Improved Precipitation Rates and Data Quality by Using Polarimetric Measurements

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1 Introduction

The additional information provided by polarisation diversity radar has the potential to remove many of the ambiguities and uncertainties present when only the conventional reflectivity (Z) and Doppler information are available. In this chapter we emphasise the application in an operational environment with a typical one degree beamwidth radar and a dwell time of about one sixth of a second so that it completes a PPI every minute and provides rainfall estimates with a spatial resolution of 2km or better. An excellent and very comprehensive book on polarimetric Doppler weather radar has recently appeared by Bringi and Chandrasekar (2001) and we refer the reader to this book for detailed derivations and analysis which we will simply quote. Essentially, polarimetric radar provides information on the shape and orientation of the radar targets. In section 2 we define the polarisation parameters, summarise their applications and discuss the theoretical and practical limits to the accuracy with which they can be estimated. The variation of raindrop shape with size and the typical raindrop size spectra are reviewed in section 3. The following sections consider how to exploit the shape and orientation information from a polarimetric radar which can transmit and receive horizontally and vertically polarised radiation as described by Gekat et al. (2003, this book).

- *Recognition of anomalous propagation and ground clutter (Section 4).*
  Precipitation particles are essentially an array of quasi-spherical scatterers which are usually small enough to Rayleigh scatter and so can be distinguished from ground echoes. Ground clutter and 'anaprop' are essentially non-Rayleigh scatterers and are far from spherical having very different scattering cross sections for horizontally and vertically polarised radiation.
- *Improved rainrate estimates (R) when rain alone is present (Section 5).*
  Empirical $Z(R)$ relationships of the form $Z = aR^b$ result in errors of R of up to a factor of two. They arise because of the variability in the raindrop size spectrum: $Z = \sum ND^6$ for rain, where N is the concentration and D the raindrop diameter, but $R \simeq \sum ND^{3.67}$. Because raindrops are oblate to a degree which depends upon their size, polarisation observations have the potential to estimate mean raindrop size and so provide better rainrates.
- *The use of combined and integral polarisation parameters (Section 6).*
  High resolution polarisation parameters tend to be very noisy. An alternative approach is to exploit their path integrated properties which can be

estimated much more accurately and use them as a constraint to provide, for example, absolute calibration of the radar reflectivity, an estimate of the constant 'a' in $Z = aR^b$, and a measure of the attenuation.

- *Improved rainfall estimates when ice is present (Section 7).*
  Large hailstones lead to very high values of Z which can be interpreted as spuriously high rainfall rates, but may be identified because they are oblate and tend to tumble as they fall. Melting snowflakes are responsible for the enhanced radar return at the melting layer known as 'the bright band' and can easily be recognised because they are wet, oblate and fall with a characteristic rocking motion.
- *Correction for attenuation (Section 8).*
  Attenuation for the C-band (5.6cm wavelength) radars used in Europe and Japan is a serious problem in heavy rain. Polarisation parameters which are immune to attenuation can be used to correct for such effects.
- *Identification of hydrometeors (Section 9).*
  For ice hydrometeors other than hail and melting snow, identification using polarisation is more difficult but polarisation can yield useful information.

In Section 10 the potential for implementation of polarisation parameters is summarised for an operational environment both for the 5.6cm wavelength (C-band) as generally used in Europe and Japan and 10cm (S-band) used in the USA. The use of S-band for polarisation observations has two major advantages: firstly, we have Rayleigh scattering for nearly all meteorological targets, and secondly, there are normally negligible propagation or attenuation problems. Conventional radar networks yield estimates of Z and rainfall rate with a spatial resolution of about 2km and an update time of 5 (or perhaps two and a half) minutes. If polarisation techniques require longer dwell times then the poor sampling could negate any increased accuracy of specific rainfall estimates.

2 The Polarisation Parameters

2.1 Introduction

We shall restrict the discussion to the use of linear polarisation in which the radar can transmit pulses which are alternately polarised in the horizontal (H) and vertical (V), and can measure both the two co-polar returns Z_H, Z_V; and, if the radar is Dopplerised, the phase of the horizontally and vertically polarised returns, ϕ_h and ϕ_v. We shall consider the following four parameters: the differential reflectivity (Z_{DR}), the co-polar correlation (ρ_{hv}), specific differential phase shift (K_{DP}), and the linear depolarisation ratio (LDR). In principle, these four parameters can also be derived from measurements made with circular polarisation but nearly all recent work has used linear polarisation. We will also briefly discuss transmission of pulses polarised at 45° with reception at H and V, which has the advantage that no rapid polarisation switching between transmitted pulses is needed.

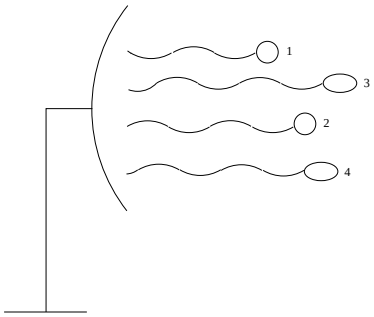

Fig. 1. Signals backscattered from precipitation particles arrive at the antenna with random phase.

Recall from Fig. 1, that the amplitudes of the backscattered waves from individual hydrometeors arrive at the antenna with random phase, so that as the scatterers move relative to one another, the observed intensity Z will fluctuate. The fractional standard error of the mean Z estimate is given by $1/\sqrt{n}$ for large n, where n is the number of independent samples of Z. The time to independence is given by $\lambda/(4\sigma_v\sqrt{\pi})$ where σ_v is the Doppler width of the target. Assuming that at a low beam elevation the precipitation has a Doppler width of 0.5m s^{-1}, then the time to independence at C-band is about 16ms, and at S-band about 28ms. If we need one PPI a minute then the dwell time is 0.166s and averaging over six adjacent gates, each typically 150m long (1μs transmitted pulse), provides a range resolution of 1km to match the one degree beamwidth azimuthal resolution at a range of 57km. This leads to n of about 63 at C-band and 36 at S-band, so the statistical sampling error of Z will be about 13% (0.53dB) and 17% (0.68dB), respectively. System errors are discussed by Gekat et al. (2003, this book).

2.2 Z_{DR} - Differential Reflectivity

Z_{DR} (=10 log(Z_H/Z_V))is a measure of mean particle shape. Z_{DR} is particularly useful for rain, because small raindrops are spherical but larger ones become increasingly oblate. An improved rainfall rate, $R(Z, Z_{DR})$, can be obtained from Z and Z_{DR}. Assuming an exponential Marshall-Palmer (1948) raindrop size distribution

$$N(D) = N_0 \exp(-3.67D/D_0) \tag{1}$$

where D_0 is the median volume drop diameter, then the value of D_0 can be estimated from Z_{DR}. The advantage of Z_{DR} is that it is a ratio which is independent of the concentration parameter, N_0. Once D_0 is known then the value of N_0 is fixed by the observed value of Z. An empirical Z-R relationship is approximately equivalent to assuming we have a Marshall-Palmer distribution with a variable D_0 but N_0 constant and equal to 8000 m^{-3} mm^{-1}. The use of Z and Z_{DR} to fix both N_0 and D_0 should result in more accurate rainfall estimates (Seliga and

Bringi, 1976). For a given value of Z_{DR} (fixed D_0), the rainfall rate and the value of Z both scale linearly with N_0, so we expect relationships of the form:

$$Z = cRf(Z_{DR}) \tag{2}$$

Ice particles have a lower dielectric constant than liquid water and so even if they are oblate they tend to have low values of Z_{DR}, particularly if, as is the case of snow, they are a low density mixture of air and ice. If oblate ice particles become wet, the value of Z_{DR} increases, and so the melting snow in the bright band is associated with high values of Z_{DR}. In vigorous convection supercooled raindrops can be recognised by narrow vertical columns of positive Z_{DR} extending above the freezing level (e.g. Illingworth et al., 1987). Hail usually tumbles as it falls and so should be associated with a Z_{DR} of 0dB. At low altitudes, where rain is to be expected, a Z_{DR} value of 0dB accompanied by a high value of Z can be used to identify hail. Unfortunately, hail is often mixed with rain - so the result will be an intermediate value of Z_{DR}, which is lower than it should be for the rain, but the presence of hail will lead to higher than expected values of Z. In section 5 we shall show that for improved rainfall estimates using $R(Z, Z_{DR})$, Z must be accurately calibrated and Z_{DR} estimated to within 0.2dB. Z_{DR} can be calibrated to 0.1dB by observing precipitation at vertical incidence which we know has a value of 0dB, or, less satisfactorily, looking at the statistics of Z_{DR} values in very light drizzle. Bringi et al. (1983) show that for individual observations to be accurate to 0.2dB requires about 60 independent samples of Z and a correlation coefficient of the H and V time series (see next subsection) above 0.98. These values should be satisfied by an operational radar scanning in rain with a resolution of about 1-2km, but in practice other factors may limit the accuracy of Z_{DR}:

- *Reflectivity Gradients.*
  In the presence of reflectivity gradients a significant fraction of the power may be received through the sidelobes which are mismatched in their polarisation characteristics. Herzegh and Carbone (1984) report spurious values of Z_{DR} of up to 10dB in low reflectivity regions adjacent to intense echoes. The sidelobes of operational radars may well quite commonly lead to the introduction of errors in Z_{DR} of 0.2–0.5dB.
- *Mismatched beams.*
  If the H and V beam patterns are not well matched then the two beams will not be sampling the same volume of precipitation and the correlation may fall below 0.98.
- *Triple scattering echoes.*
  Triple scattering echoes (Illingworth and Caylor, 1988; Hubbert and Bringi, 2000) can occur in a weak echo region behind an intense storm because some of the radiation scattered by the main echo is reflected by the ground and then scattered a third time by the main echo back to the antenna. No vertically polarised power is scattered to the ground, so such echoes are associated with anomalously high values of Z_{DR}.

- *Differential attenuation.*
  This leads to a much more severe problem at C-band. The large oblate raindrops attenuate the horizontally polarised radar beam more than the vertical one, and this differential attenuation leads to increasingly negative values of Z_{DR} with range. One example is show by Collier and Hardacre (2003, this book). To demonstrate the severity of this problem, note that at C-band a differential attenuation of 0.2dB results from only 1km of rain at 65 mm hr^{-1} or 2km of rain at 40 mm hr^{-1}. Upton and Fernandez-Duran (1999) report values as low as -5dB behind intense echoes, such attenuation must be corrected to an accuracy of 0.2dB.
- *Hail.*
  Finally, as described above, the presence of hail mixed with the rain will lead to a reduction in Z_{DR} so that application of the $R(Z, Z_{DR})$ algorithm will lead to the inference of a spuriously high rainfall rate.

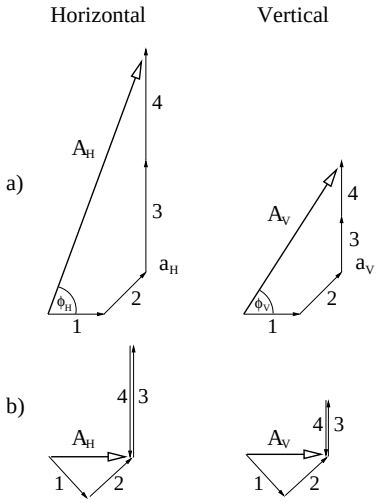

Fig. 2. Argand diagram showing that the addition of backscattered amplitudes from the mixture of particles shapes in Fig. 1, a) at time t=t$_1$ and b) at a later time t=t$_2$, leads to a lowering of the correlation of the H and V returns and fluctuations in the differential phase.

2.3 Co-polar Correlation Coefficient

The co-polar correlation coefficient (ρ_{hv}) is the correlation of the time series of successive estimates of Z_H and Z_V and is a measure of the variety of shapes of hydrometeors present. The importance of this parameter is demonstrated in Fig. 2 which shows schematically how the amplitude and phase of the signal

returned from four targets, two spheres and two oblate spheroids (see Fig. 1), add to give a fluctuating return as the targets reshuffle in space between two times, t=t_1, and t=t_2. In the figure the spheres and the oblate spheroids all have the same backscatter amplitude (a_v) for vertical polarisation, but for horizontal polarisation the spheres have $a_h = a_v$ whilst for the oblate spheroids $a_h = 2a_v$. The figure shows that as the particles reshuffle, the relative phases change and the resultant amplitudes A_h and A_v at the two polarisations will fluctuate. In this case because the ratio of a_h to a_v is not the same for all the scatterers, then as the particles reshuffle the ratio of A_h to A_v will fluctuate and so the time series of Z_H and Z_V (proportional to A_h^2 and A_v^2) will not be perfectly correlated. As the correlation falls then the estimate of Z_{DR} also becomes less accurate. The correlation coefficient, (ρ_{hv}), will only be unity if the ratio of a_h to a_v is the same for all the particles, that is to say the particles all have the same shape and orientation in space. The values of correlation can be used to identify targets as follows:

- *Raindrops*
  Raindrops are nearly spherical and so the correlation is at least 0.98.
- *Hail and the Bright Band*
  The correlation drops to about 0.9 when a wide mixture of hydrometeor shapes is present. This occurs in the bright band or when hail particles are large enough to result in non-Rayleigh scattering.
- *Clutter and Anaprop*
  Returns from the ground, whether as clutter or anomalous propagation, have a correlation which is essentially zero because the returns in the two polarisations have almost random amplitudes and phases.

In practice the observed values of ρ_{hv} will be lowered somewhat because the H and V beams can never be perfectly matched in space and also because H and V are not sampled simultaneously. The effect of staggered H and V samples can be corrected by either assuming a Gaussian Doppler width of the target or, more accurately, by interpolating the two time series. For very precise estimates of ρ_{hv} the dwell times must be much longer than those allowed in an operational system.(Illingworth and Caylor, 1991).

2.4 K_{DP} - Specific Differential Phase.

The velocity of a horizontally polarised radar wave propagating through a region containing oblate raindrops is slightly less than that of a vertically polarised wave, so, as shown in Fig. 3, the phase of the horizontal return (ϕ_h) lags progressively behind the phase of the vertical polarised return (ϕ_v). As a result the differential phase, $\phi_v - \phi_h = \phi_{DP}$, normally increases monotonically with range and K_{DP}, the rate of change of ϕ_{DP} with range (measured in $^\circ$ km^{-1}), should be positive. K_{DP} should not be affected by tumbling hailstones, but should increase with rain rate. $R(K_{DP})$ relationships (Sachidananda and Zrnic, 1986, 1987) have been proposed of the form:

$$R = aK_{DP}{}^b \tag{3}$$

with a value of $b = 0.866$. This technique appears to be very attractive, and has the following advantages (e.g., Blackman and Illingworth, 1995; Ryzhkov and Zrnic, 1996)

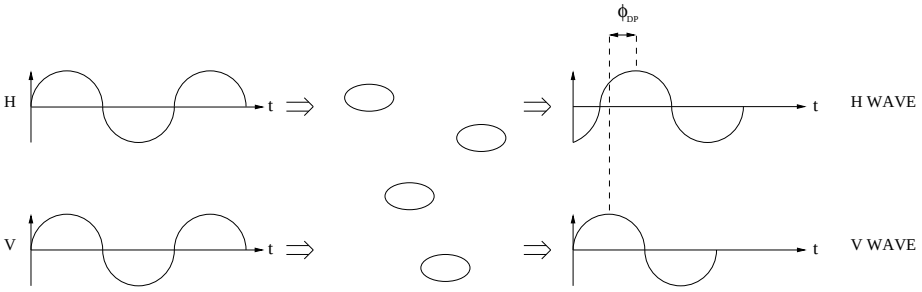

Fig. 3. Differential phase shift, ϕ_{DP}, introduced because the H wave propagates more slowly than the V wave through a region of oblate raindrops.

- *Calibration and Attenuation*
  The measurement of phase is immune to calibration problems and is unaffected by attenuation.
- *Hail*
  The phase measurement should not be affected by hail.
- *Linearity*
  If the value of 'b' in Eqn 3 is 0.866 then K_{DP} has a more linear dependence on R than conventional $R(Z)$ relations. Bringi and Chandrasekar (2001) show that K_{DP} should be less dependent upon the precise form of the drop shapes and, assuming Rayleigh scattering, is approximately proportional to the product of the liquid water content times D_0, or to the fourth moment of the drop size spectrum and so should be closer to R than Z.
- *Integrated rainfall over a catchment.*
  If Eqn 3 is nearly linear then the total integrated differential phase shift along the path of the radar beam (Φ_{DP}) will be a measure of the integrated rainfall along the path.
- *Use of partially blocked beams.*
  If ϕ_{DP} can be observed in low elevation beams which are partially blocked so that Z values are considerably reduced, it may be possible to estimate R from K_{DP} and so use these low elevations beams which will remain in the rain out to greater distances.

A difficulty arises because the phase shifts are quite small. If these advantages are to be realised, then ϕ_{DP} should be estimated to 1° or better, but unfortunately the phase measurement can be quite noisy. Firstly, there is a theoretical limit to the accuracy of the ϕ_{DP} estimate:

- *Sequential Pulses*
  ϕ_{DP} is usually measured from sequential pulses transmitted with alternate horizontal and vertical polarisation, but the phase itself is continually changing due to the mean Doppler velocity of the targets and so interpolation is required to estimate ϕ_{DP}. The accuracy of this interpolation is limited by the Doppler width of the target. Ryzhkov and Zrnic (1998b) show that for 60 pulse pairs and a normalised spectral width ($= 2\sigma_v t_s/\lambda$, where t_s is the time between pulses) of 0.1, the standard error of ϕ_{DP}, (σ_ϕ), should be about 1°.
- *'Hybrid' Transmission at 45°*
  Transmission of a pulse polarised at 45° and simultaneous reception at H and V (Bringi and Chandrasekar, 2001) avoids the interpolation problem, but the correlation of the target is slightly less than unity and this introduces noise into the ϕ_{DP} estimate which still leads to a σ_ϕ of about 1°. The mechanism by which fluctuations in the estimate of ϕ_{DP} are introduced by the low correlation can be seen in Fig. 2.

This theoretical accuracy of ϕ_{DP} is not achieved in practice. Ryzhkov and Zrnic (1995) cite a typical σ_ϕ at S-band of 3°, Hubbert et al. (1993) and Keenen et al. (1998) quote similar figures for their C-band radars. This degradation is probably caused by the following factors:

- *Gradients of reflectivity.*
  Gradients in reflectivity across the beam cause one part of the beam to experience a larger ϕ_{DP} than another (Ryzhkov and Zrnic, 1998a) and can give rise to negative K_{DP}. Gradients along the beam lead to biases in inferred K_{DP} (Gorgucci et al., 1999a) whereby, for example, a sudden change in Z of 30dBZ can bias the value of K_{DP} by up to 30%.
- *Extreme sensitivity to clutter.*
  If the random phase of a ground clutter signal is added to a precipitation return which has an amplitude ten times larger, then the resultant signal will have a phase noise of 5°. This will render the value of ϕ_{DP} virtually useless, even though the ground clutter will have a Z (i.e. an intensity) which is 20dB below (i.e. 1% of) the Z of the precipitation.
- *Sidelobes.*
  ϕ_{DP} is much more sensitive to signals from mismatched sidelobes than Z_{DR}. If the phase of the sidelobe signal is mismatched in H and V, then again all we need is sidelobe signal which is 20dB below the precipitation return from the main lobe to introduce a 5° noise in ϕ_{DP}.
- *Differential phase shift on backscatter.*
  If the particles or rain or hail are large enough for non-Rayleigh scattering then a differential phase shift on backscatter, δ, will be introduced as a transient superposed on the monotonic increase in ϕ_{DP} due to propagation and can lead to apparent negative values of K_{DP}. For rain, Testud et al. (2000) show that the effect is negligible at S-band, but at C-band δ is about 1°, 4.5°, and 9.5° for Z_{DR} =2dB, 3dB, and 4dB, respectively. Ryzhkov and Zrnic (1996) suggest removing the δ transient by 'light' filtering over a distance of 2.4km for heavy rain and 'heavy' filtering over 7.2km for light rain.

The result of the above effects is a noisy trace of ϕ_{DP} not always increasing monotonically with range. Filtering is difficult because the extent of the δ region may not be known and it leads to an unacceptable loss of spatial resolution. Ryzhkov and Zrnic (1996) suggest using the modulus of K_{DP} in Eqn 3 to avoid the problem of negative values, but this does not seem physically justifiable.

2.5 LDR - Linear Depolarisation Ratio

LDR is defined as 10 log (Z_{VH}/Z_H), where Z_{VH} is the cross-polar return at horizontal polarisation for a vertically polarised transmission. Only oblate particles falling with their major axis at an angle to the vertical or horizontal direction yield a cross-polar return. LDR is able (Frost et al., 1991; Straka et al., 2000) to identify melting snowflakes associated with the bright band, on occasion it can differentiate ice from rain. Finally, ground clutter has anomalous LDR signals.

Fig. 4a demonstrates that raindrops result in negligible depolarisation because they fall with their major axis aligned in the horizontal; the horizontal field excites a single dipole along the major axis which re-radiates with horizontal polarisation. Contrast this with the oblate non-aligned hydrometeor in Fig. 4b; the incident field excites two dipoles along the major and minor axis, and if the polarisabilities along the two axes are not the same then the resultant induced dipole is no longer parallel to the induced field, and so the re-radiated field has an orthogonal or cross-polar component.

The difference in polarisability is largest for wet oblate particles, so the highest values of LDR of about -15dB are associated with melting snowflakes, because such particles rock and roll as they fall. The LDR of melting hail and dry high density ice crystals is in the range -20 to -26dB. Rain has an LDR generally below -30dB but the observed value is often limited by the antenna isolation. The cross and co-polar signals are essentially uncorrelated, so if the error in Z is about 0.5dB, then value of LDR can be estimated to 0.7dB. Masking out of echoes where LDR > -10dB removes much of the ground clutter (see Section 4).

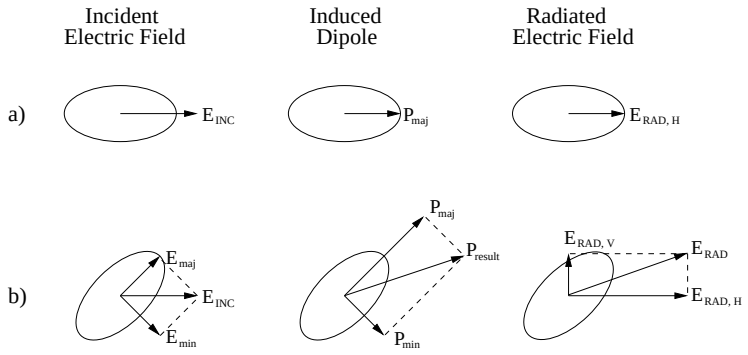

Fig. 4. Linear Depolarisation Ratio. a) No depolarisation of incident wave by aligned raindrops. b) Depolarisation by non-aligned oblate spheroids.

2.6 An example of a PPI scan through vigorous convection

An S-band low elevation PPI scan though vigorous convective showers which illustrates the signatures of Z, Z_{DR} and ϕ_{DP} is displayed in Fig. 5 which also appears as a colour plate. The observations were made with the Chilbolton 0.28° beamwidth radar in the UK on 28 July 2000 at 1308UT at 0.5° elevation, so that the beam is dwelling in the rain. Because of the narrow beamwidth we believe that this image is not affected by any of the artifacts discussed above such as sidelobes and reflectivity gradients either across or along the main beam. From Fig 5 we see that once Z is above about 35dBZ then Z_{DR} reaches about 1dB, but that the high phase shifts generally result from regions of Z above 45dBZ extending over a few km. Two such regions producing large phase shifts are clearly visible: one at 90km E and -12km N, the other at 115km E and about -8km N. Even for such large phase shifts the estimates of K_{DP} from differentiating the phase are quite noisy and have a spatial resolution which is much less than the 300m of the Z and Z_{DR} and observations. Note that the highest values of Z_{DR} are about 4dB but are quite localised and often not coincident with the highest Z; they do not produce any phase shift and we believe that they result from low concentrations of anomalously large rain drops.

3 Raindrop Shapes and Size Spectra

3.1 Introduction

Both Z_{DR} and K_{DP} essentially provide information on the mean shape and hence the size of raindrops. If this is to lead to improved rainfall rate estimates, then the shape of raindrops as a function of size must be known precisely, but there is currently some uncertainty over these shapes. In addition, if the size sensed by the radar is to be related to the appropriate mean drop size for the rainfall rate then we need to know the range of size spectra for naturally occurring rainfall. Again there is some uncertainty over these spectra.

3.2 Raindrop Shape Model

Until very recently, a simple 'linear' relationship between axial ratio, r, and drop diameter, D in mm, was in widespread use (Pruppacher and Pitter, 1971):

$$r = 1.03 - \beta D \tag{4}$$

with β having a value of 0.062 mm^{-1} for drops larger than 0.5mm diameter; drops smaller than 0.5mm are assumed to be spherical. However, in some of the earliest observations Goddard et al. (1982) found that observed values of Z_{DR} in the range 0dB to 1.5dB were from 0.5 to 0.2dB lower than the values of Z_{DR} computed using the linear shapes from a raindrop disdrometer at the ground 200m below the radar beam. They could only obtain agreement if they empirically adjusted the shape of drops smaller than 2.5mm to be more spherical

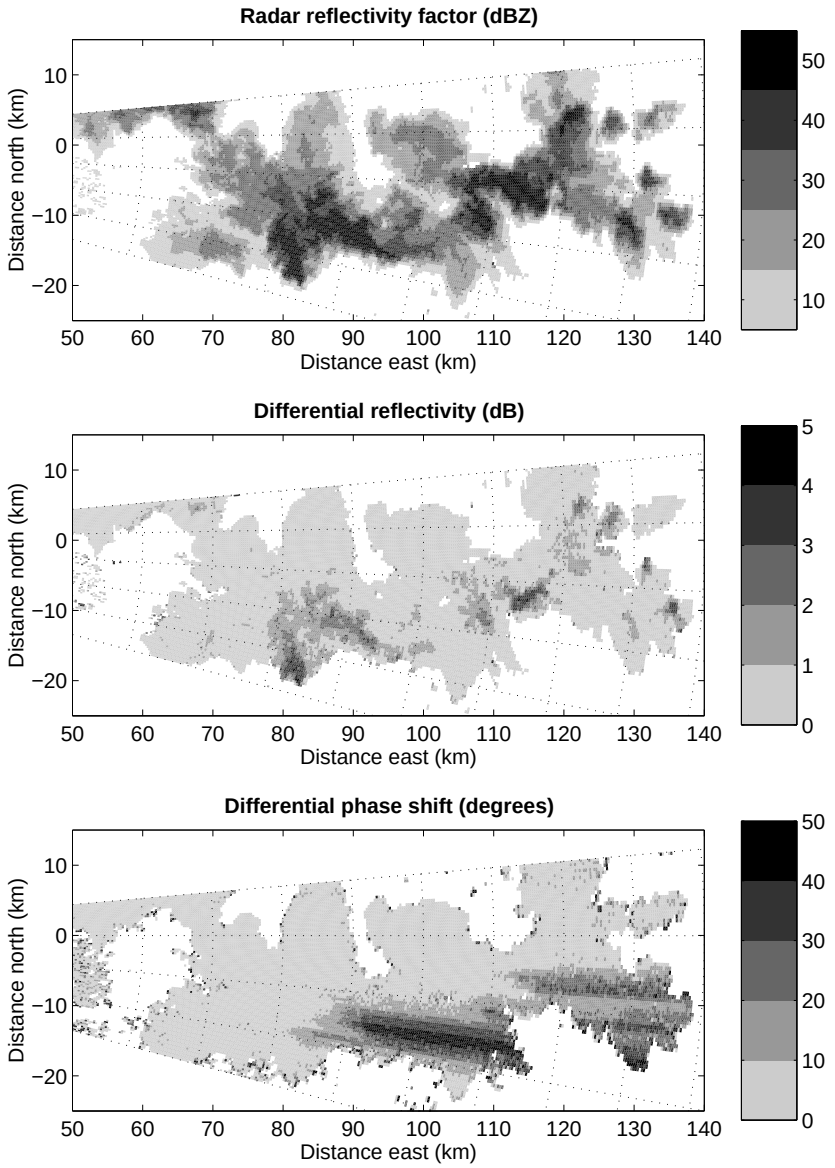

Fig. 5. An example of Z, Z_{DR} and ϕ_{DP} observed with the narrow beam S-band Chilbolton radar in the UK on 28 July 2000. The data are from a low elevation (0.5°) PPI scan, so the beam is dwelling in the rain. Even for these large values of ϕ_{DP} the derived values of K_{DP} are rather noisy with much poorer range resolution than Z and Z_{DR}. Figure courtesy of R.J.Hogan (U of Reading). Radar data kindly supplied by RCRU, Rutherford Appleton Laboratory.

than given by Eqn 4. They proposed (Goddard et al., 1995) a new drop shape model:

$$r = 1.075 - 0.065D - 0.0036D^2 + 0.0004D^3 \tag{5}$$

for drops larger than 1.1mm with smaller drops again assumed spherical. Confirmation of such drop shapes is not simple. If the raindrop shape is measured close to the ground then the drops have just experienced atypically high shear as they fall through the boundary layer. Similarly, drop shapes measured with in-situ aircraft probes have been subjected to unusual stresses as they are deflected around the aircraft wing. It is only recently that careful experiments in long wind tunnels such as those carried out by Andsager et al. (1999), and references therein, have proposed the following polynomial:

$$r = 1.012 - 0.01445D - 0.01028D^2 \tag{6}$$

for D in the range 1-4mm. The three drop shape models are compared in Fig. 6. Eqns 5 and 6 are very similar and predict virtually the same values of K_{DP} and Z_{DR}. This remarkable independent confirmation of the empirical adjustment gives us confidence that the 'linear' shapes are indeed an oversimplification and that these 'new' drop shapes are probably correct. The last two Eqns lead to negligible differences in the predicted polarisation parameters; when we refer to 'new' drop shapes we will be using those of Eqn 5 as opposed to the 'linear' shapes of Eqn 4. Eqn 6 is equally valid.

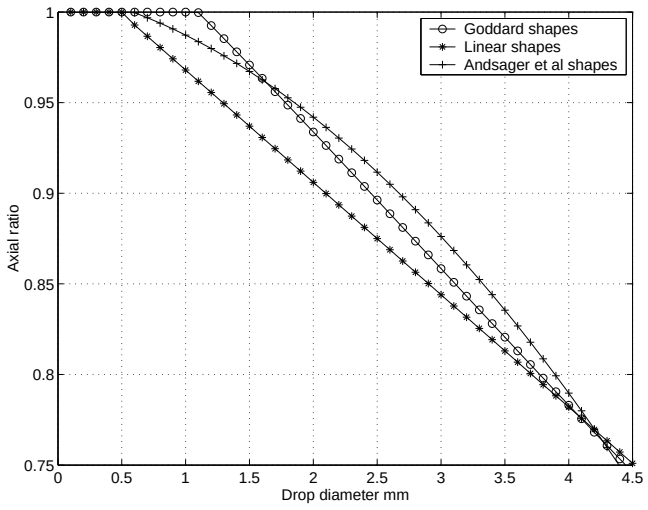

Fig. 6. Different raindrop shapes models.

The linear drop shapes used in the past lead to two effects. Firstly, the linear drop shapes are more oblate than those occurring naturally leading to predicted

values of Z_{DR} and K_{DP} which are too high so that values of 'c' in Eqn 2 and 'a' in Eqn 3 are too small; this may account for the underestimates of rainfall using K_{DP} reported, for example, by May et al. (1999) and Petersen et al. (1999) who used the 'linear' raindrop as discussed by Bringi and Chandrasekar (2001). Secondly, for drops in the range 1-3mm the 'new' drop shapes predict a larger value of β than the 0.062 in Eqn 4; this means that the increase in the values of Z_{DR} and K_{DP} as the rainfall rate increases will be more rapid than predicted from the linear drops, so that in reality the value of 'b' in the $R(K_{DP})$ exponent will be lower than the value of 0.866 predicted using linear shapes (Illingworth and Blackman, 2002).

3.3 Raindrop Size Spectra

To gauge the performance of polarisation techniques in improving rainfall estimates we need to know how variations in the drop size spectra affect the representativeness of the mean drop size inferred from the polarisation parameters. Rather than the simple exponential representation in Eqn 1, natural raindrop size spectra are well represented by a normalised gamma function of the form

$$N(D) = N_w f(\mu) \left(\frac{D}{D_0}\right)^{\mu} \exp\left(-\frac{(3.67+\mu)D}{D_0}\right) \tag{7}$$

where

$$f(\mu) = \frac{6}{(3.67)^4} \frac{(3.67+\mu)^{\mu+4}}{\Gamma(\mu+4)} \tag{8}$$

which when $\mu = 0$ reduces to the simple exponential form of Eqn 1 with $N_w \equiv N_0$. This form has the advantage that the three variables are independent, with D_0 being the median volume drop diameter, μ being the shape of the drop spectrum, and N_w is a normalised drop concentration so that the liquid water content remains constant even if μ changes, and is preferable (Illingworth and Blackman, 2002) to the non-normalised gamma function of Ulbrich (1983),

$$N(D) = N_0 {D_0}^{\mu} \exp\left(-\frac{(3.67+\mu)D}{D_0}\right) \tag{9}$$

We have followed the convention of Bringi and Chandrasekar (2001); N_w is the same as the N_L used by Illingworth and Blackman (1999, 2002) and the N_0^* used by Testud et al. (2001, 2003 - this book), with 'w' and 'L' both indicating normalisation with respect to liquid water content. These equations are sometimes expressed in terms of D_m, the mass weighted mean diameter, in which case D_m is equal to D_0 times the factor $(4+\mu)/(3.67+\mu)$.

Ulbrich (1983) derived the expected range of values of N_0 and μ in the non-normalised Eqn 9 in natural rainfall by comparing expressions for R and Z from the appropriately weighted integral of Eqn 9 with the 69 $Z(R)$ relationships of Battan (1973); he deduced the range of μ was from -1 to 5. The mathematical validity of this approach for deriving the 'Ulbrich' range of values of N_0 and μ has

been questioned by Illingworth and Blackman (2002), but it has become common practice to derive relationships between R and the polarimetric variables by cycling over these 'Ulbrich' values of N_0 and μ, calculating Z, Z_{DR}, K_{DP} and R and performing a non-linear regression.

We now explore other methods of deriving the range of values of μ, N_w and D_0 in naturally occurring rainfall. One approach is to use observed raindrop spectra. A least squares fit is not appropriate because it gives equal weight to small and large rain drops, whereas for radar and rainfall the larger drops are much more important. Kozu and Nakamura (1991) and Illingworth and Johnson (1999) both equated the sixth, fourth and third moments of observed raindrop size distributions to the appropriately weighted integral of the gamma function and deduced values of μ in the range 0 to 15 with a mean value of about 5 or 6. Illingworth and Johnson (1999) found that in the UK the mean value of the normalised raindrop concentration, N_w, was close to the Marshall-Palmer value of 8000 m^{-3} mm^{-1} with a standard deviation of a factor of three. Bringi and Chandrasekar (2001) found a similar spread of N_w when they analysed an entire season of rainfall spectra from Darwin, Australia.

These high values of μ derived from fitting the higher moments have been criticised because they are very dependent upon maximum drop size in the spectra which are poorly sampled by disdrometers. The values obtained depend upon the moments chosen for the fit. Higher moments are appropriate for relationships between Z and R and lead to higher values of μ. Testud et al. (2001) and Bringi and Chandrasekar (2001) infer μ values closer to unity; this may be because they first derive N_0 and D_0 in an exponential spectrum by fitting moments, but then choose μ to minimise a least squares fit to the observed spectrum. This procedure for fixing μ assigns equal weight to drops of all sizes and may lead to values of μ which are inappropriate for the higher moments involved in radars studies. Ulbrich and Atlas (1998) used truncated moments by limiting the experimental spectra to D_{max} and found that this reduced μ from about 4 to a median value of 0 with a mean of 1.6.

The disadvantage with the above techniques is the use of a disdrometer which has poor sampling of the larger drops which are important for the higher moments. This can be avoided by appealing to radar measurements themselves; because of the much larger radar sampling volume the large drops are sampled satisfactorily. Wilson et al. (1997) reported values of μ derived from the 'Differential Doppler Velocity', DDV, the difference in the Doppler velocity of rain for horizontally and vertically polarised radiation with the radar beam dwelling at a finite elevation. They showed that the value of DDV as a function of Z_{DR} depends upon the value of μ and found a range of μ between 2 and 10 with a mean value of 5. There is also some evidence from observations of the Doppler spectral width of rainfall at vertical incidence which indicate (pers. comm., D. Bouniol) that the width is significantly narrower than would be observed for an exponential and is consistent with a μ of around 5.

3.4 Implications for $Z = aR^b$ Relationships

It appears that the range of naturally occurring raindrop size spectra is accurately described by the normalised gamma function in Eqn 7 with values of μ near 5 and values of N_w which have a standard deviation of factor of three. We now follow the treatment of Bringi and Chandrasekar (2001) and show that the normalised gamma function leads to $Z(R)$ relationships of the form $Z = aR^{1.5}$ with different values of 'a' reflecting changes in N_w. There is however an additional dependence of 'a' upon the value of μ. Integrating the appropriately weighted normalised gamma function produces expressions of the form:

$$Z = F_Z(\mu) N_w D_0^7, \tag{10}$$

and assuming the terminal velocity varies as $D^{0.67}$,

$$R = F_R(\mu) N_w D_0^{4.67}; \tag{11}$$

eliminating D_0 gives

$$\frac{Z}{N_w} = H(\mu) \left(\frac{R}{N_w} \right)^b \tag{12}$$

where b = 7/4.67 ≈ 1.5 or:

$$Z = H(\mu) {N_w}^{1-b} R^b = \frac{H(\mu) R^{1.5}}{\sqrt{N_w}}. \tag{13}$$

It is usually assumed that H(μ) is a slowly varying function of μ. This is so if we restrict μ to the range 0 to 2, but earlier we argued that μ has an average value of 5, and can vary between 0 and 10. As μ increases from 0 to 5 and 10, the value of H(μ) falls by a factor of 1.55 (2dB) and 1.79 (2.5dB).

Accordingly we can associate changes in the multiplicative factor 'a' in $Z = aR^b$ not only with changes in N_w but also with μ. For example, a standard deviation of N_w of a factor of three will lead to a change in 'a' of a factor of 1.7, and a change in rainfall rate derived from Z of ±40%. To this we must add a similar change in 'a' and Z from the range of μ; if the two are uncorrelated then the error in R will be about 64%. If the range of values of N_w is a factor of ten: this will lead to changes in 'a' of a factor of three, and a factor of two error in the rain rate. This gives us a good estimate of the error expected by a simple $Z = aR^b$ relationship because of variability in drop spectra. In Section 6 we will discuss a simple polarimetric method of estimating N_w and hence gauging the value of 'a', leaving only the variability due to the unknown μ.

4 Identification of Ground Clutter and Anomalous Propagation

Anomalous Propagation can be one of the most difficult signals to reject using conventional radar (e.g. Pamment and Conway, 1998) but fortunately their identification via polarisation returns is unambiguous. Both clutter and 'anaprop'

give rise to non-Rayleigh scattering so that the both the amplitudes and phases of the H and V returns are essentially uncorrelated. Accordingly, these signals can be identified by:

- The most reliable technique relies upon a copolar correlation of essentially zero as opposed to values above 0.9 for precipitation echoes (Caylor and Illingworth, 1992; Ryzhkov and Zrnic, 1998b).
- A noisy ϕ_{DP} profile (Ryzhkov and Zrnic, 1998b)
- Higher values of LDR than occur in natural precipitation (Wilson et al., 1995; Hagen, 1997).
- Noisy estimates of Z_{DR} which fluctuate between ±3dB from gate to gate (Hall et al., 1984).

5 Improved Rainfall Rates using Polarisation Parameters

5.1 Introduction

In this section we discuss the improved accuracy in rainfall rates by using Z_{DR} or K_{DP}. The error in R derived from Z alone can be up to a factor of two; the goal is to reduce this to ±25%. We shall initially consider cases when we know that the beam is dwelling in rain and there is no ice present. This will apply to summer time situations in the mid-latitudes out to a range of about 75km when the melting layer is above 2km and no hail is present. The situation is more common in the tropics because the melting level is much higher and large hail is rarer.

5.2 $R = f(Z, Z_{DR})$

The procedure outlined in Section 2.2 for calculating rainfall from Z and Z_{DR} seems straightforward. Bringi and Chandrasekar (2001) suggest that at S-band $D_0 = 1.529{Z_{DR}}^{0.467}$ (with Z_{DR} in dB) and that we should be able to estimate D_0 from Z_{DR} to within 0.1mm, but plots of D_0 as a function of Z_{DR} computed for normalised gamma functions (Fig. 7) with the 'new' drop shapes are initially discouraging, showing great sensitivity to μ. From Fig. 7, if $Z_{DR} = 1.5$dB and $\mu = 0$ then we have $D_0 = 1.5$mm but for the narrower spectrum with $\mu = 5$ and the same Z_{DR}, the absence of larger drops in the tail of the spectrum means that we need $D_0 = 2.2$mm. However, the goal is to measure R not D_0, and R is a function of both D_0 and μ. Fortunately, for D_0= 2.2mm and $\mu = 5$, both the rainfall rate and the value of Z are about five times higher than for $D_0 = 1.5$mm and μ=0. As we shall see in the next paragraph, this means that the value of Z per unit R as a function of Z_{DR} for the normalised gamma functions has a much reduced dependence upon μ, with the curves for $\mu = 0$ and $\mu = 5$ only 1dB apart.

In Section 2.2 we argued that Z/R should be a function of Z_{DR} (Eqn 2). Values of Z/R as a function of Z_{DR} for the 'linear' (Eqn 4) and 'new' (Eqn 5) drop shapes are plotted in Fig. 8. These curves can be interpreted as the Z

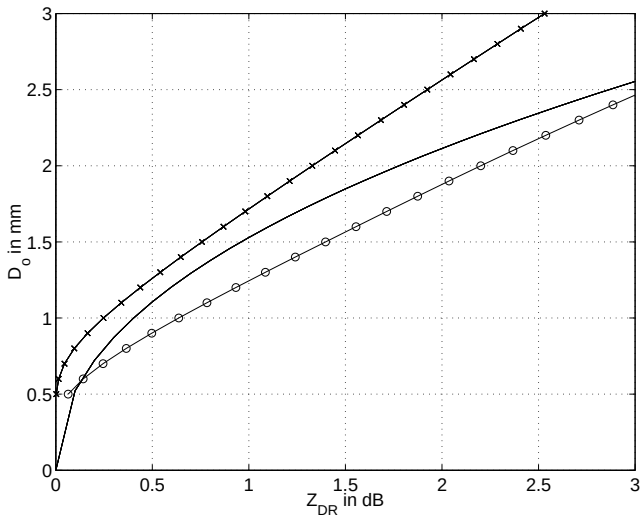

Fig. 7. Plot of D_0 against Z_{DR} for the new drop shape models at S-band with $\mu = 0$ (o) and $\mu = 5$ (x). The solid line is $D_0 = 1.529{Z_{DR}}^{0.467}$ suggested by Bringi and Chandrasekar (2001)

.

which would give R of 1 mm hr^{-1} for a particular value of Z_{DR}. The rain rate scales with Z, so, if for a given Z_{DR}, the observed value of Z is 'x'dB above the line then the rain rate is 'x'dB above 1 mm hr^{-1}.

It was argued in Section 3.3 that μ of 5 is the average value for natural rainfall. A third order polynomial fit of Z/R as a function of Z_{DR} for $\mu = 5$, gives the following expression for Eqn 2 (Illingworth and Blackman, 2002) :

$$Z/R = 21.48 + 8.14Z_{DR} - 1.385(Z_{DR})^2 + 0.1039(Z_{DR})^3 \tag{14}$$

at S-band which is accurate to 0.5dB. The equivalent expression at C-band is:

$$Z/R = 21.50 + 8.35Z_{DR} - 1.89(Z_{DR})^2 + 0.1976(Z_{DR})^3 \tag{15}$$

This assumes that the values of Z_{DR} have been corrected for any attenuation to an accuracy of 0.2dB.

An alternative approach has been widely adopted in the literature (see discussion in Bringi and Chandrasekar (2001)) in which the requirement for Z to scale with R for constant Z_{DR} is dropped and a relationship of the form

$$R = c_1 {Z_H}^{a_1} 10^{0.1 b_1 Z_{DR}} \tag{16}$$

is used. The coefficients are chosen by allowing the N_w (or N_0), D_0 and μ to cycle over the 'Ulbrich' (Secn 3.3) range of values, calculating Z, Z_{DR} and R and performing a non-linear regression. The curve proposed by Bringi and

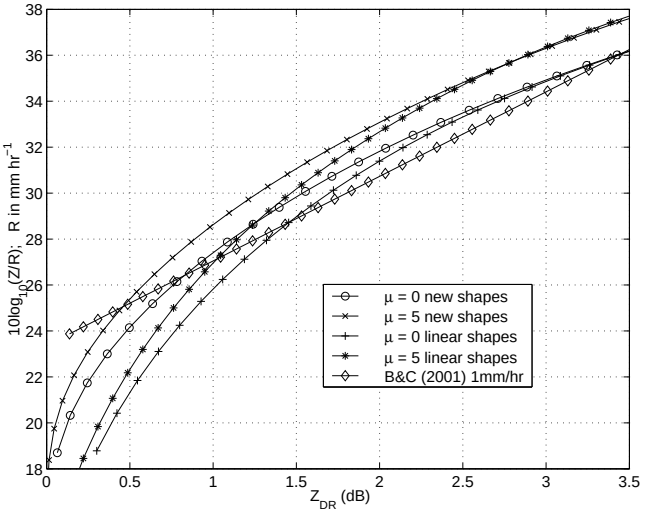

Fig. 8. Plot of Z/R against Z_{DR} for the linear and new drop shape models at S-band

Chandrasekar (2001) with $c_1 = 0.0067, a_1 = 0.93, b_1 = -3.43$ at S-band using the 'new' drop shapes for $R = 1$ mm hr^{-1} is also plotted in Fig.8 and predicts rainfall rates 2dB higher than Eqn 14 once Z_{DR} is above 1dB. For a higher rain rate of 10 mm hr^{-1} the non-linearity with Z introduced by $a_1 = 0.93$ means that the curve would be 0.7dB higher, but the rainfall overestimate would be still over 1dB. Physically, it is difficult to justify using a value of a_1 which is not unity, unless the heavier rainfalls are systematically associated with higher values of μ. Earlier relationships based on Eqn 16 (Gorgucci et al., 1994; Chandrasekar and Bringi, 1988; Chandrasekar et al., 1990) used the linear drop shapes and the 'Ulbrich' range of N_0 and μ (Illingworth and Blackman, 2002) leading to curves about 1-2dB lower than the values of Bringi and Chandrasekar (2001) and a probable overestimate of the rainfall rate by up to a factor of two.

If we seek a value of R accurate to $\pm 25\%$ (1dB) then from Fig. 8 we note:

- *Accuracy of Z_{DR} estimate*
  A 25% accuracy in R implies that Z/R should be known to 1dB, which from the slope of the curves imposes a limit of 0.2dB for the accuracy of the Z_{DR} estimate for R > 10 mm hr^{-1}. For a rainfall rate of 3 mm hr^{-1} an accuracy of 0.1dB is needed. As we saw in Section 2.2 this is very difficult to achieve.
- *Calibration of Z*
  R is only as accurate as the value of Z, so a 25% error in R implies that Z must be calibrated to 1dB. This is a serious constraint and will be discussed further in Section 6.
- *Sensitivity to drop shapes*
  The correct choice of drop shapes is crucial. The 'new' drop shapes will lead to rainfall rates from 1 or 3dB lower than the traditional 'linear' shapes.

Using the Andsager shapes (Eqn 6) rather than Goddard shapes (Eqn 5) changes the values of Z/R by less than 0.5dB.

- *Theoretical limit to accuracy of R estimate*
  Supposing that the drop shapes are known, then the naturally occurring variability of μ limits the fundamental accuracy of the technique. If the mean value of μ in rain is 5, but varies between 2 and 10, then this will introduce an uncertainty in the R estimates of about 0.5dB or $\pm 12\%$. A greater variability of μ with values down to 0 implies an error of 1dB or 25%.
- *Sensitivity to truncation of drop spectrum*
  A value of $\mu = 5$ in rain introduces a natural truncation of the drop spectrum so that the curves in Fig. 8 are insensitive to the choice of maximum drop size providing it is larger than 8mm.

5.3 $R = f(K_{DP})$

The potential advantages of deriving R from K_{DP} were summarised in Section 2.4 tempered by remarks on the noisy character of observed values of K_{DP}. Plots of the values of K_{DP} (one-way) as a function of R at S-band for the 'linear' and 'new' drop shape models using the normalised gamma function with $\mu = 0$ and 5 with constant $N_w = 8000$ m^{-3} mm^{-1} are displayed in Fig.9 and demonstrate:

- *Sensitivity to drop shape model.*
  For a value of K_{DP} of 0.55° km^{-1}, the linear drop shapes with $\mu = 0$ predict a rainfall of about 20 mm hr^{-1} rather than about 25 and 30 mm hr^{-1} for the 'new' drop shapes with $\mu = 0$ and 5, respectively. This may account for the persistent underestimates of rainfall using K_{DP} with these drop shapes (e.g. most recently, Brandes et al. (2001)) from the widespread use of the Eqn:
  $$R = 40.56 {K_{DP}}^{0.866} \tag{17}$$
  Bringi and Chandrasekar (2001) suggest using a coefficient of 50.7 rather than 40.56 is more appropriate for the 'new' drop shapes.
- *Increased non linearity and sensitivity to drop concentration and μ*
  We have argued that $\mu = 5$ is more appropriate than $\mu = 0$; over the range 10-100 mm hr^{-1} with the new drop shapes at S-band (9.75cm) this yields:
  $$R = 50.1 {K_{DP}}^{0.7} \quad \text{or} \quad \mathrm{K_{DP}} = 0.00417 \mathrm{R}^{1.4} \tag{18}$$
  Note that the index for R of 1.4 in Eqn 18 is almost as large as the value of 1.5 derived for a $Z(R)$ relationship in Section 3.4 and identical to the index for the NEXRAD default Z(R) relationship ($Z = 300R^{1.4}$) quoted in Brandes et al. (1999). This suggests that apart from the immunity to hail, many of the advantages claimed for the $R(K_{DP})$ approach may be difficult to realise. Increases of N_w of a factor of ten will lead to the inferred rainfall rate being underestimated by a factor of two for all values of R. Changing μ from 5 to 0 reduces R by about 30% for values of R of about 20 mm hr^{-1}.

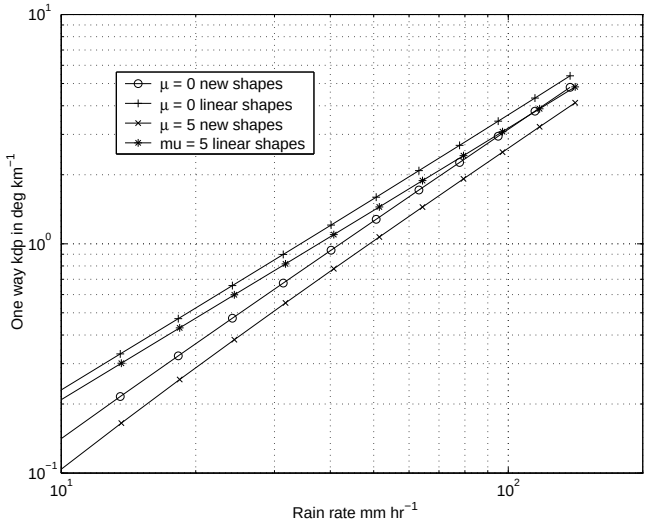

Fig. 9. Plot of R against one-way K_{DP} at S-band for the linear and new drop shape models with $N_w = 8000$ m^{-3} mm^{-1} and $\mu = 0$ and 5.

The $R(K_{DP})$ dependence at C-band is similar to that shown in Fig.9 but with the values of K_{DP} approximately scaled by the increased frequency. The appropriate equation for the 'new' drop shapes and $\mu = 5$ at C-band (5.35cm) is:

$$R = 31.4{K_{DP}}^{0.7} \quad or \quad K_{DP} = 0.00802R^{1.4} \tag{19}$$

Matrosov et al. (1999) present differential phase observations at X-band (3.2cm) and suggest an equation similar to Eqn 19 for rain rates up to 15 mm hr^{-1} with a coefficient of 14.2 and an exponent of 0.85 for the linear shapes in Eqn 4 with β of 0.062 mm^{-1}, but found better agreement with rain gauges if they used much less oblate drops with a rather unrealistic β of 0.044 mm^{-1} which gave a coefficient of 20.5. For these rain rates they calculated that the maximum value of δ was an insignificant 1°. Although ϕ_{DP} is larger at X-band, in heavy rain the bigger drops will Mie scatter, so δ will be larger and the correlation lower; both effects lead to an increasingly noisy ϕ_{DP}.

Thus far we have only discussed the form of the $R(K_{DP})$ dependence. The next stage is to consider the error in the derived values of R, σ_R, resulting from the error in the differential phase estimate σ_ϕ which can in theory be as low as 1° but in practice is at least 3°. The value of the one-way K_{DP} is derived by a least squares fit through the N gates each of length $\Delta(r)$ to give a range resolution of $L = N\Delta(r)$ and has an error given by:

$$\sigma_{KDP} = \frac{\sigma_\phi}{L}\sqrt{\frac{3}{N-(1/N)}} \tag{20}$$

The errors in K_{DP} predicted from this Eqn appear quite encouraging (Bringi and Chandrasekar, 2001); with a gate length of 150m, a resolution of 2.4km, and σ_ϕ of 1 and 3° the value of σ_{KDP} from a fit over the 16 gates is 0.18 and 0.54° km^{-1}, respectively. However, Blackman and Illingworth (1997) extended the analysis to the context of an operational radar and derived the error in the rainfall rate, σ_R, with an azimuthal resolution L to match the range resolution, for a radar with a beamwidth of θ_b at a range r, and derived an expression:

$$\sigma_R = \frac{\sigma_\phi}{abR^{b-1}L^2}\sqrt{12\theta_b r\Delta(r)} \tag{21}$$

where θ_b is in radians, R is the rainfall rate and a and b are the constants in the expression $K_{DP} = aR^b$. This leads to the following errors in rainrate estimates:

- *Resolution of 2.4km, range of 25km, and σ_ϕ of 1°*
  An R of 10mm hr^{-1} can be estimated to 60% and R=50mm hr^{-1} to 6%.
- *For a more realistic σ_ϕ = 3° and range = 100km*
  The accuracy for R = 10 and 50 mm hr^{-1} is 361% and 40%.
- *Performance at C-band*
  At C-band the value of b is about half that at S-band, so these percentages would be reduced by a factor of two, provided low values of σ_ϕ can be achieved at C-band.
- *Performance at 4.8km resolution rather than 2.4km*
  The fractional errors are reduced by a factor of four, but at a range of 100km the S-band error at 10 mm hr^{-1} would be still 90% and 10% for 50mm hr^{-1}.

The remaining possibility is to multiply the dwell time by four which would halve the phase error, σ_ϕ and the rain rate error, σ_R, but lead to an unacceptably slow scanning rate. The use of the integrated differential phase shift, Φ_{DP}, along a path length of 30km (Ryzhkov et al., 2000) and 10km (Bringi et al., 2001a) and exploiting the assumed linearity of the $R(K_{DP})$ relationship to provide an integrated average rainfall over a catchment has been proposed but, for most applications, this approach leads to an unacceptable loss of spatial resolution. We are forced to conclude that the noisy characteristics of ϕ_{DP} mean that K_{DP} cannot provide sufficiently accurate rainfall rates at the scales of 2km required in an operational environment.

6 Improved Rainfall Rate Using Integrated Polarisation Parameters

In this section we analyse what information can be retrieved from the three parameters Z, Z_{DR} and K_{DP}. First we look at their use in an operational environment and conclude that their accuracy is probably insufficient to provide an improved rainfall rate using all three parameters on the 2km scale, but that their integrated properties can provide a valuable constraint. We then consider how the non-independence of the three parameters can be used to provide an auto-calibration of Z to within 0.5dB every time there is heavy rain. Next we

discuss an alternative approach whereby the radar is assumed calibrated but the non-independence of the parameters is used to derive a drop shape model. Finally we consider the 'ZPHI' technique, whereby the value of the integrated phase shift is used to fix the value of 'a' in the $Z = aR^b$ relationship.

6.1 R from Z, Z_{DR} and K_{DP}

In the previous section deriving R from Z and Z_{DR} or from K_{DP} were discussed. A natural progression would seem to be to use all three parameters, Z, Z_{DR} and K_{DP} from which one might derive the three parameters of the normalised gamma function, N_w, D_0 and μ and hence an improved rainfall rate. Several authors have published equations of the form:

$$R(K_{DP}, Z_{DR}) = cK_{DP}{}^a Z_{DR}{}^b \tag{22}$$

with the values of a, b, and c obtained by a regression analysis of values obtained by scanning over the 'Ulbrich' values of N_0, D_0 and μ. Bringi and Chandrasekar (2001) provide a table of the coefficients in Eqn 22 with, for example, at S-band c= 90.8, a = 0.89 and b=-1.69 (for Z_{DR} in linear units). Four comments may be made concerning this approach:

- *Physical arguments would lead us to expect that* $a = 1$
  For a given value of Z_{DR}, both R and K_{DP} should scale linearly with N_w.
- *The Z information is not used at all*
  Z can be estimated more accurately than either Z_{DR} or K_{DP}, but of course Z must be calibrated whereas Z_{DR} and K_{DP} are 'self-calibrating'.
- *Insensitivity to hail*
  Although K_{DP} is very noisy, it has the unique advantage that it is insensitive to hail, but in the above equation this advantage is lost. Hail will lead to a reduced value of Z_{DR} and an erroneous inferred rainfall rate.
- *Non-independence of Z, K_{DP} and Z_{DR} in rainfall*
  Once Z and Z_{DR} are known then the value of K_{DP} can be derived.

6.2 Auto-calibration of Z using Polarisation Redundancy.

Calibration of Z is sometimes neglected but is crucial if accurate rainfall rates are to be derived from Z alone or from both Z and Z_{DR}. Traditional methods rely on characterising the link-budget or comparisons with rain gauges but are often only good to a factor of two. Goddard et al. (1994) showed (Fig 10) that for rain K_{DP}/Z_H is a unique function of Z_{DR} which is virtually independent of μ and proposed that this redundancy could be used to provide an automatic calibration of Z to within 0.5dB each time there was moderately heavy rainfall. The technique is as follows: The observed value of Z and Z_{DR} at each gate along a ray are used to predict the value of K_{DP} at that gate and so the predicted phase shift at each gate can be calculated and added up to give the theoretical total phase shift along the ray, Φ_{DP}. This total phase shift can be compared

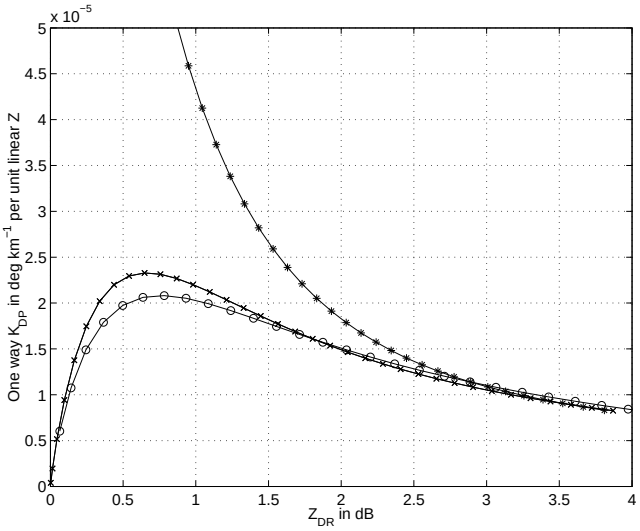

Fig. 10. Autocalibration technique for Z at S-band. Values of one-way K_{DP}/Z for the normalised gamma function as a function of Z_{DR}. 'New' shapes: 'o' $\mu = 0$: 'x', $\mu = 5$. 'Linear' shapes: '$\star$' $\mu = 5$.

with the observed total phase shift and the value of Z scaled until the computed value agrees with the observed value as illustrated in Fig. 11. The technique has the following advantages:

- *The total phase shift is a robust observation*
  The technique avoids deriving K_{DP} by differentiating an observed noisy ϕ_{DP} profile to give an even noisier gradient, but instead uses the two almost constant values of ϕ_{DP} before and after the heavy rain..
- *Calibration accuracy of 10% or 0.5dB*
  This can be achieved if an observed Φ_{DP} of 10° can be estimated to 1°accuracy. The calibration can be repeated for many rays so that any anomalous rays with very noisy Z_{DR} or ϕ_{DP} can be recognised and rejected.
- *Insensitivity to errors in Z_{DR}*
  Most of the phase shift occurs with Z_{DR} of between 1 and 2dB, so from the slope of the curve in Fig 10 we only need estimate Z_{DR} to about 0.4dB to achieve the required calibration accuracy of 0.5dB.
- *The drop shape model provides a limit to the attainable accuracy*
  The use of the 'linear' drop shapes rather than the new ones changes the calibration typically by about 1dB (see Fig 10).
- *The effect of attenuation*
  Le Bouar et al. (2001) express concern over attenuation of Z and Z_{DR}. The technique does indeed fail at X-band, but if Φ_{DP} at C-band is limited to 10°, then the Z attenuation is less than 0.5dB and does not pose a problem.

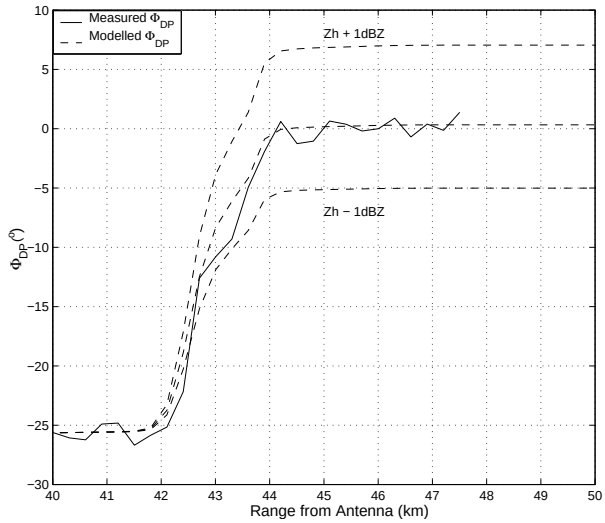

Fig. 11. Example of autocalibration. Different calibrations of Z lead to thee different traces for Φ_{DP}. Comparison with the observed Φ_{DP} in rain fixes the calibration of Z to 0.5dB (10%).The initial value of ϕ_{DP} is an arbitrary system offset.

Gorgucci et al. (1999b) propose a rather similar calibration technique based upon examination of the consistency between the two polarimetric rain-rate estimates, one using Z and Z_{DR} the other using K_{DP}. Because this method involves deriving K_{DP} directly from the observations, the noisy value of K_{DP} introduces considerable scatter into the data, making the comparison rather difficult.

6.3 Use of Polarisation Redundancy to Derive Drop Shape Model

A rather different approach in exploiting the non-independence of the polarisation variables in rainfall is to assume that the radar is independently calibrated but allow the raindrop shape versus size relationship to be a free variable which is fixed by the observations. Gorgucci et al. (2000) consider the raindrop spectrum to be described as a gamma function and calculate values of Z, Z_{DR}, and K_{DP} by cycling randomly over the 'Ulbrich' (Secn 3.3) range of N_0, D_0 and μ to be expected and also allow a variation of the values of the slope, β, in Eqn 4 for the linear model of drop shape. A non-linear regression analysis yields:

$$\beta = 2.37\ {Z_H}^{-0.377}\ {K_{DP}}^{0.396}\ 10^{0.093 Z_{DR}} \tag{23}$$

They suggest the β can be estimated to 10%. Inferred values in one storm range from 0.04 to 0.08, with average values falling from 0.061 to 0.053 for the Z ranges 40-45dBZ to above 53dBZ which they ascribe to raindrop oscillations. Gorgucci et al. (2001) use this approach and report rainfall estimates which are immune

to changes in the raindrop shape-size relation. Recently, Matrosov et al. (2002) report that at X-band the use of a variable β leads to improved rainfall rates than from K_{DP} alone. The immunity to drop shape models provided by this approach is attractive, but the following caveats apply:

- *Apparent change of β with rainfall rate.*
  Eqn 23 is virtually identical to the Goddard et al. (1994)) calibration curve; the indices of Z_H and K_{DP} are nearly equal and opposite, so for a given β the Eqn reduces to a single curve of Z_H/K_{DP} against Z_{DR}. Rather than use a non-linear regression, the values of Z_H, K_{DP} and Z_{DR} as D_0 varies have been calculated for a normalised gamma function with $\mu = 5$, and the apparent value of β in Eqn 23 plotted out in Fig. 12 as a function of Z_{DR} for the 'new' drop shapes. From Fig. 12 we now see that the apparent fall of β in heavy (high Z_{DR}) rain arises naturally from the fall in the K_{DP}/Z ratio with increasing Z_{DR} in the calibration curve Fig. 10, rather than appealing to drop oscillations (Gorgucci et al., 2000).

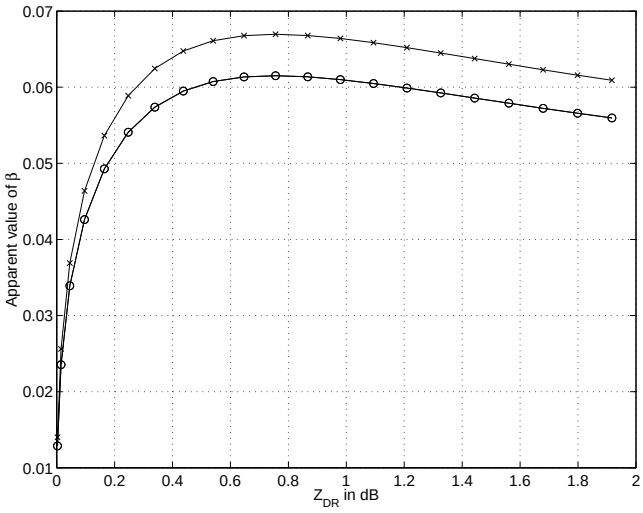

Fig. 12. The apparent value of β as a function of Z_{DR} if the rain drop spectrum is described as a normalised gamma function with $\mu = 5$ (lower curve (o)). In the heavier rain D_0, and hence Z_{DR}, increases and leads to an apparent fall in β. The upper curve (x) shows the change in apparent β if the Z calibration changes by 1dB.

- *Sensitivity to calibration.*
  Because the redundancy is being used to fix β it is assumed that Z is perfectly calibrated by some independent means. An small error in Z calibration of 1dB, will change β by about 10% (Fig 12).
- *Drop shape model.*
  The use of a single free variable β may not be sufficient to define the drop

shape model; the intercept is also a variable (see Fig 6) and the curve is not necessarily a straight line.

- *Sensitivity to noisy K_{DP}.*
  This approach involves the implicit computation of K_{DP} from the observations with its associated uncertainty, rather than the approaches above which use the more accurate integrated phase shift.

6.4 The ZPHI technique

The 'ZPHI' technique is described by Testud et al. (2000) and encouraging results at C-band in heavy tropical rainfall are reported by Le Bouar et al. (2001). This method uses the total phase shift, Φ_{DP} recorded along a path along which Z is observed to provide a constraint and fix the value of 'a' in the $Z = aR^b$ relationship. We now describe the principle of the technique at non-attenuating wavelengths such as S-band. At shorter wavelengths, C-band and below, attenuation can be appreciable, and the full technique which also uses the phase shift to correct for the attenuation is described in Testud et al. (2003, this book). In Section 3.4 it was shown that if naturally occurring raindrop spectra can be described by normalised gamma function then the value of b should be 1.5. Recall Eqn 12:

$$\frac{Z}{N_w} = H(\mu)\left(\frac{R}{N_w}\right)^{7.67/4.67} = H(\mu)\left(\frac{R}{N_w}\right)^{b}. \tag{24}$$

Fig. 13 shows that, for a normalised gamma function, a log-log plot of K_{DP}/N_w against Z/N_w with varying D_0 is also very nearly a straight line which is only weakly dependent upon μ. Thus, with we may write:

$$K_{DP}/N_w = f(Z/N_w)^g \quad \text{or} \quad \mathrm{K_{DP}} = \mathrm{f{N_w}^{1-g}Z^g} \tag{25}$$

where f and g are constants. Integrating along the path, r, we have:

$$\Phi_{DP} = f{N_W}^{1-g}\int Z^g dr \tag{26}$$

Accordingly, since f and g are known (Fig. 13) the value of N_w in Eqn 26 can be derived from the observed total phase shift, Φ_{DP} and the path integral of the reflectivity raised to the power 'g'; the value of N_w then fixes 'a' in $Z = aR^{1.5}$. Le Boar et al. (2001) report much improved rainfall rates at C-band, both as a result of the attenuation correction and from using the appropriate values of N_w. The technique has the following characteristics:

- *The total phase shift is a robust observation*
  As for the calibration method described above, the technique avoids the use of a noisy value of K_{DP} obtained by differentiating the ϕ_{DP} phase profile, but instead reduces the noise by integration. Note the sensitivity to the drop shape model. For the new drop shapes, when Z_H/N_W is below 1 (e.g. $Z =$ 39dBZ and $N_W = 8000$ m^{-3} mm^{-1}), g increases and approaches unity, but this region does not correspond to the regions of heavy rainfall with large phase shifts.

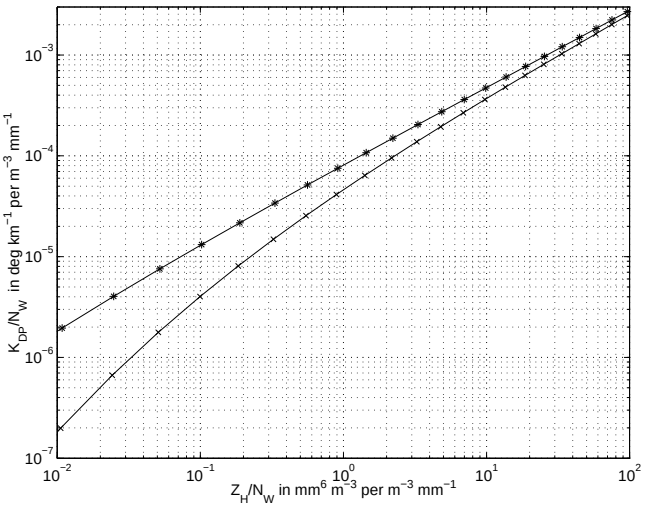

Fig. 13. K_{DP}/N_W for rain at S-band as a function of Z_H/N_W for a normalised gamma function with μ=5: 'x' - New shapes. '$\star$' - Linear shapes. The ZPHI technique relies on this curve being a straight line with constant slope 'g' which is not unity.

- *Need for accurate calibration of Z*
  Any calibration error in Z will directly affect the derived value of N_w and the value of 'a' in the $Z = aR^b$ relationship. Le Bouar et al. (2001) suggest that the method can be used to calibrate Z, based on the 'climatological' stability of N_w. They also propose a calibration technique which relies on the consistency of deriving rainfall rate from 'ZPHI' and from an attenuation corrected Z_{DR} and the inferred attenuation coefficient; in other words the method exploits the integrated phase information, Z_{DR} and Z, and is equivalent to the calibration techniques in Section 6.2 in another guise.
- *Sensitivity and operation in moderate rain rates*
  At C-band Φ_{DP} must be above 6°at C-band if R is to be estimated to 32% and above 32° for 21% error. At this frequency 10 mm hr^{-1} has a one way K_{DP} of 0.2° km^{-1} which would need to extend over 15km to produce a total differential phase shift of 6°. For 50mm hr^{-1} the distance would be 2km, and for 3mm hr^{-1} 75km. All distances need to be multiplied by 5 to reduce the rainfall error to 21%.
- *Constraint of constant N_w along the path*
  This is quite restrictive, but the path may be divided into segments provided the total differential phase change per segment is at least 6°.
- *Interference from hail*
  A means of recognising hail is needed, otherwise hail will increase Z but not affect Φ_{DP} and so lead to errors in inferred values of N_w.

The ZPHI technique is clearly a very powerful constraint for providing accurate rainfall rates in heavy rain. It is particularly applicable to the tropics where

hail is most unlikely. Based on the results of Matrosov et al. (1999) there is scope for extending the ZPHI technique to X-band for moderate rain rates.

7 Improved Rainfall Rates when Ice may be Present

In this Section we will consider two situations when the radar beam may be scanning a region containing icy hydrometeors. Firstly, in the summer time hail in convective clouds can occur at lower levels where the temperature is above freezing. Secondly, in the winter time and at longer ranges in all seasons the beam will be dwelling regions above the freezing level and can expect to encounter ice. In many papers only the first situation is addressed (e.g. Brandes et al., 1999) but at 200km range even in the summer the lowest elevation beam of a NEXRAD radar will be sampling ice.

7.1 Estimating rainfall when hail may be present

Apart from the use of K_{DP} alone, the techniques described in Section 6 for estimating rainfall rely upon an independent means of identifying the presence of hail. Hail cannot be identified unambiguously from the observed values of Z and Z_{DR}, but its presence will raise Z and depress Z_{DR}. The calibration technique of Goddard et al. (1994) relies on the consistency of the observed values of Z, Z_{DR} at each gate and the total observed phase shift Φ_{DP} along each ray within rain. Smyth et al. (1999) took this argument one stage further, suggesting that the best technique is to identify hail is to continually monitor these three variables along each ray, and use the failure of the consistency as an indication of the presence of hail. When hail is present then the only approach is to use one of the $R(K_{DP})$ equations to estimate the rainfall.

7.2 Estimating rainfall when the beam is at or above the freezing level

Essentially we are dealing with the problem of correcting for the vertical profile of reflectivity. This is a major operational problem and is dealt with in this book by Germann and Joss for mountainous regions and by Koistinen et al. for a cold climate. Can polarisation help in this problem? We must distinguish between the stratiform case, when there is an enhanced reflectivity, ΔZ, of up to 10dB to 13dB associated with the melting of low density snowflakes and a rapid fall off of reflectivity in the ice above the bright band, and the convective case, when the higher density graupel particles are not associated with a bright band and the fall of reflectivity with height is less pronounced.

The bright band can easily be identified by its very obvious polarimetric signature; the values of LDR are generally in the range -14 to -18dB and far higher than that of other precipitation particles. The enhanced value of Z in such regions, ΔZ is about 10dB higher than in the rain below. Melting graupel is associated with lower values of LDR of about -22 to -26dB and values of

Z similar to that in the rain below (ΔZ=0dB). This suggests that the value of LDR measured at the height of the 0° isotherm (obtained with an accuracy of 1° or so from an operational model) can be used as the basis of a bright band correction system. Before this is done the following aspects must be addressed:

- *Is there a reliable relationship between the LDR value and ΔZ?*
  How reliable is it and what is the quantitative performance of a Z correction scheme based on it?
- *Effect of range on the observed bright band*
  The bright band is typically about 600m deep and so at larger ranges will not fill the beam of a 1° radar. This means that the observed ΔZ will be less. LDR should also be lower, but how well related are the two variables?
- *Absence of a bright band in convection*
  How reliable is the value of LDR which indicates the presence of graupel, and the absence of a bright band?
- *Propagation effects at C-band*
  As the beam propagates through regions of finite LDR, the targets depolarise the incident beam, and the apparent value of LDR rises with range. The effect is negligible at S-band but needs to be quantified at C-band. Is a correction scheme needed at C-band and can one be developed?

Deriving rainfall at the ground once the radar beam is dwelling only in the dry ice above the melting layer is a challenging problem which is very important for operational radars at most ranges in the cold season and for cold climates as extensively discussed in this book by Koistinen et al. The snow occurring at these heights does not have any very marked polarisation characteristics which can be exploited.

8 Correction for Attenuation

Gate by gate correction schemes are notoriously unstable (Hitschfeld and Borden, 1978) and very sensitive to small calibration errors (Hildebrand, 1978). Differential phase shift provides a very powerful technique to correct for attenuation. In heavy rain attenuation at C-band is a severe problem and affects both Z and Z_{DR}. For example, at 40 mm hr^{-1}, the attenuation, A_H, is about 0.5 dB km^{-1} and the differential attenuation, A_{DP}, is about 0.1dB km^{-1}. In the literature there are many expressions of the form

$$A_H = \alpha \, {K_{DP}}^b \tag{27}$$

and

$$A_{DP} = \beta {K_{DP}}^b. \tag{28}$$

In the Rayleigh limit A_H is proportional to LWC (liquid water content) and for drop shapes which vary linearly with size, K_{DP} is proportional to the product of LWC and D_0, so at S-Band the value of 'b' in Eqn 27 is about 0.84. At C-band, the larger drops Mie scatter and absorb more, so that Eqn 27 is very

nearly linear. This suggests that correction for attenuation using K_{DP} or Φ_{DP} at C-band should be reasonably straightforward. In practice the following factors need to be considered.

- *Temperature effects.*
  Attenuation depends upon the imaginary part of the dielectric constant which changes with temperature, but K_{DP} is a function of the real part which is constant. Attenuation is about twice as large at 0°C than at 20°C. Operational models provide a reliable temperature structure so, in principle, this effect can be corrected for.
- *The precise value of α and β is a matter of some dispute.*
  Bringi et al. (1990) quote 0.054 and 0.0157 dB $^{\circ}C^{-1}$, respectively, at 15°C, but Carey et al. (2000) find that reported values range over a factor of two.
- *Variation with D_0.*
  Changes in the value of D_0, and hence the degree of Mie scattering, are probably the cause of this spread (Smyth and Illingworth, 1998; Carey et al., 2000). One method of resolving this difficulty would be an iterative approach, whereby the attenuation-corrected value of Z_{DR} is used to estimate D_0 and hence the correct value of α and β.

Our aim is to correct attenuations or Z to 1dB and Z_{DR} to 0.2dB (or better), so clearly, a quite complex attenuation correction is needed in the heaviest rain. There are two alleviating factors. In heavy attenuating rain, the value of Φ_{DP} is large and it may be possible to use the 'ZPHI' method to correct for Z and avoid the more stringent requirement for correcting Z_{DR}. Smyth and Illingworth (1999) suggested using an additional constraint; the value of Z_{DR} on the far side of a storm where Z is low should be close to 0dB. Any negative values of Z_{DR} provide a path integrated constraint for the total path integrated value of A_{DP}. A_H (the total path attenuation) can then be derived from A_{DP} through a simple multiplicative factor, because both have the same temperature dependence. Bringi et al. (2001b) have combined these two constraints, Φ_{DP} and Z_{DR} on the far side of the rain cell, so that the uncertain values of α and β in Eqns 27 and 28 are no longer prescribed but are derived along with values of D_0; the results are remarkably good and the technique is confirmed by the agreement with disdrometer values of D_0.

9 Identification of Hydrometeors

Thus far we have discussed identification of hydrometeors in terms of hail and melting snow impacting upon rainfall estimation. Hydrometeor identification itself is important for identifying severe weather (hail), for evaluation of the representation of different hydrometeors both in NWP models and in cloud resolving models (CRMs) and ultimately for assimilating hydrometeor information into such models. Particle classification also leads to a deeper understanding of the processes involved in precipitation formation and growth, so that conceptual

and numerical models can be improved. This is especially important in understanding deep convective systems as discussed in this book by Meischner et al. Straka et al. (2000) provide a very thorough review of the range of values in Z, Z_{DR}, K_{DP}, LDR and ρ_{HV} expected for hail, graupel, rain, rain wet hail mixtures, and snow crystals and aggregates. For example, hail can be identified by its high value of Z which is accompanied by lower values of Z_{DR} and K_{DP} than would be expected for rain. Liu and Chandrasekar (2000) have extended this classification using a fuzzy logic scheme; the in-situ validation data are very encouraging.

One difficulty that remains is that of identifying different forms of ice crystals. Matrosov et al. (2001) analysed the variation of the depolarisation ratio with elevation angle in terms of the ice particle aspect ratio and found the use quasi-circular polarisation provided the best discrimination. In theory pristine crystals have high density and so should give distinctive polarisation signatures. However, from an extensive analysis of in-situ data in stratiform clouds, Korolev et al. (2000) concluded that pristine crystals were rare and that 85% of ice particles were of irregular form. The problem is that the occasional large low density aggregates, which look spherical to the radar, dominate the radar return and mask the returns from the smaller crystals. This is confirmed by Wolde and Vali (2001) who found that in theory different crystal habits could be distinguished using a 94GHz polarimetric cloud radar, but that for real clouds polarimetric signatures were found for only a very few per cent of observations. Ryzhkov et al. (1998) suggested that the IWC should be related to a function of K_{DP}/Z_{DR} which should be insensitive to particle shape. Results were encouraging for regions containing pristine crystals, but the difficulty is that in the more common regions of aggregates, the values of the parameters are low and noisy and the IWC function is ill-defined. There is some scope for identifying ice particles by their signatures when they melt; for example, pristine plates will give very high Z_{DR} when they get wet, but after falling 100m or so will melt completely. We conclude that the polarimetric radar properties can be used to identify the larger hydrometeors but are usually unable to differentiate between the smaller types of ice particles.

10 Conclusion

This review has identified some powerful techniques based upon polarimetric observations. In the following discussion we shall assume that 'new' raindrop shapes of Eqn 5 are accurate, and that natural raindrop size spectra can be represented by a normalised gamma function with a mean shape factor of 5 and a range of 1 -10, and an average normalised concentration N_w of 8000 m^{-3} mm^{-1} with a range of up to a factor of ten both larger and smaller. This range leads to a relationship of the form $Z = aR^{1.5}$ with 'a' proportional to $1/\sqrt{N_w}$ and errors of up to a factor of two in rain rate. If we are to estimate rainfall rates to better than 25%, then it may be necessary to account for the increased terminal velocities ($\approx$ 10% at 2km) aloft (Matrosov et al., 2002).

- *Ground clutter and anomalous propagation*
  They can be recognised at each gate using polarimetric parameters. Particularly powerful is the use of the copolar correlation coefficient as a data quality check.
- *Rain rates from Z and Z_{DR}*
  Z should scale with R as for a given Z_{DR} (or D_0) as N_w changes; this leads to a formula of the form $Z/R = f(Z_{DR})$ given by Eqn 14 at S-band and 15 at C-band. For rainrates to be accurate to 25%, Z must be calibrated to better than 25% (1dB). Z_{DR} must be accurate to 0.2dB for $R > 10$mm hr^{-1} and 0.1dB for R > 3 mm hr^{-1}. Such Z_{DR} accuracies are difficult to achieve in practice on a scale of 2km.
- *Rain rate from K_{DP}*
  This technique is attractive because of its immunity to both calibration errors and to attenuation and to its insensitivity to hail. However, the exponent in $K_{DP} = aR^b$ is about 1.4 and not very different from those in $Z - R$ relations, so the R derived from K_{DP} has almost the same sensitivity to changes in raindrop spectra as R derived from Z, and the scope for using path integrated phase shift to infer rainfall over a catchment is limited. In addition, the differential phase measurement is inherently noisy and in practice is also extremely sensitive to small amounts of ground clutter or mismatched sidelobe signals. As a result the R estimates from K_{DP} are not accurate enough for operational work at 2km resolution unless hail is present.
- *Use of integrated polarisation parameters*
  We conclude that at the 2km scale needed for an operational system, the additional information from Z_{DR} and K_{DP} is insufficiently accurate to improve rainfall estimates, but that the use of integrated polarisation parameters can provide a valuable constraint.
- *Auto-calibration of Z and identification of hail*
  Z can be accurately calibrated to 0.5dB (10%) by exploiting the fact that Z_{DR} and K_{DP} are not independent in rain, but that K_{DP}/Z is a unique function of Z_{DR}. The technique is to compute the predicted total theoretical differential phase shift along a ray (Φ_{DP}), by integrating the value of K_{DP} predicted from the observed values of Z and Z_{DR} at each gate, and then scaling the Z values so that the predicted value of Φ_{DP} agrees with the observed one. This technique avoids using the noisy observed K_{DP}. The presence of hail can be inferred from the failure of the redundancy exploited in the calibration technique.
- *Drops shape constraint from polarisation redundancy?*
  An alternative approach is to exploit this redundancy by choosing the value of β in the drop shape (Eqn 4) so that the observed Z, Z_{DR} and K_{DP} are consistent, but the accuracy is limited because this approach uses the very noisy observed values of K_{DP}. It seems more sensible to accept the extensive measurements of drop shape and use the redundancy to fix the Z calibration.
- *The 'ZPHI' technique to fix 'a' in $Z = aR^{1.5}$ in heavy rain*
  The observed path integrated differential phase shift, Φ_{DP}, together with the integrated value of Z along the ray can be used in the 'ZPHI' technique

to provide a constraint to N_w and hence an estimate of 'a' in $Z = aR^{1.5}$ provided there is no hail present. The technique is limited to rather heavy rainfall because Φ_{DP} at C-band must be at least 32° if the error in R is to reduced to 21%. Increased phase shifts at X-band may be counteracted by the extra noise introduced by Mie scattering effects, such as differential phase shift on backscatter and a lowering of the co-polar correlation coefficient.

- *Use of ϕ_{DP} to correct Z and Z_{DR} for attenuation*
  ϕ_{DP} is nearly linearly related to attenuation at C-band but the constant changes with T and D_0. Correction of attenuation using the 'ZPHI' technique is discussed by Testud et al. (2003, this book). Further improvements to the attenuation correction may be possible when the 'ZPHI' Φ_{DP} constraint is combined with the requirement that the total differential attenuation is consistent with the values of Z_{DR} expected on the far side of a heavy rain echo. There is scope for using the ZPHI method at X-band (3.2cm) provided the rainfall is not too heavy so there is total loss of signal.
- *Integrated Z/Z_{DR} technique in regions of low rainfall*
  In regions of moderate or light rain where the phase shifts are too low for the ZPHI technique, the average value of N_w over a domain can be inferred from the many estimates of N_w from Z and Z_{DR} made at each gate. The random errors associated with noisy Z_{DR} should average to zero, and providing Z is accurately calibrated, the mean value of N_w can be estimated, and hence accurate (25%) rainfall estimates obtained down to perhaps $R = 3$ mm hr^{-1}, but this still remains to be demonstrated.
- *Hydrometeor Identification*
  Combined polarisation parameters can be used to distinguish the different types of the larger hydrometeors and this should be useful in evaluating parameterisation schemes in NWP (Numerical Weather Prediction) models and the performance of CRMs (Cloud Resolving Models), and may ultimately be used for assimilation into such models (see Macpherson et al., 2003, this book).

To summarise, we find that polarisation radar has the potential to remove most of the ambiguities which arise when Z alone is measured with a conventional radar. However, the difficulty of estimating rainfall rates at the ground at long ranges when the beam is dwelling in ice above the melting layer remains. In all these applications it is important to use the latest 'new' raindrop shapes. It may be that the use of the 'linear' drop shapes is responsible for some of the disappointing rainfall estimates reported in the literature. The requirement for pulse to pulse switching of polarisation can be avoided by the use of 'hybrid' (45°) transmission, but this approach will probably not reduce the noise of the differential phase measurement. Ground clutter and 'anaprop' can be unambiguously recognised, but generally, the parameters cannot be estimated accurately enough to contribute independent information to improve rainfall estimates on the operational scale of 2km. However, when the parameters are integrated over a domain, they can correct for attenuation in C-band systems, provide absolute calibration of Z to 10%, and reduce errors in inferred rainfall rates from

a factor of two down to about 25%, effectively by fixing the value of 'a' to be used in $Z = aR^{1.5}$ over a domain. Finally, to accompany these improved rainfall estimates the polarisation parameters can also provide an estimate of the error which is essential for use in data assimilation.

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