Electrification Life Cycle of Incipient Thunderstorms

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Key Points:

- The majority of first cloud-to-ground flashes are preceded by intracloud lightning
- Initial positive differential reflectivity anomaly is present aloft, associated with supercooled raindrops
- Reduced differential reflectivity is present before and during the time of the initial lightning flash
Index terms

3304 - Atmospheric Processes: Atmospheric electricity
3314 - Atmospheric Processes: Lightning
3324 - Atmospheric Processes: Convective processes
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Keywords

Polarimetric Radar; Lightning Localization System; First Cloud-to-Ground Lightning Flash; First Radar Echo, Graupel; Supercooled Raindrops, Differential Reflectivity
Abstract

This work evaluates how clouds evolve to thunderstorms in terms of microphysical characteristics and produces the first intracloud (IC) and cloud-to-ground (CG) lightning flashes. Observations of 46 compact isolated thunderstorms during the 2011/2012 spring-summer in Southeast Brazil with the X-band polarimetric radar and two- and three-dimensional Lightning Location Systems demonstrated key parameters in a cloud’s vertical structure that produce the initial electrification and lightning activity. The majority (98%) of the first CG flashes were preceded (by approximately 6 min) by intracloud (IC) lightning. The most important aspect of the observations going into this paper, which came originally from the visual examination of a large number of thunderstorms, is that an initial positive differential reflectivity ($Z_{DR}$) (associated with supercooled raindrops) evolved to reduced $Z_{DR}$ (and even negative values) in the cloud layer between 0° and to -15°C before and during the time of the initial lightning, suggesting evolution from supercooled raindrops to frozen particles promoting the formation of the conical graupel. An enhanced negative specific differential phase ($K_{DP}$) (down to -0.5 ° km$^{-1}$) in the glaciated layer (above -40 °C) was predominantly observed at the time of the first CG flash, indicating that ice crystals, such as plates and columns, were being aligned by a strong electric field. These results demonstrate that the observations of $Z_{DR}$ evolution in the mixed layer and negative $K_{DP}$ in the upper levels of convective cores may provide useful information on thunderstorm vigor and lightning nowcasting.
Lightning is recognized as an important atmospheric component; however, the knowledge about thunderstorms electrification process and its life cycle is still limited. Understanding this process and the thunderstorm life cycle will help develop nowcasting tools and lightning parameterization in numerical weather models. The common aspect provided by experimental simulations and observational studies is that collisions between graupel and small ice crystals inside an environment dominated by supercooled water and strong updrafts is the primary process of thunderstorm electrification [Reynolds et al., 1957; Takahashi, 1978]. Cloud microphysical estimations from weather radar have provided a significant improvement in the understanding of hydrometeors characteristics and the production of lightning in thunderstorms [Workman and Reynolds, 1949; Reynolds and Brook, 1956; Goodman et al., 1988; Buechler and Goodman, 1990; Hondl and Eilts, 1994; Jameson et al., 1996]. Nevertheless, the understanding on how the clouds evolve, from the first cloud droplets to full thunderstorms and how the first intracloud (IC) lightning and cloud-to-ground (CG) lightning flashes are produced using an large sample of compact and isolated thunderstorms is still lacking.

Observations provided by single and dual-polarization radar have shown an organized progression from the first development of supercooled raindrops to the initial electrification [Workman and Reynolds, 1949; Reynolds and Brook, 1956; Krehbiel, 1986; Goodman et al., 1988; Ramachandran et al., 1996; Jameson et al., 1996; Bringi et al., 1997; Lopez and Aubagnac, 1997; Carey and Rutledge, 2000; Bruning et al., 2007; MacGorman et al., 2008; Woodard et al., 2012; Mecikalski et al., 2015; Stolzenburg et al., 2015]. Workman and Reynolds [1949] evaluated the initial electrification in twelve summer storms in New Mexico and reported the occurrence of the first IC lightning...
approximately 13 min following the first radar echo and coincident with the time that
the radar echo begins to descend. Meanwhile, Reynolds and Brook [1956] showed that
initial electrification in thunderstorms is associated with a rapid vertical development of
the initial radar-precipitation echo. Goodman et al. [1988] used polarimetric radar and
documented the first IC lightning event approximately 4-6 min after water at the top of
a supercooled column froze and radar-inferred graupel particles had formed. On the
other hand, Ramachandran et al. [1996] analyzed several convective cells within two
Florida storms with in situ aircraft and polarimetric radar observations. They observed
the first CG lightning 15 min after the first radar echo and after the development of ice
phase precipitation via the freezing of supercooled rain within the upper portion of the
updraft region, followed by the appearance of an electric field of tens of kilovolts per
meter close to -6°C isotherm. Additionally, Jameson et al. [1996] showed that the onset
of electrification coincided with the appearance of significant volumes of differential
reflectivity (Z_{DR}) above the -7 °C level. More recently, Stolzenburg et al. [2015]
examined the initial electrification of New Mexico thunderstorms and documented the
initiation of the first IC flash at temperatures between -10° and -20°C and 5-8.6 min
after of the earliest deflection of electric field at surface. The authors documented that
the initial electrification became evident at the surface after stronger reflectivity Z_{H} (40
dBZ) developed above the -5°C isotherm, and a rapid growth of the surface electric
field was observed after this Z_{H} extended above the -15°C isotherm. The results
suggested also that the radar data with higher time resolution could improve the lead
time by 2-7 min in detecting the onset of initial electrification compared with the
electric field measurements at the surface.

Indeed, the aforementioned studies show the importance of the regions with larger
raindrops above the melting layer (defined as +Z_{DR}-columns) for the formation of
graupel embryos, that in turn, such mixed-phase hydrometeors (graupel, ice crystals and supercooled raindrops) are fundamental for the early cloud electrification. $Z_{DR}$-columns and their vertical extent are well correlated with the updraft intensity [Hall et al., 1980, 1984; Caylor and Illingworth, 1987; Illingworth et al., 1987; Bring et al., 1991; Herzegh and Jameson, 1992; Conway and Zrnic, 1993; Ryzhkov et al., 1994; Hubbert et al. 1998; Smith et al. 1999; Picca et al., 2010; Kumjian et al., 2012; Kumjian et al., 2014; Homeyer and Kumjian, 2015; Snyder et al., 2015]. The pioneering work presented by Hall et al. [1984] documented a narrow column of strong $Z_H$ ($\sim 48$ dBZ) and positive $Z_{DR}$ ($\sim +2.5$ dB) reaching 1.5 km above the melting layer, which was interpreted as supercooled water carried upward by strong updraft. Illingworth et al. [1987] also documented positive $Z_{DR}$ ($\sim +3$ dB) and moderate $Z_H$ ($\sim 30$ dBZ) extending up to the -10°C level in developing convective clouds and attributed this observation to a low concentration of large (> 4 mm diameter) supercooled raindrops. Their observations suggested also that $Z_{DR}$-columns persist for less than ten minutes with drop diameters between 1 and 2 km. The authors argued that such a low concentration of large raindrops could be efficient as hail embryos, growing to large hailstones after the freezing process due to the lack of competition for cloud water. Hubbert et al. (1998) evaluated polarimetric radar measurements and ground observations of hail in Colorado and documented that 30-40 % of the hailstones contained frozen drop embryos. This result showed the importance of mixed-phase particles in the upper portion of the $Z_{DR}$-columns for the formation of hail. In addition, Smith et al. [1999] evaluated $Z_{DR}$ signatures and linear depolarization ratio (LDR) using 3-cm radar and compared with in situ photographs and shadowgraphs of precipitation particle observations by the T-28 aircraft in a Florida convective storm. $Z_{DR}$-columns presented large positive $Z_{DR}$ (+2-3
dB) and large LDR (-24 and -30 dB) close to -10°C level, possibly indicating the initial freezing of rain.

More recently, Homeyer and Kumjian [2015] analyzed the behavior of the polarimetric signature in three different overshooting convective thunderstorms: organized convection, discrete ordinary convection, and discrete supercell convection. Their results revealed deep columns of highly positive $Z_{DR}$ and specific differential phase (KDP) representing the lofting of liquid hydrometeors within the convective updraft and above the melting level. The formation of $Z_{DR}$-columns is frequently attribute for the condensation and coalescence process through of updrafts regions [Illingworth et al., 1987] and alternatively associated the existence of ultragiant nuclei serving to the growth of large drops [Caylor and Illingworth, 1987]. These regions are a probable source area for graupel embryos, produced by raindrops frozen inside those columns, which are fundamental for early electrification and lightning [Conway and Zrnic, 1993; Carey and Rutledge, 1998]. Woodard et al. [2012] showed that the $+Z_{DR}$-columns are a useful parameter for radar-based operational forecasting algorithms for lightning initiation.

Many works have demonstrated that electric fields in clouds can also influence the orientation of ice crystals with oblate and prolate shapes [Weinheimer and Few, 1987; Metcalf, 1993, 1995; Krehbiel et al., 1996; Caylor and Chandrasekar, 1996; Metcalf, 1997; Foster and Hallett, 2002; Ventura et al., 2013; Mattos et al., 2016]. In conditions of strong electric field ice crystals are mostly vertically-oriented. When detected by polarimetric radars these hydrometeors traditionally present negative values for differential propagation phase ($\varphi_{DP}$) [Caylor and Chandrasekar, 1996; Ryzhkov and Zrnic, 2007; Carey et al., 2009] and negative KDP [Caylor and Chandrasekar, 1996; Carey et al., 2009; Hubbert et al., 2014a; Mattos et al., 2016]. For example, Caylor and
Chandrasekar [1996] found that $\varphi_{DP}$ and $K_{DP}$ gradually decrease and then abruptly increase after lightning. The authors documented changes of $5^\circ$ in $\varphi_{DP}$ and a systematic minimum peak in $K_{DP}$ of approximately $-0.9 \,^\circ\text{km}^{-1}$ in the height range 7-14 km just before lightning occurrences. More recently, Mattos et al. [2016] documented the polarimetric behavior as a function of the lightning density and observed that negative $K_{DP}$ is observable only in conditions with the strongest lightning frequency. However, the larger unaligned graupel-ice crystals mixture could mask the electrical alignment of ice crystals as indicated by the change in $K_{DP}$ measurements as discussed in previous studies [Marshall et al., 2009; Carey et al., 2009]. Carey et al. [2009] evaluated the polarimetric signatures of ice particles using a T-Matrix approach and found that in regions with $Z_H > 40-45 \,\text{dBZ}$, the $|K_{DP}|$ of graupel can be on the order of the $|K_{DP}|$ of vertically-oriented plate ice crystals. These results suggested that in a graupel-plate mixture, horizontally-oriented graupel can mask the electrical alignment signature that would otherwise be present in the $K_{DP}$ of vertically-oriented plates alone. Recently, Hubbert et al. [2014a] evaluated the nature of the ice crystals causing negative $K_{DP}$ in polarimetric measurements from S-band polarimetric radar. The results suggested two types of ice crystals: 1) smaller aligned ice crystals (columns or plates) with relatively small $Z_H$ and 2) larger aggregates or graupel randomly oriented with larger $Z_H$ that masks the $Z_{DR}$ of the smaller aligned ice crystals. The interpretation of these results has been supported by satellite measurements. Satellite measurements at microwave frequencies (85 GHz) have shown negative differences in brightness temperature (TB) in the 85 GHz channel (TBV-TBH) associated with strong lightning frequency [Prigent et al., 2005; Mattos and Machado, 2011]. Mattos and Machado [2011] documented that CG lightning rate increases linearly with polarization reduction, suggesting that such
polarization differences could be explained by relatively large and non-spherical particles that are mostly vertically oriented. Some works showed also that stratiform regions of summer squall lines can exhibit strong electric fields sufficient for lightning initiation [Chauzy et al., 1980; Engholm et al., 1990], which could promote negative $Z_{DR}$. On the other hand, snowstorms with moderate snowfall rates probably do not have strong electric fields. For example, Williams et al. [2015] documented predominantly positive $Z_{DR}$ in snowstorms and warm season stratiform systems, indicating that the vertical electric fields they contain are not sufficient to reorient ice crystals and change the sign of $Z_{DR}$.

Other works have documented that negative $Z_{DR}$ could be caused also by graupel particles [Wiens et al., 2005; Dolan and Rutledge, 2009; Evaristo et al., 2013; Homeyer and Kumjian et al., 2015; Stolzenburg et al., 2015; Oue et al., 2015; Bringi et al., 2016]. For example, Dolan and Rutledge [2009], using a T-matrix scattering model for several different hydrometeor types, suggested the presence of negative $Z_{DR}$ associated with graupel particles. High-density graupel may have negative $Z_{DR}$ values with relatively large reflectivity. Consistent with this picture, Evaristo et al. [2013] showed a linear decrease in $Z_{DR}$ with the cone apex angle of conical graupel; for instance, a $Z_{DR}$ of around -1.2 dB was obtained for 30°. Bringi et al. [2016] documented negative $Z_{DR}$ (~$-0.3$ to $-0.7$ dB) values along a vertical column in a winter storm at high plains region of Colorado. The results suggested that conical graupel were more prevalent in the 3.5-4.0 km Mean Sea Level (MSL) height layer of the echo cores where $Z_{DR}$ tended to be slightly more negative. More recently, Homeyer and Kumjian [2015] documented a near-zero $Z_{DR}$ minima in organized and discrete supercells close to and at altitudes higher than the updraft column features, indicating the presence of large hail. Additionally, minimum $Z_{DR}$ and negative $K_{DP}$ was observed throughout the portion of
the convective cores of organized convective systems that overshot the tropopause, suggesting signatures from small hail and/or lump or conical graupel.

Understanding the relationships between the characteristics of polarimetric radar in the early development of clouds and the production of the first lightning could be very useful for lightning nowcasting and lightning parameterization. The basic aspects of cloud electrification have been described in several studies; however, cloud evolution during the electrification life cycle, from the first radar echo until the moment of the first IC and CG flashes, has not been completely defined. What is different for this work from prior studies is the extensive documentation of the first IC and CG lightning flashes using two- and three-dimensional lightning location systems together with time-resolved polarimetric observations for a special subset of storms with lower attenuation. The thunderstorms studied in this work are very compact clouds (~ 9 km of diameter) and this procedure enabled a simpler physical interpretation of early thunderstorm development and with reduced radar attenuation operating in the X-band. In addition, this study aims to evaluate the signatures from conical graupel using X-band radar with signatures strongly tied to the initial electrification of these thunderstorms. Therefore, this study aims to cover these aspects based on case studies and a large statistical analysis of 46 isolated compact thunderstorm life cycles.

Section 2 presents the general aspects of the CHUVA-Vale campaign and the radar and lightning observations that were employed. Section 3 presents the methodology, in section 4 are discussed in details three case studies and in section 5 the statistical analysis of the 46 thunderstorms. The discussion and the main conclusions are presented in section 6.
2 The CHUVA-Vale Campaign

The CHUVA (Cloud Processes of the Main Precipitation Systems in Brazil: A Contribution to Cloud Resolving Modeling and to the Global Precipitation Measurement) project’s main scientific motivation was to contribute to the understanding of cloud processes, which represent one of the least understood components of the weather and climate system. During five years (from 2010 to 2014) the CHUVA project has conducted several field campaigns in Brazil [see Machado et al., 2014 for a detailed description]. The CHUVA-Vale campaign’s objective was to understand the cloud processes that evolve when clouds transform into thunderstorms. The CHUVA-Vale campaign took place during the Brazilian spring-summer (from November 2011 to March 2012) in São José dos Campos, in Southeast Brazil. The essential datasets used in this study were the polarimetric variables from the XPOL radar, the return stroke information provided by the Brazilian Lightning Detection Network (BrasilDAT) and the Very Low Frequency (VHF) radiation sources from the São Paulo Lightning Mapper Array (SPLMA). Figure 1 shows the location of the XPOL radar (gray diamond) and the lightning sensors from the BrasilDAT (blue stars) and SPLMA (red filled circles) networks. Asterisks represent the locations of the 46 incipient thunderstorms at the time of the first CG lightning flash.

2.1. XPOL radar and Data Post-processing

To identify and track the polarimetric signatures in the thunderstorms we utilized all the volume scans from the mobile XPOL (9.345 GHz) polarimetric radar. We evaluated the horizontal reflectivity ($Z_H$, dBZ), differential reflectivity ($Z_{DR}$, dB), specific differential phase ($K_{DP}$, $^\circ$ km$^{-1}$) and the correlation coefficient ($\rho_{HV}$) [see Straka et al., 2000 for a detailed description of these variables]. The XPOL strategy was consistently performed
every 6 minutes, and included a 4 minute standard volumetric scan with 13 elevations
from 1.0° to 25.0° and radar samples resolved to 150 m in range and 1.0° in azimuth.
The strategy also included a scan at 89° elevation with full azimuthal rotation for
purposes of Z_{DR} calibration (offset check) and specific Range Height Indicator (RHI)
scans.
The radar data were post-processed following several steps to correct the raw data for
attenuation and the Z_{DR} offset. To correct the attenuation in Z_{H}, the so-called Rain
Profiling Algorithm (ZPHI) proposed by Testud et al. [2000] was applied. The
reflectivity field was compared with a nearest S-band radar (distant 50 km far from
XPOL radar). The XPOL radar captured very well the storms structure in comparison
with the S-band radar. However, for higher XPOL reflectivity values (~ 40-60 dBZ) a
negative BIAS (around -5 dBZ) was found as well as for the comparison with the
reflectivity simulated through the T-matrix method using information from the
disdrometer. For higher reflectivity values the XPOL radar attenuation correction
scheme does not fully correct the reflectivity field. Section 3 discusses this effect in the
results and the limitations on the storm selection to reduce these uncertainties.
Differential attenuation in Z_{DR} is experienced when radiation moves through
populations of oblate raindrops; this was corrected using the method of linear \( \Phi_{DP} \),
which considers the total differential attenuation to be linearly proportional to \( \Phi_{DP} \)
[Bringi et al., 2007]. The Z_{DR} average and median corrections were 0.3 and 0.2 dB,
respectively. The offset in Z_{DR} due to the imbalance of the horizontal and vertical
channels was determined for three periods of the CHUVA-Vale campaign: -0.27 dB
(period before the exchange of the radome), -0.33 dB (period after the exchange of the
radome) and -0.59 dB (period after the calibration/substitution of components)
[Sakuragi and Biscaro, 2012].
To compute $K_{DP}$ an iterative finite impulse response (FIR) range filter developed by Hubbert and Bringi [1995] was applied to the raw $\phi_{DP}$ data. This filter smoothes $\phi_{DP}$ and an interactive methodology was performed to remove $\phi_{DP}$ deviations caused by backscatter differential phase shift ($\delta$) from large oblate particles. The $K_{DP}$ values were derived by a least-squares regression of $\phi_{DP}$ over several range gates. The average $\delta$ observed for the dataset was 0.3°. Comparisons between the polarimetric variables from the T-matrix method using a Joss disdrometer and the variables estimated by the XPOL radar showed average disagreements (radar-disdrometer) of -0.04 dB ($Z_{DR}$) and of -0.07 °km⁻¹ ($K_{DP}$). Further details of the data post-processing in the CHUVA experiment can be found in Schneebeli et al. [2012]. In the foregoing analysis, these corrections were included in the $Z_{H}$ and $Z_{DR}$ radar measurements.

2.2. Lightning Observations

Lightning information was provided by two independent Lightning Location Systems (LLSs). The IC and CG return strokes were provided by the Brazilian Lightning Detection Network (BrasilDAT) (Figure 1, blue stars). This lightning network employs technology from Earth Networks and operates in a large frequency range (from 1 Hz to 12 MHz), and locates lightning using the time-of-arrival method. During this study, 56 sensors from BrasilDAT covered the southeastern, southern, central and a portion of the northeastern regions of Brazil. Additional sensors from BrasilDAT were located close to the CHUVA-Vale region to improve the data quality. The reprocessed data pertaining to location, time of occurrence, and polarity of the IC and CG return strokes were used in this study. A preliminary evaluation of BrasilDAT using high speed cameras showed a detection efficiency of about 88 % for CG flashes [Naccarato et al., 2012]. More recently, Williams et al. [2016] analyzed the same thunderstorms selected in this study.
and concluded that the majority of lightning flashes detected by BrasilDAT was also
detected by the Brazilian Integrated Lightning Detection Network (RINDAT) network.
A good agreement was found between the CG stroke-multiplicity and peak current
estimated by BrasilDAT and RINDAT lightning networks. Additionally, a second LLS,
operating in the VHF range, was used to determine the initiation region of the initial IC
and CG return strokes detected by BrasilDAT. Named the São Paulo Lightning Mapper
Array (SPLMA) (Figure 1, red filled circles), this network was developed by the New
Mexico Institute of Mining and Technology and installed in a collaborative effort
between NASA (National Aeronautics and Space Administration), the University of
Alabama in Huntsville, INPE (National Institute of Space Research), and USP
(University of São Paulo). During the CHUVA-Vale campaign, the SPLMA was
composed of 12 stations operating in the frequency band of TV channels 8 (180-186
MHz) and 10 (192-198 MHz) [Blakeslee et al., 2013; Bailey et al., 2014; Albrecht et al.,
2014]. The dataset was reprocessed to provide the time, latitude, longitude and altitude
of the VHF radiation sources from all lightning detected during the campaign. The
mean chi-square $\chi^2$ and mean number of stations per solution were 1.3 and 7,
respectively. The SPLMA data were used for three thunderstorms for the analysis of
lightning initiation (section 4). These thunderstorms occurred very close to the SPLMA
center (less than 30 km distance), favoring a region with higher detection efficiency and
enabling the detection of the majority of lightning activity from these thunderstorms.

3 Identification of Thunderstorms and Co-location with Lightning Observations

In this study, thunderstorms were identified and tracked manually with observations
from the XPOL radar. These thunderstorms represent isolated precipitating cells and
were chosen for further study if no additional thunderstorms were obstructing the path
between the radar and the respective thunderstorm cell. This procedure enabled a simpler physical interpretation of early thunderstorm development and reduced radar attenuation effects characteristic of radars operating in the X-band. A minimum distance of 20 km from the radar was selected to avoid thunderstorm cases with limited-top due to the lower beam height close to the radar. Based on the aforementioned considerations, the life cycle of thunderstorms were typically sampled using a 6-minute scan strategy and the azimuth angle and distance limits from the radar were used to determine the thunderstorm’s boundaries.

From the radar perspective, we considered thunderstorm initiation to be when a radar echo with any value of reflectivity above the reflectivity noise level was first detected in any height in any Plan Position Indicator (PPI) scan. The majority of thunderstorms were identified between 20 and 60 km range from radar, with minimum detectable radar reflectivity of approximately -4 dBZ and 9 dBZ, respectively. This procedure aimed to identify the initial development of the thunderstorms without restrictions. Based on these constraints, 46 thunderstorm life cycles were selected for this study.

The lightning information was co-located to each time step of the 46 thunderstorm life cycles. First the return stroke observations from BrasilDAT were grouped into flashes using a temporal and spatial criterion of 0.5 seconds and 20 km, respectively. The 0.5 sec time threshold is close to that employed by McCaul Jr. et al. [2009] and Goodman et al. [2005] (0.3 sec) and Nelson [2002] (0.5 sec). The criterion of 20 km corresponds to the fact that the majority of the 46 storms had diameters smaller than 20 km (see Figure 2 found in Williams et al., 2016). Williams et al. [2016] used this same data set and spatial criterion and showed that the stroke-multiplicity and peak current from BrasilDAT was very similar to that from RINDAT network. In fact, the 20 km spatial criterion is a good choice for this study, because the storms are isolated and presents
low flash rates; however for larger and more complex storms, other values for this spatial criterion should be evaluated. Afterward, IC and CG flashes were accumulated every 6-minutes and assigned to respective thunderstorms by using the area boundaries as a constraint. Finally, to determine the initiation height of these flashes, VHF sources from SPLMA were linked with the IC and CG flashes from BrasilDAT. The same spatial-temporal criteria used to combine strokes into flashes (i.e., 0.5 s and 20 km) were used to find the VHF sources associated with every IC and CG flash. In this way, a dataset was created for the life cycle of each thunderstorm, from the polarimetric variables of the initial radar echo through the time of the first IC and CG flashes.

4 Thunderstorms Case Studies

In order to follow the behavior of the polarimetric variables on a case-by-case basis and evaluates the initiation region of the first lightning flashes, was selected and discussed in detail three thunderstorms with different total (intracloud+cloud-to-ground) lightning flash rate: i) relatively weak (1.5 flashes min\(^{-1}\)), ii) relatively moderate (1.9 flashes min\(^{-1}\)) and iii) relatively strong (2.3 flashes min\(^{-1}\)) lightning flash rate. These thunderstorms occurred close to the SPLMA center in a region with good coverage by the XPOL radar which made it possible to evaluate the region of initiation of the first flashes in terms of lightning and polarimetric radar signatures with good efficiency. Note that the designations relatively weak, moderate and strong lightning flash rate are based on the specific initial lighting flashes rate for the small storms used in this study; a procedure that differs from many other published studies. These thunderstorms present lower lightning flash rates likely due to their small size (< 20 km in diameter) and weaker updraft. Moreover, these definitions are used in this study as a reference and only to classify the lightning flash rates of these thunderstorms. In the context of the natural
variability of the 46 selected thunderstorms, these three cases give a representative
description of the studied thunderstorm population.

4.1. Case #1: The Thunderstorm at 1800 UTC on 20 February 2012

Time-height plots of selected radar parameters were computed for the thunderstorms
life cycle based on the PPI-volume scans. For this purpose, the maximum $Z_H$ in each
PPI and the coincident value of $Z_{DR}$, $K_{DP}$ and $\rho_{HV}$ in the cell were selected. Each volume
scan of the thunderstorm was represented by single vertical profile with 13 altitude
levels, one for each PPI. Employing the methodology described above, the vertical
sampling cannot be considered perfectly vertically-aligned. However, a statistical
evaluation of the degree of verticality of each profile showed a median value of 250 m
offset between the vertical radar gates. This small effect is probably attributable to the
compact and isolated thunderstorms that were selected. This procedure aimed to
illustrate the dominant behavior of the polarimetric parameters. Figure 2 shows the
time-height plots of $Z_H$, $Z_{DR}$, $K_{DP}$ and $\rho_{HV}$ from the first radar echo until the time of the
first CG flashes for the thunderstorm with relatively weaker total lightning flash rate
(1.5 total flashes min$^{-1}$) observed on 20 February 2012.

The thunderstorm developed during late afternoon at 1800 UTC (1600 LT) and showed
the first radar echo (at time of 0 min) at 3 km height and with 15 dBZ. The first IC (CG)
flash was registered 28 (30) min after the first radar echo time. During this interval
(from 0 to 30 min), the reflectivity $Z_H$ (Figure 2a) evolves from weak (15 dBZ) to
strong (56 dBZ) value at 0 °C level, indicating the formation of large hydrometers. For
simplicity, the term ‘large hydrometers’ is referred to those hydrometers that are in the
Rayleigh regime (< 2 mm in diameter), hereafter. The time-height plot for $Z_{DR}$ (Figure
2b) indicates an initial concentration of raindrops close to melting layer which is shifted
to higher altitudes when approaching to the time of the first IC flash (at 28 min). At the
time of the first CG flash (~ 30 min), a negative value of $K_{DP}$ (-0.3 ° km$^{-1}$, Figure 2c)
and a strong value for $\rho_{HV}$ (> 0.99, Figure 2d) were found in upper levels (13 km
altitude), indicating ice particles are being aligned by strong electric field. In fact,
polarimetric variables show important characteristics between 0° and -40°C levels.

Figure 3 shows the evolution of the minimum values for $Z_H$, $Z_{DR}$, $K_{DP}$ and $\rho_{HV}$ observed
in the layer between 0° and -40°C. The reduction of $Z_{DR}$ and $K_{DP}$ before the first CG
flash, followed by the increase of $\rho_{HV}$ in the mixed-layer (0° and -15°C), is the main
remarkable behavior. It probably corresponds to the signature of the freezing of ice
particles and the formation of graupel, fundamental conditions for the non-inductive
electrification mechanism and lightning production.

In order to evaluate the polarimetric behavior of the entire vertical structure of this
thunderstorm, vertical cross sections were computed following the centroid of the
thunderstorm (i.e., Lagrangian analysis) from the first radar echo until the time of the
first CG flash. Figure 4 shows the vertical cross section as well as the projected
locations of the flashes. The centroid location of the first 10 % of the total sources in a
flash was considered to be the origin location for the first IC and CG flashes, as in Lang
and Rutledge [2008] and Lund et al. [2009]. The thunderstorm initiated with the
formation of small raindrops with weaker $Z_H$ (20 dBZ) in the warm cloud layer (Figure
4a) at 10 and 20 km distance-range and after six minutes the supercooled raindrops
reach the -16 °C isotherm (Figure 4b) at 13 km distance-range. The weaker $Z_H$ (20 dBZ)
and the moderate $Z_{DR}$ (+1.5 dB) observed close to the -25 °C level suggest the initial
formation of ice crystals. In the following image (Figure 4c) strong $Z_H$ (50 dBZ) and
$Z_{DR}$ (+3.5 dB) is observed close to the -15 °C level at 13 km distance-range, indicating
the intensification of updrafts. Six min later (Figure 4d) this region intensifies and
reaches higher levels and also reaches the ground, indicating the formation of large precipitation particles near the ground. Note the existence of a deepened tower reaching up to -16°C with strong $Z_H$ and including a narrow tower of positive $Z_{DR}$ above the melting layer, indicating the lofting of supercooled raindrops by strong updraft. Ice crystals and graupel particles are evident in the upper layers (> 8 km) of this column, inferred by the negative $Z_{DR}$ and $K_{DP}$ in this region. In the following time step (Figure 4e) the $Z_H$ value above the -30 °C level centered at 13 km distance-range decreases by 5-10 dBZ and a large region with more negative $Z_{DR}$ and $K_{DP}$ is formed simultaneously with the reduction of the $+Z_{DR}$-column, promoting the occurrence of the first IC flash close to the -20 °C level. The freezing of the supercooled raindrops aloft when the $+Z_{DR}$-column is collapsing is the probable causes for the diminished $Z_H$, since the dielectric constant for ice is smaller than for liquid water particles [Battan, 1973]. The $+Z_{DR}$-column collapse associated with the formation of graupel is evident in the image of the first CG flash (Figure 4f), where large conical graupel is suggested close to the melting level. In the follow image (Figure 4g), conical graupel particles are predominant above the melting layer when the IC flashes rate increases.

4.2. Case #2: The Thunderstorm at 1536 UTC on 22 January 2012

The thunderstorm with relatively moderate total lightning flash rate (1.9 flashes min$^{-1}$) developed during the afternoon period at 1536 UTC (1336 LT), and 20 min after of the first radar echo (Figure 5), it registered its first IC flash (indicated by the black arrow). Three minutes after of the first IC flash, the first CG flash was registered (indicated by the blue arrow). The first radar echo is observed at the 4.3 km altitude with weaker $Z_H$ (Figure 5a, 15 dBZ), positive $Z_{DR}$ (Figure 5b, +2 dB) and strong $\rho_{HV}$ (Figure 5d, 0.96) suggesting initial raindrops. From the thunderstorm initiation time until the radar scan
time (from 0 to 18 min) that marked the first IC and CG flashes, the radar observations show an increase in $Z_H$ up to 51 dBZ, a decrease in $Z_{DR}$ down to -0.1 dB and an increase in $\rho_{HV}$ (0.99) at the 8 km altitude (-23 °C level), suggesting the formation of graupel particles and ice crystals. Positives $K_{DP}$ values (Figure 5c, +0.2 ° km$^{-1}$) at the 10 km altitude (~ -34 °C) show the existence of large concentration of iced hydrometers. Note that around 12 min after the first radar echo a deep column of $Z_{DR}$ (+2.7 dB) reached 6.5 km altitude (-13°C), suggesting a strong supply of supercooled raindrops before the first IC and CG flashes times. Similar to the previously discussed thunderstorm, this feature of supercooled liquid particles lofted by the thunderstorm’s updraft in the mixed phase is now well recognized as a $+Z_{DR}$-column [Hall et al., 1980, Caylor and Illingworth, 1987; Bring et al., 1991; Conway and Zrnic, 1993; Ryzhkov et al., 1994; Hubbert et al. 1998; Smith et al. 1999; Kumjian et al., 2012; Kumjian et al., 2014; Snyder et al., 2015]. In fact, these changes are more conspicuous in the mixed-layer (between 0° and -15 °C). Figure 6 shows an increase of $K_{DP}$ (up to +0.4 ° km$^{-1}$), while $Z_{DR}$ decreases (down to 0 dB) prior to the IC flash time.

It is noted that the convective process was initiated by two adjacent $+Z_{DR}$-columns (Figure 7b, 1542 UTC) localized at ranges of 6 km and 11 km, likely associated with two convective updraft regions. In this distance-range (6 km and 11 km) very small pockets of low $\rho_{HV}$ (0.85-0.90) are observed at the top of the $+Z_{DR}$-columns. These observations are qualitatively consistent with previous studies [Herzegh and Jameson, 1992; Conway and Zrnic’, 1993; Jameson et al., 1996; Bringi et al., 1997; Hubbert et al., 1998] and are likely regions containing a mixed phase with supercooled drops, partially freezing drops and frozen hydrometeors. The two $+Z_{DR}$-columns merged into a single column and descended close to the -10 °C isotherm in the 10 km distance-range and negative $K_{DP}$ (-0.5 ° km$^{-1}$) observed between 0° and -15°C suggest the formation of
graupel particles (Figure 7c, at 1548 UTC). This demise of the $+Z_{\text{DR}}$-column produced mixed-phase hydrometeors such as graupel and supercooled raindrops as indicated by the low $\rho_{HV}$ (0.85) at the 12 km distance-range observed between the 0° and -15°C level. In the following moments (Figure 7d, at 1554 UTC) a large heterogeneous horizontal distribution of hydrometeors was observed from 5 to 17 km distance-range more pronounced between 0°C and -15 °C levels and the first IC was recorded (white circle); three minutes later (as described in Figure 5) the first CG flash (white cross) was recorded. The first IC flash initiated near the -30 °C level at 8 km distance-range, close to a region with vertically-oriented ice crystals, and the first CG flash was recorded over the region close to the -3 °C level at 13 km distance-range, with strong reflectivity at the top of region with near-to-zero $Z_{\text{DR}}$ with positive $K_{\text{DP}}$ (~ +0.5 ° km$^{-1}$) and at transition region with moderate-strong $\rho_{HV}$ (~ 0.86-0.95), separated by a region with large negative $Z_{\text{DR}}$, which was likely dominated by rimed graupel. The notable region with very low very $\rho_{HV}$ (~0.80) between the 0 °C and -15 °C levels centered at 11 km distance-range is likely associated with nonuniform beam filling (NBF) effects.

According to Ryzhkov [2007] large cross-beam gradients of $\varphi_{\text{DP}}$ may cause noticeable decrease of $\rho_{HV}$. Ryzhkov [2007] documented negative bias of approximately 0.2. Although this bias exist, correcting $\rho_{HV}$ for such biases is not practical because the biases cannot be estimated with sufficient accuracy.

4.3. Case #3: The Thunderstorm at 1818 UTC on 7 February 2012

Figure 8 shows the time-height plots for the thunderstorm that showed a relatively strong total lightning flash rate (2.5 flashes min$^{-1}$) observed on 7 February 2012. This thunderstorm developed during the late afternoon at 1818 UTC (1618 LT) and produced its first IC (CG) flash 31 min (38 min) after the initial radar echo. Actually, one can note
a rapid intensification of the thunderstorm between 9 and 20 min after the initial radar
echo with strong reflectivity (46 dBZ) and moderate $Z_{DR}$ (+2.3 dB) reaching up to the 8
km altitude (-23 °C level). The formation of the initial ice hydrometers is indicated by
the appearance of the negative $Z_{DR}$ (-0.1 dB) at 20 min close to the cloud top (10 km
altitude, -38 °C). When the first IC flash occurred (between 29-31 min after the first
radar echo) is observed a strong $Z_{DR}$ (+3.5 dB) and $K_{DP}$ (+3 ° km$^{-1}$) close to 7 km
altitude (-16 °C), suggesting a rapid increase in the concentration of supercooled
raindrops, which have contributed for the initial formation of the ice crystals and
graupel in this layer. The ice crystals formation occurs through activation of ice nuclei
within the updraft or associated with ice multiplication processes. The Hallet-Mossop
(H-M) mechanism [Hallett and Mossop 1974] is largely accepted as a dominant ice
multiplication process and occurs as small ice splinter from graupel growing by riming
of supercooled droplets in temperature-range between -3°C and -8°C.

After that time, and leading up to the time of the first CG flash (at approximately 35
min), the IC flash rate increased. At 37 min time a column of negative $K_{DP}$ (Figure 8c,
minimum of -1.0 ° km$^{-1}$ at 6 km altitude, ~ -9 °C) was observed in the altitude-range
between 6 and 14 km (between -9 ° and -64 °C), while the $\rho_{HV}$ value (Figure 8d) close
to 10 km altitude decreased dramatically from 0.97 to 0.83, between the time of the first
IC flash (at 30 min) and the time of the first CG flash (at 35 min), indicating the
freezing of hydrometers and a mixture with supercooled raindrops. In fact, in the mixed-
layer, $Z_{DR}$ and $K_{DP}$ reached their minimum values (+0.2 dB and -1.1 ° km$^{-1}$) and strong
reflectivity (> 55 dBZ) prior to the CG flash time (Figure 9), probably associated with
signatures from conical graupel.

The aforementioned observations are more evidenced in the vertical cross sections
(Figure 10). We can note a slow increase in $Z_H$ in the thunderstorm during the first 13
min (Figures 10a-d) before the convection became intense. Two convective towers with large $Z_H$ and $Z_{DR}$ indicating strong updrafts were observed at distance-range 12 km and 17 km, respectively, extending up to approximately -15 °C (Figure 10e). At 1843 UTC (Figure 10f), these convective towers began to merge, and a deep and narrow column with positive $Z_{DR}$ (+3.5 dB) and moderate $\rho_{HV}$ (~ 0.95) in the 14 km distance-range formed, demonstrating the lofting of supercooled raindrops by strong updrafts. The first IC flash was recorded in a radar image at 1847 UTC (Figure 10g) at approximately the -10 °C isotherm at time that the $+Z_{DR}$-column decreased and the cells with strong $Z_H$ completely merged. Additionally, the strong $Z_H$ (55-60 dBZ) up to the -40 °C level suggested the existence of strong updrafts promoting the lofting of raindrops, which are freezing on the graupel surface.

These regions with strong updrafts are inferred indirectly from observations of $+Z_{DR}$-columns with strong $Z_H$. At the time of the first CG flash, at 1852 UTC (Figure 10h), a large region with negative $Z_{DR}$ and $K_{DP}$ down to the -15 °C level, suggests the freezing of supercooled raindrops and the formation of conical graupel. Probably the graupel grows to a size too large by be sustained by updrafts and then begin to fall and ultimately melt, promoting the formation of large raindrops close to the surface. Although the initiation regions of both IC and CG flashes show evidence for similar microphysical conditions, one could suggest that a large region with strong negative $Z_{DR}$ and $K_{DP}$ in the altitude range above the -15 °C isotherm is a predominant signature at the time of the first CG flash.

Although the last two cases evidenced distinct and separate double reflectivity maxima (which indicated the existence of two distinct updraft maxima regions), the merging of two updrafts columns in this case seems not a physical requirement for lightning occurrence. The majority (80 % of 46 cases) of the thunderstorms in this study showed
only one reflectivity maximum (not shown). These results are consistent with Goodman et al. [1988] that documented the first IC flash after that hail was initially indicated by radar, during a period of rapid vertical development as the cloud top neared its maximum height and the first CG flash when the maximum reflectivity core descended. Instead, the pronounced positive $Z_{DR}$ in the layer between 0° and -15°C, followed by a decrease in the $Z_{DR}$ in this region is the most consistent characteristic of the storm evolution to lightning occurrence documented in this study.

5 Statistical Evaluation of the 46 Thunderstorms

This section presents the general behavior of the 46 thunderstorm life cycles selected in this study. Initially, the time lag between the first IC and the first CG flashes (Figure 11a) and the time lag between the first radar echo and the occurrence of the first IC and CG flashes (Figure 11b) were computed. For the large majority (98 %) of the thunderstorms, an IC flash preceded the first CG flash. Note that the mean time difference between IC and CG flashes was approximately 6 min and the median time difference was about 4 min. Only one thunderstorm exhibited a CG flash as its first flash. This overall behavior bears a close similarity to the earlier results (considering the mean values), particularly Workman and Reynolds [1949] (6 min), Goodman et al. [1988] (5 min), Williams et al [1989] (6 min), Harris et al. [2010] (4.7-6.9 min), Seroka et al. [2012] (2.4 min) and Stolzenburg et al. [2015] (4.6 min). These results suggest that the predominance of the IC lightning in early stages is likely due to the vertical velocity and growth of ice particles and radar reflectivity above the negatively charged center. On the other hand, the occurrence of CG lightning has been documented in previous studies close to regions with graupel and hail descending below the negative charge center [Goodman et al., 1988; Carey and Rutledge, 1996; López and Aubagnac, 1997;
The three case studies previously presented (i.e., Figures 4, 7 and 10) corroborate these findings, showing the first IC preceding the first CG flash at the boundary of the negative $Z_{\text{DR}}$ layer. Figure 11b shows IC flashes occurring, on average, approximately 29 min after the initial radar echo, while CG flashes were most frequently delayed by approximately 36 min. For some thunderstorms, the first flash occurs as much as 50 min later, or more. Studies using satellite only [Harris et al., 2010] or a combination of satellite and radar [Mecikalski et al., 2013] have documented a time elapsed for the lightning initiation on the order of 30-60 min. Since sensors on geostationary satellites detect clouds rather than precipitation, larger time differences are expected in comparison to those values documented in our study. Other studies employing radar reflectivity values at various heights have documented lead times for lightning initiation of about 10-20 min [Dye et al., 1989; Buechler and Goodman, 1990; Hondl and Eilts, 1994; Gremillion and Orville, 1999; Vincent et al., 2004; Yang and King, 2010; Mosier et al., 2011]. The time elapsed to the lightning initiation depends on the criterion used to define the first radar echo or cloud initiation and the environmental instability conditions. In our case we have considered as the first radar echo any reflectivity value (any value above the local noise floor of the radar), at any height. The detection of this earlier first echo could be responsible for the long time observed. In addition, the thunderstorms selected from CHUVA-Vale campaign are small compact and isolated, and grew in an environmental with low-to-moderate Convective Available Potential Energy (CAPE) and low wind shear, indicating a slow growth of precipitating cells and hydrometers. Consistent with this approach, Mecikalski et al. [2013] have compared two groups of storms, with smaller (1458 J kg$^{-1}$) and greater (2512 J kg$^{-1}$) CAPE and found that the smaller CAPE storms are characterized by slower and steadier or development rates. The storms with lower CAPE likely possess weaker updrafts and
display earlier development of warm rain processes. We believe, when organized mesoscale system is considered, we should expect a shorter time. This supports the idea on the existence of the slower thunderstorm development associated with an early warm rain phase in the developing cloud. As consequence of this slow process, the graupel and ice crystals in the mixed-layer takes a longer time to grow and activate the lightning initiation. Moderate CAPE supports the likely occurrence of moderate updrafts in these small thunderstorms, leading to a moderate incloud charging process. In fact, for the great majority (85%) of the thunderstorms, the first echoes are warm rain echoes (Figure 12a) with mean height around 2 km and have weaker $Z_H$ (25 dBZ) related to moderate $Z_{DR}$ (up to +4 dB) (Figure 12b). These observations are consistent with the results presented by Tuttle et al. [1989] that documented the storm’s first echo below the melting level, suggesting precipitation development was through warm rain processes (e.g., accretion and coalescence growth). In contrast for these results, Martinez [2001] documented the first heights in the cold part of the clouds, and many at 7-8 km height in a dry environmental during the STEPS at eastern Colorado. These results suggest that the air in which the thunderstorms are growing is relatively moist and clean, or the ascent speed close to the cloud base height is on the low side (to enable more time for warm rain coalescence).

In order to evaluate the statistical distribution of the polarimetric variables, whisker plots were compiled for all thunderstorms, for each cloud’s altitude range at different life cycle times. The thunderstorm life cycle was studied at four specific times: i) the time of the first radar echo, ii) the intermediate time between the first radar echo and the first IC flash, iii) the time of the first IC flash and iv) the time of the first CG flash. Hereafter the analysis will focus on these four times, which are named: #1Echo, Int., #1IC and #1CG. Consistent with the procedure followed by Mattos et al. [2016], radar
vertical profiles of the thunderstorm can be separated into four altitude layers: i) warm (below 0 °C), ii) mixed 1 (from 0 ° to -15 °C), iii) mixed 2 (-15° to -40 °C) and iv) glaciated phase (from -40 ° to -65 °C). These layers were chosen due to the different physical behaviors related to the thunderstorm electrification process observed by Mattos et al. [2016]. Figure 13 presents the whisker plots for all four thunderstorm layers in the four different lifetimes.

Generally, the initial radar echo in the warm layer of thunderstorms shows $Z_H$ greater than 10 dBZ, suggesting the cloud growth associated with initial updrafts in the life cycle stages (Figure 13a, red boxes, #1Echo). We should remember that the maximum $Z_H$ value was chosen by PPI in this study, and this certainly influences the higher $Z_H$ values found in the analysis. We note that $Z_H$ in the warm layer (Figure 13a, red boxes) exhibits the major differences among the polarimetric variables over the thunderstorm life cycle (from #1Echo to #1CG). This result indicates that hydrometeors with different sizes can be inferred prior to the first CG flash. Note that $\rho_{HV}$ (Figure 13d, red boxes) in the warm layer shows a certain variability; it is larger at the time of the first radar echo, probably associated with the large differences in droplet sizes, and then decreases slightly as the thunderstorm evolves to the first CG flash. Although, very large raindrops (> 2 mm in diameter) represent a only a small percentage (< 4 %) of the raindrops observed in the warm layer for this study (not shown), resonance effects and non-Rayleigh regime produced by these raindrops could be contributing to the strong positive $Z_{DR}$ (above +5 dB) and low $\rho_{HV}$ (below 0.90) values observed in the warm layer [Ryzhkov and Zrnic, 2005].

The distribution of $K_{DP}$ (Figure 13c, red boxes) presents a spreading leading to the time of the first CG flash in the warm layer, suggesting the formation of a strong concentration of flattened raindrops. In part, it is possible that the dispersion observed in
K\textsubscript{DP} in the warm layer is linked to nonuniform beam filling effects \cite{Gosset2004, Ryzhkov2005}. However, important to note that comparisons between the polarimetric variables from the T-matrix method using a Joss disdrometer and the variables estimated by the XPOL radar showed a median disagreement (radar-disdrometer) in K\textsubscript{DP} of -0.07 ° km\textsuperscript{-1}, indicating that NBF effects likely has reduced impact in this study.

The K\textsubscript{DP} distribution at the time of the first IC and CG flashes in the mixed 1 layer (Figure 13c, light gray boxes) were similar; i.e., both distributions show large positive K\textsubscript{DP} (up to +1.5 ° km\textsuperscript{-1}). A large variability in the intermediate time in this layer was also noted. This is possibly related to the +Z\textsubscript{DR}-column, which is better defined at the intermediate stage, followed by the +Z\textsubscript{DR}-column collapse, resulting in an average behavior for this layer. The spreading of the $\rho$\textsubscript{HV} (Figure 13d, light gray boxes) for low values is noteworthy and is only observed at the time of the first CG flash, indicating that mixture between supercooled raindrops and freezing hydrometers is predominant at this time. On the other hand, in the mixed 2 layer (Figures 13b-c, dark gray boxes), there is evidence for the freezing of large concentrations of supercooled raindrops, indicated by a narrowing of the Z\textsubscript{DR} distribution and by Z\textsubscript{DR} approaching near-to-zero values, while the K\textsubscript{DP} distribution reached negative values before the time of the CG flash. In the glaciated layer (Figures 13c, blue boxes) the K\textsubscript{DP} distribution was narrow and predominantly negative (down to -0.6 ° km\textsuperscript{-1}), and Z\textsubscript{H} was as strong as 55 dBZ at time of the first IC and CG flashes. This demonstrates that large concentration of ice crystals with different sizes, such as plates or columns, were oriented vertically by a strong electric field. Indeed, the existence of columnar crystals between -40 ° to -70 °C is the likely cause of the negative Z\textsubscript{DR} and K\textsubscript{DP}, which was suggested by an actual ice habit diagram \cite{Bailey2009}. However, some large positive and outlier
values (> +1.5 dB) for Z_{DR} in the glaciated layer could in part is associated with cross
coupling effects of the horizontally and vertically transmitted waves caused by
vertically-oriented ice crystals in the mixed layer [Hubbert et al. 2014b]. The subtle
differences in terms of polarimetric variables between the time of the first IC and CG
flashes in the mixed and glaciated layers are often a result of the initial IC and CG
flashes belonging to the same radar volume scan.

To build a description of the average thunderstorm life cycle, the mean polarimetric
variables were computed for the different cloud altitude ranges (Figure 14). In the warm
layer (Figure 14a), an increase in Z_H, Z_{DR} and K_{DP} was observed up to the time of the
first IC flash (from #1Echo to #1CG), suggesting the formation of large raindrops by
this time. The average ρ_{HV} in the warm layer decreased from the intermediate stage to
the time of the first CG flash, suggesting the existence of melting graupel or a mixture
of melting graupel and large raindrops in this layer. In the mixed 1 layer (Figure 14b),
K_{DP} and Z_H dramatically increased, indicating the intrusion of supercooled raindrops in
this layer, and at the same time ρ_{HV} decreased, indicating the mixing of hydrometeors as
the thunderstorms evolve to the time of the first CG flash. In this layer (mixed 1 layer,
Figure 14b), Z_{DR} shows the most notable behavior, an increase up to the intermediate
stage, followed by a sharp decrease at the time of the first IC flash, indicating the
freezing of supercooled raindrops aloft. In the mixed 2 layer (Figure 14c), it is clear that
the formation of negative Z_{DR} and K_{DP} indicate that graupel is likely forming via the
accretion of supercooled cloud water in this layer.

In the glaciated layer, K_{DP} (Figure 14d) shows a striking characteristic; i.e., K_{DP}
dramatically decreased to negative values during thunderstorms evolution, suggesting
rapid formation of ice particles and an electric field capable of orienting these
hydrometeors vertically. Negative K_{DP} was documented by several works, such as
This result shows that the mean polarimetric information has typical signatures for each lifetime step of the thunderstorms electrification process. In the mixed 1 layer, both $K_{DP}$ and $Z_{DR}$ have remarkable signatures, while in the glaciated layer the $K_{DP}$ behavior could be a good indicator of the time of the first CG flash.

This general statistical analysis considered only the polarimetric variables separately. Therefore, it is important to evaluate the relationships between the polarimetric variables simultaneously in the layers. The joint interpretation of two or more polarimetric variables from the same radar gate is much more effective and less prone to uncertainty and non-uniqueness than single parameter analysis. This procedure often provides the best clues for inferring cloud processes and precipitation properties. Figure 15 shows the scatter plots relating $Z_{H}$ and $Z_{DR}$ (left panels) and $Z_{H}$ and $K_{DP}$ (right panels) for the four pre-defined altitude ranges (i.e., warm, mixed 1, mixed 2 and glaciated layers) in the thunderstorms and for the four lifetime stages. The observations in the warm phase layer (Figures 15a-b) provide a confident identification of the main microphysical characteristics outlined in the previous analysis, indicating that the largest $Z_{H}$ values (up to 67 dBZ) are associated with highly positive $Z_{DR}$ (up to +6 dB) and $K_{DP}$ (up to +6.5 ° km$^{-1}$), especially near the time of the first CG flash. This finding demonstrates that the largest difference in the polarimetric fields remain confined to the regions of deeper and larger updrafts. However, Figure 15b also reveals that the largest $Z_{H}$ values may have a large range of $K_{DP}$ values from 0 to +6.5 ° km$^{-1}$ in the warm layer, which suggests in part a resonance effect [Ryzhkov and Zrnic, 2005] or nonuniform beam filling effects [Gosset, 2004].
The mixed 1 layer (Figures 15c-d) revealed a slightly positive relationship between $Z_H$, $Z_{DR}$ and $K_{DP}$. Notably, the initial stage of the mixed 1 layer was characterized by a $Z_H$ above 20 dBZ with moderate $Z_{DR}$ (up to +4.5 dB) and $K_{dp}$ (up to +1.0 ° km$^{-1}$), indicating the formation of initial supercooled drops. On the other hand, weaker differences are noted between the time of the first IC and CG flash. However, we observed that the largest positive $K_{DP}$ (+4.5 ° km$^{-1}$) and $Z_H$ (60 dBZ) occurs at time of the first CG flash, suggesting an evolution from supercooled liquid water to frozen drops. Additionally, it is noted that a minimum in $K_{DP}$ (down to -1 ° km$^{-1}$) was coincident with a near-to-zero $Z_{DR}$ associated with strong $Z_H$ (up to 60 dBZ), which is consistent with the presence of graupel and hail. The population of low $Z_{DR}$ observed in the mixed 1 layer could be related to small hail, lump and/or conical graupel, or snow aggregates [Aydin and Seliga, 1984; Evaristo et al., 2013]. As discussed by Kumjian et al. [2014], aggregates are probably not the primary target because they are formed in weaker updrafts that allow larger crystals to fall and collect smaller ice crystals. In this study, it is evident that the first flash occurs in a region with deeper updrafts, which are favorable to the formation of hail and graupel promoting the non-inductive cloud electrification mechanism.

Notably, in the mixed 2 layer (Figures 15e-f), near-to-zero $Z_{DR}$ were predominant and associated with $Z_H$ from 20 to 55 dBZ and negative $K_{DP}$ (down to -2 ° km$^{-1}$) at the time of the first CG flash. This demonstrates that hail and graupel are dominant signatures in this layer. In contrast, negative $K_{DP}$ (down to -1 ° km$^{-1}$) with moderate $Z_H$ (from 25 to 45 dBZ) are predominant signatures in the glaciated layer (Figures 15g-h). Larger $Z_H$ (35-45 dBZ) values with negative $K_{DP}$ (down to -1 ° km$^{-1}$) are consistent with signatures from conical graupel [i.e., Evaristo et al., 2013; Bringi et al., 2016]. However, since these reflectivity values represent the maximum $Z_H$ extracted in each elevation angle,
this approach is masking some of the strongest negative $K_{dp}$ and $Z_{dr}$ values associated with vertical ice crystals. Consistent with this picture, the vertical cross sections (Figures 4, 7 and 10) showed that both negative $K_{dp} (-0.3^\circ km^{-1})$ and $Z_{dr} (-2$ dB) values are more prevalent at low reflectivity (20-30 dBZ) than at higher values in the region above the -40°C level in the cloud. Therefore, although conical graupel probably is present (dominating the $Z_H$ signatures and which may is masking the signatures from vertical ice crystals) the existence of vertically-aligned ice crystals by strong electric field is prevalent and notable when considering the whole glaciated layer. As suggested by Weinheimer and Few [1987] and by an actual ice habit diagram in Bailey and Hallett [2009], these ice particles are likely plates or columns; although columnar crystals between -40 ° to -70 °C are much more likely to align than plate-like crystals.

6 Discussion and Conclusions

This study describes the polarimetric characteristics as a function of the life cycle in different cloud layers to estimate thunderstorm microphysical properties from the time of the first radar echo until the production of the first IC and CG lightning flashes. Observations of 46 thunderstorms during the 2011/2012 spring-summer in Southeast Brazil with an XPOL radar and two- and three-dimensional Lightning Location Systems demonstrated the key parameters in different layers related to the initial electrification process in these thunderstorms. A discussion of three case studies in details revealed the main characteristics of the thunderstorms life cycle. Time-height plots and vertical cross sections of thunderstorms lifecycle evolution were the basis for this analysis. The study cases showed highly positive $Z_{dr}$ and $K_{dp}$ columns extending up to the -15 °C isotherm prior to the first IC and CG flashes, suggesting a lofting of supercooled raindrops by strong updrafts.
feeding the production of hail or conical graupel. We observed that these $+Z_{DR}$-columns extended to higher levels, with a region aloft characterized by a negative $Z_{DR}$ signature, indicating the likely existence of highly charged ice and graupel hydrometeors. The first IC flash was observed at the top of the $+Z_{DR}$-column in the central dipole region close to the -16 °C (7 km) isotherm, followed by the first CG flash observed below of this layer. The thunderstorm configuration at this time showed a strong heterogeneous horizontal distribution of hydrometeors. These characteristics were clearly observed on a case-by-case basis as well as in the statistical analysis. There was notably a minimum $\rho_{HV}$ on the boundary of the positive $Z_{DR}$ and negative $Z_{DR}$ regions (Figures 4, 7 and 10) indicating a freezing zone, as a result of the mixture of ice and liquid particles. This result is consistent with other observations of reduced $\rho_{HV}$ or even enhanced linear depolarization ratio (LDR) near the tops of $Z_{DR}$-columns [Bringi et al. 1997; Hubbert et al. 1998; Smith et al. 1999; Kumjian et al. 2014; Snyder et al. 2015]. Hubbert et al. [1998] documented through polarimetric radar measurements that regions with low $\rho_{HV}$ (94-96) were coincident with regions of strong LDR ($\geq -22$ dB) and strong $Z$ (40–50 dBZ) at the top of the positive ZDR column, consistent with a mix of supercooled drops, partially frozen drops, and asymmetric wet graupel. In addition, Smith et al. [1999] presented comparisons between T-28 aircraft measurements and LDR in 3-cm radar observations in mixed-phase regions that showed the presence of drops in the process of freezing in regions with enhanced LDR signatures atop $Z_{DR}$-columns.

The composite analysis considering all thunderstorms was largely consistent with the aforementioned cases. The statistical analysis of the 46 cases showed contrasts in the polarimetric signatures throughout the thunderstorm life cycle. The decrease of $K_{DP}$ to negative values in the glaciated layer, from the time of the first development of the glaciated layer up to the time of the first CG flash, was clearly observed. This is likely
related to the high concentrations of ice crystals, such as like plates and columns, being
vertically-aligned by a strong electric field. The most important aspect of the
observations going into this paper, which came originally from the visual examination
of a large number of thunderstorms, is that an initial \( +Z_{DR} \) (associated with supercooled
raindrops) evolved to reduced \( Z_{DR} \) (and even negative values) in the mixed 1 layer
before and during the time of the initial lightning, suggesting an evolution from
supercooled raindrops to frozen particles, and the formation of graupel.

Based on the above description, it is possible to develop a conceptual model of earlier
electrification for isolated thunderstorms (Figure 16). The initial stage (Figure 16a) of
thunderstorms is dominated by small raindrops growing by coalescence in the updrafts
in the warm layer. If the upward air motion is sufficient, some of these raindrops may
reach a height above the melting level and start to grow rapidly through collection of
small droplets. The updraft intensifies and a well-defined \( +Z_{DR} \)-column (with mean
value +1-2 dB) containing supercooled liquid raindrops is produced after a mean time of
15 min (Figure 16b). The eventual reduction in reflectivity of the \( +Z_{DR} \)-column is
associated with the phase change in freezing raindrops and the attendant change in
dielectric constant, and is followed by the appearance of negative \( Z_{DR} \). The existence of
a mixture of supercooled drops and partially frozen drops, with a variety of shapes and
dielectric constant, is likely responsible for the low \( \rho_{HV} \) observed at the top of the \( Z_{DR} \)-
columns, and further supports the importance of mixed-phase hydrometeors in this
region. Atlas [1966] and Lhermitte and Williams [1985] recognized this location as the
“balance level” (6-7 km MSL), where the particle mean terminal velocity is equal in
magnitude to the upward air motion. This level is favorable to the formation of
relatively large precipitation particles (graupel and small hail, associated with radar
reflectivity in the range 45 and 50 dBZ) suspended in an updraft. By this time, the

34
strong updrafts promotes strong collision rate between the graupel and ice particles leading to an increase in the initial cloud electrification. After 29 min (Figure 16c), the conical graupel atop the $+Z_{\text{DR}}$-column in the updraft grow to sizes too large to be suspended, and accordingly they begin to fall out. As they descend through the updrafts, more supercooled raindrops are collected and they are dragged down through the melting level. The descent of the conical graupel and the progressive freezing of the suspended supercooled raindrops marks the demise of the $+Z_{\text{DR}}$-column. The gravitational separation of charged ice crystals at higher levels and the oppositely charged graupel in lower levels promotes the formation of a strong electric field and the first IC flash is registered. Therefore, the $+Z_{\text{DR}}$-column shows a well-defined lifecycle consistent with plausible microphysics and suggests a potential for lightning nowcasting. The maximum intensity of the $+Z_{\text{DR}}$-column is reached before the first IC flash (~15 min before), followed by a weakening of the column at the moment of the first IC flash. By the mean time around 36 min (Figure 16d) after the first radar echo, the strong electric field is sufficient to align the ice crystals close to the cloud-top (above the -40°C isotherm) and the conical graupel are dominant between the melting layer up to -15 °C. Ice crystals of many kinds (i.e., like columns and plates) are likely aligned vertically by the electric field in the glaciated layer, as suggested by Weinheimer and Few [1987]. At this time the strong electric field is intensified and the first CG flash is registered. The large negative $K_{\text{DP}}$ in the glaciated layer and the decrease of $Z_{\text{DR}}$ in the mixed 1 layer are systematic signatures before and during the time of the first CG flashes. The supercooled raindrop characteristic of the $+Z_{\text{DR}}$-column is a systematic feature observable before the first IC flash of a developing thunderstorm. Despite the fact that $Z_{\text{DR}}$-columns in the mixed phase and negative $K_{\text{DP}}$ in the glaciated layer have been investigated in many studies in association with lightning occurrence,
many of the details of their development and their relationship with initial lightning were not previously investigated in detail. Differently from previous studies, this work provides an extensive documentation of the first IC and CG lightning flashes (for 46 thunderstorms) using two- and three-dimensional Lightning Location Systems together with the time-resolved polarimetric observations for a special subset of storms with lower attenuation. These thunderstorms are very compact clouds (avoiding mesoscale effects) and this procedure enabled a simpler physical interpretation of early thunderstorm development and with reduced radar attenuation characteristic of radars operating at X-band. In addition, this study was carried out in another geographical location, in another hemisphere in the tropics and the results presented in this work provide statistical confirmation of the previous studies but for isolated tropical clouds, giving additional information about the thunderstorm electrification life cycle.

Important practical applications are highlighted here. This model can be used to design a nowcasting tool. Changes in the size and height of the $+Z_{DR}$-column provide the potential for the formation of graupel and ice crystals; which are fundamental conditions for the lightning occurrence. This feature can also be used to estimate thunderstorm strength and maturity, increasing the lead-time used for lightning nowcasting. Additionally, a systematic observation of conical graupel (negative $Z_{DR}$) in the mixed layer and vertically-oriented ice crystals (negative $K_{DP}$) in the upper levels of thunderstorms may provide helpful information concerning thunderstorm vigor and its lightning diagnostic. The description of the thunderstorm life cycle could also open new opportunities for microphysical and lightning parameterization in cloud resolving models, as well as for the testing of numerical models that describe the life cycle. Future analyses should consider different meteorological contexts, the degree of baroclinicity,
the effect of seasons, and different thunderstorm sizes to evaluate this conceptual model for isolated thunderstorms.

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**FIGURE CAPTIONS LIST**

**Figure 1.** Region of the CHUVA-Vale campaign with the localization of the X-band radar (gray shaded region), and the SPLMA (red filled circles) and BrasilDAT (blue stars) lightning sensors. Asterisks represent the locations of the 46 thunderstorms at the time of the first cloud-to-ground (CG) lightning flash. The gray dashed lines represent the distance rings (20, 60 and 100 km) from the radar. The minimum detectable radar reflectivity at 20 km and 60 km range were -4 dBZ and 9 dBZ, respectively.

**Figure 2.** Time-height plot of (a) $Z_H$ (dBZ), (b) $Z_{DR}$ (dB), (c) $K_{DP}$ ($^\circ$ km$^{-1}$) and (d) $\rho_{HV}$ for the thunderstorm observed at 1800 UTC on 20 February 2012. The horizontal dashed line marks the 0°C level as determined by the sounding data. The black vertical lines in the figures represent the times of the intracloud lightning flashes and the blue lines represent the cloud-to-ground lightning flashes. Arrows indicate the first intracloud (black) and the first cloud-to-ground (blue) flashes.

**Figure 3.** Maximum value of $Z_H$ (dBZ) (black line) and minimum values of $Z_{DR}$ (dB) (blue line), $K_{DP}$ ($^\circ$ km$^{-1}$) (orange line) and (d) $\rho_{HV}$ (red line) in the cloud layer between $0^\circ$ to $-40^\circ$C for the thunderstorm evolution documented in Figure 2. The black and blue dashed lines mark the times of the first intracloud and cloud-to-ground flashes, respectively.

**Figure 4.** Vertical cross sections of the polarimetric variables ($Z_H$, $Z_{DR}$, $K_{DP}$ and $\rho_{HV}$) of the thunderstorm evolution documented in Figure 2. The locations of the initiation points for the intracloud flashes are indicated with black circles and for the cloud-to-
ground flashes by blue crosses. Symbols in white color indicate the first intracloud
(Figure 4e) and cloud-to-ground flashes (Figure 4f).

**Figure 5.** Time-height plot of (a) $Z_H$ (dBZ), (b) $Z_{DR}$ (dB), (c) $K_{DP}$ ($^\circ$ km$^{-1}$) and (d) $\rho_{HV}$ for the thunderstorm observed at 1536 UTC on 22 January 2012. The horizontal dashed line marks the 0°C level as determined by the sounding data. The black vertical lines in the figures represent the times of the intracloud lightning flashes and the blue lines represent the cloud-to-ground lightning flashes. Arrows indicate the first intracloud (black) and the first cloud-to-ground (blue) flashes.

**Figure 6.** Maximum value of $Z_H$ (dBZ) (black line) and minimum values of $Z_{DR}$ (dB) (blue line), $K_{DP}$ ($^\circ$ km$^{-1}$) (orange line) and (d) $\rho_{HV}$ (red line) in the cloud layer between 0° to -40 °C for the thunderstorm evolution documented in Figure 5. The black and blue dashed lines mark the times of the first intracloud and cloud-to-ground flashes, respectively.

**Figure 7.** Vertical cross sections of the polarimetric variables ($Z_H$, $Z_{DR}$, $K_{DP}$ and $\rho_{HV}$) of the thunderstorm evolution documented in Figure 5. The locations of the initiation points for the intracloud flashes are indicated with black circles and for the cloud-to-ground by blue crosses. In Figure 7d the symbols in white color indicate the first intracloud and cloud-to-ground flashes.

**Figure 8.** Time-height plot of (a) $Z_H$ (dBZ), (b) $Z_{DR}$ (dB), (c) $K_{DP}$ ($^\circ$ km$^{-1}$) and (d) $\rho_{HV}$ for the thunderstorm observed at 1818 UTC on 7 February 2012. The horizontal dashed line marks the 0°C level as determined by the sounding data. The black vertical lines in
the figures represent the times of the intracloud lightning flashes, and the blue lines represent the cloud-to-ground lightning flashes. Arrows indicate the first intracloud (black) and the first cloud-to-ground (blue) flashes.

**Figure 9.** Maximum value of $Z_H$ (dBZ) (black line) and minimum values of $Z_{DR}$ (dB) (blue line), $K_{DP}$ ($^\circ$ km$^{-1}$) (orange line) and $\rho_{HV}$ (red line) in the cloud layer between $0^\circ$ to $-40^\circ$ C for the thunderstorm evolution documented in Figure 8. The black and blue dashed lines mark the times of the first intracloud and cloud-to-ground flashes, respectively.

**Figure 10.** Vertical cross sections of the polarimetric variables ($Z_H$, $Z_{DR}$, $K_{DP}$ and $\rho_{HV}$) of the thunderstorm evolution documented in Figure 8. The locations of the initiation points for the intracloud flashes are indicated with black circles and for the cloud-to-ground by blue crosses. Symbols in white color indicate the first intracloud (Figure 10g) and cloud-to-ground flashes (Figure 10h).

**Figure 11.** Relative frequency distribution of the (a) time difference in minutes between the first intracloud (IC) and the first cloud-to-ground (CG) lightning flash and (b) elapsed time between the first radar echo and the first intracloud (black line) and the first cloud-to-ground (blue line) lightning flash for the 46 thunderstorms.

**Figure 12.** (a) Relative frequency distribution of the height (km) of lowest radar gate at time of the first radar echo and (b) scatter plots between $Z_H$ (dBZ) and $Z_{DR}$ (dB) for the lowest radar gate at time of the first radar echo.
Figure 13. Box and whiskers plots for the (a) $Z_H$ (dBZ), (b) $Z_{DR}$ (dB), (c) $K_{DP}$ (° km$^{-1}$) and (d) $\rho_{HV}$ for the glaciated (from -40 ° to -65 °C, blue boxes), mixed 2 (-15° to -40 °C, dark gray boxes), mixed 1 (from 0 ° to -15 °C, light gray boxes) and warm (below 0 °C, red boxes) phase layers. For every layer, the four boxes represent the following stages of the thunderstorm life cycle: (i) the time of the first radar echo ($#1Echo$), (ii) the intermediate time between the first echo radar and the first intracloud lightning flash (Int.), (iii) the time of the first intracloud lightning flash ($#1IC$) and the (iv) time of the first cloud-to-ground lightning flash ($#1CG$).

Figure 14. Mean values of $Z_H$ (dBZ) (black line), $Z_{DR}$ (dB) (blue line), $K_{DP}$ (° km$^{-1}$) (orange line) and $\rho_{HV}$ (red line) for the (a) warm, (b) mixed 1, (c) mixed 2 and (d) glaciated phase layers as a function of the four life cycle stages of thunderstorms: (i) $#1Echo$, (ii) Int., (iii) $#1CG$ and (iv) $#1CG$.

Figure 15. Scatter plots between $Z_H$ (dBZ) and $Z_{DR}$ (dB) (left panels) and between $Z_H$ (dBZ) and $K_{DP}$ (° km$^{-1}$) (right panels) for the (a-b) warm, (c-d) mixed 1, (e-f) mixed 2 and (g-h) glaciated phase layers for the four life cycle stages of thunderstorms: (i) $#1Echo$ (black dots), (ii) Int. (green dots), (iii) $#1CG$ (blue dots) and (iv) $#1CG$ (red dots).

Figure 16. Conceptual model of the thunderstorm electrification life cycle. It is showed the evolution from the first radar echo up to the time of the first cloud-to-ground flash: (a) the time of the first radar echo ($#1Echo$, $t_1=0$ min), (b) the intermediate time between the first echo radar and the first intracloud lightning flash (Int., $t_2=15$ min), (c) the time
of the first intracloud lightning flash (IIC, t3=29 min) and the (d) time of the first cloud-to-ground lightning flash (1CG, t4=36 min).
Figure 1. Region of the CHUVA-Vale campaign with the localization of the X-band radar (gray shaded region), and the SPLMA (red filled circles) and BrasilDAT (blue stars) lightning sensors. Asterisks represent the locations of the 46 thunderstorms at the time of the first cloud-to-ground (CG) lightning flash. The gray dashed lines represent the distance rings (20, 60 and 100 km) from the radar. The minimum detectable radar reflectivity at 20 km and 60 km range were -4 dBZ and 9 dBZ, respectively.
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Figure 4. Vertical cross sections of the polarimetric variables ($Z_H$, $Z_{DR}$, $K_{DP}$ and $\rho_{HV}$) of the thunderstorm evolution documented in Figure 2. The locations of the initiation points for the intracloud flashes are indicated with black circles and for the cloud-to-ground flashes by blue crosses. Symbols in white color indicate the first intracloud (Figure 4e) and cloud-to-ground flashes (Figure 4f).
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Figure 6. Maximum value of $Z_H$ (dBZ) (black line) and minimum values of $Z_{DR}$ (dB) (blue line), $K_D$ ($^\circ$ km$^{-1}$) (orange line) and (d) $\rho_{HV}$ (red line) in the cloud layer between 0° to -40 °C for the thunderstorm evolution documented in Figure 5. The black and blue dashed lines mark the times of the first intracloud and cloud-to-ground flashes, respectively.
Figure 7. Vertical cross sections of the polarimetric variables (ZH, ZDR, KDP and ρHV) of the thunderstorm evolution documented in Figure 5. The locations of the initiation points for the intracloud flashes are indicated with black circles and for the cloud-to-ground by blue crosses. In Figure 7d the symbols in white color indicate the first intracloud and cloud-to-ground flashes.
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Figure 14. Mean values of $Z_H$ (dBZ) (black line), $Z_{DR}$ (dB) (blue line), $K_D$ ($^\circ$ km$^{-1}$) (orange line) and $\rho_{HV}$ (red line) for the (a) warm, (b) mixed 1, (c) mixed 2 and (d) glaciated phase layers as a function of the four life cycle stages of thunderstorms: (i) #1Echo, (ii) Int., (iii) #1CG and (iv) #1CG.
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Figure 1.
Region of the CHUVA-Vale Campaign

Latitude
Longitude

20 km
60 km
100 km

Radar

SP

Storms
XPOL
SPLMA
BrasilDAT
Figure 2.
Figure 3.
Figure 4.
Figure 5.
Figure 6.
Figure 7.
Figure 8.
Figure 9.
Figure 10.
Figure 11.
a) Time Difference Between the First IC and the First CG Flash (min)

- N.Flashes = 46
- Min. Diff. = −6 min
- Max. Diff. = 38 min
- Mean Diff. = 6 min
- Std. Diff. = 7 min
- Median = 4 min

b) Time Elapsed Between the First Radar Echo and the First Flash (min)

- IC Flash
  - Min. Time = 6 min
  - Max. Time = 68 min
  - Mean Time = 29 min
  - Std. Time = 12 min
  - Median = 28 min

- CG Flash
  - Min. Time = 18 min
  - Max. Time = 70 min
  - Mean Time = 36 min
  - Std. Time = 13 min
  - Median = 35 min
Figure 12.
Min. Height=1 km
Max. Height=5 km
Mean Height=2 km

(a) Relative Frequency (
(b) $Z_H$ (dBZ) vs $Z_{DR}$ (dB)
Figure 13.
a) $Z_H$ (dBZ) distribution for different phases:
- Glaciated Phase (-40° to -65°C)
- Mixed Phase 2 (-15° to -40°C)
- Mixed Phase 1 (0° to -15°C)
- Warm Phase (> 0°C)

b) $Z_{DR}$ (dB) distribution for different phases:
- Glaciated Phase (-40° to -65°C)
- Mixed Phase 2 (-15° to -40°C)
- Mixed Phase 1 (0° to -15°C)
- Warm Phase (> 0°C)

c) $K_{DP}$ (° km$^{-1}$) distribution for different phases:
- Glaciated Phase (-40° to -65°C)
- Mixed Phase 2 (-15° to -40°C)
- Mixed Phase 1 (0° to -15°C)
- Warm Phase (> 0°C)

d) $\rho_{HV}$ distribution for different phases:
- Glaciated Phase (-40° to -65°C)
- Mixed Phase 2 (-15° to -40°C)
- Mixed Phase 1 (0° to -15°C)
- Warm Phase (> 0°C)
Figure 14.
Figure 15.