Analysis of Pre-Monsoon Convective Systems over a Tropical Coastal Region Using C-Band Polarimetric Radar, Satellite and Numerical Simulation

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Abstract: Analysis of pre-monsoon convective systems over the southern peninsular India has been performed using C-band radar and numerical simulation. Statistics on the radar polarimetric measurements show that the distribution of differential reflectivity (Z_{dr}) and specific differential phase (K_{dp}) have much higher spread over convective regions. The distribution of K_{dp} is almost uniform across the vertical over the stratiform regions. The mean profile of Z_{dr} over stratiform regions shows a distinct local maxima near melting level. A comprehensive analysis has been done on an isolated deep convective system on 13 May 2018. Plan position indicator (PPI) diagrams and satellite measured cloud top temperature demonstrate that pre-monsoon deep convective systems can develop very rapidly within a very short span of time over the region. Heavy precipitation near the surface is reflected in the high value of K_{dp} (>5° km⁻¹). High values of Z_{dr} (>3 dB) were measured at lower levels indicating the oblate shape of bigger raindrops. A fuzzy logic-based hydrometeor identification algorithm has been applied with five variables (Z_h, Z_{dr}, ρ_{hv}, K_{dp}, and T) to understand the bulk microphysical properties at different heights within the storm. The presence of bigger graupel particles near the melting layer indicates strong updrafts within the convective core regions. The vertical ice hydrometeor signifies the existence of a strong electric field causing them to align vertically. Numerical simulation with the spectral bin microphysics (SBM) scheme could produce most of the features of the storm reasonably well. In particular, the simulated reflectivity, graupel mixing ratio and rainfall were in good agreement with the observed values.

Keywords: pre-monsoon; convective systems; polarimetric radar; hydrometeor identification; southern India

1. Introduction

Thunderstorms are severe mesoscale weather phenomena that develop mainly due to intense convection over the heated landmass and are accompanied by heavy rainfall, lightning, and, sometimes, hail. They have a spatial extent of a few kilometres to a few hundred kilometres and a life span of less than an hour to several hours [1–3]. Numerous thunderstorms occur daily across the globe [4], a major fraction of which is over the tropical belt. In the case of the Indian subcontinent, most of the thunderstorms occur during the pre-monsoon (March–April–May) season [5]. They are locally known as Kalbaisakhi in West Bengal, Bordoiichila in Assam and Andhi in north-west India. A large amount of precipitation, particularly during the pre-monsoon season, occurs due to thunderstorm events [2,6]. The satellite data show the lightning climatology across the globe and reveals different hotspots, especially over the tropical region [7]. Five lightning hotspots have been identified during pre-monsoon over the Indian region and one of them is over the southern...
Analysis of data from different observatories across India shows that, the highest annual thunderstorm frequency is observed over Assam and sub-Himalayan West Bengal in the east, Jammu region in the north, and over Kerala, where the frequency of thunderstorm is higher, in the southern peninsula [9]. The thunderstorm frequency peaks in the month of May over southern India [10]. A study by Unnikrishnan et al. [11] on lightning activity using TRMM-LIS data and ground-based lightning detection network shows strong lightning activity over south India, particularly over the Kerala region. The effect of orography, along with an abundant supply of moisture from the sea and the presence of a land–sea breeze are some of the important factors that favour the occurrence of thunderstorms over the southwest peninsular region [12,13].

Thunderstorms cause damage to crops, properties and even human lives every year. It is estimated that between 1500 and 2800 deaths occurred annually due to thunderstorms/lightning during 2001–2017 across India [14]. Heavy rainfall and high winds from these weather systems cause an interruption in connectivity among different places and infrastructure in general. Hence, there is an increasing demand for better nowcasting of such weather systems. Several attempts have been made to predict such systems using a statistical approach [15–17], satellite-based nowcasting [18,19], numerical simulations [20–26] and even artificial intelligence [27–29], but because of their small-scale nature and innate underlying nonlinearity, prediction of such systems is far from the desirable accuracy. More observations are required to understand the features and internal structures of these systems, which, in turn, will help their forecasting.

Most of the thunderstorm-related studies in India were on pre-monsoon thunderstorms (Nor’wester) occurring over east and north-east parts of India [1,3,21,30]. A few studies have been conducted on the thunderstorm occurrences over the southern peninsular India, particularly over Kerala, which is one of the potential lightning hotspots in the southern peninsular India [22,31,32]. Proximity of the Arabian Sea backed by the towering Western Ghats orography influences the formation and development of clouds and thunderstorms in the region.

The Doppler weather radar (DWR) is one of the most relevant and reliable instruments to monitor these weather events in three-dimension, starting from their genesis to dissipating stage. Radars have been used in numerous studies to understand the structure and evolution of thunderstorms [22,23,30,32,33], but most of these studies mainly use radar reflectivity and sometimes radial velocity also. However, studies using polarimetric radars are rare, particularly over the Indian region mainly because of scarcity of such data. Radars with polarimetric capabilities could provide much more information about the precipitating systems, e.g., about size and shape of the hydrometeors within the system.

Polarimetry has two major advantages viz. polarimetric measurements improve the retrieval of microphysical parameters, such as mean drop size and rainfall estimation [34–37] and polarimetric clutter-detection techniques help in the removal of non-meteorological echoes [38–41]. Since polarimetric measurements contain information on the shape and size of the hydrometeors, they can be used for better retrieval of hydrometeor types. Fuzzy-logic based hydrometeor identification (HID) is a very efficient and popular method for identifying hydrometeors within the radar scan volume [42–47]. Such studies give valuable information about different ice hydrometeors present at different heights within a precipitating system. Unlike raindrops, it is not easy to obtain information about ice particles using remote sensing techniques, mainly because of their irregular shapes and varying densities. Hydrometeor identification algorithms provide an indirect way to obtain information on ice particles. Such information can help us understand the charge separation and subsequent lightning in thunderstorms as detailed in different laboratory studies [48–50]. These studies suggest that the non-inductive charge separation due to rebounding collision between graupel and ice crystals in the presence of super-cooled water droplets is the main mechanism of thunderstorm charging. Hence, hydrometeor identification is particularly important during thunderstorm events. The spatial structure of the Ockhi cyclone and implemented HID algorithm using polarimetric Doppler weather radar observations at the west coast.
of the southern peninsular India provided information about polarimetric signatures of rain-bearing clouds [51]. However, hydrometeor identification in thunderstorms using polarimetric observations has not been done yet over the Indian region, mainly because of the lack of radars with polarimetric capabilities.

C-band polarimetric Doppler weather radar data and several other observation data are used in this study to understand the features of pre-monsoon thunderstorms over the southern peninsular India. A hydrometeor classification algorithm has been applied to obtain information on hydrometeor types. The paper is organized as follows, apart from the introduction (Section 1), the data from different instruments and methodology are described in Section 2; the results and discussions are presented in Section 3; and Section 4 summarises the major findings/conclusions drawn from the study.

2. Materials and Methods

We have identified eleven convective events over the southern peninsular India during the pre-monsoon period (i.e., March to May) of 2018 using C-band radar reflectivity field. For these events convective-stratiform separation has been done to obtain statistics on radar polarimetric variables over convective and stratiform regions. A prominent convective event on 13 May 2018 has been analysed in detail for better understanding of such systems. Besides the DWR data, we have used brightness temperature data from INSAT satellite, rain drop size distribution (DSD) data from disdrometer, cloud base height (CBH; m) data from ceilometer, ERA5 reanalysis data and also radiosonde measurements. A disdrometer and ceilometer were installed over the rooftop of the National Centre for Earth Science Studies (NCESS; 8.5228 N, 76.9097 E). Locations of the DWR and NCESS along with the topography of the surrounding area are shown in Figure 1. A brief description of the instruments and data are summarised in Table 1.

**Figure 1.** Terrain height (m) over the study area and the locations of the C-band DWR and NCESS observatory are given. Concentric circles represent distance from the radar.
### Table 1. Overview of the instruments and the data used in this study.

| Data Source                  | Parameters Used in the Study                                                                 | Spatial Resolution                      | Temporal Resolution |
|------------------------------|-----------------------------------------------------------------------------------------------|-----------------------------------------|---------------------|
| C-band polarimetric Doppler  | Reflectivity at horizontal polarisation (dBZ), differential reflectivity (dB),               | 150 m along radial and 1° along azimuth| ~15 min            |
| weather radar                | differential propagation phase (deg.), cross-correlation                                        |                                         |                     |
| Disdrometer (OTT parsivel)   | Rain rate (mm h\(^{-1}\)), concentration of precipitation particles in diameter classes:     | In-situ                                 | 1 min              |
|                              | 0.2–25 mm (m\(^{-3}\) mm\(^{-1}\)).                                                        |                                         |                     |
| Ceilometer (CHM15k)          | Cloud base height (m), cloud cover (oktas), cloud penetration depth (m)                       | -                                       | 15 s               |
| INSAT-3DR                    | Brightness temperature (K)                                                                     | 4 × 4 km                                | 30 min             |
| ERA-5                        | Temperature (K), mixing ratio (g kg\(^{-1}\)), wind speed (m s\(^{-1}\)), wind direction (deg.), CAPE (J kg\(^{-1}\)) etc. | 0.25° × 0.25°                           | 1 h                |
| Radiosonde                   |                                                                                              | -                                       | -                  |

The optical disdrometer (model: OTT Parsivel, manufactured by OTT Hydromet, Germany) is a laser-based system that detects all types of precipitation at the surface [52,53]. It measures rain DSD and fall velocity distribution in 32 size and velocity classes and also provides rain rates (R; mm h\(^{-1}\)) and radar reflectivity (dBZ). The size of measurable liquid precipitation particles ranges from 0.2 to 8 mm and it varies from 0.2 to 25 mm for solid precipitation particles. It can measure the particles fall velocity from 0.2 to 20 ms\(^{-1}\). The temporal resolution of this data is 1 min. The disdrometer used in this study was installed over the rooftop of the NCESS.

The ceilometer (model: CHM15k-Nimbus manufactured by Lufft Mess-und Regeltechnik GmbH) is a ground-based remote sensing device that uses standard lidar method to determine the cloud base height (CBH) from the altitude profile of backscattered signals. It can provide cloud thickness where the cloud layers do not totally attenuate the laser beam, but the signals get attenuated in a rainy situation depending on the number concentration and size of raindrops, and hence, signal to noise ratio of the ceilometer decreases with increasing rain rate [54]. Technical details of CHM15k can be obtained from the previous studies by Heese et al. [55] and Sumesh et al. [56]. The CHM15k was operated with a vertical resolution of 15 m and a temporal resolution of 15 s.

Brightness temperature data (Infrared Brightness Temperature, IRBT) have been used as a proxy for the cloud top height. These data were obtained from INSAT-3DR, which is a multi-purpose geosynchronous spacecraft and provides data with spatial resolution of 4 km × 4 km and temporal resolution of 30 min, of mesoscale phenomena in the visible and infrared (IR) spectral bands (0.55–12.5 µm) over the Indian region. These data are freely available through the https://www.mosdac.gov.in/server (accessed on 6 September 2021).

The synoptic circulations over the study region were analysed using the geopotential (m\(^2\) s\(^{-2}\)), u-wind (m s\(^{-1}\)) and v-wind (m s\(^{-1}\)) variables from ERA5 reanalysis hourly data having spatial resolution of 0.25° × 0.25°. Radiosonde measurements from India Meteorological Department (IMD), Thiruvananthapuram have been utilised to analyse the Convective available potential energy (CAPE; J kg\(^{-1}\)), vertical profiles of temperature (K), mixing ratio (g kg\(^{-1}\)), wind speed (m s\(^{-1}\)) and wind direction (deg).

### 2.1. DWR Data and Quality Control

The C-band polarimetric Doppler Weather Radar (DWR), installed at VSSC, Thiruvananthapuram (8.5374 N, 76.8657 E, 27 m above mean sea level), operates at a frequency of 5.625 GHz and has a peak transmitting power of 250 kW. The radar performs a volumetric scan of the surrounding atmosphere within a radius of 240 km at 11 elevation angles (0.5°,
1°, 2°, 3°, 4°, 7°, 9°, 12°, 15°, 18° and 21°) with azimuthal and radial resolutions of 1° and 150 m, respectively. One full volume scan takes around 15 min. The radar provides base products, such as reflectivity at horizontal polarization (Zب), differential reflectivity (Zد), differential propagation phase (Φد), cross-correlation (ρح), radial velocity (V_r) and spectral width (σ). Zد is the difference between reflectivities (in decibel) at horizontal and vertical polarisation, Φد is the phase difference between the horizontally and vertically polarised pulses [57,58]. A comprehensive detail about the radar is given in Mishra et al. [59]. The validation of the radar data with other instruments showed that the DWR reflectivity agrees quite well with GPM satellite measurements and also the radar retrieved precipitation have a good correlation (0.89) with ground based in situ measurements [60].

Received signal by radar is often contaminated by signals reflected from non-meteorological objects, such as hills and birds etc., anomalous propagation, and also attenuation of the electromagnetic wave by different types of hydrometeors [39,41,61,62]. Even though the radar signal processor takes into account many factors to give reasonably accurate base products from the return signal, still the data need certain quality control measures. The use of simple thresholds for different variables can be quite useful in removing unwanted echoes [41,63]. The following quality control measures were considered for this study—(i) pixels with Zب > 70 dBZ or ρح < 0.7 were ignored, (ii) topography data from Shuttle Radar Topography Mission (SRTM) were used to remove ground clutter from hills present towards 40 km east of the radar as seen in Figure 1 [62,64]. Figure 2 shows the radar reflectivity during an event on 13 May 2018 before quality control (Figure 2a) and after quality control (Figure 2b). The clutter due to hills is present in the reflectivity field before quality control, which was removed nicely after applying the above-mentioned quality control measures. Other variables (Zد، Φد and ρح) were processed similarly.

![Figure 2. PPI diagrams of radar reflectivity at 2° elevation (a) before quality control and (b) after quality control. PPI diagrams of (c) folded Φد and (d) unfolded Φد at 18:54:12 IST, 13 May 2018.](image-url)
2.2. Convective-Stratiform Separation

Several studies have been done for the classification of precipitation into convective and stratiform parts using in situ measurements [65–67] and weather radars [68–72]. Convective and stratiform parts of the cloud systems exhibit significantly different behaviours in terms of dynamics as well as microphysics [73]. Vertical air motions within these two portions of a cloud system differ significantly; convective parts are mainly driven by large narrow updrafts (5–10 m s\(^{-1}\) or more), while stratiform portions are governed by gentler mesoscale ascents (<3 m s\(^{-1}\)). Thus, microphysical processes responsible for particle growth within the convective and stratiform parts are very different. Particles within convective cores regions mainly grow by riming or accretion (collection of supercooled liquid water droplets onto the ice particle surface), which leads to large/dense hydrometeors, whereas in the stratiform region vapour deposition and aggregation are dominating processes that lead to smaller and less dense ice hydrometeors (though large aggregates may exist).

The convective-stratiform separation method by Steiner et al. [68] is based on the texture of the radar reflectivity field adopted for the present study and is widely used by the radar community. This method basically checks for two criteria viz. intensity or peakedness criteria on the horizontal reflectivity field at 3 km height, to identify a grid point (pixel) as a convective centre. Any grid point with reflectivity at least 40 dBZ (intensity criteria) or greater than a fluctuating threshold (peakedness criterion) depending on the area-averaged background reflectivity (\(Z_{bg}\) calculated within a radius of 11 km around the grid point), is considered as a convective centre. For each pixel identified as a convective centre, all surrounding pixels within a certain radius of influence are also included as convective pixels. This radius of influence is dependent on \(Z_{bg}\). Once all the convective pixels are identified, the rest of the pixels with non-zero reflectivity values are assigned as stratiform pixels.

2.3. \(\Phi_{dp}\) Data Processing and \(K_{dp}\) Calculation

The differential propagation phase (\(\Phi_{dp}\)) is the phase difference between the horizontal and vertical polarised pulses on traversing through the atmosphere. The differential propagation phase is proportional to the water content along a rain path. Since most of the hydrometeors in the atmosphere are aligned with their major axis in the horizontal plane and it is a range cumulative parameter, the value of \(\Phi_{dp}\) increases with propagation path. Now, the unambiguous range of \(\Phi_{dp}\) usually is 180° in the alternate H/V transmission mode and 360° in the simultaneous H/V transmission mode. Hence, for a long propagation path in rain, \(\Phi_{dp}\) values can easily exceed the unambiguous range and then the \(\Phi_{dp}\) will be wrapped/folded, which is usually manifested as a sudden jump in the range profiles of \(\Phi_{dp}\). This issue with \(\Phi_{dp}\) is known as phase wrapping/folding [74,75]. The unfolding of these phases has been done by adding appropriate phase offset whenever such jumps are found [75]. So, even after the quality control steps mentioned in the previous section, \(\Phi_{dp}\) needs this extra processing before it can be used for further analysis. In Figure 2c, such a situation of phase wrapping is observed towards 15 km west of the radar during a convective event. Then, the phases are unfolded nicely and the unfolded \(\Phi_{dp}\) is shown in Figure 2d.

Specific differential phase (\(K_{dp}\)) is defined as the slope of range profiles of \(\Phi_{dp}\) [58,76,77] and is defined as follows.

\[
K_{dp}(r) = \left[ \frac{\Phi_{dp}(r + \Delta r) - \Phi_{dp}(r - \Delta r)}{2\Delta r} \right]
\]

\(K_{dp}\) is an important parameter for meteorological applications as it is closely related to rain intensity. More importantly it is insensitive to signal attenuation during propagation, radar calibration, partial beam blockage and the presence of hail [78,79]. This makes the specific differential phase very useful for precipitation estimation at heavy rain intensity or during partial beam blockage. Though the estimation of \(K_{dp}\) seems quite simple, it requires
further processing of $\Phi_{dp}$ range profiles before calculating the slope. $\Phi_{dp}$ is known to be a very noisy parameter, particularly in regions with low rain rates, and the process of differentiation increases this noise even further. To tackle this, we have applied a low-pass Butterworth filter [80,81] of order 10 with a cut-off scale of 2 km to reduce the statistical fluctuation, but keeping the overall features intact. Similar filters with similar cut-off scales were used in previous studies [74,82]. Figure 3a shows the $\Phi_{dp}$ plan position indicator (PPI) at 2° elevation angle during the convective event on 13 May after quality control and unfolding. Then, the low-pass filter has been applied on this $\Phi_{dp}$ and obtained a smoothed $\Phi_{dp}$ (Figure 3b). Small scale fluctuations in the $\Phi_{dp}$ field were removed in the filtered $\Phi_{dp}$. With this smoothed $\Phi_{dp}$ field, $K_{dp}$ has been estimated using Equation (1) and is shown in Figure 3c. Another $K_{dp}$ estimate using the slope of the linear regression line [83] has also been calculated. Both the methods gave similar $K_{dp}$ values. The $K_{dp}$ field shows high values close to 9° km$^{-1}$ at a distance of 5 to 15 km westward from the radar, indicating the presence of heavy precipitation. The blue line in this plot represents the 281° azimuth. Along this direction original $\Phi_{dp}$ (dot-dashed blue curve), filtered $\Phi_{dp}$ (solid blue curve), and estimated $K_{dp}$ (red curves) are shown in Figure 3d. The ranges of $K_{dp}$ values obtained here agrees quite well with previous studies on convective cases [47,74].

Figure 3. PPI diagrams at 2° elevation of (a) unfolded $\Phi_{dp}$, (b) filtered $\Phi_{dp}$ and (c) estimated $K_{dp}$. Blue line in (b,c) represents 281° azimuth. (d) Variation of $\Phi_{dp}$ (blue) and $K_{dp}$ (red) along 281° azimuth at 18:54:12 IST, 13 May 2018.

2.4. Hydrometeor Identification

A hydrometeor identification (HID) algorithm by Dolan et al. [47] is used to identify the types of hydrometeors present at different heights within a convective system. This is a fuzzy logic-based algorithm in which fuzzification of the inputs is done by calculating values of the membership functions corresponding to each fuzzy set (here, different hydrometeor types). Then, a fuzzy score for each fuzzy set is calculated using Equation (2). This step is called aggregation. Then, defuzzification is done by choosing the fuzzy set corresponding to the maximum fuzzy score, i.e., the hydrometeor with the highest fuzzy
logic score is the most probable hydrometeor type at that grid point within the radar scan volume.

$$\mu_i = \left[ \frac{W_{Z_{dr}} \beta_{Z_{dr},i} + W_{K_{dp}} \beta_{K_{dp},i} + W_{\rho_{hv}} \beta_{\rho_{hv},i}}{W_{Z_{dr}} + W_{K_{dp}} + W_{\rho_{hv}}} \right] \beta_{T,i} \beta_{Z_{dr},i}$$  \hspace{1cm} (2)

$$\beta = \frac{1}{1 + \left[ \frac{x-m}{a} \right]^b}$$  \hspace{1cm} (3)

Here, $\mu_i$ is the fuzzy logic score for the $i$th hydrometeor type. $\beta_{j,i}$ is the membership function for $i$th hydrometeor types and $j$th variable (Equation (3)). $W_j$ is the weight factor for the $j$th variable. The values of these membership function parameters and the weights are taken as in [47], which are obtained from simulations at C-band. Five variables viz. $Z_h$, $Z_{dr}$, $K_{dp}$, $\rho_{hv}$ and temperature ($T$) are used to calculate the fuzzy logic score. Seven types of hydrometeors have been considered viz. drizzle (DZ), rain (RN), ice crystals (CR), aggregates (AG), low-density graupel (LDG), high-density graupel (HDG), and vertically oriented ice (VI). Graupels are ice particles with a diameter of 2–5 mm, which grow mainly due to the riming process, i.e., collection of supercooled water droplets onto the surface of ice crystals and subsequent freezing. The temperature for the HID scheme has been obtained from radiosonde measurements by IMD Thiruvananthapuram at 5:30 IST (Indian Standard Time). Radar data interpolated on a 0.5 km $\times$ 0.5 km $\times$ 0.5 km grid have been used for the HID analysis.

2.5. Numerical Simulation

In the present study, we have used the state-of-the-art mesoscale model—Weather Research and Forecasting (WRF) model version-3.9 for the simulation of the thunderstorm observed over the southern peninsular India on 13 May 2018. We have considered three nested domains (D1, D2 and D3) of 9 km, 3 km and 1 km grid resolutions, as shown in Figure 4. For better simulation of the thunderstorm event, the innermost domain (D3) with 1 km resolution has been considered, as suggested by Rajeevan et al. [22]. The model used 50 vertical levels. It was initialised at 00 UTC of 13 May 2018 with the NCEP global final analysis data having spatial resolution of 0.25 degree. Furthermore, the boundary conditions were taken from the same analysis data.

The Kain–Fritsch cumulus parameterisation scheme [84] was used for D1 and D2, whereas explicit convection was allowed for D3 given its higher spatial resolution (1 km). For microphysics, the fast version of the spectral bin microphysics (Fast-SBM) scheme [85] was used. Fast-SBM includes four hydrometeor categories: water drops, ice/snow, graupel/hail and aerosol. In contrast to bulk microphysics schemes, which assumes a size distribution for hydrometeors, Fast-SBM uses 33 mass bins to explicitly describe the size distributions of each type of hydrometeors. Even though it is computationally much more expensive than bulk schemes, studies [86] show that Fast-SBM produce more realistic results than bulk schemes, particularly for deep convective systems. Among other physical parameterisation schemes that were used are: RRTM long-wave scheme, Dudhia shortwave scheme, Noah LSM scheme for land surface processes etc.
2.5. Numerical Simulation

In the present study, we have used the state-of-the-art Weather Research and Forecasting (WRF) model version 3.9 for the simulation of the thunderstorm on 13 May 2018. The model used is shifted slightly towards the higher value of 0.2 g cm⁻³ for the simulation of the thunderstorm event, as suggested by Rajeevan et al. [22]. The model used was the RRTM longwave scheme, Dudhia shortwave scheme, Noah LSM scheme for land surface processes etc. Besides, the microphysics schemes that were used are Fritsch cumulus parameterization [84], RRTM shortwave scheme, Noah LSM scheme for land surface processes etc. The convective-stratiform separation algorithm was implemented for all the volume scans available for all of these events. Within these radar volumes, the number of convective and stratiform pixels were found to be 4,09,049 and 14,39,977, respectively, which is equivalent to a convective and stratiform fraction of 22% and 78%, respectively. Figure 6a–c shows the contour frequency by altitude diagram (CFAD) of Z_dp over Convective and Stratiform Regions. The convective core is seen near 3 km height (Figure 6a), though such a feature is not present over stratiform regions (Figure 6d). The spread in the distribution of Z_dr and K_dp is much higher over convective regions compared to stratiform regions. Particularly, the spread is much higher at lower levels (below 5 km height). The CFADs of Z_dr and K_dp obtained in this study have quite similar features as the ones obtained by Machado et al. [87], e.g., almost uniform distribution across the vertical; although, their findings show that the centre of the CFAD of K_dp is shifted slightly towards the higher value of 0.2 g cm⁻³ as

3. Results and Discussions

3.1. Statistics of Z_h, Z_dr and K_dp over Convective and Stratiform Regions

An implementation of the convective-stratiform separation algorithm is depicted in Figure 5 during the convective event on 13 May 2018. Figure 5a shows the reflectivity field averaged between 2.5 and 3.5 km height. The convective-stratiform separation algorithm was then applied on this horizontal reflectivity field and the results are shown in Figure 5b. The red and blue pixels are identified as convective and stratiform precipitation, respectively. Not only high reflectivity regions, but also other regions with a strong gradient have been identified as convective regions.

Figure 6 shows the corresponding CFADs over the stratiform regions. The spread in the distribution of Z_dr over stratiform regions is much higher compared to convective regions. Particularly, the spread is much higher at lower levels (below 5 km height). The CFADs of Z_dr and K_dp obtained in this study have quite similar features as the ones obtained by Machado et al. [87], e.g., almost uniform distribution across the vertical; although, their findings show that the centre of the CFAD of K_dp is shifted slightly towards the higher value of 0.2 g cm⁻³ as

Figure 4. Nested domain configuration used for the simulation of the thunderstorm on 13 May 2018. Grid resolutions of the three domains (D1, D2 and D3) are 9 km, 3 km and 1 km, respectively. Colour bar shows the terrain height (m) over the simulation region.
compared to 0.1° km$^{-1}$ in our study. The reason behind the differences could be attributed to the fact that they used X-band radar data instead of C-band.

**Figure 5.** (a) PPI diagrams of radar reflectivity at 2° elevation averaged between 2.5 and 3.5 km height at 18:54:12 IST, on 13 May 2018 and (b) the identified convective (red) and stratiform (blue) regions.

**Figure 6.** Contour frequency by altitude diagram (CFAD) of $Z_{dr}$, $Z_{dp}$ and $K_{dp}$ for convective (a–c) and stratiform (d–f) regions. (g–i) Mean vertical profile with 1-σ error bars of $Z_{dr}$, $Z_{dp}$ and $K_{dp}$ over convective (red) and stratiform (blue) regions. (j–l) Frequency distribution of $Z_{dr}$, $Z_{dp}$ and $K_{dp}$ at 3 km height. The black dotted line represents rain rate of 10 mm h$^{-1}$. 
The mean profiles of $Z_{th}$, $Z_{dr}$ and $K_{dp}$ along with standard deviations are shown in Figure 6g–i. In the convective case (red), mean reflectivity values gradually increase with height from the ground and reach a maximum at ~3 km height and then gradually decrease towards upper levels. The peak value of the reflectivity is about 32 dBZ. Similar features in the reflectivity profile were found over the tropical region by Zipser and Lutz [88]. On the other hand, in the stratiform case (blue), mean reflectivity remains almost uniform (~18 dBZ) at lower levels and then increases gradually and forms a distinct peak near 5 km height. This peak in the reflectivity signifies the bright band (caused by enhanced reflectivity from melting ice particles near 0 °C level) over stratiform regions. Above the melting layer, the reflectivity decreases monotonically. The spread in the distribution is quite high for the convective case compared to the stratiform case, particularly at higher levels. The mean profile of $Z_{dr}$ over convective regions decreases monotonically towards upper levels and remains almost uniform at 6 km height. On the other hand, over stratiform regions, mean values of $Z_{dr}$ decrease gradually at lower levels, but from 3 km height it starts increasing and forms a distinct local maxima near 5 km height and again decreases towards higher heights. This local maxima in the $Z_{dr}$ may be utilised to identify the melting layer as an alternative to the peak in reflectivity profile. The mean values of $Z_{dr}$ are very close to each other for convective and stratiform cases over 5 km in height. Higher mean values of $Z_{dr}$ at lower levels over convective regions are due to the higher oblateness of larger raindrops, which were found at lower levels as a result of the coalescence process or due to the melting of bigger graupels. The spread in the distribution of $Z_{dr}$ is much higher at lower levels (below 4 km height), particularly over convective regions. The mean $K_{dp}$ profile remains almost uniform with small values (~0.2° km$^{-1}$) throughout all levels over stratiform regions. On the other hand, over convective regions, the mean $K_{dp}$ values are much higher at lower levels, which signifies the presence of strong precipitation over convective regions. The spread in the distribution of $K_{dp}$ is quite large over convective regions due to the larger spread in the raindrop size distribution as well as turbulent conditions.

Figure 6j–l shows the distribution of $Z_{th}$, $Z_{dr}$ and $K_{dp}$ over convective (red) and stratiform (blue) regions at 3 km height. The peaks in the distributions of $Z_{th}$ over the convective and stratiform regions are well separated; although, there is an overlap between the two distributions. The peaks for convective and stratiform regions occur at 34 dBZ and 20 dBZ, respectively. The dashed vertical line represents the reflectivity corresponding to the rain rate of 10 mm h$^{-1}$. Here, we have used $Z = 168R^{1.4}$ relation, which was obtained from another study by Jash et al. [89] over this region using micro rain radar data. This result clearly shows that the use of a rain rate threshold (e.g., 10 mm h$^{-1}$) to separate convective and stratiform rain is questionable, though such a simple classification method is often useful in many studies [65,66,90]. The distributions of $Z_{dr}$ over convective and stratiform regions are almost symmetric with a maximum occurrence frequency at 0.6 dB and 1.0 dB, respectively. For $K_{dp}$, the maximum occurrence frequency is at 0.4° km$^{-1}$ and 0.1° km$^{-1}$ over convective and stratiform regions. The distribution over convective regions have a much longer tail crossing beyond 3° km$^{-1}$, which resulted in a large error-bar in Figure 6i. Even though numerous studies on convective systems are available in the literature, such statistics on radar polarimetric variables ($Z_{dr}$, $K_{dp}$) are rare, particularly for thunderstorms over the Indian region. Hence, this analysis provides valuable quantitative information on the signatures/characteristics of convective and stratiform precipitation from radar polarimetric measurements.

3.2. Case Study of a Deep Convective System—13 May 2018

An in-depth analysis is performed on the convective event on 13 May 2018 for a detailed understanding of the evolution and structure of pre-monsoon convective systems over the southern peninsular India. It was an isolated deep convective system. Favourable synoptic and thermodynamic conditions help in the organisation of convective storms to develop into severe ones [30]. High moisture, atmospheric instability, vertical wind shear and a lifting mechanism are the different necessary conditions for the development
of thunderstorms. Hence, an overview of the synoptic conditions before and during the event will give more insights into its development. Geopotential height anomaly and wind data from ECMWF Reanalysis v5 (ERA5) at 12 UTC and vertical profile of equivalent potential temperature ($\theta_e$), mixing ratio, wind speed, wind direction from radiosonde measurements by IMD, Thiruvananthapuram at 00 UTC (i.e., 05:30 IST), were used to look into the environmental conditions for the event.

A low-pressure area formed in the south-west Arabian Sea (Figure 7a) on 13 May 2018, which was evident from the minimum geopotential height anomaly at 700 hPa levels between 55–65 E and 4–10 N. Even though it was far from the present study region, under the influence of this low-pressure area, the mean wind was from the Bay of Bengal to the Arabian Sea in the easterly direction (Figure 7a) converging towards the low-pressure centre. A strong negative gradient of the $\theta_e$ profile (Figure 7b (red)) up to 3 km height shows the instability present in the lower atmosphere during morning time. The mixing ratio profile (Figure 7b (blue)) indicates the presence of a moist layers between 2 and 6 km levels, and also suggests the existence of favourable atmospheric conditions for the formation of a thunderstorm. Figure 7c shows the profiles of wind speed (red) and wind direction (blue). They changed abruptly along the vertical, which is due to the turbulence associated with the unstable atmospheric conditions. Heavy rainfall in isolated places were reported over Kerala and Tamil Nadu by IMD. These conditions led to the formation of a convective system over the inland region on 13 May 2018 in the afternoon hours between 16:00 and 22:30 IST.

![Figure 7.](image)

**Figure 7.** (a) Horizontal wind vectors overlaid with geopotential height anomaly on 13 May 2018 at 12 UTC using ERA5 dataset. (b) Vertical profiles of $\theta_e$ (red) and mixing ratio (blue), (c) vertical profiles of wind speed (red) and wind direction (blue) on 13 May 2018 at 12 UTC from radiosonde measurements.

### 3.2.1. Evolution of the Storm

The development of this convective system on 13 May 2018 is captured in the plan position indicator (PPI) diagrams of radar reflectivity field at consecutive times during the event (Figure 8). The convective clouds started developing over the land around 25 km east of the radar location at 16:00 IST and then gradually it started moving westward. This movement of the system was due to the prevailing easterly wind (Figure 7a) as discussed previously. The cloud system passed over the NCESS location around 18:00 IST (Figure 8d).
3.2.1. Evolution of the Storm

The development of this convective system on 13 May 2018 is captured in the plan position indicator (PPI) diagrams of radar reflectivity field at consecutive times during the event (Figure 8). The convective clouds started developing over the land around 25 km east of the radar location at 16:00 IST and then gradually it started moving westward. This movement of the system was due to the prevailing easterly wind (Figure 7a) as discussed previously. The cloud system passed over the NCESS location around 18:00 IST (Figure 8d).

As soon as it reached over the NCESS location, extremely heavy rainfall started, which was observed in the rain rate measured by the disdrometer (Figure 9a). The rain rate crossed 100 mm h\(^{-1}\) and sustained in that range for over an hour. Gradually the rain intensity declined to a range of 0.1–1 mm h\(^{-1}\), which was basically the stratiform precipitation following the main convective activity. The rain DSD obtained by the disdrometer showed an abundance of bigger raindrops (diameter > 3 mm) during this intense convective spell, followed by smaller drops at the later stage of the event. Cloud base height measured by ceilometer shows (Figure 9b) the presence of multilevel clouds. Before 17:00 IST mostly high-level clouds were detected (CBH ~7 km) and then, just before the precipitation started, all three cloud layers had cloud bases below 2.5 km. Such a low cloud base height and high cloud top measures the significant depth of the cloud system.

The deep convective cloud system eventually moved over the Arabian Sea around 30 km westward from the radar location. Meanwhile, it turned into a stratiform system (Figure 8g–i). The IMD weather report also mentioned the rainfall during these hours. This event was associated with the rapid development of deep convective clouds, as observed in the evolution of the cloud top infrared brightness temperature (IRBT) measured from satellite (INSAT-3DR). A lower brightness temperature signifies a higher cloud top height [91]. Figure 10a–e shows the spatial and temporal evolution of the brightness temperature during this event. Around 18:00 IST, much of the region had a brightness temperature below 200 K revealing the occurrence of deep clouds over most of the region. Figure 10f shows the temporal evolution of the brightness temperature over the NCESS location (averaged over a 12 × 12 km area centred at NCESS). A rapid decrease in the brightness temperature started at 15:45 IST and reached a minimum value of 185 K at 17:45 IST, which exemplifies how fast such a deep system can develop within such a short span of time.
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Figure 9. (a) Time series of rain rate (red curve) overlaid with rain DSD (colour bar) using disdrometer data and (b) time series of cloud base height (m) of layer 1 (pink), layer 2 (cyan) and layer 3 (blue) clouds during the convective event on 13 May 2018 using ceilometer data.

Figure 10. (a–e) Spatial–temporal evolution and (f) time series of infrared brightness temperature (K) over NCESS during the convective event on 13 May 2018 using INSAT-3DR satellite data.

The CAPE value of 1713 J kg\(^{-1}\) was observed from the nearest radiosonde measurements in the mooring hour (05:30 IST), which was indicative of already existing moderate instability in the atmosphere which built up further and eventually led to strong updraft during evening hours.
3.2.2. Vertical Structure of the Storm

The vertical structure of the storm in terms of DWR polarimetric measurements and associated hydrometeor identification are shown in Figure 11. An averaged reflectivity between 2.5 and 3.5 km height during the rapid initial development stage of the storm reveals active convective regions (Figure 11a). Then, a vertical cross section along the convection line AB has been considered to analyse the vertical structure of the storm. Figure 11b shows the vertical cross section of reflectivity at horizontal polarisation ($Z_{hv}$) along the convection line AB. The x-axis represents the distance from point A towards point B. Reflectivity values greater than 30 dBZ, reaching up to 10 km height, signifies the existence of strong updraft within the convective core region. This strong updraft can keep the larger hydrometeors (bigger raindrops, graupels etc.) float aloft for a longer period, giving them more time to grow further by the collision-coalescence process for raindrops and by riming process for ice particles [92]. Since reflectivity is proportional to the 6th power of the particle diameter [58], these larger particles produce such strong reflectivity values even at higher altitudes.

![Figure 11](image_url)

**Figure 11.** (a) Radar reflectivity averaged between 2.5 and 3.5 km height on 13 May 2018. Vertical cross section of (b) reflectivity, (c) $Z_{dr}$, (d) $\rho_{hv}$, (e) $K_{dp}$ and (f) identified hydrometeor types along AB convection line at 17:07:59 IST.

Figure 11c shows the vertical cross section of differential reflectivity ($Z_{dr}$) along the convection line. The $Z_{dr}$ value gives a measure of the oblateness of precipitation particles, and hence could be useful in distinguishing between larger raindrops, hail, and graupel due to differences in shape and orientation. Since raindrops (diameter > 1 mm) are deformed into an oblate spheroid shape due to aerodynamic forces [93] with a preferred orientation of their major axes in the horizontal direction (and therefore $Z_{h} > Z_{e}$), $Z_{dr}$ is positive and increases with raindrop size. This increase in the value of $Z_{dr}$ with raindrop size is shown in terms of a polynomial fit by Bringi et al. [36] between observed $Z_{dr}$ and mean drop diameter measured by disdrometer. In our analysis, $Z_{dr}$ values greater than 2 dB were observed, which indicates the presence of bigger raindrops or melting bigger ice particles [94] below 4 km height. Bigger raindrops were also observed in the disdrometer measurements of rain DSD (Figure 9a). $Z_{dr}$ values are much smaller at higher altitudes (above 0° isotherm ~5 km height) as the ice particles, such as aggregate, graupel, and hail, tend to be spherically symmetric or tumble while falling, causing low values of $Z_{dr}$. The lower value of dielectric
constant for ice compared to water is another factor behind the lower value of $Z_{dr}$ for ice particles. Within the strong convective regions at heights above the melting layer, a higher value of $Z_{dr}$ along with high value of $K_{dp}$ indicates supercooled liquid drops above freezing level [95].

$\rho_{hv}$ shows high values (>0.95) throughout the entire cross section (Figure 11d) and $\rho_{hv}$ depends on several factors, such as eccentricity, distribution of canting angle, irregular shape and mixture of different types of hydrometeors. Relatively lower values of $\rho_{hv}$ at the central region and at higher altitudes within the cross-section, could be attributed to the mixture of ice particles with rain.

The estimated $K_{dp}$ (Figure 11e) shows that the spatial pattern of $K_{dp}$ is in tandem with that of reflectivity, though there are differences. High values of $K_{dp}$ (greater than 5 $^\circ$ km$^{-1}$) below melting level suggest the presence of intense convective precipitation with bigger raindrops formed due to the coalescence process or due to the melting of graupel. As drop eccentricity increases with diameter, the differential propagation phase increases, causing higher values of $K_{dp}$ within regions of intense convective precipitation. A similar structure of $K_{dp}$ within convective regions is reported by Ryzhkov et al. [61]. Higher values of $K_{dp}$ above freezing level suggests prevalence of supercooled droplets, which can help in the formation of graupel particles via the riming process.

Identified hydrometeor types are shown in Figure 11f. Below the melting level, it is mainly dominated by rain (RN) and above the melting level, ice aggregates (AG) are the dominating hydrometeors. At heights between 4.5 and 8 km, within the convective core regions, graupel (HDG) particles are abundant. Similar findings were obtained in Dolan et al. [47], in which HDG was found close to the melting level, as in our study, and LDG at higher heights. Within such convective cores reaching up to 10 km height, liquid droplets are pushed to heights much above the freezing level and they stay there as unstable supercooled droplets. Upon contact with ice-aggregates they immediately freeze onto the surface forming bigger ice particles viz. graupel. The strong updraft can sustain these graupels in air for longer, helping them grow even further. The presence of vertical ice indicates the existence of an electric field, which forces these particles to orient vertically and this could be due to the charging via the collisions between graupels and smaller ice crystals, as confirmed by different laboratory experiments [48–50].

3.2.3. WRF Simulation of Reflectivity, Graupel Mixing Ratio and Rainfall

The radar reflectivity field was obtained through post processing of the WRF model outputs using the ARWpost package. The simulated radar reflectivity structure during the thunderstorm on 13 May 2018 is shown in Figure 12a–c. The spatial distribution of the radar reflectivity at 3 km height (which was previously considered for the convective-stratiform separation) at 19:15 IST during the mature stage of the storm is shown in Figure 12a. Strong radar reflectivity values greater than 40 dBZ shows the most active convective regions. There are many similarities with the observed reflectivity fields (Figure 8d) from DWR, but at an earlier time (18:08 IST), i.e., in the model the storm developed almost an hour late compared to what was observed. A 5 $\times$ 5 km box (white rectangle) over an active convective region was considered to study the vertical structure of the simulated storm. Figure 12b shows the time–height cross section of the reflectivity averaged over the box. The peak reflectivity occurred around 19:15 IST, showing an hour delay in the development of the storm compared to the observation. Figure 12c shows the east–west vertical cross section along the dotted line passing through the centre of the box. The main convective region is spread over a distance of ~25 km surrounded by lower reflectivity regions. The reflectivity core reached beyond 12 km height showing the severity of the storm. Such strong reflectivity cores were also seen in the DWR observation (Figure 11b).
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Figure 12. (a–c) WRF model simulated reflectivity—(a) at 3 km height, (b) east–west vertical cross section along the horizontal dotted line at 19:15 IST and (c) time–height cross section of reflectivity averaged over the white box area. (d–f) Simulated graupel mixing ratio—(d) at 6 km height, (e,f) same as in panel (b,c), but for graupel mixing ratio. (g,h) Simulated rainfall—(g) rain rate at 19:15 IST and (b) time series of the rain rate averaged over the white box area.

The occurrence of graupel within the storm is shown in Figure 12d–f in terms of the graupel mixing ratio. Graupels are often abundant in thunderstorms because of strong updraft, which help in their growth. Graupels are particularly important while studying thunderstorms as they play a crucial role in the occurrence of lightning. Figure 12d shows the spatial distribution of the graupel mixing ratio at 6 km height. This particular height was considered as graupels are mostly found around this height. High values of graupel mixing ratio were seen over the active convective regions. Then, the same box (as in Figure 12a) was considered to study the structure of the graupel occurrence across the vertical. Figure 12e shows the time–height cross section of the graupel mixing ratio. High values of graupel mixing ratio started to be seen around 19:00 IST at heights beyond 5 km. An abundance of graupel was seen up to 20:30 IST. Figure 12f shows the east–west vertical structure of the graupel mixing ratio along the black dotted line. Graupel started to show up from 5 km height and highest mixing ratio was observed near 8 km height. Such structure of graupel occurrence across the vertical was observed from the hydrometeor identification analysis (Figure 11f), which showed graupel was identified between 5 and 9 km height.

Rainfall is probably the most important parameter in meteorology. So, the study will be incomplete without exploring how well the model captured the rainfall. Figure 12g shows the spatial distribution of the surface rain rate (mm h$^{-1}$). Strong convective rainfall greater than 10 mm h$^{-1}$ was produced over different regions, particularly over the convectively active regions. Then, the same box (as in Figure 12a) was considered to obtain the rainfall time series during the course of the event. Figure 12h shows the time series of the rain rate averaged over the box area. The model data has a time resolution of 5 min for the innermost domain (D3). The rainfall started around 18:30 IST and reached the maximum value of 52 mm h$^{-1}$ around 19:15 IST. Then, it started to decrease gradually and stopped at around 22:30 IST. So, the model simulated rain rate time series agrees quite well in terms of intensity as well as overall evolution, with the observed rainfall (Figure 9a) from the
The WRF model could capture the main features of the storm well as compared to the observations. In particular, the horizontal patterns of reflectivity as well as the vertical profiles were captured nicely as were seen in the DWR observations. The presence of graupel across the vertical was captured well, as the same was identified via hydrometeor identification analysis. The rain rate time series was beautifully simulated as confirmed by the disdrometer rain rate observation. The model could capture all of these features with the spectral bin microphysics scheme. We tried with a few other configurations with bulk microphysics as well (not shown here), but those results were far from what SBM could capture as shown here. This may be because of the isolated deep nature of this particular event considered here.

4. Conclusions

The present study is focused on the structure of pre-monsoon convective systems over a tropical coastal region in the southern peninsular India. Statistics on radar polarimetric variables for pre-monsoon convective systems have been obtained for the first time over the Indian region. Using the quality controlled DWR data, 11 convective events have been identified by inspecting the radar reflectivity fields. Out of which, a prominent convective event, which occurred on 13 May 2018, has been analysed in detail to understand the development and structure of a typical pre-monsoon convective system. Convective-stratiform separation has been done for all of the events. The following are the major conclusions of the study:

1. The distribution of differential reflectivity ($Z_{dr}$) and specific differential phase ($K_{dp}$) have much higher spread over convective regions, particularly below 5 km height. The distribution of $K_{dp}$ is almost uniform across the vertical over the stratiform regions. The mean profile of $Z_{dr}$ over stratiform regions shows a distinct local maximum near melting level, which could be utilised to identify stratiform precipitation.

2. The percentages of convective and stratiform pixels were found to be 22% and 78%, respectively. The distributions of the reflectivity values over convective and stratiform regions show that a single threshold for reflectivity or rain rate may not be useful for convective-stratiform separation as used in many studies.

3. The analysis of the thunderstorm on 13 May 2018 clearly exemplifies that pre-monsoon deep convective systems can develop rapidly within a very short span of time and cause heavy precipitation. Satellite-based cloud top temperature reveals the development of much deeper cloud.

4. Vertical structures inside the storm during the rapid development stage have been obtained by taking vertical cross sections of reflectivity through major convective regions. Convective cores reaching 10 km in height have been observed due to the strong updraft. High values of $Z_{dr}$ at lower levels were observed due to the oblate spheroid shape of the bigger raindrops. The structure of the $K_{dp}$ field is quite similar to that of reflectivity. High values of $K_{dp}$ reveals the presence of intense rainfall, as $K_{dp}$ is mainly dominated by bigger raindrops.

5. The implementation of fuzzy logic-based hydrometeor identification showed the presence of graupel at middle levels within the convective core regions, revealing the presence of strong updrafts. Ice aggregates and rain were the dominant hydrometeors above and below melting level, respectively. The presence of vertical ice signifies the presence of an electric field inside the storm. Such an electric field may be generated due to non-inductive charging via collision between graupel and smaller ice crystals.

6. Numerical simulation using the WRF model with the spectral bin microphysics (SBM) scheme could produce most of the features of the storm reasonably well. In particular, the simulated reflectivity, graupel mixing ratio and rainfall were in good agreement with the observed values. These results show the capability of the SBM scheme in simulating deep convective clouds.
7. It would be worth studying the observed lightning activity (if any) during these events, as the presence of vertical ice indicates the presence of a strong electric field. If major lightning activity occurred during these events, then it would support the collision charging mechanism, as graupels were identified within the convective core regions.

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