Chapter V

Vertical structure of Shallow and Deep Convective Precipitating Clouds
In this chapter, temporal variations of vertical structure of precipitating clouds have been studied. To understand diurnal and seasonal variations of the occurrence of [Deep Convection (DC), Shallow Convection (MC), Mixed (convection-stratiform) (MC) and Stratiform (SF)] precipitating cloud systems four years of data utilized. In this chapter, a new form of gamma drop size distribution (DSD) model is employed with two arbitrary moments as free DSD parameters to enhance the flexibility in studying the characteristics of gamma DSD model to fit the WPR spectrum for stratiform precipitating clouds.

5.1 Introduction

The capability of radio waves to penetrate cloud and precipitation has placed atmospheric wind profiler radars (WPR) in an unchallenged position for remote sensing of the convective atmosphere (Gage et. al., 1994). The UHF/L-band WPRs are designed to operate unattended, and for long-term monitoring of winds. Because, their wavelengths [915MHz (33 cm), 1290 MHz (23 cm) and 1357.5 MHz (22 cm)] are short compared to other wind profiling radars, and much more sensitive to precipitation than other wind profilers (Balsley et. al., 1982; Ralph, 1995; Rao et. al., 1999; Reddy et. al., 2002) that typically operate at ~50 and 404 MHz. The UHF wind profilers are capable of revealing details about the vertical structure of precipitating clouds in the tropics (Gage et. al., 1996; Atlas et. al., 1999; Renggono et. al., 2001). Since the vertical distribution of diabatic heating depends on the vertical structure of the convective system, it is important to study the vertical structure of the precipitating clouds occurring in the tropics. The vertical structure of tropical mesoscale convective systems (MCS) has been studied using various instruments (including wind profiler) during many international campaigns, including GARP Atlantic tropical experiment (GATE) in the eastern Atlantic Ocean (Williams et. al., 1997), equatorial mesoscale experiment (Webster and Houze, 1991) (EMEX), winter MONEX in South China Sea (Chang and Lau, 1980), Down Under Doppler and Electricity Experiment (DUNDEE) (Cifelli and Rutledge, 1994) and tropical ocean global atmosphere-coupled ocean-atmosphere response experiment (TOGA-COARE) (Webster and Lukas, 1980). As noted by Gage et. al., (1994), the installation of UHF profilers (i.e., vertically pointing Doppler radars)
in the tropics will provide a means for the ongoing description of precipitation and vertical-velocity structure within tropical MCSs, thus complementing the longer-standing function of VHF profilers in monitoring Elnino Southern Oscillations (ENSO) and other sources of variability in tropospheric wind profiles. A study of precipitating cloud systems in the tropics using a 915 MHz wind profiler had been carried out by Gage et. al., (1996) and Williams et. al., (1995) at Manus Island. Ohno et. al., (2000) used similar radar systems in Biak and Christmas Island to study the occurrence of precipitating clouds. In north western Pacific Ocean region, radar remote sensing of the tropical monsoon cloud systems is sparse. As the tremendous increased amounts of precipitation becomes more and more frequent, understanding of the vertical structure of the convective precipitating cloud system over Oceanic region becomes important issue and needs further investigations.

Measurement of the raindrop size distribution (DSD) is important for many reasons and is also crucial for determining rain attenuations at microwave and millimeter wavelengths. Three different distribution functions are used to describe raindrop size spectra, namely, the Marshall and Palmer (Marshall and Palmer 1948) type of exponential distribution, the gamma distribution (Ulbrich and Atlas 2007) and the lognormal distribution (Ajayi and Olsen, 1985; Feingold and Levin, 1986). Large number of measurements have shown the inadequacy of the exponential distribution model, particularly when the data are averaged over short intervals. It is generally agreed that the exponential distribution is valid only for data averaged over long periods of time (Joss and Gori 1976), or over large volumes of space. Raindrop spectra often tend to have a mono-modal distribution, which can be modeled by the gamma distribution function.

In recent years, use of the wind profiler radars for DSD measurements has proliferated. This is because the ability of wind profiler radars to provide frequent vertical profiles of horizontal air motion under almost any weather condition has made them a unique and valuable tool. It has been established that the Doppler power spectra measured by wind profiler radars are used to determine both vertical air motion and hydrometeor terminal velocity simultaneously, in the conditions of stratiform precipitation. For the first time, using the 46.5 MHz middle and upper atmosphere...
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(MU) wind profiler radar located near Kyoto, Japan, Wakasugi et al., 1986 resolved both the Bragg (turbulence) and Rayleigh (precipitation) scattering components in the Doppler spectrum. Later, Sato et al., 1990, developed a computer algorithm to find the initial guess involved in the process of fitting a curve directly from the original spectra under precipitation conditions. Currier et al., 1992, corrected for the effect of turbulence and mean vertical velocity on the precipitation spectra observed by a 915 MHz radar and the information obtained by 50 MHz radar and generalized the model drop size distributions to a sum of gamma distributions (Williams, 2002) reviewed existing DSD models and also retrieved simultaneous ambient air motion and raindrop size distributions from 915 MHz vertical incident wind profiler radar observations. A number of different techniques have been utilized to separate the clear-air vertical motion from the hydrometeor fall speeds in wind profiler radar spectral data. Some of the procedures include first-moment analysis of the Doppler spectra (Chu and Lee, 1994), least squares fit of the Doppler spectra with one or two Gaussian approximations (Rajopadhyaya and Vincent et al., 1993, Rao, et al., 1999) and non-linear least squares fit of the Doppler spectrum with a DSD approximation for the hydrometeor component of the spectra and a Gaussian distribution for the turbulent component (Chu and Lee, 1991, Rajopadhyaya et al., 1993, Rao and Raghavan, 1999, Lucas et al., 2001). Drop size densities also are influenced by the precipitation formation processes in clouds. Since maritime and continental aerosols are significantly different leading to differences in cloud drop size spectra, one may assume that maritime and continental rain drop size densities (DSDs) are different, too. In fact, only little is known about rain drop size distribution over sea due to a lack of suitable data over the Ocean. To fill this gap, measurements were performed during the recent 10 years at different Oceanic region viz., over the Baltic Sea, the North Sea, Atlantic, Indian, and Pacific Oceans by different research groups. For subsets of the existing data, Wind Profiler Radar information is available to decide whether precipitation is of convective or stratiform nature. Drop size densities may also differ for prevailing stratiform or convective precipitation (Uijlenho et al., 2003). This enables to investigate differences in drop size spectra of convective and stratiform precipitation.
5.2 Instrumentation and Data Base

The primary instrument used for estimation of vertical profiles of DSD parameters is L-band wind profiler radar operated at 1290 MHz. At this frequency, the wind profiler is more sensitive to precipitation echoes than clear-air echoes that are dominant in profiler Doppler spectra of VHF frequencies (i.e., 50 and 404 MHz). One vertical beam and two off-vertical beams in the orthogonal directions are currently used and the off-vertical beams are tilted 15 degrees from the zenith direction. Measured radial velocities along each beam are used to calculate wind speed and direction. Three moments of signal-to-noise ratio (SNR), Doppler velocity, and spectral width are also calculated from observed Doppler power spectra (256-FFT points) at each range gate height. The Doppler spectra data observed along the vertical beam were only used for vertical structure of precipitating clouds and retrievals of DSDs/rainfall parameters in this study. Time and height resolution of the data is 3 min and ~200 m, respectively. At high mode, the maximum height is ~11.0 km msl. The details of the WPR parameters are listed in Table 5.1.

Table 5.1: Operational parameters of the 1290-MHz Wind profiler

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Radar Wavelength</td>
<td>23 cm</td>
</tr>
<tr>
<td>Beam width</td>
<td>6°</td>
</tr>
<tr>
<td>Pulse width</td>
<td>1400 ns</td>
</tr>
<tr>
<td>Inter Pulse Period</td>
<td>71000 ns</td>
</tr>
<tr>
<td>Number of coherent integrations</td>
<td>40</td>
</tr>
<tr>
<td>Number of spectral averages</td>
<td>32</td>
</tr>
<tr>
<td>Number of FFT points</td>
<td>256</td>
</tr>
<tr>
<td>Height resolution</td>
<td>~200</td>
</tr>
<tr>
<td>Number of range gates</td>
<td>55</td>
</tr>
</tbody>
</table>

Radar reflectivities measured at 300 m msl from a micro rain radar collocated with the wind profiler were used for calibrating profiler Doppler spectra and reflectivities. The micro rain radar operates at 24 GHz frequency (i.e., Ka band), providing data such as radar reflectivity, rain rate, liquid water content, and Doppler spectra at a height resolution of 200 m up to 5 km msl (below bright band). The wind profiler is almost insensitive to attenuation due to its large wavelength (~23 cm), while the micro rain radar is sensitive to attenuation at moderate to high rainfall rates. Rain
rates from the micro rain radar at 300 m msl and an Disdrometer at the surface were compared with those retrieved from profiler Doppler spectra in order to validate the profiler retrievals.

In this study, simultaneous observations from L-band Wind profiler radar, Ceilometer, Disdrometer, Micro Rain Radar, Radiosonde and surface meteorological data collected during four years from April 2003 to March 2007 at Aimeliik, Palau are used to understand the vertical structure of the precipitating clouds characteristics and retrieval of Raindrop Size Distribution profiles over Palau in the Pacific Ocean during Easterly and Westerly monsoon.

5.3 Vertical structure of Precipitating Clouds

In this section, the capability of UHF/L-band wind profiler radar (WPR) has been illustrated to observe the vertical structure of mesoscale precipitating cloud systems. WPR is sensitive to fluctuation in the radar refractive index or turbulence in clear air as well as the presence of hydrometeors. These hydrometeors are detected at ranges resolved by the wind profiler (Eclund et. al., 1995) yielding the vertical structure of the precipitation system. The advantage of the UHF wind profiler is sensitive to hydrometeors but it does not suffer attenuation in rainfall unlike microwave frequencies over 10 GHz. Another advantage of the wind profiler is to measure directly the vertical wind component within a convective environment (Gage et. al., 1994).

On 19 September 2003, a widespread stratiform region with short-lived convective cores in a precipitation system that developed near the observational site was observed by the wind profiler radar. Figure 5.1 shows time-height sections of radar reflectivity (Z), Doppler velocity (W_dop), and spectral width (i.e., three moments of WPR spectra) and time series of radar reflectivity & rain rates (R) retrieved from the Disdrometer at the surface during the analysis period. During this period, the precipitation system was not well organized and widespread based on the composite radar images but high rainfall rates up to ~40 mm/hr (as shown in were produced as short-lived convective cells pass over the observational site. It is shown in Figure 5.1a that a bright band is observed at around 4.5 km msl during most of the analysis period and its intensity and thickness markedly change with time. In Figures 5.1a and 1b, weak
to moderate $W_{dop}$ (negative downward) was observed below a weak bright band from 1850 to about 2130 UTC.

Fig. 5.1: Time-height sections of (a) radar reflectivity (dBZ), (b) Doppler velocity (m s$^{-1}$), and (c) spectral width (m s$^{-1}$) and (d) time series of Radar Reflectivity and rainfall rate (mm hr$^{-1}$) retrieved from the Disdrometer at the surface on 19 September 2003.
In Figure 5.1c, spectral widths show a notable contrast between the rain and ice regions except for thin layers of large spectral width near the bright band. The large spectral widths exceed about 4.5 m s\(^{-1}\) and are centered near 4.7 km, which appears to be a consequence of wide distributions in fall speeds of snow aggregates. The narrow region of small spectral widths less than 2 m s\(^{-1}\) near the level of maximum \(Z\) (4.0-4.5 km) between 1100 and 1200 LT suggests much less breakup and more uniform distributions of fall speeds relative to that at the top of the melting layer.

Following this, a disrupted bright band was observed from about 0850 to 1620 LT when high rainfall rates were observed near the surface as shown in Figure 5.1d. Considering the relation, \(w = W_{\text{dop}} - V_f\) where \(V_f\) is terminal fall speed (all the signs are negative downward), \(w\) increases as \(W_{\text{dop}}\) increases for a given \(V_f\) of a raindrop. Thus, the disrupted or less well-defined bright band is probably attributed to updrafts through the melting layer up to ~1 km above where relatively smaller \(W_{\text{dop}}\) is observed. In addition, large spectral widths up to ~5 m/s near the melting layer indicate that there were comparatively higher turbulent motions associated with the updrafts.

5.3.1 Wind Profiler Radar measurements of vertical structure of a Mesoscale Convective Precipitating Cloud Systems

The occurrence of deep convection in the tropical ocean plays an important role in the global circulation, since it transports heat, water vapour and so on, from the MBL to the upper troposphere. The vertical distribution of diabatic heating depends on the vertical structure of the convective system; hence it is important to study the vertical structure of the precipitating clouds occurring in the tropics. Wind profiler radar observations yield time height cross-section of reflectivity, Doppler velocity, and spectral width that illustrate the evolution of precipitating clouds. An example of the ability of the WPR was to give detailed information on the vertical structure of mesoscale convective systems. Figure 5.2 illustrates the time-height section of a convective precipitating cloud system passed overhead of WPR site on 22 & 23 June 2006. Figure 5.2 [(a),(b) and (c)] shows the equivalent radar reflectivity factor (dBZ), Doppler vertical velocity and spectral width of hydrometeors and clear air echos. The letters DC, MC, SC and SF indicate deep convective, mixed convection, shallow convection and stratiform precipitating cloud systems, respectively.
Fig. 5.2: Time-height sections of (a) radar reflectivity (dBZ) and (b) Doppler velocity (m s\(^{-1}\)) and Spectral width (m s\(^{-1}\)) observed on 22 & 23 June 2006 from Convective precipitating clouds passage overhead of the Wind Profiler Radar site.
5.3.2 Classification of vertical structure of Precipitating Clouds

The proposed technique for identification of precipitation types uses reflectivity, Doppler Velocity and Spectral width from wind profiler radar. Four precipitation types are prescribed by analyzing vertical structures of reflectivity and radial velocity from measurement in the vertical beam of wind profiler radar. They are (1) stratiform (SF), (2) mixed convection (MC) (3) deep convection (DC) and (4) shallow convection (SC) (based on classification of Williams et. al., (1995)). These types were mainly characterized by existence and strength of melting layer signature, reflectivity structure above or below bright band and vertical Doppler velocity. The results reveal very discernible characteristics among four precipitation types. The classification of precipitation is based on presence or absence of bright band structures, the depth of atmosphere occupied by hydrometeor and high value of SNR of back-scattered signal (Pan et. al., 2010).

![Radar reflectivity and Doppler velocity](image)

**Fig. 5.3** Vertical variations of spectral power density (dBZ/m s\(^{-1}\)) observed during deep convection, shallow convection, mixed (convective/stratiform) and Stratiform precipitating cloud fractions on 16 May 2003.
The vertical structure of diabatic heating is an important factor, linking large scale atmospheric circulation systems to mesoscale tropical convection. Moreover, the diabatic heating rate profile is different from stratiform regions and deep convection in mesoscale convective systems. It is important to differentiate the different types of precipitating convective systems that are present in tropical India. The Doppler spectra are power-weighted distributions of the radial velocities of the scatters within the radar resolution volume. Typical examples of the Doppler spectra observed by the WPR on 22 & 23 June 2006 during deep convection, shallow convection, mixed, and stratiform periods of a mesoscale convective system (MCS) are shown in Figure 5.3[(a)-(d)]. The positive Doppler velocity indicates upward motion. The convective region is characterized by Doppler vertical velocities, which are greater than the typical fall speeds of ice crystals or snow [greater than approximately 1-3 $\text{ms}^{-1}$, Figure 5.3 (a)]. Precipitating particles grow primarily by accretion of liquid water. Ice particles in the upper levels of convective clouds often grow by rimming in the convective updrafts. Due to the relatively strong updrafts in convective regions the radar reflectivity shows well-defined vertically oriented cores of maximum reflectivity. The relatively large velocities lead to intense turbulence. The Doppler spectra in the mixed (convection/stratiform) [Figure 5.3(c)] shows two echoes up to a height of ~1 km. The echo near the zero Doppler shifts corresponds to that of the background air, while the echo on the negative side of the Doppler spectrum is due to the hydrometeors. The hydrometeor radar reflectivity factor and Doppler velocity are found to be smaller during the transition period compared to the convection. In contrast to the convective region, stratiform precipitation [Figure 5.3(d)] occurs when the vertical air velocity is much less than the terminal fall velocity of snow particles. With this condition, ice particles in the upper levels of clouds must fall and all the growth of the precipitating particles must occur while falling. At high levels the ice particles grow mainly by vapour deposition. When they descend to within about 2.5 km of the freezing level (about 4.5 km), aggregation and rimming can occur. At upper levels the ice crystals, which have formed in a region of predominantly super-cooled water, undergo a Bergeron process. In a Bergeron process, ice crystals increase in size by diffusion of water vapour from neighbouring water droplets due to the lower saturation vapour pressure of ice. The
melting layer in the stratiform region is often evident as a horizontal band of high radar reflectivity approximately 0.5 km thick.

### 5.3.3 Diurnal variation of Precipitating Clouds

The WPR can provide the climatology of vertical structure of the precipitating cloud systems from long-term observations over Palau. During easterly and westerly monsoon seasons convective systems occur in several forms with varying degrees of organization including isolated single cell convection, cloud clusters, monsoon depressions and tropical cyclones. Each of these systems has different vertical structural characteristics (as observed by the WPR), such as their distribution of stratiform and convective precipitation. Figure 5.4 shows the observation results of the precipitating cloud systems over Aimeliik. Diurnal variation of convection seems frequently occurs in the night time over Palau. Several researchers observed overnight enhancement in convection over the oceans (Gray and Jacobson 1977; Albright et. al., 1985; Chen and Houze 1997; Hall and Vonder Haar 1999; Zuidema 2003). Gray and Jacobson (1977) and they proposed that convection over oceans was modulated by daily variations in the horizontal divergence field, which are generated by differential radiative heating between cloud-free areas and areas where convection is present.

![Fig. 5.4: Diurnal variation of precipitating clouds over Aimeliik.](image-url)

**Fig. 5.4: Diurnal variation of precipitating clouds over Aimeliik.**
Another mechanism, discussed by Dudhia (1989), Tao et al., (1996), and Dai (2001), proposes that the overnight enhancement of convection over oceans is caused by a decrease in night time cloud entrainment due to the impacts of night time long-wave cooling on relative humidities. Randall et al., (1991) concluded that the observed diurnal cycle over the oceans could be qualitatively accounted for by the direct radiation–convection interactions (i.e., daytime absorption of shortwave radiation by the upper portions of the convective clouds increases the static stability in cloudy regions and weakens vertical motions, while night time long wave cooling decreases the static stability and enhances convection). The occurrence of stratiform clouds is different from that of convective clouds. The highest percentage of occurrence appears during 1200-2000 hrs LT. The peak of stratiform precipitating clouds has a smaller value and that too after the convective clouds. The time delay between the peak of the stratiform and convective precipitating clouds corresponds to the life cycle of the MCS. The occurrence of the diurnal cycle of the precipitating cloud systems over Palau is caused by diurnally convective boundary layer patterns established by the monsoon thermal circulations (as discussed in Chapter IV).

5.3.4 Seasonal variation of Precipitating Clouds

![Seasonal variation of precipitating clouds over Aimeliik.](image_url)

*Fig. 5.5: Seasonal variation of precipitating clouds over Aimeliik.*
Figure 5.5 shows the observation results of the precipitating cloud systems over Palau. From the observational results it is found that the occurrence of the stratiform and mixed precipitating cloud systems is more frequent during the westerly monsoon, while the occurrence of the convective precipitating cloud systems is predominant during easterly monsoon. Moreover, during the westerly monsoon, a higher occurrence of precipitating cloud systems was observed around Palau Islands. This is caused by a number of factors, which locally or regionally interfere with the general pattern. The main characteristics are: (i) differences in land/sea temperatures, (ii) location within the Inter Tropical Convergence Zone (ITCZ) region and (iii) intense convective storms. The most effective of these is convection, which can increase the amount of rainfall locally. Seasonal change (easterly or westerly monsoon) in the wind direction creates large differences in the precipitating clouds. These differences are often intensified by the characteristics of the air masses that are uplifted, where the two monsoon winds bring entirely different air masses. As a consequence, different precipitating clouds systems are developed with varying rainfall intensity.

5.4 Microphysical Contents

Figure 5.6 shows a schematic illustration for understanding of vertical structure of precipitating clouds and Rain drop size distribution (rain integral parameters) by using ground-based meteorological instrumentation and modeling studies of Melting Layer/Bright Band. Rain drop size distribution below the melting layer is estimated by using a new form of Gamma DSD model. This model employs two arbitrary moments as free DSD parameters to enhance the flexibility in studying the characteristics of Gamma DSD model to fit the Doppler radar spectrum. Stabilized Gauss-Newton method is used to obtain the solution to the non-linear least squared (NLLS) problem. The validity of a DSD model is evaluated in terms of the stability in solving NLLS problem and estimating the accuracy in DSD moment. The algorithm is validated using 1290-MHz wind profiler radar data collected during stratiform precipitation.

In the melting layer region, Non-coalescence-Non-break up model has been employed to estimate radar quantities have been compared with those measured by vertically-pointing Wind profiler radar.
5.5 Rain Drop Size Distribution Methodology

The character of precipitation detected at the surface is the final product of many microphysical interactions in the cloud above, the combined effects of which may be characterized by the observed drop size distribution (DSD). This necessitates accurate retrieval of the DSD from remote sensing data, especially wind profiler radar as it offers high vertical resolution, rigorous quality control and testing. Wind Profiler radar...
Doppler velocity spectra are used to estimate the DSD aloft and Ground truth is provided by the disdrometer.

L-band wind profiler radar detects both Bragg scattering from the radio refractive index of turbulence and Rayleigh scattering from hydrometeors. Bimodal Doppler spectra showing clear-air and precipitation echoes as illustrated in Figure 5.7 are often observed during moderate to light precipitation. The ability to measure both the clear-air velocities and the precipitation fall speeds simultaneously represents a large step forward for precipitation research.

**Fig. 5.7:** Radar Reflectivity dBZ/(ms\(^{-1}\)) observed on 23 Oct. 2005 [Upward motions are to the left of the zero value].

### 5.5.1 Gamma DSD model and its parameter estimation

In this thesis, DSD models are considered which are to be used for parametric estimation of DSD, employing a non-linear least-squares (NLLS) fitting procedure. Gamma model given by the following eq. (5.1) has been used most often (Ulbrich, 1983).

\[
N(D) = N_0 D^\mu \exp(-\Lambda D) \tag{5.1}
\]

where \(N(D)\) is the number concentration of particles which diameters range between \(D\) and \(D+dD\). \(N_0\) (m\(^{-3}\) mm\(^{-1}\)), \(\mu\), and \(\Lambda\) (mm\(^{-1}\)) are the intercept, shape, and slope parameters of the size distribution, respectively. Let \(v(D)\) be the terminal velocity of a drop, \(C\) be the radar system constant.

Then Doppler spectrum of precipitation \(S_p(v)\) is given by
S_p(v) = CN(D) D^6 | dv(D)/dD|^{-1} \quad \ldots (5.2)

Let \( \rho \) and \( \rho_o \) be the atmospheric densities at ground and aloft, then \( v(D) \) can be expressed as (Gunn & Kinzer, 1949)

\[
V(D) = -[9.65 - 10.3 \exp(-0.6 D)] (\rho/\rho_o)^{0.04} \quad \ldots (5.3)
\]

where the unit of \( D \) is mm.

Atmospheric echo spectrum \( S_t(v) \) is assumed to be Gaussian (Sato & Woodman, 1982). Therefore,

\[
S_t(v) = P_o \exp[(v - w)^2 / 2 \sigma^2] \quad \ldots (5.4)
\]

Where \( P_o, w, \sigma \) are the power at the spectrum peak the mean and standard deviation of Doppler spectrum. The normalized atmospheric echo spectrum \( S_o(v) \)

\[
S_o(v) = \frac{1}{\sqrt{2\pi\sigma}} \exp[(v - w)^2 / 2 \sigma^2] \quad \ldots (5.5)
\]

Let \( W(v) \) be the window function in FFT, \( P_n \) be the noise power density, and “*” be the operator for the convolution, the model function of wind profiler precipitation spectrum is

\[
S(v) = [S_t(v) + S_p(v) * S_o(v) + P_n] \quad \ldots (5.6)
\]

Equation (5.6) is an observation equation containing \( P_o, w, \sigma, P_n, N_o, \Lambda, \mu \) as 7 unknown parameters which can be estimated from observed spectra. \( \mu \) can be treated as a fixed parameter without significant degradation in the estimation accuracy of rain parameters.

Due to finite data length and fast Fourier transform (FFT) analyses the Doppler spectra are distorted by the window function \( W(v) \) as expressed as

\[
S'(v) = S(v) * W(v) \quad \ldots (5.7)
\]

Where \( W(v) \) is an inverse Fourier transform of a triangular auto-correlation function of the rectangular window. The theoretically estimated \( S'(v) \) contain information about DSD of precipitation and atmospheric turbulence. In order to deduce information from the observed spectra a spectral fitting method was devised by Wakasugi et. al., 1986, which is to find a parameter set that minimizes the difference to \( \varepsilon \) defined by

\[
\varepsilon = \sum_{i=1}^{N_{FFT}} [S_{obs}(vi) - S'(vi)]^2 \quad \ldots (5.8)
\]
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Where \( v_i \)'s are discrete velocity points corresponding to \( i^{th} \) discrete frequency of FFT and \( N_{\text{FFT}} \) is the number of fitted points in the periodogram.

The parameters to be obtained by the fitting are \( P_0, w, \sigma, N_0, \Lambda, V_{\text{max}} \) and \( P_n \). The method works very well when a good initial guess for the parameters is provided. For large volume of data giving this initial guess manually makes it difficult to obtain the parameters. This is overcome by providing initial guess automatically by the approach adopted by Sato et. al., (1990) using non-linear least square fitting. The initial values for \( P_0, w, \sigma \) are estimated from the observed turbulent (clear air) Doppler spectra by taking the lower three order moments and mean noise level \( P_n \). It is experimentally proved that the maximum diameter attainable by raindrop is around 6 mm and growing more than this size will break the drops into smaller ones. Using Eq.(5.3) the maximum fall velocity attainable for largest drop is around 9.6 ms\(^{-1}\), which is used as \( V_{\text{max}} \) in the fitting.

**5.5.2 Moment expression of the Gamma DSD model**

It is not necessary to use \((N_o, \Lambda, \mu)\) as the parameter set. In general, it may be important to choose a parameter set in which the parameter elements are independent as much as possible. It should be noted that the precipitation Doppler spectrum is weighted with \( D^6 |d\nu(D)/dD|^{-1} \) and small raindrops would not contribute much to form the spectrum shape. Therefore, DSD parameters that affect much to the DSD shape at larger raingdrop regions would be effective to improve the fitting performance. If we use \((N_o, \Lambda, \mu)\), then \( \Lambda \) is relatively sensitive to the large drop region and \( N_o \) to the small drop region. In order to study the relation between the choice of DSD parameters and the performance of the least-squares fitting, we propose a different parameterization that treats the moments of DSD as a DSD parameter. The \( x^{th} \) moment, \( M_x \), is expressed as

\[
M_x = N_o \Gamma(\mu + x + 1) / \Lambda^{(\mu + x + 1)} \quad \ldots (5.9)
\]

From Eq. (5.9), by fixing \( \mu \) and using \( M_x \) and \( M_y \), DSD parameters \((N_o, \Lambda)\) can be expressed by \((M_x, M_y)\). There are several ways to express DSD using the moments. In this study, we keep \( \Lambda \) and use \( M_y \) as a parameter related to the magnitude of DSD. The parameter \( M_x \) is used to obtain \( \Lambda \) by combing this with \( M_y \). Actual DSD expression is made by using the normalized moments, \( m_x \) and \( m_y \), which are given as
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\[ m_x = M_x / \Gamma(\mu + x + 1), \quad m_y = M_y / \Gamma(\mu + y + 1) \]  \hspace{1cm} \ldots (5.10)

\[ N(D) = M_y \Lambda^{(\mu + y + 1)} D^\mu \exp(-\Lambda D) \]  \hspace{1cm} \ldots (5.11)

Where \( \Lambda \) is given by

\[ \Lambda = (m_x/m_y)^{1/(y-x)} \]  \hspace{1cm} \ldots (5.12)

Using Eqs. 5.9 and 5.10 and the observed spectrum an initial value for \((N_0, \Lambda)\) can be obtained. Once the fitting is done, the estimated parameters can be made use to derive values of all other moments using Equation 5.10.

5.6 Estimation of vertical profiles of Rain integral parameters

\[ \text{Fig. 5.8. Time-height section of radar reflectivity (extended to the melting layer) on 08 May 2005.} \]

A pronounced bright band signature was observed by the WPR from 1700 to 2240 hrs LT (with an intermittent gap from 2100 to 2150 hrs LT) as depicted in Figure 5.8. Based on the characteristics in vertical structures and bright band intensities with time, two representative times at 2038 and 2206 hrs LT (hereafter referred as to T1 and T2) were selected to examine the DSD characteristics of the observed Doppler spectra.

Figure 5.9 displays the Doppler spectra of vertical beam observed at 1900 LT on 08 May 2005. The vertical resolution is 200 m. From this figure one can notice that below about 5 km two remarkable spectral peaks can be discriminated clearly. One spectral peak corresponds for precipitation echoes and the other one is responsible for refractivity returns (turbulence echoes). For weak to moderate rainfall, wind profiler radar was employed to obtain the double peak in Doppler spectra caused by the
precipitation and turbulence echoes, allowing the estimation of spectral/rain integral parameters.

**Fig. 5.9**: Vertical variations of spectral power density (dBZ/m s^{-1}) observed on 08 May 2005 at 19:10 hrs LT [Upward motions are to the left of the zero value].

**Fig. 5.10**: Estimated Raindrop concentration Vs drop diameter at different heights of Fig. 5.9.
5.6.1 Comparison between DSDs derived from WPR and MRR

The results in Figure 5.10 are verified by using Micro Rain Radar. The WPR and MRR are located 7 m apart in an open ground and can be said to be exposed to the same rain conditions. The instruments are operated based on different physical principle, thus providing independent measurement of the same rain event. A comparison of Rain drop concentration measurements taken by WPR and that from MRR the 1.5 km level obtained from MRR show. From the comparison of DSD spectra found that the values are within the limit of the stratiform precipitation. The results may be useful for understanding rain structures over Ocean region.

Fig. 5.11: 10-min averaged N(D) verse D derived from WPR spectrum and measured by the MRR (at altitude of 1.5 km) [To estimate DSD parameters gamma model is used when (a) \( \mu = 0 \) on 08 May 2005 from 1900 hrs to 2000 hrst, LT as indicated horizontal line in Fig.5.8.]
5.6.2 Vertical variations of observed Rain Drop Size Distributions

![Graph A](image1.png)

![Graph B](image2.png)

**Fig. 5.12:** 10-min integrated rain drop size distribution measured by the WPR (at altitude of 2.1 km) and Disdrometer. [To estimate DSD parameters gamma model is used when (a) $\mu = 0$ and (b) $\mu = 6$] on 08 May 2005.

Figure 5.12(a) shows the drop concentration $N(D)$ [m$^{-3}$ mm$^{-1}$] derived from the wind profiler radar measurements (using gamma model with $\mu = 0$), disdrometer and also from Marshall-Palmer (M-P) distribution. The shape of $N(D)$ derived from radar observations, particularly, at 2.1 km, is very closely followed as M-P distribution and
also has reasonably good agreement with disdrometer estimated $N(D)$ on the ground. To validate the present model, a relative comparison has been made with the existing lognormal Indian climate model. Wind profiler radar derived DSD parameters. Figure 5.12 [(b) with $\mu = 6$] agree closely with lognormal model as compared to M-P distribution. The present results suggest that gamma model is also useful for the estimation of DSD parameters in the tropical Oceanic, especially, from wind profiler radar observations.

Figure 5.13 shows vertical variations of DSDs derived from the observed profiler spectra at T1 and T2. It is shown in Figure 5.13b that the retrieved mass-weighted mean diameter (Dm) values were small (less than 1 mm) over a depth of ~1 km below the melting layer at T1, suggesting that there are a large number of small drops. This may be attributed to melting of small rimed particles resulted from accretion of ice particles and super-cooled liquid drops within updrafts. Huggel et. al., (1996) also noted that radar bright band is less well-defined in precipitation with heavily rimed particles than precipitation with dominant aggregation. Thus, we speculated that rimed particles associated with relatively strong updrafts played a role in making the bright band less prominent at T2 in Figure. 5.8a. In Figure 5.13b, below ~3 km msl, the Dm values started to increase considerably with decreasing height probably by collision-coalescence process.

Fig. 5.13: Vertical variations of raindrop size distributions at (b) 2038 LT and (c) 2206 LT that were derived from the WPR Doppler spectra (In figure 4.8 represented as T1 and T2).
5.6.3 Properties of rainfall parameters and correlations with regard to vertical air motion

Figure 5.14 shows time-height sections of Z, R, Dm, and w from 1 to 4 km msl below the melting layer. R and Dm values were those retrieved from the profiler Doppler spectra by using the SAM model and w values were determined from $w = W_{\text{dop}} - V_f$ where $V_f = 3.5Z^{0.084}$ (Atlas et al., 1973). It is shown that R is overall positively correlated with Z and this relation tends to be more obvious in the regions of the updrafts particularly during the 2230-2300 UTC and the 0040-0130 UTC periods in Figure 4d. In Figures 4b and 4c, R tends to be negatively correlated with Dm in the regions of the downdrafts during the 2340-0020 UTC period. The large Dm and small R values during this period could be overestimated and underestimated, respectively, within the downdrafts below the strong bright band since R is underestimated within downdrafts as downdrafts shift a hydrometeor spectrum to a large drop range (Rajopadhyaya et al., 1999).

Since the retrieved rainfall parameters showed close relations with w, we divided w into three categories of updrafts ($w > 0.5 \text{ m s}^{-1}$), neutral ($-0.5 \text{ m s}^{-1} < w < +0.5 \text{ m s}^{-1}$), and downdrafts ($w < -0.5 \text{ m s}^{-1}$) in order to examine physical relations between the rainfall parameters in each w category. In particular, the neutral w category was additionally made to minimize misclassified w points into the updrafts or downdrafts due to w biases related to the Z calibration and the choice of a Z-Vf relation in this study. Regarding Z-Vf relations, there are several empirical Z-Vf relations for raindrops documented in previous literatures. In this study, $V_f = 3.5Z^{0.084}$ (Atlas et al., 1973) was used because the w values calculated from this Z-Vf relation showed better agreement with those retrieved from the WPR Doppler spectra by using our algorithm.

The fitting algorithm and deducing the precipitation parameters fails when the clear air and precipitation echoes are merge together. Due to this reason it is difficult to find DSD distribution around melting layer and to understand physical phenomena such as generation, aggregation and successive process associated with precipitation. However it helps to classify the events distinctly and derive the DSD below the melting layer.

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Fig. 5.14: Time-height sections of (a) reflectivity (dBZ), (b) rain rate (mm/hr), (c) mass-weighted mean diameter (mm), and (d) vertical air motion (m/s) from 1.0 to 4.0 km msl below the melting layer. The gray-shaded area in (b) indicates rainfall rates greater than 10 mm/hr.