Meteorology has ever been an apple of contention, as if the violent commotions of the atmosphere induced a sympathetic effect on the minds of those who have attempted to study them

…Joseph Henry

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Chapter II
Meteorological Instruments, Theory and Data Processing Techniques
Radar remote sensing techniques used for probing the earth’s atmosphere are discussed in brief. The scattering mechanisms such as Bragg scattering, Fresnel scattering/reflection and Rayleigh scattering relevant to the radar remote sensing of the atmosphere and specific to the present study are outlined. A brief description of Wind Profiler Radar (WRP), Ceilometer, Joss Waldvogel Disdrometer (JWD), Micro Rain Radar (MRR), Automatic Weather Station (AWS), Tropical Rainfall Measuring Mission (TRMM), Radiosonde and Ordinary Rain Gauge (ORG), which are used, for this study are also described briefly.

2.1 Introduction

To study the lower atmosphere, several wind profiler radars have been installed in various places around the world in the last two decades (Rogers et. al., 1993; May and Wilczak, 1993; Angeline et. al., 1994 and 1998; Hashiguchi et. al., 1995b; Reddy et. al., 2001; Yang et. al., 2005), following the first successful operation of 915-MHz wind profilers specially designed for the boundary layer (Ecklund et. al., 1988). It is widely recognized that the dynamics of the Marine Boundary Layer (MBL) during easterly and westerly monsoon are very important to understand the short-range forecasts and meteorological now-casting, including the prediction of boundary-layer evolution and pollution dispersion in the East Asia. In addition, many aspects of the structure and dynamics of the MBL observed during monsoon season have not yet been clarified. The MBL is the interface between surface and free atmosphere, and it strongly influences vapor transport, sensible heat flux over the surface, and convection precipitating cloud systems. Clarifying MBL development mechanisms on diurnal and seasonal time scales will help to elucidate regional hydrological budgets.

The advantage of the wind profiler Radar (WPR) is to measure directly the vertical wind component within a convective environment (Gage et. al., 1994). Several researchers (Gage et. al., 1994, 1996; Atlas et. al., 1999; Rao et. al., 2001; Renggono, et. al., 2001; Reddy et. al., 2002; Atlas and Williams, 2003) have used
WPR to reveal details about the vertical structure of precipitating cloud systems. 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 during monsoon period over Palau in the western north Pacific Ocean. Lower atmospheric profilers operating at several frequencies are now used routinely as research tools to profile atmospheric winds in field campaigns and to monitor winds and vertical structure of precipitating cloud systems in support of operational missions (Gage et al., 2002). Radar has increasingly been used to observe horizontal winds and convective systems during Asia monsoon. This chapter will focus on pulsed Doppler radar widely known as the wind profiler Radar, Joss-Waldvogel Disdrometer, Micro Rain Radar, Radiosonde and Automatic Weather Station.

2.2 History of radar remote sensing

Atlas (1990) gives a good review of the history of radar use in meteorology. Radar was originally developed to detect and determine the range of aircraft. Observations of the atmosphere using radio waves can be traced back to 1924 where reflections from the ionosphere were first detected. The invention of the resonant magnetron in 1940 gave the ability to transmit power at centimeter wavelengths. Before the magnetron radar, generally operated at frequencies of around 200 MHz (wavelength ~1.5 m). Shorter wavelengths allowed the detection of precipitation. Originally, echo’s from precipitation were considered clutter, as they masked targets such as aircraft. The ability to observe precipitation and the movement of storm systems was, however, the beginning of radar meteorology. Radar meteorology had its greatest advance during World War II. The US military recognized the importance of weather radar and began a significant development programme at the Fort Mammoth labs in 1945. The early weather radar systems were adapted from military radar systems. Watson and Watt developed the first 10 cm weather radar in 1940 at the General Electric Labs. A convective storm was tracked in 1941 using a 10 cm wavelength radar in England, and multicellular storms were first looked, when they studied vertical wind shear and showed that a new cell grew up-shear of an old one. Most early radars were continuous-wave
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(CW), relying on the interference between the direct signal received from the transmitter and the Doppler-shifted signal from the moving target for detection. The CW radar determined the velocity of a target using its range and angle in successive illuminations. A pulsed Doppler radar is essentially continuous wave radar where the wave is also amplitude modulated. It allows the measurement of radial velocity through the determination of a Doppler spectrum. Pulses are used to remove range ambiguity and the principles of pulsed Doppler radars are discussed in the next section.

2.3 Wind Profiler Radar (WPR)

Meteorological radar originally relied on the presence of precipitation for useful observations. The observation of clear air echo’s was a major development in radar meteorology, giving the ability to observe atmospheric properties such as wind speed in the absence of precipitation. The ability to make observations in clear air also gave rise to the development of wind profiling radars. From 1935 to 1949 several types of clear air echos were observed and reported. Friend (1949) showed that the reflections observed by the ionospheric radars in the troposphere were regions of temperature inversion with large gradients of refractive index. Also the first observations of insects and birds made which were previously unknown causes of “Angle Echos” on radar screens. Work on propagation beyond the radio horizon (over the horizon) by scattering of radio waves by turbulent tropospheric structures marked the second period. This work was important as it is also a factor in the radar back scattered energy. In this period, the importance of isotropic scattering and reflection were discussed. For vertically pointing radar beams, echos from horizontal structures with large refractivity gradients would be important. However, for off vertical beams, isotropic scattering is important.

The third period from 1963 to 1972 showed increased activity in observing atmospheric structures. The first radar observations of convective processes in the boundary layer were made, and evidence that clear air echos from insects and birds occurred at wavelengths of 10 cm or less was found during this period. At longer wavelengths, reflection occurs from refractive index gradients rather than from particles. Radars operating at wavelengths of 3 cm or less predominantly
observe clear air echos from insects and birds. At wavelengths greater than 10 cm
the echos result from strong refractive index gradients. Konrad and Kropfli
observed clear air convection over the sea and reported that the radar echo pattern
moves with the mean wind in the convective layer.

From 1973 to 1982, there was increasing use of VHF wavelength radars
which were constructed mainly as large phased array antennas that operated
unattended (e.g., Woodman and Guilled 1974). The Sunset radar operated in
Boulder, Colorado was the first VHF radar built for observation of the lower
atmosphere. It was operated at a frequency of 40 MHz, with the first results
published by Green et. al., (1975). By the 1980’s several VHF and UHF radars
designed specifically for the troposphere (wind profilers) were being designed and
deployed as networks (Strauch et. al., 1984). The NOAA Aeronomy laboratories
were pioneering this work (Gage and Balsley 1978). The UHF and VHF radars
have shown capabilities for measuring, horizontal and vertical velocities,
turbulence, momentum flux and precipitation (Gage 1990). The wind profilers
have fixed beam directions which can be switched rapidly. By using three beam
directions, one vertical and two inclined from the vertical orthogonal to each
other, profiles of the horizontal wind can be determined. The Colorado wind
profiler network, described by Strauch et. al., (1984), was a first step towards
making the transition from the research to the operational sector. The first wind
profiler observations routinely incorporated in analysis were from Christmas
Island in the central Pacific Ocean. These wind profilers were developed for
probing the free Mesosphere, Stratosphere and Troposphere and hence were called
MST radar. They have been been used for investigating measurements included
momentum fluxes, phase velocities and modes and energy associated with internal
gravity waves (e.g., Clifford et. al., 1994; Larsen and Röttger 1982). Work on the
tropospheric wind profilers led to the development of the boundary layer wind
profiler (Ecklund et. al., 1988). The boundary layer wind profiler has become an
important tool for the measurement of atmospheric boundary layer parameters.
Measurements can be made at high temporal resolution and with greater height
coverage than meteorological towers or clear air measurements with Doppler
weather radar. In the mid 1980’s a 915 MHz wind profiler was developed by NOAA Aeronomy Laboratories specifically for measurements within the ABL (Ecklund et. al., 1988). The boundary layer wind profiler (BLWP) was designed to complement the existing 50 MHz wind profiler network in the tropics by providing high resolution wind measurements in the boundary layer (Ecklund et. al., 1988). It was shown by Basely and Gage (1982) that to obtain a height resolution of 100 m or better, with fast system recovery (providing measurements at least as low as 100 m), a frequency near 1000 MHz is required. The original wind profiler antenna developed by Ecklund et. al.,(1988) was mechanically steered to four beam positions. The newer antennas are electronically steered, using phase-shifted antenna elements.

Boundary layer wind profilers have been developed to operate at various frequencies including 1357.5 MHz radar that has been in operation since 1992 (Hashiguchi et. al.,1995; Reddy et. al., 2002). Doppler radar studies of the Atmospheric boundary layer (ABL) have included observation of momentum fluxes, wind velocities and temperature profiles associated with ABL development (e.g., May and Wilczak 1993; Angevine 1994) the sea breeze (e.g., Banta et. al., 1998) the nocturnal jet and cold fronts (e.g., Shapiro et. al., 1984; May et. al., 1990a).

2.4 Operating principles for pulsed Doppler radar

With a pulsed Doppler radar, a high power amplifier is turned on and off by a pulse modulator to transmit a train of pulses which have a set duration $t$ and a pulse repetition time (PRT) of $T_s$ and angular frequency $\omega$. Part of the pulse $\omega e^w$ is scattered back towards the radar by fluctuations in the refractive index. If the transmitter and receiver antennas are co-located, a transmit/receive switch modulates between a transmit time and receive time. When the signal is received, it is mixed to base band (i.e., frequency about DC) with known relative phase (Doviak and Zrnic, 1993). A quadrature detector converts the signal to an in phase component and a component 90 degrees (quadrature) out of phase.

The phase angle of the returned signal can be determined, and the time evolution shows whether the radial velocity is towards or away from the radar.
The phase change between pulses resulting from a moving target is extremely small. As a result, pulses that are close to each other in time (a period over which phase change is assumed negligible) can be averaged to increase the signal by reducing the random noise. A block diagram of a typical wind profiler is shown in Figure 2.1. The antenna and VHF/UHF transmit/receive unit are located in the field. A computer located in the field laboratory controls the radar operating parameters, processes the returned signal, and displays and stores the resulting data.

A wind profiler antenna usually consists of a number of antenna elements in an array. In the case of Doppler weather radar this is usually replaced by a parabolic dish. The wind profiler antenna designed by Ecklund et. al., (1988) consisted of 16 radiating elements. As coherent waves are transmitted from each of these elements, the waves will constructively or destructively interfere with each other. As a result, the power transmitted will depend on the angle from the antenna bore sight, with most of the power being transmitted in the direction of constructive interference. If the antenna elements are appropriately spaced, and the phases of the transmitted waves from these elements are varied with respect to each, then the beam can be focused in a particular direction. If we assume a linear array of elements, the superposition of waves shows that a $\sin(\theta)/\theta$ power distribution is generated (Figure 2.2).

The beam pattern shows a maximum gain in the main direction. However side lobes will also occur and in radar design it is beneficial to minimize these side lobes. Side lobes of the transmitted beam are reflected from ground targets including towers, topography and forests leading to an unwanted component in the returned signal.
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Fig. 2.1: Block diagram of a wind profiler radar.

Fig. 2.2: Beam geometry of a linear array of radiating elements forming a sinθ/θ pattern. Two alternate display methods are shown: Amplitude on a polar plot (left), and using decibels on a Cartesian plot (right).
The maximum observable height is limited by the power of the radar, as the power of the return pulse is reduced with range. It is known that for volume scattering, the return power is inversely proportional to the square of the range. The return signal (S) strength is related to the background noise level (N) through the SNR (Signal to Noise Ratio). The SNR expressed in decibels (dB) may be calculated as

$$\text{SNR} = 10 \log_{10} \left( \frac{S}{N} \right)$$ \hspace{1cm} \text{..... 2.1}

The noise component is spread over a greater range of frequencies than the signal and hence the signal may be detectable even at SNR down to approximately -10 dB.

The backscatter of radio waves depends on atmospheric refractive index variations which are caused by variations of air temperature and moisture. The potential refractivity gradient (M) can be used to model the radar return based on the moisture and temperature contributions to the refractive index and is given by

$$M = -77.6 \times 10^{-6} \frac{P}{T} \left( \frac{\delta \ln \theta}{\delta z} \right) \left[ 1 + \frac{15550q}{T} \left( 1 - \frac{1}{2} \frac{\delta \ln q}{\delta z} \right) \right]$$ \hspace{1cm} \text{..... 2.2}

Where $P$ is the pressure (hPa), $T$ is the absolute temperature (K), $\theta$ is the absolute potential temperature (K), $q$ is the specific humidity, and $z$ is the height (m).

The echo power $P_v$ is given by

$$P_v = \frac{P_t \lambda^2}{16\pi^2 r^2} G^2 C E(2k) M^2$$ \hspace{1cm} \text{..... 2.3}

where $P_t$ is the transmitted power, $\lambda$ is the radar wavelength, $r$ is the range from which the signal is backscattered, $G$ is the antenna gain, $C$ is a constant.
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depending on the radar wavelength and height resolution and $E(2k)$ is the energy density of fluctuations with a vertical scale of half the radar wavelength (Tsuda et al., 1988).

Wind profilers often show horizontal bands of enhanced reflectivity within the atmosphere. These can be a result of refractive index variations associated with the vertical profiles of temperature and moisture. The top of the boundary layer is often marked by an increase in temperature and decrease in moisture. The equation for potential refractivity gradient suggests that the wind profiler should be able to observe the top of the boundary layer as a region of enhanced signal return. Enhanced signal return also occurs within clouds as a result of turbulent mixing.

2.5 A Brief Description of L-band Wind profiler at PALAU

Palau wind profiler radar (WPR) is a LAP-3000 built by Vaisala Corporation (formerly Radian Corporation) in Boulder, Colorado, with post processing software from Sonoma Technology Inc. The design is the commercialized version of the systems designed in the NOAA Aeronomy laboratory. It operates at a radio frequency of 1290 MHz.

The WPR has an electronically steered phased array antenna capable of producing five beams. Nominally the five beam directions are north, south, east, west, and vertical. The off-vertical beams are at an elevation of 74.5 degrees (15.5 degrees down from vertical). The effective antenna area of 6.8 m$^2$ (9 x 0.87 m x 0.87 m) by symmetric arrangement of 3 x 3 antenna panels, thus using a pulse peak power of 800 W with a peak power-aperture product of 340 Wm$^2$. The WPR antenna is enclosed in a clutter screen and covered by a radome. Figure 2.3 shows the wind profiler antenna with 4-Radio acoustic sounding system (RASS) speakers around it. The transmitter is capable of producing pulses of four lengths: 400, 700, 1400 and 2800 ns. These correspond to vertical resolutions of 60, 105, 210 and 420 meters. The inter-pulse period (pulse repetition frequency) is fully controllable, so the maximum range is limited only by the strength of the returned signals (depends on the background meteorological conditions). Sampling of the returned signal (i.e. the range gates) can be done at intervals that are multiples of
the pulse lengths. The transmitting/receiving unit is installed below the wind profiler antenna and converts the 1290 MHz signal; both transmitted and received signals to an intermediate frequency of 60 MHz. The radar processor calculates the spectra, moments, wind and temperature data and stores in the radar computer. For detailed WPR-3000 description, refer to Ecklund et al., (1988) and Carter et al., (1995).

The WPR computer (with external hard disk) offers the possibility to save the raw data (i.e., averaged Doppler spectra) as well as the moment data computed with the single peak algorithm and Intermittent Clutter Reduction Algorithm (ICRA) (Merritt, 1995). The WPR was always operated in a 3-beam set-up and data were collected in low and high mode alternatively [Table 2.1]. The winds are derived from a consensus average of the radial velocities over a chosen averaging interval (typically 30-60 min). The consensus window width was set to 3 m/s for the oblique beam and 2 m/s for the vertical beam. The threshold percentage of data points within the window needed to compute a valid consensus average was set to 60%. An average signal-to-noise ratio (SNR) is associated with each consensus average by averaging the SNR for all the records that were accepted for the radial velocity consensus average. After the consensus-averaged radial velocities (corrected for vertical wind component in the off-vertical beam radial velocities) are computed for each radar beam position (Carter et al., 1995). From these corrected and combined radial velocity values, the wind components u, v and w (or speed and direction) are computed. The 30-min or 1-hr u, v and w data were quality controlled with a continuity quality control algorithm. This was performed off-line and data of one single day were processed as one continuous block. The parameters for the continuity quality control were set to values which (subjectively judged) performed satisfactorily. These signal-processing methods work well for most of the time; however, there are limitations due to unrealistic wind data (Lambert et al., 2003).
Fig 2.3: (a) Wind Profiler Radar (WPR) installed at Aimeliik state of Babeldaob Island (7.45° N, 134.47° E) of Republic of Palau in the North-Western Pacific Ocean.
(b) 5-beam configuration of WPR.
During RASS mode, the acoustic signal was transmitted continuously with its frequency changed for every 15 ms and selected randomly from the interval of 2025-2130 Hz, corresponding to a temperature interval of 5 - 35 °C. The fundamental vertical resolution, as determined by the radar pulse duration, was 105 m, though the RASS signals were sampled at an interval of 0.4 μs to give spacing of 60 m. The time resolution, determined by the amount of coherent and spectral integration, was 22 sec.

Typical RASS operation parameters are listed in Table 2.1. Generally, the speed of sound and the vertical velocity are measured simultaneously by a power spectral analysis using 2048 complex time series points using with a Nyquist velocity of approximately 350 m/s. Thus, the clear-air signals near zero velocity as well as the RASS signal near the speed of sound (330 m/s) are measured simultaneously. Hence, a correction of the influence of vertical wind on the measured sound velocity is possible. This simultaneous radio-acoustic sounding system usually estimates virtual air temperature (T_v) profiles with lower height coverage than simple wind measurements owing to the smaller height coverage of the acoustic emission. RASS and wind height coverage depend on the atmospheric conditions.
### Table 2.1: Typical Wind Profiler Radar (WPR) operating parameters:

<table>
<thead>
<tr>
<th>WPR parameters</th>
<th>Low mode</th>
<th>High mode</th>
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</thead>
<tbody>
<tr>
<td>Frequency (Wavelength)</td>
<td>58</td>
<td>202</td>
</tr>
<tr>
<td>Peak power</td>
<td>70</td>
<td>55</td>
</tr>
<tr>
<td>Antenna aperture (m)</td>
<td>4.2</td>
<td>12</td>
</tr>
<tr>
<td>Max. Radial velocity (m/s)</td>
<td>20</td>
<td>20</td>
</tr>
<tr>
<td>Spectral points</td>
<td>256</td>
<td>256</td>
</tr>
<tr>
<td>Spectral averages</td>
<td>60</td>
<td>60</td>
</tr>
<tr>
<td>Dwell time (sec)</td>
<td>-30</td>
<td>-30</td>
</tr>
<tr>
<td>Number of beams</td>
<td>Vertical, north &amp; east with 15° Zenith angle</td>
<td></td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>RASS parameters</th>
<th></th>
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</tr>
</thead>
<tbody>
<tr>
<td>Frequency</td>
<td>2 – 3 KHz</td>
<td></td>
</tr>
<tr>
<td>Acoustic power</td>
<td>60 W</td>
<td></td>
</tr>
<tr>
<td>Antenna type</td>
<td>Parabolic reflector</td>
<td></td>
</tr>
<tr>
<td>Effective antenna</td>
<td>1.8 m²</td>
<td></td>
</tr>
<tr>
<td>Beam width</td>
<td>10°</td>
<td></td>
</tr>
<tr>
<td>Direction</td>
<td>Zenith</td>
<td></td>
</tr>
<tr>
<td>Number of spectral points</td>
<td>2048</td>
<td></td>
</tr>
<tr>
<td>Sample spacing</td>
<td>60 m</td>
<td></td>
</tr>
<tr>
<td>Averaging time</td>
<td>3 minutes (adj.)</td>
<td></td>
</tr>
</tbody>
</table>

### Other parameters

- Data recording: WPR/RASS computer hard disk
- Electric power consumption: 4 kw
- Electric power voltage: AC 220 V
- Recording media: External Hard disk connected to the Radar Computer
- Data items recorded:
  1. System parameters (Binary data)
  2. Doppler spectrum (Binary data)
  3. Moments (Binary data)
  4. Wind and temperature (ASCII data)
  5. Wind and temperature (Microsoft Access data base)
2.6 Signal processing in pulsed-Doppler radar

The following descriptions are based on parameters used for the 1.290 GHz lower atmospheric (boundary layer) wind profiler [LAWP, wind profiler or wind profiler radar] at PALAU. The wind profiler consists of an antenna, a transmit/receive unit, a radar controller and signal processor and a recording device. For a typical measurement, a 0.7 µs pulse is transmitted by the radar with a peak power of 400 W. As the pulse travels through the atmosphere, it is scattered due to refractive index gradients and turbulence (Reddy et. al., 2001). Scattering occurs most strongly from turbulence on scales of half the radar wavelength ($\lambda/2 = 23.3$ cm for the 1.290 GHz profiler). This process is known as Bragg scatter. Since turbulent parcels are adverted by the wind, the wave scattered back towards the radar will have a Doppler shift related to the mean radial wind component. If the radial wind direction is towards the radar, there will be an increase in frequency (a positive Doppler shift). If the wind direction is away from the radar, there will be a decrease in frequency (a negative Doppler shift). The range ($r$) from which the pulse arrived is determined by the return time ($t$) as

$$r = \frac{ct}{2}$$

Where, $c$ is the velocity of light. The radar pulse undergoes back-scatter throughout its propagation. By matching receiving gates on the radar to the return time of the pulse at specified ranges, a single pulse can provide data for a range of heights (Figure 2.4). A series of pulses is transmitted (e.g., 54 pulses) and these pulses are averaged to produce an average complex signal (coherent integration). Averaging pulses increases the ability to detect the signal for low SNR by reducing the noise (random errors) variance (e.g., Woodman and Guillen 1974). A Doppler spectrum is then calculated using 128 averaged points.
Fig. 2.4: Range/time diagram. In this diagram, \( r_{\text{min}} \) is the minimum observable height which is equal to the pulse length \( \Delta r \) or equivalently in time \( \Delta t \). The maximum unambiguous range \( r_{\text{max}} \) is such that the pulse travel time \( t_1 \) is less than the inter pulse period (IPP). Hence \( T_1 \) is the first pulse \( T_2 \) is the time of the second pulse a period equal to the IPP later (Rottger, 1985).

A spectrum of Doppler shifts is observed due to variations in the radial velocity component of the advected turbulent parcels within a pulse volume. The pulse volume is determined by the pulse length (0.7 µs; which can be converted to a pulse length in meters by Equation 2.4) and the beam width (9 degrees). It has been found that this return Doppler spectrum is Gaussian in shape (Woodman 1985). Sixty successive power spectra are calculated and averaged to increase the statistical stability of the Doppler and power estimates.

The three parameters which define the Doppler power spectrum (\( S \)) are (Woodman, 1985):

1) The total power (\( P \)) (zeroth moment of the Doppler spectrum) given as the integral of the power spectrum amplitude for all Doppler frequencies \( \omega_d \) (area under the power spectrum):

\[
P = \int S(\omega_d) \, d\omega_d
\]

...... 2.5
2) The average frequency shift ($\Omega$) (first moment of the Doppler power spectrum) is given by the Doppler frequency component weighted by the spectral power at that frequency:

$$\Omega = \frac{1}{P} \int \omega_d S(\omega_d) d\omega_d$$  \hspace{1cm} ..... 2.6

3) The spectral width $\sigma_W$ (second moment of the Doppler spectrum) is given by the square of the frequency components deviation from the mean weighted by the spectral power at that frequency and is equivalent to the variance of the distribution:

$$\sigma_W^2 = \frac{1}{P} \int (\omega_d - \Omega)^2 S(\omega_d) d\omega_d$$ \hspace{1cm} ..... 2.7

If the return spectrum is Gaussian in shape, these three parameters contain all the information obtainable from the radar echo’s (Woodman 1985). The physical significance of the total spectral power, mean Doppler frequency and spectral width are depicted in Figure 2.5.
Fig.2.5: (a) Elements of the Doppler power spectrum. (b) Shown are the mean noise level, the Zeroth (Power), first (mean) and second (spectral width) moments of the power spectrum. The ratio of the signal ($S$; above the noise level, to the noise underneath the signal peak ($N$) is the signal to noise ratio.
The three parameters contain important information on the physical properties of the medium through which the radar pulse propagates. The total power gives information on the intensity of the turbulence, the mean frequency shift corresponds to the mean radial velocity and the spectral width refers to the velocity dispersion (Woodman 1985).

The pulse pair technique is also often used in the estimation of Doppler velocity, particularly for Doppler radar. This method estimates the first two moments of the Doppler spectrum from estimates of the auto-correlation function at one sample lag. The Doppler power spectrum is a Fourier transform pair with the auto-correlation function (Wiener-Khinchin Theorem). The auto-correlation at one sample lag can be estimated as

\[ R(T_s) = \frac{1}{M} \sum_{m=0}^{M-1} V^*(m) V(m+1) \]  

\[ \text{..... 2.8} \]

Where \( V \) is the signal at the \( m \) the sample each of which is spaced at \( T_s \), and \( M \) is the total number of samples used in the calculation (Doviak and Zrnic 1993). In the pulse pair technique, the mean change of phase over time is estimated by the argument of the auto correlation function at one lag and gives the mean Doppler shift

\[ v_r = -\frac{\lambda \omega_d}{4\pi} \]  

\[ \text{..... 2.9} \]

The complex auto correlation function has an amplitude and phase. This phase angle expressed as Arg \( R(T_s) \) is the mean change in phase between two successive samples in the sample interval \( T_s \). Ground clutter generally introduces a symmetrical spectral peak close to the zero Doppler frequency. This is only an issue for targets at a distance greater than the first range gate although nearby objects may contribute to beam blockage. The effect of ground clutter is to bias radial velocity estimates towards zero.
Many techniques have been applied to reduce the impact of ground clutter both before measurement and after measurement. Ground clutter contamination can be reduced by sitting the profiler in a favorable topographic environment or by using a clutter fence to block waves travelling along the surface towards the radar.

Post measurement clutter removal techniques have varied from simple n-point interpolation about the zero Doppler frequency (Strauch et. al., 1984), and half plane subtraction, and DC removal, to the application of neural networks and wavelet analysis on the time series (Jordan et. al., 1997). Intermittent clutter from airborne targets (e.g., birds or aircraft) and communications equipment also interferes with the returned signal. Techniques for reducing this interference have been developed by researchers including Merritt (1995) and Jordan et. al., (1997).

The radial velocity estimates can be used to derive horizontal and vertical velocity. The horizontal winds are calculated using the beam geometry and an assumption of statistical homogeneity between the beams (Figure 2.6).

![Fig. 2.6 Beam geometry. The Horizontal wind is determined by the vertical velocity (w) and the radial velocity [V(θ,R)] from a beam inclined θ from the vertical. The height of the observation is given by Z and the range by R.](image-url)
An example for a three beam system with a vertical beam and two beams inclined from the zenith by $\theta$ towards the east and north, the zonal velocity ($u$) is given by

$$u = \frac{V_E - w \cos\theta}{\sin\theta} \quad \text{..... 2.10}$$

where $V_E$ is the east beam radial velocity, $w$ is the vertical beam radial velocity.

The meridional component ($v$) can be derived by substituting the north beam radial velocity

$$v = \frac{V_N - w \cos\theta}{\sin\theta} \quad \text{..... 2.11}$$

These equations assume a spatially uniform wind field over the beam separation. Since the beam measurements are not taken simultaneously, a suitable averaging period is needed.

2.7. Ground-based Instrumentation at Palau

At Aimeliik observatory (7.45°N, 134.47°E) several ground based remote sensors [Wind profiler radar (WPR) with Radio acoustic sounding system (RASS), Ceilometers, Micro Rain Radar (MRR), JW Disdrometer (JWD), and Automatic Weather Station (AWS)] are operated fairly continuously. The instruments were placed within a few meters of each other; thus, to the extent possible their measurements describe the same atmospheric column. The operational status of the three remote sensors is summarized in Figure 2.7, allowing gaps of up to one minute to be considered within normal operation. Overall, the observations are fairly continuous with great overlap between the three remote sensors, spatially as well as temporally.
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Fig. 2.7 Ceilometer, JW Disdrometer, Automatic Weather Station with Tipping Rain Gauge installed nearby Wind Profiler Radar at Aimeliik, Palau.

2.7.1 Ceilometer

The Vaisala single-lens ceilometer CT25K measures the optical backscatter intensity of the air at a wavelength of 905 nm (near infrared) as shown in Figure 2.8. Its laser diodes are pulsed with a repetition rate of 5.57 kHz. The lens has a focal length of 377 mm and an effective diameter of 145 mm. Laser beam full divergence and field-of-view divergence of the receiver are 1.4 mrad each. Because of the monostatic optical system and the small divergence multiple scattering effects are negligible and the Mie scattering with scattering angles between 179.9° and 180.1° is dominant. Additional technical characteristics are given in Table 2.2.
The CT25K lidar ceilometer samples the return signal every 100 ns from 0 to 50 μs, providing a spatial resolution of 15 m from the ground up to an altitude of 7500 m. The backscatter intensity depends mainly on the particulate concentrations in the air. As the size of particles varies with their moisture content, the reflectivity is influenced by atmospheric humidity, too. Clouds, fog

---

**Fig. 2.8**: Vaisala CT25K laser ceilometer installed at Aimeliik Observatory.

**Table 2.2**: Technical properties of the CT25K ceilometer

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Specification</th>
</tr>
</thead>
<tbody>
<tr>
<td>Wavelength</td>
<td>905 nm</td>
</tr>
<tr>
<td>Laser</td>
<td>InGaAs MOCVS laser diode</td>
</tr>
<tr>
<td>Pulse Peak Power</td>
<td>16 w typical</td>
</tr>
<tr>
<td>Average Power</td>
<td>8.9 mw</td>
</tr>
<tr>
<td>Repetition rate</td>
<td>5.57 kHz</td>
</tr>
<tr>
<td>Pulse width</td>
<td>100 ns</td>
</tr>
<tr>
<td>Range</td>
<td>15 m</td>
</tr>
<tr>
<td>Measurement range</td>
<td>0 – 7500 m</td>
</tr>
</tbody>
</table>
and precipitation inhibit measurements. The performance of the CT25K ceilometer is sufficient for analysing boundary-layer structures. Compared to more sophisticated Lidar systems commonly used for these investigations it has several advantages, including the low first range gate, its ability to operate eye safe and maintenance-free for several years in any climatic environment with just some regular window cleaning, and its comparably low price. Main disadvantage due to the low emitted power is its relatively low maximum range, but for mixing layer studies (mostly below 3 km) this does not present a problem.

The sensitivity of the ceilometer on molecular or particles backscattering depends on the wavelength of the laser. Backscattering in the molecular range is almost negligible at 900 nm. Strong backscattering is caused by high aerosol concentration (clouds or Sea splash) using a laser with a wavelength of 900 nm. Below a radius of 1 µm the scattering efficiency decreases significantly. The Palau-ceilometer is able to detect lower cloud layers reliably. Transitions of aerosol concentrations give the opportunity to detect boundaries in the atmosphere, because the vertical distribution of particles is heavily influenced by the thermal structure of the atmosphere.

The backscatter signal intensity $P(r)$ of a near infrared lidar is a function of the instrument function $C(r)$, the backscatter coefficient $\beta$ and background signal $P_0$. This is also known as the lidar equation.

$$P(r) = \frac{C(r)}{r^2} \beta(r)T(r) + P_0$$

..... 2.12

where $r$ is the round-trip transmission factor. All parameters are wavelength dependent, except of distance $r$. Extinction $\alpha$ and volume back scattering $\beta$ coefficients are a combination of aerosol and molecular components.

2.7.2 JWD Disdrometer

Joss-Waldvogel Disdrometer (JWD) measures raindrops continuously and automatically and is considered as one of the standard instruments for the measurement of DSD. JWD records number and size of raindrops that span from 0.3 to 5.3 mm in 20-drop size classes. JWD consists of two units: a sensor, which
is, exposed to rain and a processor to process the signal coming from the sensor.

The pictures of disdrometer sensor (as shown in figure 2.7) and processor at Aimeliik, Palau, and block diagram and Photograph of the JWD are shown in Figure 2.9.

JWD Sensor consists of an electromechanical unit and a feedback amplifier housed commonly. When a raindrop hits the surface of a conical Styrofoam body (having a surface area of 50 cm$^2$), the two magnetic coils together with the Styrofoam body move downwards in a magnetic field and a voltage is induced in the sensing coil. This voltage is amplified using an amplifier. The output of the amplifier is a measure of the size of the drop that caused it. To improve the sensitivity of the Styrofoam body to next falling drop, the sensing coil voltage after amplification is applied to driving coil, which gives a counteract force to the Styrofoam body. The drop diameters ranging from 0.3 to 5.3 mm produce pulse amplitudes in the range of 0.3 mV - 10 V with a dynamic range of 90 dB.

![Figure 2.9](image)

**Fig. 2.9** (a) Block diagram and (b) photographs of the sensor of the Joss Waldvogel (RD80) Disdrometer installed at Palau.
Processor consists of power supply, signal processing circuit and test signal generator. The signal processing circuit consists of noise reduction filter, dynamic range compressor, signal recognition circuit and non-linear analog to digital converter. The noise reduction filter is an active band pass filter, whose frequency response is designed to give an optimum ratio between the signal from raindrops and the signal from acoustic noise affecting the sensor. The dynamic range compressor consists of an amplifier with a voltage dependent feedback network. It alters the amplitude response of the system to the desired characteristics. The signal recognition circuit distinguishes the signal pulses caused by drops hitting the sensor and the more uniform oscillations caused by acoustic noise. When the pulses caused by raindrops exceed the oscillations caused by acoustic noise, a gate passes the pulses to the analog to digital converter. It generates a 7-bit code at the RS-232 interface of the processor for every drop measured by the sensor. The measured diameters are initially divided into 127 classes and later sorted into 20 classes with varying diameter intervals to get statistically meaningful number of drops in each class interval. The test signal generator is used to check the performance of the system. If the sensor is not properly connected or if any part of the processor is not working properly, LED number 4 will not glow.

The accuracy of the measured drop diameter depends somewhat on the location of a drop impact on the sensitive surface of the Styrofoam cone. The drops of equal diameter hitting on different locations of the sensitive surface form a distribution of pulse amplitudes around the average amplitude. If the drops are evenly distributed over the sensitive surface, the standard deviation of the pulse amplitude distribution, which is transformed into drop diameters, is approximately ±5%. Hence the error in the measured raindrop diameter is ±5%. The specifications of the sensor and processor are given in Table 2.3.
Table 2.3: Specifications of the JWD Sensor and Processor

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Range of drop diameter</td>
<td>0.3 mm to 5.3 mm</td>
</tr>
<tr>
<td>Sampling area</td>
<td>50 cm$^2$</td>
</tr>
<tr>
<td>Error</td>
<td>$\pm 5%$ of measured drop diameter</td>
</tr>
<tr>
<td>Resolution</td>
<td>127 size classes distributed more or less exponentially over the range of drop diameters</td>
</tr>
<tr>
<td>Output format</td>
<td>According to RS-232-C standard, 7 data bits, even parity, 1 stop bit</td>
</tr>
</tbody>
</table>
| Relation between $D$ and amplitude of output pulse Ucompr. | $U_{\text{compr.}} = 0.94 \times D^{1.27}$  
(Ucompr. in Volts, $D$ in mm) |
| Operating temperature range    | 0 to 40°C for processor  
0 to 50°C for sensor                                                    |
| Standard length of sensor cable| 20 m                                                                 |

Rain integral parameters

JWD was originally designed for measurement of DSD, but utilized in many fields of meteorology and atmospheric sciences. The recorded number of rain drops in 20 drop size classes are utilized to estimate $N(D)$, in turn, the rain integral parameters. The $N(D_i)$ at channel number $i$ is estimated using the formula:

$$N(D_i) = \frac{10^4 n_i}{Ftv(D_i)\Delta D_i} \text{ (m}^3\text{mm}^{-1})$$  

.... 2.13

$$R = 6\pi \frac{1}{F_t} \sum_{i=1}^{20} (n_i D_i) \text{ (mm hr}^{-1})$$  

.... 2.14

(or)

$$R = 3.6 \times 10^3 \frac{\pi}{6} \sum_{i=1}^{20} [N(D_i)D_i^3 v(D_i)\Delta D_i] \text{ (mm hr}^{-1})$$  

.... 2.15

$$W = 10 \frac{\pi}{6} \rho \sum_{i=1}^{20} \left[ \frac{n_i}{v(D_i)} D_i \right] \text{ (g m}^{-2})$$  

.... 2.16

(or)

$$W = 10^{-3} \frac{\pi}{6} \rho \sum_{i=1}^{20} [N(D_i)D_i^3 \Delta D_i] \text{ (g m}^{-2})$$  

.... 2.17
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\[ Z = 10^4 \frac{1}{F \tau} \sum_{i=1}^{20} \left[ \frac{n_i}{v(D_i)} D_i^6 \right] \quad (\text{mm}^6 \text{m}^{-3}) \]  

\[ Z = \sum_{i=1}^{20} [N(D_i) D_i^6 \Delta D_i] \quad (\text{mm}^6 \text{m}^{-3}) \]  

\[ D_m = \frac{\sum_{i=1}^{20} [N(D_i) D_i^4 \Delta D_i]}{\sum_{i=1}^{20} [N(D_i) D_i^3 \Delta D_i]} \quad (\text{mm}) \]

Where number of drops measured in drop size class \( i \); \( D \) is the average diameter (mm) of the drops in class \( i \); \( F \) is the surface area (50 cm\(^2\)) of the Styrofoam body; \( \tau \) is the sampling time (60 s); \( v(D_i) \) is the fall velocity (m s\(^{-1}\)) of the drop with diameter \( D_i \); \( \Delta D_i \) is the diameter interval (mm) of the drop size class \( i \); and \( \rho \) is the water density (= 1 g cm\(^{-3}\)).

Limitations

JWD measurements were affected by the recovery time (dead time) of the system, acoustic noise, and winds (both vertical and horizontal). As \( R \) increases, the concentration of larger diameter drops also increases and hence, the number concentrations of smaller drops are significantly underestimated due to the dead time of the instrument. The channel counts were corrected for dead time of the instrument due to the ringing of the Styrofoam cone after impact by each drop. The corrected channel count is given by:

\[ n_i^- = n_i \exp \left[ \frac{0.035}{\Delta T} \sum_{D_i=0.85D_i}^{D_{\text{cut}}} n_i \ln \left( \frac{D_i}{0.85(D_i - 0.25)} \right) \right] \]

\[ \text{..... 2.21} \]

Where \( n_i \) and \( n_i^- \) are the number of drops in channel \( i \) before and after correction, respectively. When the channel count is small, the statistical sampling error is large and corrected count has the same relative uncertainty as the
uncorrected count. Counts for zero are not changed by this correction and it is impossible to know the drops count, if the dead time has caused drops not to be detected. This gives an error of < 3%, while calculating the rain integral parameters (Tokay and Short 1996) are the number of drops in channel $i$ before and after correction.

Table 2.4 Properties of the Joss-Waldvogel Disdrometer (JWD)

<table>
<thead>
<tr>
<th>Type</th>
<th>Impact-Type</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sampling area</td>
<td>50 cm$^2$</td>
</tr>
<tr>
<td>#bins internal</td>
<td>128</td>
</tr>
<tr>
<td>#bins output</td>
<td>20</td>
</tr>
<tr>
<td>Drop size range</td>
<td>0.35 – 6 mm</td>
</tr>
<tr>
<td>Sampling time</td>
<td>60 sec</td>
</tr>
</tbody>
</table>

Table 2.5 Processed Joss-Waldvogel Disdrometer (JWD) output data

<table>
<thead>
<tr>
<th>Quantity</th>
<th>Unit</th>
</tr>
</thead>
<tbody>
<tr>
<td>Total drops</td>
<td># admin</td>
</tr>
<tr>
<td>Concentration</td>
<td>1/m$^3$</td>
</tr>
<tr>
<td>LWC</td>
<td>g/m$^3$</td>
</tr>
<tr>
<td>Rain rate, R</td>
<td>Mm/hr</td>
</tr>
<tr>
<td>Reflectivity, Z</td>
<td>dBz</td>
</tr>
<tr>
<td>$D_m$</td>
<td>mm</td>
</tr>
</tbody>
</table>

The drop diameter measured by a JWD is sensitive to the drop momentum and in turn to the fall velocity. The presence of vertical winds changes the force with which the drop impacts the sensing surface. For instance, the downdrafts (updrafts) increase (decrease) the impact of the drop as well as the number concentration by increasing (decreasing) the fall velocity of the drops. The horizontal winds also affect the JWD measurements because the sensor surface is slightly conical. The drop count is underestimated in the presence of strong winds that produce turbulence at the edges of the sensor.

The instrument will suppress signals caused by acoustic noise, but drop signals not exceeding the level of the noise signal will also be suppressed together.
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with the noise signal. The output voltage produced by the smaller diameter raindrops is in the vicinity of the acoustic noise voltage, in particular in heavy rain. Therefore, this instrument under estimates smaller drops in heavy rain.

Though there exists some limitations, the instrument has shown it’s capability in measuring DSD continuously and regarded as one of the standard instruments for the measurement of DSD.

2.7.3 Micro Rain Radar (MRR)

Micro Rain Radar (MRR) is a vertically pointing Frequency Modulation (FM) - Continuous Wave (CW) Doppler radar at 24 GHz which measures the Doppler spectrum from 0 to 12 m/s. The Micro Rain Radar (MRR) short description and its functionality are given here, more detail can be found in Löffler-Mang et. al., (1999) and Peters et. al., (2002). Micro Rain Radar (MRR), a vertically point Doppler radar, is a very useful instrument to measure vertical profiles of precipitation particle size distributions and structures (Löffler-Mang et. al., 1999). The complete performance and quantitative measurements of rain by MRR was explained by Löffler Mang et. al., (1999). The antenna assembly and specification of the MRR installed nearby WPR is shown in Figure 2.10 and Table 2.6, respectively.

Conventional weather radars use the pulse radar mode while MRR uses a frequency modulated Gunn-diode-oscillator with integrated mixing diode. Thus, MRR can measure precipitation particle size distributions. It operates with electromagnetic radiation at a frequency of 24.1 GHz with a modulation of 0.5 – 15 MHz according to the height resolution (e.g. 10 m – 300 m). The CW-operation makes optimum use of the available transmit power. Thus a stable and reliable Gunn oscillator with only 50 mW output power can be used for transmitter. The MRR retrieves quantitative rain rates, drop size distributions, radar reflectivity, fall velocity of hydro meteors and other rain parameters simultaneously on vertical profiles up to several kilometers above the radar antenna.
Fig. 2.10 Micro Rain Radar Trans-Receiver.

The radiation is transmitted vertically into the atmosphere where a small portion is scattered back to the antenna from rain drops or other forms of precipitation. Due to the falling velocity of the rain drops, at the antenna there is a frequency deviation between the transmitted and the received signal (Doppler frequency). The standard real-time processing uses the relation given by Atlas et al., (1973), to attribute drop diameters to Doppler velocities. Mie theory is used to calculate drop numbers from the spectral volume reflectivity. Corrections for oblate drops and lower air densities leading to higher falling velocities in high altitudes are applied. The DSD is calculated for falling velocities from 0.78 to 8.97 m/s in 43 intervals, corresponding to drop diameters from 0.245 to 4.53 mm, which is the range where the signal to noise ratio is considered adequate. A list of all standard MRR data products is given in Table 2.7.
Table 2.6: The operating specifications of the Micro Rain Radar.

<table>
<thead>
<tr>
<th>Sl. No.</th>
<th>Parameter</th>
<th>Specification</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.</td>
<td>Radar Frequency</td>
<td>$f = 24.1$ GHz</td>
</tr>
<tr>
<td>2.</td>
<td>Radar Wavelength</td>
<td>$\Lambda = 1.25$ cm</td>
</tr>
<tr>
<td>3.</td>
<td>Modulation</td>
<td>FM-CW</td>
</tr>
<tr>
<td>4.</td>
<td>Beam Direction</td>
<td>Vertical</td>
</tr>
<tr>
<td>5.</td>
<td>Antenna</td>
<td>Offset Parabola</td>
</tr>
<tr>
<td>6.</td>
<td>Radom</td>
<td>NO</td>
</tr>
<tr>
<td>7.</td>
<td>Range Resolution</td>
<td>10 - 1000 m.</td>
</tr>
<tr>
<td></td>
<td></td>
<td>30 – 100 m (Typical values)</td>
</tr>
<tr>
<td>8.</td>
<td>Time Resolution</td>
<td>10- 3600 s</td>
</tr>
<tr>
<td>9.</td>
<td>Number of Range gates</td>
<td>30</td>
</tr>
<tr>
<td>10.</td>
<td>Spectral Velocity Resolution</td>
<td>0.191 m/s</td>
</tr>
<tr>
<td>11.</td>
<td>Nyquist velocity Range</td>
<td>0-12.3 m/s</td>
</tr>
<tr>
<td>12.</td>
<td>Number of power spectra</td>
<td>$\sim 10^3$</td>
</tr>
<tr>
<td>13.</td>
<td>Min. detectable radar reflectivity</td>
<td>-2 dBZ</td>
</tr>
</tbody>
</table>

Table 2.7: Micro Rain Radar standard output data

<table>
<thead>
<tr>
<th>Quantity</th>
<th>Unit</th>
</tr>
</thead>
<tbody>
<tr>
<td>Reflectivity, Z</td>
<td>dBZ</td>
</tr>
<tr>
<td>Height above ground</td>
<td>m</td>
</tr>
<tr>
<td>Rain rate R</td>
<td>mm/h</td>
</tr>
<tr>
<td>LWC</td>
<td>g/m$^3$</td>
</tr>
<tr>
<td>Fall velocity, V</td>
<td>m/s</td>
</tr>
<tr>
<td>DSD, $n(d)$</td>
<td>m$^{-1}$mm$^{-1}$</td>
</tr>
<tr>
<td>Doppler spectrum</td>
<td>dBn</td>
</tr>
</tbody>
</table>

An attenuation correction necessary in moderately high rain rates is done by calculating Mie extinction from the derived DSD. Rain rate, LWC, and Rayleigh reflectivity Z are calculated from the DSD, while mean falling velocity (first Doppler moment) and integral reflectivity (zeroth Doppler moment) are calculated directly from the measured Doppler spectrum. The MRRs range resolution can be set from 10 to 200 m in 30 height intervals. Attenuation at 24
GHz prevents the use of ranges higher than 6 km. The averaging time for one measurement can be set from 10 s up to several hours.

**Retrieval of Microphysical (Rain integral) Parameters**

The retrieval of range-resolved Doppler spectra follows the method described by Strauch and Merrem (1976). The MRR output products include the vertical structure of the radar reflectivity (Z in dBZ), liquid water content (LWC in g/m$^3$), rain rate (mm/hr) and falling velocity (m/s). Briefly how the output products are obtained is explained below.

The rain rate (RR) is obtained by integration over the precipitation particle size distribution:

$$ R = \frac{\pi}{6} \int_{D_{\text{min}}}^{D_{\text{max}}} N(D) \nu(D) D^3 dD $$

….. 2.22

The precipitation particle size distribution ($N(D)$) is given

$$ N(D) = \frac{\eta(D)}{\sigma(D)} , $$

….. 2.23

Where $D$ is raindrop diameter, $\sigma(D)$ is the backscattering cross-section, and $\eta(D)$ is the spectral reflectivity as a function of $D$, which is related to $\nu$, as

$$ \eta(D) = \eta(\nu) \frac{\partial(\nu)}{\partial(D)} . $$

….. 2.24

Where $\eta(\nu)$ is the spectral volume reflectivity as a function of $\nu$. The falling velocity is estimated by the spectra volume reflectivity,

$$ \nu = \frac{\lambda}{2} \int_{0}^{\infty} \eta(f) f df $$

….. 2.25

Where $\lambda$ is wavelength.
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The relation between falling velocity and drop diameter has been expressed by appropriated analytical form by Atlas et al., (1973)

\[ v(D) = 9.65 - 10.3 \times \text{Exp}(-0.6 \times D) \] , \hspace{1cm} (2.26)

Where \(0.109 \text{ mm} \leq D (\text{mm}) \leq 6 \text{ mm} \).

The \(Z\) and LWC are defined by the precipitation particle size distribution, \(N(D)\),

\[ Z = \int_{0}^{\infty} N(D) D^6 dD \] , \hspace{1cm} (2.27)

\[ \text{LWC} = \rho_w \frac{\pi}{6} \int_{0}^{\infty} N(D) D^3 dD \] , \hspace{1cm} (2.28)

Where \(\rho_w\) is the density of water.

To estimate the impact of using Mie theory for computing the backscatter cross-section (Löffler-Mang and Kunz, 1999), the backscattering cross-section \(\sigma_{\text{Mie}}\) of a dielectric sphere for a plane electromagnetic wave is given by

\[ \sigma_{\text{Mie}} = \frac{\lambda^2}{4\pi} \left| \sum_{n=1}^{\infty} (-1)^n (2n + 1)(a_n - b_n) \right|^2 \] , \hspace{1cm} (2.29)

Where \(a_n\) and \(b_n\) are derived from Bessel and Henkel function.

2.8. Radiosonde

The radiosonde (as shown in Figure 2.11) is a balloon-borne instrument platform with radio transmitting capabilities. Radiosondes are capable of making direct in-situ measurements of atmospheric temperature, pressure and humidity at different altitudes. It is capable of measuring upper air atmospheric parameters up to an altitude of about 30–35 km. The measured data are transmitted immediately to the ground station using a radio transmitter located within the instruments package.
The receiver at the ground receives the telemetry signals. The range, elevation and azimuth angles are determined by tracking the balloon at the ground using tracking radar. Presently the global positioning sensors are doing the job of tracking radars. As the balloon drifts with the wind, the position of the balloon at different altitudes can be used to obtain the wind speed and direction. The sensors used are thermistors, hygristors and aneroids to measure temperature, humidity and pressure, respectively. The temperature measuring range for the thermistor lies between approximately $+50 \, ^\circ C$ to $-90 \, ^\circ C$. The aneroid is designed to register pressures from 1014 hPa to 10 hPa or less. The hygrostat is designed to record the ambient relative humidity in the range of 1-100%.

Regular radiosonde launches (every 12 hours) were performed throughout the deployment to characterize the thermodynamic state of the atmosphere, as well as the wind speed and direction.

Fig. 2.11: Launch of GPS balloon at Korror-National Weather Service, Palau.
2.9. Tropical rainfall measuring mission precipitation radar

Satellite remote sensing is an efficient tool for measuring rainfall on a global scale. The Precipitation Radar (PR) on board on Tropical Rainfall Measuring Mission (TRMM) is the first space borne rain radar that can directly give the vertical distribution of rain. The radar consists of 128-element waveguide (WG) planar array arranged in an aperture of 2 m x 2 m. The operating frequency of PR is 13.8 GHz. During the normal observation mode, PR antenna beam scans in the cross track direction within ±17° from nadir, resulting a swath width of 220 km. The antenna beam width is 0.71° and there are 49 observation angle bins within the scanning angle of ±17°. The horizontal resolution (footprint size) is 4.3 km at nadir from 350 km altitude and about 5 km from 403 km orbit (from August 2001). The range resolution of TRMM-PR is 250 m, which is equal to the vertical resolution at nadir. Figure 2.1 shows the observation concept of TRMM-PR. The radar echo sampling is performed over the range gates between the sea surface and the altitude of 20 km for each observation angle bin. After the detailed statistical calculation, the minimum detectable $Z_e$ can be considered to be about 16-18 dBZ. The effective signal – to–noise ratio of 3 dB is obtained when $Z_e$ - factor is 17 dBZ.

The standard algorithms of PR are classified as Level-1 (1B21, 1C21), Level-2 (2A21, 2A23, 2A25) and Level 3 (3A25, 3A26). Level-1 and Level-2 products are the data in each scan pixel. Level-3 data provides the statistical values of rain parameters in $5° \times 5°$ grid boxes. The algorithm 2A25 retrieves profiles of the radar reflectivity factor $Z_e$, with rain attenuation correction and $R$ for each radar beam by the combination of Hitschfeld-Bordan and surface reference methods (Iguchi et. al., 2000).
Fig. 2.12: Observation concept of TRMM-PR (from TRMM-PR User guide).