases. Although we have measurements of the sign and magnitude of the electric field over extended time periods and locations in these clouds, we do not attempt to explain the structure and changes of polarity of electric field that we observed in these stratiform clouds. Our measurements were made primarily along horizontal penetrations. Spiral ascents and descents were needed to allow us to infer more about the vertical electrical structure. Some attempts were made to see if relatively simple charge distributions might explain the observed electric field arrangements in different storms, but we found the sign and location of electric fields and charges to be complex and not straightforward to explain.

3.1. 24 June 2000—Stratiform From Widespread, Multicellular Convection

The first small cumulonimbus that ultimately helped to form this deep stratiform cloud appeared about 25 km north of PAFB at about 1520 UTC with tops to about 10 km. (All times in this paper will be in Coordinated Universal Time, UTC. Subtract 4 hr for local daylight time. Times are given in the format HHMM:SS. Altitudes are referenced to mean sea level.) The first lightning occurred near 1535. By 1615 there was a line of new cells extending ~50 km to the southwest of the initial cell with lightning in several of the cells. An anvil and somewhat later the beginnings of a small stratiform region with tops to ~13 km formed to the northeast of the initial convection. Near 1715 another cluster of cells formed over PAFB and to the southeast with an anvil and the start of a stratiform region spreading to the north and northeast that merged with the previous anvil/stratiform development. There was copious lightning in both of these areas by 1730. A small stratiform area formed from the outflow of a number of cells and by 1845 extended ~75 km north of the active convection with a west-east dimension of ~60 km.
CAPPIs of the cells near 1840 and the developing stratiform area are shown in Figure 1. We will present a number of CAPPIs in this format for different times and different cases. For analysis we animate these CAPPIs to view the history, motion, and development of different storm elements. For each of these the time of the radar volume scan is shown at the top of each figure with aircraft track overlaid for the time period of the volume scan plus 2 min before and 2 min after the volume scan. The origin of these CAPPIs is the WSR74C radar at PAFB. In these figures a CAPPI at the flight level of the aircraft is shown, often in the upper right. Figure 1 shows that lightning was occurring in many different cells at this time.

The aircraft investigation started at 1700 with a penetration at 8 km above the core of a growing convective cell. $E_r$ was highly variable with values as large as -35 kV/m directly over the cell (not shown) and with abrupt changes indicative of nearby lightning.

The aircraft began to investigate the southern part of the stratiform area at about 1800 with penetrations first at 7, 8, and 9 km before climbing to near 11 km as shown in Figure 1. During these passes the magnitude of

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**Figure 1.** CAPPIs at 4-, 7-, and 11-km MSL for the 1839:27 to 1842:05 volume scan of the WSR74C radar on 24 June 2000. The Citation track from ~1837:27 to 1844:05 UTC is shown in red with the initial position shown by a small square. Small Xs show each successive minute along the track. The location of cloud-to-ground lightning is shown by red triangles, while very high frequency sources from the LDAR are shown as black pluses. The LDAR detection threshold was not very sensitive during June 2000. LDAR = Lightning Detection and Ranging; CGLSS = Cloud-to-Ground Lightning Surveillance System.
the electric field, $E_{\text{mag}}$, ranged from a few kilovolts per meter to ~40 kV/m and were highly variable both in magnitude and sign. Figure 2 shows particle concentration measurements from different instruments, aircraft parameters, radar reflectivity, and electric field measurements for the time period 1830 to 1840 near the time of the CAPPIs of Figure 1. The aircraft was ascending from 9.9 to 10.5 km ($C_0$/34 to $C_0$/40 °C) during this period. This pass was about 35–40 km from the closest convective activity. The magnitude and sign of the electric field were quite variable with magnitudes of ~1 to 15 kV/m. Rapid field changes indicative of lightning can be seen near 1831:10, 1833:00, 1835:00, and 1837:10. These variations in electric field are typical of variability that we observed in anvils or stratiform regions near convective cells in early stages of stratiform development where transport of charge from convective cells was probably occurring. Likewise, lightning may have removed or deposited charge because nearby very high frequency sources were sometimes detected by the Lightning Detection and Ranging system. The radar reflectivity structure in Figure 2 is not highly stratified and suggests the aircraft was flying over residuals of

**Figure 2.** Measurements for 1830 to 1840 on 24 June 2000. (top panel) Particle concentrations from different instruments: FSSP total concentration = light, solid line; 2D-C total concentration = bold, solid line; 2D-C concentration >1 mm = dashed line; 1D-C total concentration = dotted line. (second panel) Reflectivity at the aircraft location, bank angle of the aircraft, and ambient temperature. (third panel) Curtain of radar reflectivity above and below the aircraft constructed from 1-km pixels of the Cartesian-gridded measurements. The numbers to the right of the color scale show the lower limit of reflectivity for each color interval; bold, red line = aircraft altitude. (bottom panel) $E_z$, the vertical component of electric field, is a thin, solid line and referenced to the linear scale on the left. $E_q/E_{\text{mag}}$, shown as a dotted line, is a measure of the field due to charge on the aircraft, $E_q$, compared to the magnitude of the ambient field, $E_{\text{mag}}$. It is also referenced to the linear left scale. $E_{\text{mag}}$, the scalar magnitude of the vector field, is shown as a bold, solid line and referenced to the log scale on the right side of the panel. FSSP = Forward Scattering Spectrometer Probe.
Convective cells. The particle measurements show evidence of broad PSDs with high concentrations of smaller particles as well as particles >1-mm size. Some of the particles appear to be small graupel, but it is difficult to determine this with certainty from the 2D-C images. Unfortunately, the CPI data for this day has been lost. It is hard to distinguish the traces of the different particle instruments in the top panel but all are dominated by the concentration of the smaller particles and show similar temporal variations in total particle concentration. After 1836 there is large variability and large decreases in concentration as the aircraft turns and flies in and out of the eastern edge of the cloud.

A review of CAPPIs for this day showed that a complex arrangement of cells sustained this complex stratiform region with bright band from circa 1830 until about 2400, a duration of about 5 1/2 hr. The aircraft landed to refuel at Daytona Beach from 1955 to 2055. During the second flight, stair-stepped horizontal traverses were made along the same approximate path at increasing altitudes from 2115 to 2215 (5 to 11 km; −2 to −45 °C) in this complex, evolving stratiform region with bright band. CAPPIs including lightning of the stratiform region near 2145 when the aircraft was near 9 km are shown in Figure 3. Note the broad area with reflectivity of 20 to 25 dBZ at 7 km and the bright band with radar reflectivity of 35 to 40 dBZ at 4 km. (The concentric arcs of the bright band and in the CAPPI at 9 km are the result of radar scan gaps.) The aircraft track for 2142:38 to 2149:17 is shown. This penetration was from east to west along a path of ~35-km

Figure 3. Same as Figure 1 except for 2144:38 to 2147:17 on 24 June 2000. LDAR = Lightning Detection and Ranging; CGLSS = Cloud-to-Ground Lightning Surveillance System.
length. Lightning was occurring in the small, convective cells 40 to 50 km southeast of the stratiform area being flown in by the Citation but did not extend to the Citation’s location.

The measurements along this track as the Citation ascended from 7.7 km (−19 °C) to 9.2 km (−28 °C) are shown in Figure 4. In contrast to the highly variable electric field trace shown in Figure 2 note the relative smoothness and uniformity of the electric field trace after 2144. $E_{\text{mag}}$ ranged from 20 to 35 kV/m for most of this period with an increase to ~45 kV/m at 2145. $E_{z}$ was negative with values comparable to $E_{\text{mag}}$ indicating that the electric field was primarily vertical. (However, toward the eastern cloud edge on the left side of Figure 4, $E_{z}$ changed sign at 2143:40 from +10 to −30 kV/m.) The radar reflectivity structure was highly stratified with values of 20 to 30 dBZ from the top of the bright band up to 9 km and with decreases toward cloud top. Similarly, the microphysical measurements shortly after 1943 exhibit a high degree of uniformity.

### 3.2. 2 June 2001—Stratiform in Association With an Intense Convective Cell

On this day moderate cells to the west and southwest of PAFB entered the area at about 1750 and moved to the east. By 1900 a cluster of cells about 75 km southwest of PAFB intensified eventually growing into a closely tied cluster of cells that evolved into the intense area of convection from which the stratiform region formed. Aircraft investigations started near 1930 at 9.8 km (−36 °C) first in passes very close to the strong cells and then a little further out in the developing anvil. Electric fields were highly variable with magnitudes as large as 30 kV/m with both positive and negative polarity in different parts of the anvil. 2D-C images suggest some graupel were present. Near 2000 the system began to develop a...
stratiform structure south of the dominant cell. This stratiform area was obvious from the growth of a uniform layer of 20 to 25 dBZ reflectivity at 7-km altitude and above. A penetration at 10 km, about 25 km from the core at 2005, showed variable electric fields with magnitudes of 5 to 20 kV/m and mostly positive $E_z$. Afterward, the Citation flew near an embedded cell that grew to ~10 km at 2100 before collapsing into the growing stratiform region.

A penetration from 2140 to 2150 across the stratiform region is shown in Figure 5. Note the one dominant cell on the north side of this complex with reflectivity of 35 to 40 dBZ extending above 9 km. In the stratiform part of the system there was a large area of 20 to 25 dBZ at 7 km with a few areas of 20–25 dBZ at 9 km. At 7 km, the 20- to 25-dBZ area extended 60 to 70 km in the west to east direction. The airborne measurements corresponding to the pass shown in Figure 5 are presented in Figure 6. From 2140 to 2148 $E_z$ was uniformly +20 to 30 kV/m decreasing to ~8 kV/m before changing sign and gradually decreasing to −30 kV/m, an increase in magnitude. In this case, as with the mature 24 June 2000 stratiform cloud, the fields were relatively uniform, the radar structure was highly stratified with values of 20 to 30 dBZ from 5 to 9 km. However, this case did not show an obvious bright band. The top panel shows that the concentrations of both the smaller particles and those >1 mm size were uniform even after 2148:20 where $E_z$ changed sign. The change in sign of $E_z$ seen in Figure 6 was on the eastern side of the storm where the reflectivity structure was less stratiform-like.
3.3. 10 June 2001—Stratiform Area Mostly Detached From Anvil

On 10 June 2001 a large, strong convective cell with radar reflectivity of 25 to 30 dBZ at 10 km appeared in the research area at about 1750 with smaller cells to its south. Examination of successive CAPPIs showed that a small anvil was growing to the northeast. New cells propagated to the southeast with increasing intensity. Transport from the core into the anvil continued to the northeast but by 1845 at 10 km some of the northern part of the anvil began to develop to the southeast perpendicular to the original anvil. Development followed this same general pattern until 2025 when the aircraft began its investigation with a penetration over the top of a growing convective cell at 9 km. The vertical electric field was $-35 \text{ kV/m}$ in the anvil very close to the convective cell with high variability right over the core with values from $-2$ to $-20 \text{ kV/m}$ with clear indications of nearby lightning (not shown). An hour later near 2035 part of the anvil had evolved into a stratiform-like area of 50 by 60 km in size that was almost detached from the original development as shown in Figure 7.

The airborne measurements corresponding to the plan views shown in Figure 7 are shown in Figure 8. Although the aircraft was making turns during much of the period shown in Figure 8 and did not make a transect directly across the cloud these and following measurements in this area showed very smooth electric field traces with magnitudes as large as 35 to 45 kV/m. Radar reflectivity was highly stratified with values of 20 to 30 dBZ from 6 or 7 to 10 km extended over an area of roughly $40 \times 40$ km. Broad, similar PSDs were observed in this area. These characteristics were the same as pointed out for...
the two previously described cases. In this case, there was no bright band development and the bottom of the cloud barely extended below 0 °C.

3.4. 28 June 2001—Widespread Stratiform Layer With Marginal Electrification

We present this marginally electrified case as a contrast with the stratiform clouds that were highly electrified. A number of widely spaced, modest convective cells, some of which produced lightning, were moving to the NW in a loosely organized zone from SW to NE over the KSC area at 1600, when radar coverage began, with a broad stratiform region with embedded convection trailing behind. These cells were probably the result of the southeasterly flow associated with the tropical wave that extended from near Jacksonville to SW Florida. By 1815 the lightning producing cells and most of the embedded convection were outside of the area and more than 150 km north of PAFB. A weak, widespread stratiform layer that was probably the remnant of the moderate convective cells and embedded convection that had passed KSC earlier persisted over the KSC area at the time of the Citation takeoff at circa 1925.

The stratiform layer even at this time was deep, extending from the surface to 10 to 11 km as seen in Figures 9 and 10. However, the maximum radar reflectivity above 0 °C and at the 6-km altitude of the aircraft was <20 dBZ. $E_{mag}$, the magnitude of the electric field (shown as the bold trace in Figure 10) and other aircraft passes in this cloud were often a few hundred volts per meter with maxima of 1 to ~2 kV/m. Values of $E_z$
(missing in Figure 10) were predominantly positive with largest values of 0.2 to 0.3 kV/m. Since electric fields in clear air outside of clouds at this altitude are often 10 to 100 V/m or less, these clouds were somewhat electrified but only marginally. These measurements suggest that radar reflectivity must exceed 20 dBZ above the 0 °C level for a cloud to become electrified, but one must also consider that the concentration of small and intermediate-sized particles is reduced compared to especially the first two cases. This has an impact on ice-ice collision rate as discussed in section 4. It is also possible that greater electric fields might have been observed if the aircraft had flown above 6 km.

28 June 2001 is the only case that we have examined in which there was more than a 1-min period with limited SLW detected by the RICE probe. Simultaneously the FSSP concentration in Figure 10 show increases at ~1942:30, ~1945:00, and from 1948 to 1950. The RICE measurements also show evidence of sporadic, variable amounts of SLW of 0.004 to 0.020 g/m³. The increased FSSP concentration extend continuously from circa 1948 to circa 1958, and the RICE also shows evidence of SLW up 0.020 g/m³ during this period. Ice particle images from the CPI in some of these same periods show some riming on the diffusionally grown crystals (see section 4.3 below). Both the 2D-C and the CPI during this period show small and intermediate-sized irregular particles as well as some rimed crystals but there was no evidence of millimeter-sized graupel. A little later in the flight a few small graupel were seen by the 2D-C. Clearly, this cloud contained small amounts of SLW at −8 °C (6.0 km) in some locations. We rarely detected the presence of SLW with the RICE in these stratiform clouds and then primarily at temperatures warmer than −10 °C. We were unable to detect this limited amount of liquid water with the King hot wire probe because the variation of the signal

Figure 8. Same as Figure 2 except for 10 June 2001 for 2130 to 2140. FSSP = Forward Scattering Spectrometer Probe.
baseline exceeded the values of liquid water being measured by the RICE. This emphasizes the difficulty in making reliable liquid water measurements in these stratiform clouds especially with the large aggregates found near the melting zone and shifts in baseline caused by changes in altitude, temperature, and airspeed.

4. Observations of Microphysical Content in Deep Stratiform Clouds

In this section we present a comparison of PSDs for the four cases of stratiform clouds described in section 3 above. We also present PSDs observed at different altitudes for the 24 June 2000 case showing the great depth over which broad size distributions were observed. We then present selected images from some of the cases to show the nature of the particles. Finally, we use these measured PSDs to estimate ice particle collision rates.

4.1. Particle Size Distributions

Figure 11 shows PSDs for each of the 4 days described in section 3 averaged over 4- to 6-min periods during which the electric field and particle concentrations were relatively uniform as shown in Figures 4, 6, and 8 and in Figure 10 for part of the 28 June 2001 case. Figure 11 also includes an average PSD for the long-lived anvil case, 4 June 2001, discussed and shown in Figure 5 of Dye and Willett (2007). The PSDs in Figure 11 combine measurements from the PMS 2D-C and the SPEC HVPS. The 2D-C measurements are used for

Figure 9. As in Figure 1 but for 28 June 2001 for 1944:44 to 1947:22.
particles <1 mm in size, while for particles >1 mm the HVPS measurements were used. The lack of abrupt changes at 1 mm attest to the agreement between the two instruments. Although particles <100-μm size were measured by the 2D-C we have not included those measurements in these PSDs because there is large uncertainty in concentration.

Four of the PSDs were from aircraft penetrations at mid to upper levels (3 at 9.3 km and 1 at 8.0 km). The PSD for 28 June 2001 was from the penetration at 6 km (~8 °C) because the aircraft did not fly higher on that day. All of the five PSDs in Figure 11 are broad with many small- and intermediate-sized particles and large aggregates. However, although the PSD for the 28 June marginally electrified case was broad, the particle concentrations were almost a factor of 10 lower for much of the size range compared to the 3 days at the top of the plot. The three PSDs for the 24 June 2000, the 2 June 2001, and the anvil of 4 June 2001 are quite similar. The 10 June 2001 case has fewer small and midsized particles but has a higher concentration of particles >3 mm compared to the other PSDs. We do not have an explanation for why the PSD in this case is different.

Roughly similar PSDs were observed over a considerable cloud depth for most cases and large aggregates of several millimeters in size were observed high in these stratiform regions (for 28 June we have no measurements above 6 km). During the second flight on 24 June 2000, the Citation made stair-step penetrations along the same horizontal path from 2115 (4.5 km, ~0 °C) to 2225 (10.2 km, ~38 °C). Measurements from the penetration at 9.3 km were shown in Figures 3 and 4 in section 3. Figure 12 shows PSDs averaged

Figure 10. As in Figure 2 except for 1940 to 1950 on 28 June 2001. The trace for \( E_z \) (bottom panel) is missing from this plot, but maximum values were 0.2 to 0.3 kV/m.
over a 3- to 4-min period for each penetration from 6.4 to 10.3 km. The PSDs were quite similar over a depth of 4 km on this day with aggregates almost as large as 1 cm in size up to 10 km and also an abundance of smaller particles at 6.4 and 7.7 km. But some increase in the concentration of smaller particles at upper levels and an increase in large aggregates at lower levels can be seen in the figure. For another case, 24 June 2001, a spiral descent (not shown) was made from an anvil at 9.3 km (−30 °C) into a stratiform region with bright band to below the melting level. That descent also shows very broad size distributions with high concentrations of small- and intermediate-sized particles and aggregates as large as 8 to 9 mm throughout the descent until the 0 °C level. We suggest that these broad size distributions with large aggregates is a common feature at temperatures <0 °C throughout much of the depth of deep, precipitating stratiform clouds.

4.2. Images of the Particles

In this section we show examples of images from the CPI, 2D-C and HVPS to highlight a number of small- to intermediate-sized particles to give the reader a sense of the variety of cloud particle types. We are not presenting images from all cases because they are rather similar in the stratiform regions. The lack of evidence of graupel and rimed particles (except for some periods at −6 C in the 28 June 2000 case) in these images is notable. In other regions of these clouds such as near convective cells or not far downwind in anvils, we do observe graupel and rimed particles.

Examples of images from the CPI, the 2D-C and the HVPS for 2 June 2001 are shown in Figures 13a–13c, respectively. The CPI images in Figure 13a were taken sequentially at 9.3 km (−32 °C) during a 2-s interval beginning at 2146:49. It is evident from the CPI particles images that most of the particles are aggregates of smaller crystals, many of which appear to be ~50 to 100 μm in size. Note the sharp, angular edges indicative of active diffusional growth on many of the plates in the aggregates. This demonstrates that the particles were growing in a weak updraft since SLW was not detected. The images from the 2D-C show the highly irregular shape of the aggregates and the feathery, open structure of some of the aggregates. The HVPS images also show the irregular shapes and the propensity for elongated chains of crystals.

Images for 4 June 2001, the anvil case reported by Dye and Willett (2007) are shown in Figure 14. They show similar features as those shown in Figure 13 for 2 June 2001.

The images in Figure 15 are from the marginally electrified case of 28 June 2001. Note than the individual crystals are much larger than the individual crystals seen in Figures 13 and 14, which were from deep stratiform and anvil clouds that had strong electric fields and developed in association with deep convective cells. For the 28 June 2001 case the stratiform clouds that were investigated were probably remnants of earlier widely spaced, lightning-producing, moderate cells that moved across KSC in the southeasterly flow associated with a tropical wave. The images and measurements from the 2D-C and HVPS clearly show small- to intermediate-sized particles as well as the large crystals seen in Figure 15, but the concentrations were smaller compared to the previous cases. The ice particles seen in Figure 15 obviously had a significant amount of time to grow to the large sizes observed in these images and the presence of small amounts of SLW at times suggest that there some weak updrafts were present.

The images in Figure 15 do not show any evidence of riming. However, at times when the RICE showed and the FSSP suggested some limited SLW such as seen in figure in 10 from 1948 to 1950,

Figure 11. Average particle size distribution including measurements from both the 2D-C and High Volume Particle Sensor. The specific times of each average, the altitude, and temperature of each pass were as follows: 24 June 2000: 2157 to 2201; 9.3 km, −30 °C; 2 June 2001: 2144 to 2150; 9.3 km, −31 °C; 10 June 2001: 2134 to 2139; 8.0 km, −19 °C; 28 June 2001: 1941 to 1947; 6.1 km, −7.7 °C; and 4 June 2001 2133 to 2138; 9.3 km, −32 °C.

Figure 12. Average particle size distributions observed on 24 June 2000, flight 2, for straight leg passes at different altitudes and temperatures as indicated.
the images from the CPI clearly show evidence of riming. Examples of particles imaged near 1948:45 to 1948:52 in a region showing evidence of SLW by the RICE and the FSSP are presented in Figure 16. Heavier riming was seen a little later between 1949 and 1950 in regions in which the RICE indicated greater SLW.

Many of the aggregates imaged in regions with strong electric fields were composed of chain-like aggregates of smaller crystals (mostly plates) or with chain-like extensions protruding from more compact aggregates. Some examples can be seen in Figures 13a, 13c, 14, and 17. The shapes suggest that electric field forces had acted to enhance aggregation. A few other investigators have also observed what appears to be the effects of electric fields on aggregation. Stith et al. (2002) is one of the first to comment on chain-type aggregates and possible electric field effects on aggregation. Connolly et al. (2005) reviewed laboratory work on this topic and also presented images of chain-like aggregates in anvils in which lightning was likely to have occurred. The laboratory studies of Saunders and Wahab (1975) and Wahab (1974) showed that in a vertical electric field (i.e., the field was aligned with the direction of fall) a threshold of ice crystal concentration of 250/L and an electric field threshold of ~60 kV/m was required in order to enhance aggregation. In another study Stith et al. (2016) have observed chain-like aggregation of frozen cloud droplets in the upper parts of some

Figure 13. Images for the Cloud Particle Imager (a), the Particle Measuring Systems 2D-Cloud probe (b), and the HVPS (c) for 2 June 2001 at 9.3 km (−32 °C) taken between 1945 to 1946. The 2D-C images show particles from the first buffer for every 10 s at 1945. The width of the 2D-C array is ~1 mm. The HVPS images are from one buffer at 1945:15. The width of the array of the HVPS is 5 cm. HVPS = High Volume Particle Sensor.
anvils in association with high values of NOx in which lightning was known to have occurred. They attributed the aggregation of frozen droplets to electric field forces. None of the above studies had measurements of electric field. Although balloon measurements have shown fields of 75 to 100 kV/m in stratiform clouds and anvils, our field measurements rarely reached 50 kV/m. However, given how frequently our observations show chain-like aggregates we suspect that enhanced aggregation might be happening in weaker electric fields than suggested in the laboratory studies.

Examples of particles imaged during the 10 June 2001 case near 2132 (Figure 8) near the edge of the cloud are shown in Figure 18. The edges of the particles are rounded and without the sharp angles seen in images of crystals in Figures 13, 14, 15, and 17 and thus suggest that sublimation had occurred. The PSD (not shown) was relatively flat with fewer small- and intermediate-sized particles compared to the period from 2135 to 2139 when strong electric fields were observed. Collision rate calculations such as those described in the following section show the collision rate to have been about 50 collisions per cubic meter per second, which is more than 2 orders of magnitude less than for the densest part of the stratiform anvil region from 2135 to 2139. The electric field in this region with sublimating particles was at most only 0.7 to 0.9 kV/m as can be seen from 2130 to 2133 in Figure 8. In addition to a lower collision rate and therefore reduced charge transfer rates between colliding particles, the time scale for the electric field to decay from 50 kV/m to 0 was 500 to 600 s (~10 min) compared to 4,000 to 5,000 s in the dense part of this anvil near 2137. See

Figure 14. Cloud Particle Imager images of particles taken on 4 June 2001 at ~2133:48.

Figure 15. Images of partial crystals from the Cloud Particle Imager on 28 June 2001 at approximately 1944:01. The maximum dimension of the center crystal is about 1.1 mm.
Willett and Dye (2003) for an explanation of the field decay time constant. The weak fields indicate that active electrification had not taken place recently, if at all, in this region of sublimation.

### 4.3. Calculation of Ice Particle Collision Rates

In this section, we will use the observed ice PSDs from the 4 June 2001 anvil case of Dye and Willett (2007) to calculate an estimate of ice-ice collision rates that could occur as a parcel of air containing ice particles exits the convective region and moves downstream forming the anvil and then a mesoscale updraft. The specific collision rate, $C_{D,d}$, between particles of size $D$ and $d$ can be given by

$$C_{D,d} = \pi/4 \ N_D \ n_d \ E \ (D + d)^2 \ (V - v).$$

where $N_D$ and $n_d$ are the concentrations of large particles, $D$, and small particles, $d$, respectively; $E$ is the collision efficiency of $D$ with $d$; $(D + d)^2$ is the combined cross-sectional area; and $V$ and $v$ are the terminal velocities of the faster and slower falling particles, respectively. We assume that $E = 1$ in this calculation. We recognize that this equation applies to spherical particles not aggregates, but it is beyond the scope of this paper to treat this in detail. The main objective here is to examine the evolution of collision rates in the 4 June anvil and of the collision rates in the enhanced anvil in which Dye and Willett (2007) inferred a mesoscale updraft was acting.

![Figure 16](cloud_particle_imager_images.png)

**Figure 16.** Cloud Particle Imager images of rimed crystals on 28 June 2001 near 1948:49.

![Figure 17](cloud_particle_imager_images.png)

**Figure 17.** Cloud Particle Imager images of particles exhibiting long, quasi-linear extensions composed of many individual crystals. The maximum dimensions of the particles on the left and the right are 1.4 mm. From left to right, the particles were imaged on 2 June 2001 at 2143:01; 10 June 2001 at 2135:10; and 10 June 2001 at 2138:28.
The total collision rate, $C_T$, for a given PSD then is given by

$$C_T = \frac{1}{2} \sum_d \Sigma_D C_{D,d}.$$ (2)

For particle terminal velocity in these calculations we used the results of Heymsfield and Westbrook (2010) for aggregates adjusted for air density and temperature at 10-km altitude. A 100-μm ice particle would fall at ~40 cm/s, a 1-mm particle would fall at ~1.6 m/s, and a 5-mm aggregate would fall at ~2.7 m/s. Thus, in a mesoscale updraft in a deep stratiform cloud in a 30-min time period there could easily be 2- to 3-km vertical separation between the 100-μm size particles and millimeter-sized aggregates.

The total collision rate for observed PSDs during the long-lived anvil flight of 4 June 2001 is shown in Figure 19. As seen in the figure, if the collision rate is calculated including the estimate concentration of all observed particles greater than 50 μm in size, the total collision rate almost doubles compared to only using particles greater than 100 μm in the calculation. The collision rate is almost halved if only particles greater than 200 μm are included in the calculation. This demonstrates how greatly the collision rate depends upon the concentration of the smallest particles.

The first penetration of this storm was made immediately adjacent to the storm core at 2012 to 2017 and had the highest collision rate during this flight because of the large number of small particles. Graupel, other rimed particles, and ice crystals, most likely transported from lower levels in the convective core, are suggested by the 2D-C and CPI images on this pass at 9.3 km ($-31^\circ$ C). Cross-anvil penetrations were made from circa 2020 to 2100 at successive distances downwind of the core and show the lowest collision rates even though each was made through the high reflectivity of the anvil at that location and time. Both of the passes showed some graupel, some rimed plates, many rimed, irregular particles and many medium to large particles.

Figure 18. Cloud Particle Imager images of sublimating particles taken near 2132 on 10 June 2001.

Figure 19. The total collision rate as a function of time for the 4 June 2001 flight calculated based on the measured ice particle size distributions.
aggregates (with and without riming; Dye & Willett, 2007). Beginning near 2110 the aircraft began to make penetrations along the axis of the anvil toward and away from the core in what Dye and Willett called the enhanced anvil, a region in which they inferred there was a mesoscale updraft. In this region the CPI mostly showed particles with sharp, angular edges as seen in Figure 14, but at other times closer to the edge of the cloud some appear slightly rounded. Evidence of riming was not apparent. Note that between 2133 and 2137 the collision rate for particles of >100 or >200 μm is almost as great as the first pass, which was nearest the core and more than a factor of 3 greater than the cross-anvil passes. This is expected as Figure 8 in Dye and Willett (2007) shows that particle concentrations for particles <500 μm in size were a factor of 2 to 3 less for the 2021 cross-anvil pass compared to 2136 in the enhanced anvil. Note also that the collision rate was also relatively uniform over a period of about 5 to 6 min (~35 to 40 km at the flight speed of 120 m/s of the Citation). Figure 19 provides support for an increased collision rate, increased charge separation rate, and enhancement in electric field in the mesoscale updraft as suggested by Dye and Willett (2007) in comparison to values earlier in the anvil history.

5. Charge Transfer From Nonriming Ice-Ice Collisions

Our observations of strong electric fields, highly stratified moderate reflectivity and diffusional growth on ice particles without evidence of riming existing over large distances and for long time periods in deep stratiform and anvil clouds strongly suggest than an in situ charging mechanism without SLW was taking place in mesoscale updrafts. Laboratory studies also report charging from ice-ice collisions even without SLW being present. Most laboratory work on charge separation has been devoted to investigating ice crystal collisions with simulated riming graupel in the presence of supercooled water (e.g., Reynolds et al., 1957; Takahashi, 1978; Jayaratne et al., 1983). Williams et al. (1991) reviewed the studies conducted up to that time to examine the sign of charge transfer in both riming and nonriming conditions. Most of the laboratory studies have focused on electric charge transfer in riming conditions because the charge transfer was often found to be 2 orders of magnitude greater with riming as compared to nonriming conditions. Nevertheless studies such as those of Gaskell and Illingworth (1980), Jayaratne et al. (1983), Baker et al. (1987), Caranti et al. (1991), Keith and Saunders (1990), Saunders et al. (2001), and Luque et al. (2016) have reported weak transfer to a simulated graupel particle even without SLW being present. Takahashi (1978) only detected charge transfer if SLW was present; however, there is no mention of the charge transfer detection threshold in his paper. The number of studies showing some weak charge transfer leaves little doubt in our minds that some weak charge transfer does occur during collisions between colliding ice particles even without SLW being present.

Most of the studies of collisions between ice crystals and simulated graupel targets in nonriming conditions were done to try to understand the physical mechanism responsible for the charge transfer in riming conditions and were performed in conditions in which the simulated graupel were either heated or cooled artificially and sublimating or growing by diffusion. Consequently, the ice crystals and target were at different temperatures and in different super or subsaturated conditions with respect to ice. Most of these experiments showed that the target acquired negative charge when the target was sublimating but positive charge if the target was growing by diffusion (e.g., Jayaratne et al., 1983; Baker et al., 1987; Caranti et al., 1991; Keith & Saunders, 1990; Saunders et al., 2001).

In a mesoscale updraft both particles would experience roughly the same supersaturation with respect to ice except for differences due to ventilation (Williams et al., 1991) and thus be growing by diffusion or, if in descending air both particles, would be sublimating. Baker et al. (1987) found positive charging when the target was growing by diffusion and negative charging when it was sublimating. Interestingly, when the ice crystals and target were at the same temperature they found weak negative charging.

Until the recent work of Luque et al. (2016) environmental conditions were not carefully controlled and measured. Luque et al. examined the charge transfer between ice crystals and target in an environment that was at equilibrium and supersaturated with respect to ice for both the crystals and the target. Experiments were performed from −8 to −21 °C with a relative humidity with respect to ice of 106% to 116%. At temperatures >−10 °C the sign of charging to the target was positive. But at temperatures from −10 to −21 °C they reported that the charge transfer to the target to be predominantly negative with values varying from −0.01 to −0.3 fC per collision. They attributed this change of sign at −10 °C as an indication that they did not
succeed in maintaining the humidity below saturation with respect to water for temperatures greater than −10 °C.

In a previous study at low liquid water contents (<0.5 g/m³), Ávila et al. (2013) found roughly the same magnitude and sign of charge transfer as the results of Luque et al. They also found that the charge to the target changed sign from positive to negative at temperatures <−10 °C. The similarity in the change of sign at −10 °C in both sets of experiments is curious, making us wonder if the change in sign of the Luque et al. results were not because water saturation was reached but rather something else. There is a change in the growth habit for crystals growing by diffusion from columnar to plate growth at about −8 °C, which is near −10 °C (Bailey & Hallett, 2009).

It is clear that many issues remain and that additional laboratory studies are needed in which the humidity and temperature in the chamber is varied, controlled, and measured. Experiments are also needed in an environment in which both particles are sublimating as well as ones in which diffusional growth is occurring. But if we consider the results of those experiments in which the temperature of the target was maintained at the same temperature as that of the ice crystals (e.g., Ávila et al., 2013; Baker et al., 1987; Luque et al., 2016) they would suggest that the target would acquire a negative charge during diffusional growth in a mesoscale updraft.

The charge transfer results of Luque et al. and others cited above were obtained from collision of ice crystals growing by diffusion with rimed targets that were then exposed to diffusional growth. In order to use their results to calculate charge transfer between ice crystals and aggregates, both growing by diffusion, we must assume that the charge transfer between crystals and aggregates is the same as the targets in the Luque et al. and other experiments. When ice particles of any type are growing by diffusion, the crystal faces will be growing. We suggest that the active surface for charge transfer on the growing particles could be similar whether the particles are ice crystals, aggregates, or graupel. Saunders et al. (2001) showed that the sign of charge transfer (and therefore the nature of the surface) responded almost immediately when moist air was directed at their simulated graupel target. Even limited diffusional growth seemed to alter the sign of charge transfer and evidently the nature of the surface. Similarly, Caranti et al. (1985) found that for an initially sublimating ice target as little as 1 s of time and 1 μm of diffusional growth on the target changed the sign of charging from negative to positive. These results suggest that perhaps the magnitude of charge transfer between colliding ice crystals and aggregates might be approximately the same as between ice crystals and a graupel surface growing by diffusion but not by riming, a point of view taken by Williams et al. (1991).

The observations and laboratory studies discussed above suggest that nonriming collisional charge separation is a likely charge transfer mechanism in these stratiform and anvil clouds. However, we have insufficient knowledge from laboratory studies to be able to robustly quantify the amount of charge separated by this mechanism. As a step in trying to roughly estimate the amount of charge that might be separated we will assume that the charge transfer per collision is within the limiting values of charge separation per collision found by Luque et al. (2016). Recall that their experiments were with small ice crystal of 20 to 25 μm colliding with simulated graupel at 3 m/s. The actual charge transfer in clouds is likely to be a function of particle size, differential terminal velocity, nature of the particles, and perhaps temperature or other factors yet unknown.

Assuming that the charge separation is taking place in an area of 20 by 20 km and a depth of 2.5 km, the active volume would be 10³ km³ or 10¹² m³. We further assume that the process occurs over a time period of 30 min (1.8 × 10³ s) and a collision rate of 5 × 10⁵ collisions per second, the rate calculated in Figure 18 for the enhanced anvil of 4 June 2001. This yields the total number of collisions in this 20 × 20 × 2.5 km volume in 30 min to be 9 × 10¹⁸ collisions. We do not know what fraction of the crystals that collide with aggregates or other particles succeed in separating from the collision (or dislodge other parts of the larger particle) and in transferring charge. Keith and Saunders (1989) used rods of 0.5- to 5-mm diameter to examine the collision efficiency and probability of separation and bounce off of ice crystals of different sizes and speeds colliding with the rods. For plates >60 μm in size colliding at 3 m/s they found the collision efficiency to be ~1. From these experiments they determined event probabilities (the collision efficiency times the separation probability) that ranged from <0.1 for small plates to ~0.5 for ice crystals of 400 μm separating from a 3-mm rod. This yields a separation probability ranging from <0.1 to 0.5. For our calculations of estimated total charge transfer we have assumed the low value of 0.1, but there is great uncertainty in this value.
Using a separation probability of 0.1 and charge per collision of 0.01 fC or 0.2 fC, the range of the charge transfer rate reported in the Luque et al. experiments for 20- to 25-μm crystals, we calculate an estimated amount of total charge transfer in this volume in this 30-min time period as 9 to 180 C of total charge. A typical simple lightning flash discharges roughly 5 to 20 C of charge. Those producing transient luminous events discharge hundreds of Coulombs of charge (Lang et al., 2010), but they occur over a much larger area and mostly lower in the cloud. There are many uncertainties in this calculation, but the estimate, as rudimentary as it is, does suggest that substantial charge could be separated.

6. Discussion and Concluding Remarks

The existence of strong fields in trailing stratiform regions behind MCSs or in thunderstorm anvils has been well documented from many balloon-borne electric field measurements. See Stolzenburg and Marshall (2008) for a brief overview. From these measurements they and many others have inferred that there must be extensive layers of substantial charge that can persist for hours in deep stratiform regions. Measurements from lighting mapping arrays (LMAs) have also provided evidence of extensive, long-lasting bilevel layers of charge in deep stratiform and anvil clouds (e.g., Kuhlman et al., 2009; Tessendorf et al., 2005; Weiss et al., 2012; Wiens et al., 2005).

The observations in Florida presented here and in Dye et al. (2007) and Dye and Willett (2007) are the first combined electric field and detailed microphysical measurements reported from aircraft penetrations made at middle to upper levels (temperatures < –10 °C) in deep stratiform clouds and anvils. The measurements were made in a variety of different types of stratiform cases and anvils but the emphasis was at 7 to 10 km to explore the strength of electric fields and electric field decay in anvils. Not surprisingly in view of previous balloon measurements, they showed the existence of thunderstorm strength fields at middle and upper levels (6 to 10 km) over extensive distances in several different deep stratiform clouds. The strong electric fields were quite uniform without short time scale variations, unlike the same measurements made during our penetrations near and not far downwind of convective cores. These fields were colocated in or above regions with highly stratified, moderate radar reflectivity of 20 to 25 dBZ at 5 to 10 km (sometimes 25 to 30 dBZ at 5 to ~7 km). The PSDs in these vertically deep regions were broad with numerous small-sized and midsized particles and aggregates up to about 1 cm in size even high in the clouds. Most of our measurements were at 8- to 10-km altitude with only a few measurements at 11 km or above. One of us (A. B.) has also observed these broad size distributions containing large aggregates over a large vertical extent of deep stratiform clouds in spiraling ascents and descents over Oklahoma as shown in Figure 5 of Jensen et al. (2016). We suggest that these very broad size distributions with large aggregates are indicators of mesoscale updrafts. Additionally, images of the particles from the CPI in Figures 13a, 14, and 17 definitively show diffusional growth with no evidence of riming. This is a clear indication of upward moving air. In other regions in which the RICE and FSSP suggest some small amounts of SLW, the CPI particle images such as in Figure 16 show riming. These particles were located in regions in which the total ice particle concentration was low compared to regions in which we think a mesoscale updraft was acting.

Mesoscale updrafts in deep stratiform and anvil clouds create an environment in which ice particles can grow by diffusion and aggregation over a large area and a deep volume of the cloud for a long time period. Because mesoscale updrafts often cover an extensive area of hundreds to a few thousand square kilometers, microphysical development is similar over broad regions and over long periods of time thus leading to the development of extensive areas of moderate reflectivity of 20 to 25 dBZ. Simultaneously, as a result of the broad size distributions with large aggregates there are high ice particle collision rates, which can potentially separate substantial charge. Thus, the areas with layers of uniform moderate radar reflectivity are the same as those with strong uniform electric fields. Broad areas in stratiform clouds or anvils with layers of uniform reflectivity of 20 to 25 dBZ (about –10 to –20 °C in summertime Florida) are good indicators of the likelihood of strong electric fields and possible areas of triggered or natural lightning hazard.

Measurement from the Rosemount Icing Detector, which is very sensitive to small amounts of SLW with a detection limit of about 0.002 g/m³, showed no evidence of detectable SLW in any of the stratiform or anvil clouds at temperatures colder than –10 °C other than at the edge of the cloud or regions in which ice particle concentrations were small. Laboratory studies show that even without SLW being present, some weak charge transfer occurs when ice particles growing by diffusion collide and separate. Given the lack of...
SLW, the broad size distributions and the high collision rates in these clouds we infer that the charge transfer mechanism responsible for the strong electric fields in these deep stratiform and anvil clouds is most likely the nonrimming ice collision mechanism.

While mesoscale updrafts are commonly referenced in much of the cloud and mesoscale dynamics literature, they are normally discussed for stratiform areas in association with severe storms and mesoscale convective systems. Yet we suspect that weak, relatively long-lasting updrafts may be commonplace in many anvils and stratiform regions because our observations showed moderate, stratified radar reflectivity; uniform, strong electric fields; and broad size distributions on scales ranging from about 10 km to many tens of kilometers. Boccioppio (1996) using volume velocity processing Doppler radar techniques, which average over large areas, identified regions with weak updrafts and front to rear flow in storms too small to be considered MCSs in Florida, New Mexico, and the western Pacific. Based on microphysical retrievals and laboratory results available at that time he concluded that his results were not inconsistent with in situ charge separation.

We wonder how small particles in the broad PSDs can be maintained over a large altitude depth in the mesoscale updraft. Aggregation of the smaller ice crystals to form large aggregates should gradually reduce the concentration of small particles. Yet PSDs at different altitudes such as in Figure 12 show copious amounts of ice crystals <100 μm in size. As in the 4 June 2001 anvil case the enhanced anvil was about 100 km at the closest point from strong convection that might have carried small particles by advection. This suggests to us that new ice particle formation might be occurring in these broad mesoscale updrafts.

We found that even in anvils where cloud base barely reached the melting zone, some of our cases at middle to upper levels showed the presence of strong electric fields, highly stratiform reflectivity structure and broad PSDs with large aggregates. Examples of this were the 10 June 2001 case shown here and the long-lived anvil cases of Dye and Willett (2007). Based on these observations we infer that the in situ charging mechanism responsible for the electrification at middle to upper levels in these deep stratiform and long-lived anvils is not a result of nor dependent upon charge separation occurring near the melting zone. It seems possible that nonrimming charging might contribute to in situ charging between −10 °C and the melting level, but we have not investigated this. Weak downdrafts are often found between −10 °C and the melting zone (Protat & Williams, 2011) so the particle spectra and characteristics as well as charge transfer are likely to be different than at temperatures <−10 °C. Additionally, in situ charge separation due to the classic noninductive rimming electrification mechanism can be acting in regions with SLW.

Williams (2018) has pointed out the importance of the zone between 0 and −10 °C in winter storms, in the End of Storm Oscillation of dissipating summer convection (e.g., Marshall et al., 2009) and as the source of flashes that produce transient luminous events (e.g., Lang et al., 2010). He calls attention to the existence in these situations of layers of positive charge near and slightly above 0 °C with a layer of negative charge above. He refers to this as the “snow dipole” because large aggregates or snowflakes are observed in these zones in contrast to the “graupel dipole” of convective regions with negative charge located near −10 °C with positive charge found above. Thus the snow dipole has the opposite polarity of the graupel dipole. He postulates that in situ charging is occurring in the snow dipole by collisions between ice particles and the larger aggregates. This is similar to the nonrimming collisional charging we infer to be occurring at colder temperatures in deep stratiform and anvil clouds, but the warmer temperature zone presents complications because both SLW and rimed ice particles are sometimes found in this zone, for example, our 28 June 2001 case. It is further complicated because it is unclear if this zone has weakly ascending air (probably if SLW is found) or in other cases weakly descending air. As Williams indicated, observations of electric fields, microphysics, and dynamics are needed in these regions.

If nonrimming collisional charging was occurring in a mesoscale updraft in a deep stratiform or an anvil cloud, one would expect to see the vertical charge profile exhibit a bilevel structure of oppositely charged layers due to the gravitational separation of the smaller and the larger charged particles. For example, a layer of negative charge lying above a layer of positive charge or vice versa. The strength and the vertical profile of the mesoscale updraft would likely influence the vertical structure and perhaps strength of the electric field and the altitudes of the bilevel layers. We as a community have little information on the mesoscale updraft profile. However, we do know that as the air mass at the top of the cloud approaches the tropopause the strength of the updraft must diminish. Thus, there is likely to be a maximum in
upward motion somewhat below cloud top. This will act as a balance and accumulation level for smaller particles with terminal velocities close to that of the updraft maximum. If the particles in this size range are ones that carry a substantial amount of charge as a result of nonriming collisional interactions, charge will accumulate at those levels. Particles with terminal velocities greater than the updraft maximum will fall in the cloud. With this scenario there could be a layer of charge somewhat near the top of the cloud with a thicker layer of charge of opposite polarity below.

Many investigators have suggested the existence of screening layers at the top of anvils and stratiform regions in order to explain the observed layers of charge in the top kilometer of the cloud (e.g., Marshall et al., 1989). There is a strong physical rationale for the presence of screening layers in the top kilometer or so of electrified stratiform clouds (Brown et al., 1971; Klett, 1972). Byrne et al. (1989) argued that a layer of charge 2 to 4 km below cloud top might have formed from the sedimentation of charged particles in an older screening layer from higher in the cloud, but he pointed out that this was speculative and that observations were needed to test this hypothesis. In both of these papers in situ charge separation from nonriming collisional charging was not considered. If the nonriming collisional charging mechanism is acting to produce one sign of charge in the upper portion of a cloud with a layer containing the opposite charge below, the top layer of charge would reduce the electric field near and slightly above the top of the cloud. In effect this would provide the “screening” layer for a charged layer below.

We suggest that an alternate explanation for the upper negative layer in the top couple of kilometers from cloud top and the lower positive layer of charge and sometimes increases in electric field observed by a number of balloon-borne investigations might be the result of nonriming collisional charging in a mesoscale updraft. Likewise, studies using LMA measurements have also shown two layers of charge in anvils and stratiform regions and authors have inferred that the layer near the top of the cloud was a screening layer. For example, a paper by Kuhlman et al. (2009) reported lightning initiated several tens of kilometers out in the anvil of a storm in Oklahoma and showed an extensive layer of positive charge in the middle of the anvil with a negative charge layer above that was a couple of kilometers below radar cloud top. Similarly, Tessendorf et al. (2005) and Wiens et al. (2005) examined LMA observations in combination with polarization radar measurements to study the formation of hail and the evolution of the lightning and charge structure in a severe supercell storm in northwestern Kansas. Wiens et al. show a bilevel charge structure extending from the convective region into the thick anvil with a negative layer at 10 to 11 km above a positive layer at 8 to 9 km. Tessendorf et al. in their Figure 2 present fuzzy logic hydrometeor classification results showing a patchwork of vertical ice (i.e., ice crystals) and dry snow (i.e., dry aggregates) all the way across the stratiform region and from ~6- to 13-km altitude for at least 40 km downwind of the convective core. A vertical section of the dual Doppler wind field in that paper suggests weak upward motion from ~7.5- to 10-km altitude and perhaps higher. The hydrometeor classification results and also the weak upward motions are consistent with our microphysical measurements and our inference of a mesoscale updraft. The existence of a bilevel charge structure is consistent with nonriming collisional charging in a mesoscale updraft.

The laboratory results reviewed in section 5 tend to suggest that the larger particles, that is, aggregates, would acquire a negative charge during collisions if diffusional growth is occurring on both ice crystals and aggregates. Yet many field observations such as those of Stolzenburg and Marshall (2008), Kuhlman et al. (2009), Tessendorf et al. (2005), and Wiens et al. (2005) suggest a layer of positive charge in the main body of the anvil with negative charge above. But as seen for the Weiss et al. (2012) observations and some cases from this study, there is not a consistent pattern. We do not have a good explanation for this. Perhaps the mesoscale updraft and nonriming collisional charging develops lower in the cloud in some cases. The presence or absence of mesoscale downdrafts in lower levels (≥ −10 °C) would also complicate the charge structure.

It seems clear that additional laboratory studies, quantitative modeling simulations, and coordinated airborne and multiparameter radar observations of the kinematic, microphysical, and electrical evolution of deep stratiform clouds and long-lived anvils are needed. Because currently only aircraft can carry a comprehensive payload of microphysical instruments and electric field mills and make measurements over a larger area of 10 km or more, it seems necessary to make airborne spiral ascents and descends in deep stratiform.
clouds and anvils to address this problem. Although difficult, these flights should be coordinated with balloon-borne electric field measurements, with measurements from multiparameter, dual Doppler radars to determine the large-scale microphysical and airflow structure of the storms and from 50-MHz Doppler radar profilers. A main focus of such a field program should be to observe and determine the life cycle of mesoscale updrafts in combination with the microphysical, electrical, and radar observations.

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