CARPE-DIEM
Report for Deliverable 8.2

Use of polarimetric measurements to estimate effects of variation in drop-size distributions on the uncertainty inherent in rainfall estimates collected at different spatial and temporal scales.

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The nature of uncertainty

The term "*data quality*" can be regarded as referring very broadly to the usefulness of a set of measurements when applied in the context of a specific application and for a particular purpose. In very general terms, an application will encapsulate some form of implicit or explicit model, in the sense that it is inevitably a gross simplification of a far more complex prototype system, intended only to represent a particular set of behaviours thought to be of relevance to some real-world problem. Simplification can normally be taken as implying that processes are represented only at a particular scale or set of scales, and this will therefore underpin any definition of what is meant by data quality.

By way of example, measurements of rainfall are often used as input to flood forecasting models, which in most instances are represented as a simple "black box", attenuating a higher frequency input (i.e. the average depth or intensity of rainfall) to a lower frequency output (i.e. rate of river flow at the base of the catchment). Since a "lumped" representation of the catchment hydrology is employed, the spatial and temporal scales at which the relevant processes are being represented correspond to the spatial extent of the catchment and to the time step used in iterating the model. An ideal rainfall observation would have sampling properties that exactly matched these scales, but in practice there will always be some mismatch between this and the characteristics of the sensor employed.

A rain gauge measures the depth of rainfall at a point, but this is unlikely to be representative of rainfall over the catchment as a whole, and will become less representative for larger catchments. Similarly an instantaneous observation such as is obtained from radar, may not be representative of the average rainfall in the intervening period between measurements. Underlying this, and the reason why observations are required in the first place, is the fact that the magnitude of the atmospheric fluxes which we regard as constituting weather will vary across temporal and spatial scales. Clearly, a point measurement is less likely to be representative of rainfall over the catchment as a whole, if the intensity of rainfall is itself highly variable in space.

Regardless of any errors specific to the method or instrument used in making an observation, if the observation is not representative of the input actually required by the model, an error will result, in the sense that the output will be different to that which would be obtained in the case of an observation with ideal sampling characteristics. For reasons which will be obvious, this is sometimes described as a "representativeness error" and arises from the interaction between the sampling properties of the sensor, the spatial and temporal scales being represented in the application and the underlying variability of the phenomena being observed.

In general, the instruments used in remote sensing provide coverage which is both limited in extent and of variable quality. For surface-based observing systems this most clearly refers to spatial extent, with the quality or relevance of the data tending to decrease with increasing distance from the sensor. In the specific case of weather radar, this degradation tends to associated with factors such as reduced sensitivity, increasing beam height and a larger pulse volume. However, a similar degradation in data quality (or relevance) occurs where time is used as the reference frame, with the representativeness of an instantaneous observation tending to decline as the separation between the time when the measurement is made and the time when it is applied increases. An extreme example is in the case of non-geosynchronous satellite observations, where coverage may only be available at intervals of hours or even days, but similar considerations apply even where the interval between measurements is of the order of minutes.

Clearly, this is only a problem because what is being measured is non-stationary, this being the reason why observations are required in the first place. The atmospheric fluxes which we describe as weather vary in magnitude across different spatial and temporal scales, and so a definition of the scale of the processes we are trying to represent in any application will underpin what we regard as defining data quality. Naturally, this also applies with reference to time, with the relevance of an observation made at a particular instant decreasing away from that instant. Whereas in both cases this degradation in data quality for a single sensor might reasonably be approximated as a simple function of distance or time, for an integrated observing network, the coverage from each individual sensor is overlaid with multiple instruments providing information for the same location but of differing quality.

Radar Measurements of Precipitation

As with other remote sensing techniques, weather radar does not in itself provide a direct measurement of the quantity being sought. Instead, weather radar provides an estimate of the intensity of precipitation through some relation to the measured backscattered power. For a conventional, reflectivity based estimate, this relationship is given as

$$Z = \int [N(D_e)\, D_e^6] \quad \text{mm}^6\ \text{m}^{-3} \qquad (1)$$

, where D_e is the diameter of each droplet, $N(D_e)$ is the number of droplets per unit area in this size interval and Z is termed the reflectivity factor. It will be noted that Z is related to the sixth power of D_e, indicating a key dependence on the actual distribution of drop sizes within the sampled volume.

The size spectrum of droplets falling as naturally occurring rainfall is a result of collisions between individual droplets, and the balance between the likliehood that these collisions will result in a droplet growing through coalescence, or breaking up into a number of smaller droplets. The likliehood of breakup increases as a droplet grows in size, and this serves to limit the maximum size to which droplets can grow to around 8mm.

It is possible to attempt to parameterize these rates of coalescence and breakup for different droplet sizes, and arrive at what is termed an equilibrium drop size distribution. This is the relative concentration of different droplet sizes which will be arrived at, given sufficient time or fall depth over which the rates of coalescence and breakup can balance out. The results of this type of study suggest that, providing that an equilibrium has been reached, the relative concentrations of different droplet sizes should be the same regardless of rainfall intensity.

This is not in fact what is observed in practice, either in general or in specific instances, the key reason likely being that there is a finite fall time before rainfall will reach the ground. The time taken to reach an equilibrium is dependent on the rate at which collisions between individual droplets occur, and thus primarily on the initial concentration, or effectively, the rainfall intensity. Thus, as rainfall intensity increases, the relative number of larger droplets might be expected to gradually increase towards an equilibrium level, since the rates of coalescence and breakup will balance out more rapidly.

This is what is described in the classic, exponential model of the drop size distribution derived from the measurements of Marshal and Palmer (1954). This states that the shape of the drop size distribution will vary in response to rainfall intensity such that

$$N(D_e) = N_o \exp(-\Lambda D_e) \qquad \textbf{(2a)}$$
$$\Lambda = 4.1\ R^{-0.21}\ \mathrm{mm}^{-1} \qquad \textbf{(2b)}$$

, where N_o is a scaling parameter for concentration with the value 8000 m^{-3} mm^{-1}, and Λ, which describes the slope, is related to rainfall intensity (R, in $mmhr^{--1}$). Thus as rainfall intensity increases, the number of larger droplets will gradually increase relative to the number of smaller droplets. Assuming horizontal polarisation, and that droplets are falling at terminal velocity in still air, this gives the relation

$$Z = 200\ R^{1.6} \qquad \textbf{(3)}$$

, providing a practical basis for estimating precipitation intensity directly from the measured reflectivity factor.

Dependence on the Drop Size Distribution

Although widely used in practice, Eqn 3 is strongly dependent on Eqn. 2. Over the past few decades, extensive comparisons between the measured reflectivity and rainfall at the surface have provided a wide variety of alternative Z-R relations reported in the literature. Battan (1973) notably listed as many as 69 different Z-R relationships which had been reported in the literature to that date and as Rinehart (1991) wryly comments, many others have appeared since. Although Eqn 3. appears to represent a reasonable average case, these discrepancies cannot be explained entirely as a result of instrumental errors, and instead suggest that the Drop Size Distribution can on occasion vary markedly from that described by Marshal and Palmer. There are a variety of reasons why this might be the case.

Firstly, it should be noted that Eqn. 3 assumes that droplets are falling at terminal velocity, and that the actual rainfall intensity experienced at the ground will be altered in the presence of strong up or down draughts. In the presence of a strong up draught, such as on the leading edge of a thunderstorm, smaller droplets will be held aloft and only the larger droplets will fall out, altering the relation between the measured Z aloft and the rainfall intensity R at the ground. A similar discrepancy may occur where the precipitation generating mechanism is vertically stratified, an example being the “seeder-feeder” mechanism postulated for orographic rainfall. In this drops from a higher ‘seeder’ layer, fall through a lower ‘feeder’ layer where smaller droplets are condensing. In essence, the two drop size distributions are superimposed with the effect that smaller droplets are present at higher relative concentrations.

Secondly, a number of authors have pointed out that Marshal and Palmers original measurements relate to a maximum intensity of 25 $mmhr^{-1}$, and that particularly in the case of deep, tropical precipitation, it may be possible that equilibria is being reached. This would suggest that at some intensity, there would be a transition from the Marshal-Palmer type of DSD with D_o increasing with intensity and N_o remaining constant, to an equilibrium form of DSD where N_o increases with intensity and D_o instead remains constant (Hodson, 1986). In other words, the Z-R relationship observed might be expected to switch from an exponential to a linear form at higher rainfall intensities.

There is some evidence for this. Based on analysis of disdrometer data from sites at three widely differing latitudes, Sauvageot and Lacroux (1995) found evidence of a similar transition occurring for rainfall intensities in excess of 20 $mmhr^{-1}$. It was suggested that this point of transition was in part influenced by lattitude, tending to occur at higher intensities for shallower mid-latitude systems. In a more explicit examination of the problem, Hu and Srivastava (1995) compared the results of their simulation with measured spectra from heavy tropical rainfall, the expectation being that equilibrium was most likely to be reached at high concentrations and where the rainfall generating layer was relatively deep. They did find evidence of parallel, exponential spectra, but with a slope of 2 mm^{-1} rather than the 6.5 mm^{-1} predicted from their model. In conclusion, they felt that this was most likely a result of the Low and List formulae (Low and List, 1982) underestimating coalescence and overestimating droplet breakup, but noted that it could also be a consequence of averaging of measured drop spectra. Testud *et al* (2001) also report reasonable agreement between drop spectra measured by PMS probe, and predictions from equilibria models, but noted that surprisingly this applied equally to both convective and stratiform cases

Finally, it should be noted that in a series of recent articles (Jameson and Kostinski, 1998, 1999 and 2000; Kostinski and Jameson, 1997 and 1999; Jameson *et al.* 1999) it is argued that 'clustering' of similarly sized raindrops can occur at scales of metres or less. The starting point for this argument is the observation (also reported by Hosking and Stow, 1987 and Auf der Maur, 2001) that analysis of droplet arrival times shows a degree of temporal correlation that would not be present if the process were truly random. These correlations decay with time, indicating a period of coherence, which when compared across droplet size intervals suggest that individual forms of droplet spectra exist as quasi-physical entities. In other words, if we were seeking to average drop counts from an instrument such as a disdrometer in order to obtain a representative drop size distribution, there is a definite period over which this averaging should be performed. Averaging over a longer period would in effect be combining measurements from two or more raindrop clusters, and the net result would be the average of two or more different drop size distributions. Averaging over a shorter period would lead to the spectrum being dominated by random fluctuations, particularly at lower concentrations.

A possible explanation for this clustering effect comes from a new appreciation of the role that small-scale turbulence may play in promoting rainfall. It will be recalled that there are a number of shortcomings in currently accepted theories as to the microphysical origins of rainfall; most notably in explaining the initial broadening of the droplet spectrum following condensation and in explaining the rapidity with which the onset of falling precipitation occurs in nature. The idea that turbulence may play a role in promoting the onset of precipitation has some history (see for example, Arenburg, 1939, or East and Marshall, 1954), but it is only comparatively recently that the tools have become available to enable realistic attempts to parameterise these processes.

For example, while expressing reservations about the difficulties of interpreting the results from a wind tunnel experiment, Vohl et al (1999) suggest that droplet growth in turbulent rather than laminar flow appeared to be enhanced by 10 to 20%, this being sufficient to promote the rapid development of precipitation. Pinsky and Khain (1997, 2002) used numerical models to demonstrate that the effect of small-scale eddying motion is to create localised concentrations of droplets, this serving to greatly promote droplet interaction and coalescence. Pigeonneau and Feuillebois (2002) report the results of a similar study. An initial and localised broadening of droplet spectra, would not only accelerate growth to the point at which falling precipitation developed, but would also do so in a spatially inhomogeneous manner, creating pulses of heavier rainfall.

Thus, we have a difficulty in defining what is meant by drop size distribution; whether it is with regard to the physical grouping of droplets in certain proportions generally at smaller scales, or whether it is a statistical description of average concentration of droplets across some volume such as a radar pulse volume. In the first case, scale is largely determined by the microphysics of the precipitation process rather than by some external constraint, and it follows that this will also serve to determine the degree to which there is a real discrepancy between the two definitions. Given evidence of clustering at scales of metres, in most if not all cases the notional drop size distribution represented by the radar pulse volume will actually be the result of multiple and overlaid quasi-physical drop spectra.

Jameson and Kostinki (2001) go on to argue that whilst remaining eminently useful from a practical point of view, power-law relations between quantities such as Z and R arise from a purely statistical relationship, rather than from any physical causality. In other words, the points plotted between Z and R represent a family of separate linear relationships, and while it is possible to fit a regression line to these points, this does not imply any knowledge as to how N_o and D_o might be varying with rainfall intensity, since it is the net effect of many different variations which is being observed.

This thesis is certainly helpful in resolving the apparent discrepancies between equilibrium theory and observations (*n.b.* while not necessarily implying that equilibria spectra will exist in nature; the authors themselves believe that they do not), and does serve to highlight a fact which has perhaps been too readily accepted as an implicit assumption in much of the previous work, namely that we are dealing with averages over a volume which is not chosen to reflect the microphysics of rainfall, but rather is determined by external constraints. It also serves to suggest that the wilder variations in drop spectra parameters experienced using instruments such as disdrometers, may not occur to the same extent at the scales observable by radar.

Polarimetry and Droplet Shape Models

As well as being characterised by a phase and amplitude, an electromagnetic wave is characterised by a polarisation. This describes the oscillation of the wave amplitude in directions perpendicular to the direction of propagation. These oscillations are regular, and in the most general case the magnitude of these oscillations defines an ellipse. The terms linear and circular polarisation both refer to special instances of elliptical polarisation. In the case of linear polarisation, the wave oscillates in one plane only (i.e. the magnitude of oscillations in other directions is zero), and this could be horizontal or vertical to the direction of propagation (i.e. horizontal or vertical polarization), or could be some other angle such as +45^o or -45^o (sometimes referred to as slantwise polarisation). Polarimetric capability describes the capability of some modern weather radar systems to transmit and receive two or more polarizations. This is most usually at linear horizontal and vertical polarizations and is what is assumed here.

Practical utilization of polarimetric weather radar for quantitative precipitation measurement largely rests on the fact that as rain droplets become larger, they also tend to become less spherical and more oblate. This is a result of the increasing force exerted by the air through which a droplet is falling, which acts against the force being exerted by surface tension and which tends towards sphericity. The shape of the water droplet changes so that the horizontal axis tends to become elongated in relation to the vertical axes, this being described as an oblate spheroid. An early attempt to parameterize this relation was given by Prupacher and Pitter (1971), which suggests that on average

$$D_r = 1.030 - 0.062D_e \quad \textbf{(4)}$$

, where D_r is the ratio of the vertical to horizontal axes. D_e is the equivolumetric diameter (i.e. the diameter of a sphere having the same volume) and is assumed wherever a reference is made to droplet size.

More recent measurements such as Andsager *et al* (1999) suggest a slightly less linear relationship, with smaller droplets tending to be more spherical, with the effect that

$$D_r = 1.012 - 0.0144D_e - 0.013D_e^2 \quad \textbf{(5)}$$

, for D_e in the range 0.7mm to around 4.1mm. Examples of these model raindrop shapes are shown in Figure 1, though it should be emphasized that individual raindrops will exhibit a much wider range of shapes. This is particularly true for larger drop sizes, and although present in much lower concentrations, these larger droplets do tend to contribute disproportionately to radar measured parameters.

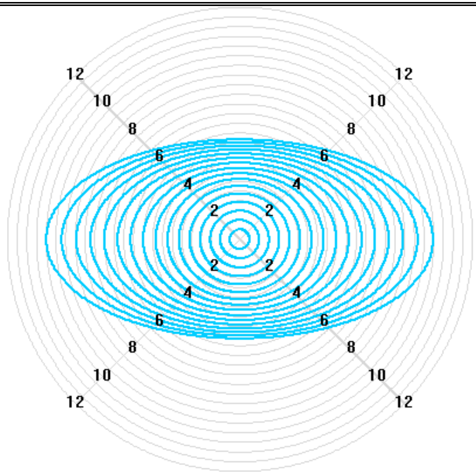


Fig. 1 : Representation of average raindrop shapes as a function of equivolumetric diameter (D_e) plotted for intervals of 0.5 mm to a maximum of 8 mm and using the drop shape model given in Andsager et al (1999). Average raindrop shape is approximated as an oblate spheroid (or ellipsoid), forming a flattened ellipse when viewed along the horizontal..

Polarimetric Estimates of Rainfall

While a drop size distribution can serve as a means to translate between a measured quantity such as reflectivity, and some desired quantity such as rainfall intensity, the accuracy of these estimates will obviously depend heavily on the accuracy of the drop size distribution on which they are based. A central problem is the fact that even in its simplest form, a drop size distribution must be specified in terms of at least two parameters: a concentration parameter (N_o) which relates to the total volume of water present; and a slope parameter (D_o) which describes the proportionality between different droplet sizes. Using only a single measured quantity, such as Z, it is only possible to deduce another quantity, such as R, by making an assumption that either N_o or D_o is constant, or that these two parameters are related in some linear fashion.

The usual practice, and the approach taken in applying the Marshal-Palmer relation, is to assume that N_o is constant, and that D_o varies with rainfall intensity. In other words, heavier rainfall is assumed to be predominantly a consequence of the proportion of larger droplets increasing. The opposite approach, discussed later in relation to equilibrium theory, is to hold D_o constant and assume that heavier rainfall is due to an increase in N_o. In other words, the total number of droplets increases in heavier rainfall, but different droplet sizes remain in the same proportions.

If it is reasonable to suppose that both N_o and D_o may vary in nature, then it naturally follows that in either of these cases, some error may result in applying a standard power-law relationship. Furthermore, the magnitude of this discrepancy will depend not only on degree to which the measured and actual droplet spectra differ, but also on the power relation between the two quantities. Whereas the rate of attenuation is largely a function of D_e^3 and thus is almost in direct proportion to R, the reflectivity factor Z is a function of D_e^6 and the magnitude of any errors will increase exponentially. Atlas and Chmela (1957) give an example from measured data, where the same value for Z could relate to a rainfall intensity of either 11 mm hr^{-1} or 33 mm hr^{-1}, a difference of 300% or more than 4.8 dB.

The possibility of using two measured quantities to better parameterise drop spectra has long been realised, dating back at least to Wexler and Atlas (1963). The basic concept that lies behind dual-measurement techniques is that, for any two parameter model of drop size distribution, it should be possible to derive values for both parameters from a set of two unrelated measurements. As a consequence, any derived quantities such as rainfall rate can be more accurately determined. Partly for technological reasons, early examples of dual-measurement techniques predominantly focused on the use of dual wavelengths; one strongly attenuating and the other weakly attenuating. A rate of attenuation could be derived by comparing Z at the two wavelengths, and this could then be used in combination with Z to obtain an improved estimate of rainfall intensity (Eccles and Mueller, 1973; Atlas and Ulbrich, 1974; Goldhirsh and Katz, 1974). However, while attractive in principle, a major problem with this approach is that using the same antennae, there will be a gross mismatch between the beam patterns at the two wavelengths, with the consequence that the sampled volumes are not identical (Rinehart and Tuttle, 1982).

Over the past two decades, attention has shifted more towards the use of dual-polarisations, rather than dual-wavelengths, to provide better specification of the drop size distribution. For the purpose of rainfall estimation, polarimetric techniques largely rely on the fact that drop shape is a function of size, with the consequence that differences between the signals at horizontal and vertical polarisation is related to the proportion of larger drop sizes. Although the practical benefits of using polarimetric techniques for quantitative rainfall estimation remain largely unproven, there is now sufficient confidence in the potential value of the qualitative information provided by polarimetry that this is likely to become increasingly viable as an operational method.

Although there are a variety of ways in which polarimetry can be used, and these touch upon some of techniques with which data quality problems can be detected and treated, attention will focus primarily on the two main methods for rainfall estimation. As the name should imply, differential reflectivity (Z_{DR}) is simply a measure of the ratio of the powers at the two orthogonal polarisations, and is therefore expressed in decibels. In contrast, specific differential propagation phase (K_{DP}) is a measure of the rate of change in propagation phase between the two polarisation along a given path, and is therefore expressed in deg km^{-1}. A related quantity in the same units is differential phase (ϕ_{DP}) which refers to the absolute difference in phase at fixed intervals

Differential Reflectivity (Z_{DR})

The use of measurements of differential reflectivity (Z_{DR}) at linear horizontal and vertical polarisations to quantify drop spectra variations was originally proposed by Seliga and Bringi (1976 and 1978). It will be recalled that as raindrops get larger they will tend to adopt an increasingly oblate shape, this resulting in a more marked difference in their backscatter cross-sections at horizontal and vertical polarisations. In heavier rainfall, where from the Marshall-Palmer relation, the proportion of larger droplets will be expected to increase, this might be expected to result in a difference in reflectivity between the two polarisations. Using the symbols Z_H and Z_V to denote reflectivity factor measured at horizontal and vertical polarisations respectively, the differential reflectivity (Z_{DR}) can be defined as

$$Z_{DR} = 10 \log_{10}\left(Z_H / Z_V \right) \text{ dB} \tag{6}$$

, or more simply, as the difference between Z_H and Z_V when reflectivity factor is expressed in dBZ. In the absence of other factors, light rainfall predominantly composed of small, spherical droplets would be expected to produce a Z_{DR} signal close to zero, whereas heavier rainfall containing larger, more oblate droplets would be expected to give a larger positive value for Z_{DR} .

Given the assumption of a two-parameter drop size distribution, and a certain relation between droplet shape and size, Z_{DR} is simply a function of the mean drop diameter D_o, and is independent of N_o. It follows that D_o can be estimated directly from Z_{DR} and that N_o can then be determined by dividing Z_H by D_o, thus allowing full specification of the drop size distribution, and in theory at least, providing a more accurate estimate of the rainfall intensity. However, some caution needs to be expressed. In particular it should be remembered that Z_H and Z_V arise through integration over a volume, and that even in intense rainfall, the larger droplets which contribute most to Z_{DR} are at relatively low concentrations. As a consequence, Z_{DR} will be less sensitive to R than might at first be assumed, particularly once allowance is made for a realistic level of measurement error.

Assuming a Marshal-Palmer drop size distribution, Figure 2 shows how Z_{DR} changes in relation to R using the four drop shape models described in the previous chapter. From this, it will be noted firstly that Z_{DR} will tend to be quite weak, with a value of less than 2 dB for rainfall intensities less than 128 mm hr^{-1} regardless of the drop shape model employed. One clear implication of this is that to be useful, Z_{DR} needs to be estimated with a high degree of precision and relatively free from noise. Since Z_{DR} is a relative measure, inaccuracies in the radar constant will not have any significance although account does need to be taken of any cross-channel biases. Quantisation errors will be significant where Z_{DR} is derived as a secondary quantity, since the 8-bit formats often used to store values for Z will give errors well in excess of the 0.1 to 0.3 dB quoted by Bringi *et al*. (1980).

Secondly, it is apparent that the choice of drop shape model does have a significant effect, though this is more pronounced at lower rainfall intensities. For example, a Z_{DR} of 1 dB would correspond to a rainfall intensity of 12 mm hr^{-1} for the original Prupacher and Pitter (1971) drop shape model, compared with 31 mm hr^{-1} for the latest Andsager et al (1999) drop shape model. The three more recent models all define smaller droplets as being more spherical, resulting in a weaker Z_{DR} signal for light to moderate rainfall, but respond more rapidly in heavier rainfall, with the result that all four models begin to converge at intensities of around 256 mm hr^{-1}, giving a Z_{DR} of around 2.3 dB.

Since Z_{DR} is independent of N_o, it can be expressed as a function of D_o and without reference to R, as shown in Figure 4.2. It should perhaps be noted that since the drop size distribution is truncated at a maximum droplet diameter of 8 mm, Z_{DR} will tend to level out at higher values of D_o, reaching a limit of around 4.3 dB at D_o = 8 mm, this roughly equating to the point at which different droplet sizes are assumed to be in equal proportions. Higher values for Z_{DR} would only be possible if it were assumed that the normal trend in an exponential drop size distribution were reversed, and that larger droplets were in fact present in higher concentrations. This does occur in nature, and an obvious situation which has been alluded to previously with regard to Z-R relations, is the situation where there is a strong vertical updraft, with only the larger droplets falling out.

Early measurements Z_{DR} of showed some promise, with values generally being in the expected range (Seliga and

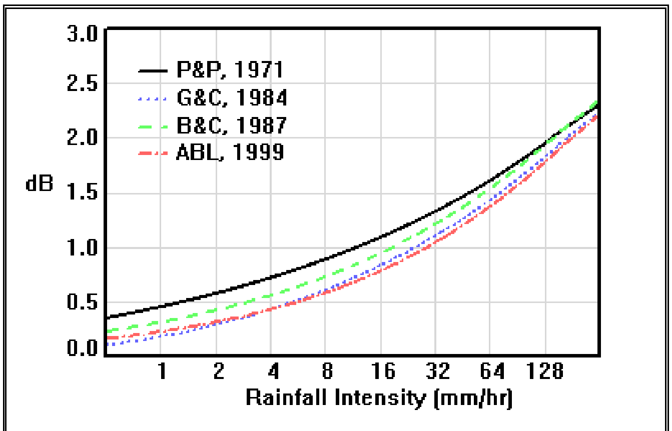


Fig. 2a : Z_{DR} plotted versus R assuming a Marshal-Palmer drop size distribution and using the drop shape models from Prupacher and Pitter (P&P, 1971), Goddard and Cherry (G&C 1984), Beard and Chuang (B&C, 1987) and Andsager et al. (ABL, 1999) respectively.

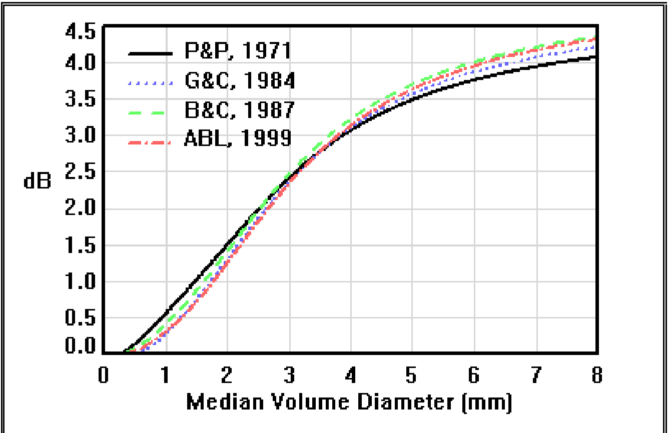
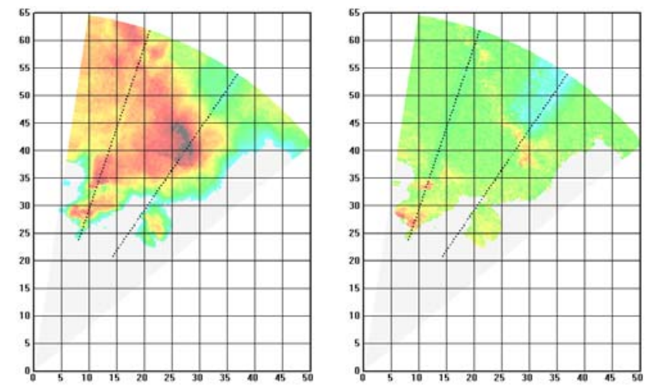
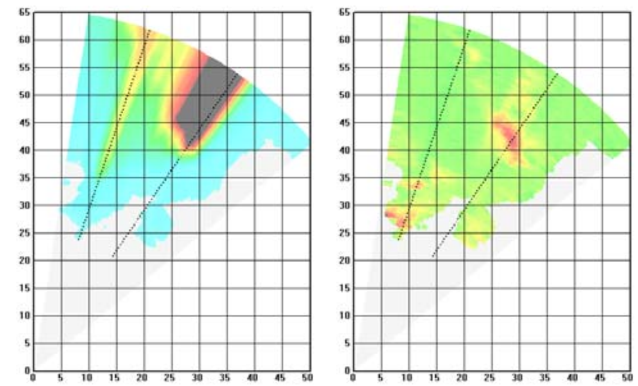


Fig. 2b : Z_{DR} plotted versus the median volume diameter (D_o) assuming an exponential drop size distribution and using the drop shape models from Prupacher and Pitter (P&P, 1971), Goddard and Cherry (G&C 1984), Beard and Chuang (B&C, 1987) and Andsager *et al.* (ABL, 1999) respectively. For a Marshal-Palmer drop size distribution, corresponding values for D_o are 0.89, 1.20, 1.60, 2.14 and 2.87 mm at rainfall intensities of 1, 4, 16, 64 and 256 mmhr^{-1} respectively.



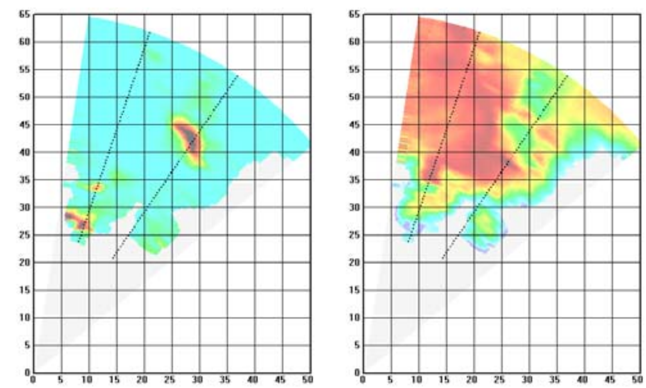
Z_H (-16 to +64 dBZ) Z_{DR} (-4 to +4 dB)

Figure 3a : Z_H (left) and Z_{DR} (right) data collected by the POLDIRAD radar for a thunderstorm in southern Germany on the 10th August 1994. As Fig. 3.2 but following removal of clutter affected areas.



ΣK (-4 to +4 dB) Z_{DR} (-4 to +4 dB)

Figure 3b : Cumulative attenuation (left) and Z_{DR} following adjustment for attenuation (right).



D_o (1 to 4 mm) N_o (10^2 to 10^6)

Figure 3c: Estimated values for parameters D_o and N_o assuming an exponential drop size distribution.

Bringi, 1978, 1980), and with rainfall estimates derived by this method comparing favourably with those derived using reflectivity at single polarisation. Seliga *et al* (1980) report that this was the case even when systematic biases were removed using data from a network of rain gauges, helping to support the conclusion that this improvement was indeed a result of variations in drop spectra being better represented. Several authors have gone on to propose power law type relationships based on polarimetric measurements. For example, Sachidananda and Zrnic (1987) give the relationship

$$R = 6.84\text{x}10^{-3}\,(Z_H)^{-3.86}\,(Z_V)^{4.86} \quad \text{mm hr}^{-1} \qquad \textbf{(7)}$$

, whereas Gorgucci *et al* (1995) suggest a more explicit relationship to Z_{DR} so that

$$R = 10^{-3}\,(Z_H)^{0.92}\,10^{(-0.369\,Z_{DR})} \quad \text{mm hr}^{-1} \qquad \textbf{(8)}$$

, both of these relationships applying for rainfall intensities in the range 20 and 50 mm hr^{-1}.

Nevertheless, there are a number of factors which may impact on measurements of Z_{DR}, and thus on the accuracy of any estimates of rainfall obtained by this method. In particular, ground clutter can exhibit a marked polarisation dependence, with scattering cross-section tending to be much larger at horizontal polarisation. However, particular objects such as towers or masts on the radar horizon can have the opposite effect, with a much stronger signal at vertical polarisation. Thus, in areas contaminated by clutter, Z_{DR} is likely to appear particularly grainy, with values which are predominantly strongly positive, but which may also be strongly negative. more difficult aspect is the fact that as a result of factors such as the positioning of the feed mechanism, the antennae illumination function will normally be quite different at the two polarisations. The volumes sampled in each case will not match exactly, and this will be exacerbated by delays in switching between the two polarisations.

Thus, measurements of Z_{DR} might be expected to become less accurate in situations such thunderstorms where the meteorology is relatively dynamic, or at higher rotation rates. Switching polarisation only after groups of pulses, rather than on alternate pulses is obviously undesirable. In addition, the position of supporting elements in the radome used to protect most operational systems, might be expected to have some effect in terms of the Z_{DR} signal, but the magnitude of this effect would be very difficult to predict and would be expected to vary with azimuth and elevation. In general, these types of errors are essentially random, and would suggest that measurements of Z_{DR} might be expected to be somewhat noisy. Given that Z_{DR} exhibits only limited sensitivity to drop spectra variations, this is obviously a concern, though it is also true that at shorter ranges, errors will tend to average out where data is scaled up to the resolutions at which a rainfall product is normally required.

However, at shorter wavelengths (X and C-band) the most serious concern is that the extinction cross section, and therefore the rate of attenuation, will be different at the two polarisations. Since extinction is cumulative and will be stronger at horizontal polarisation, the net effect is that Z_H will tend to decrease more rapidly with range leading to Z_{DR} being underestimated or becoming negative. Since strong differential attenuation might be expected in exactly the same situations where Z_{DR} is likely to be most useful, this represents a significant barrier to operational application, at least as far as directly obtaining estimates of precipitation. Obtaining a reliable and operationally robust method of correcting for the effects of attenuation has in itself proved an intractable problem, and it is difficult to hold out any hope for a robust scheme that would do so with the level of accuracy required.

As an example, Figure 3 shows the effect of strong attenuation in a thunderstorm on ZH and ZDR measurements from the C-band POLDIRAD radar in southern Germany. This is one of a number of scans collected on the afternoon and evening of the 10th of August 1994 during a period of intense thunderstorm actrivity. In this case, the parent storm is located approximately 50 km to the north-east and is moving almost directly along the radial towards the radar. There is some suggestion that the main cell is being maintained by continuous propagation from right to left, similar to a squall line but at smaller scales. In addition, a number of daughter cells are apparent in advance of the main storm and appear to have been triggered by low level outflow.

From this initial view of the measurements for ZH and ZDR, a number of features are worth commenting on. Firstly, it is noticeable that a weakening of the ZH signal to the rear of the main storm cell is barely discernible in the first picture, but becomes very apparent in the second picture where ZDR is observed to become negative. This is very clearly a result of differential attenuation, with the reflectivity signal at horizontal polarisation attenuating more rapidly with range. Furthermore, the ZDR measurements reveal a second narrow sector of strong attenuation immediately to the rear of one of the daughter cells and at an azimuth of 18.1^{o}.

While this can perhaps be taken as suggesting that at shorter wavelengths Z_{DR} can be useful in detecting the presence of attenuation, and such a technique has indeed been proposed by Upton and Fernandez (2000) among others, it is clearly unhelpful if the intention in measuring Z_{DR} is to better parameterise the drop size distribution. Even in milder cases, the effect of differential attenuation will be to cause a bias towards smaller D_o and larger N_o with range, and it becomes difficult to know if the features apparent are genuinely a consequence of distortions to the drop size distribution arising from the internal dynamics of each storm cell, or instead an artefact of the view angle. Thus, at shorter wavelengths some adjustment for the affects of attenuation is required.

Another feature worth noting from Fig. 3a is the effect of ground clutter contamination on measurements of Z_{DR}. This is apparent at shorter ranges within about 25 km of the radar where the weak echo is believed to be almost entirely due to reflections from the ground. Corresponding measurements for Z_{DR} are strongly positive, this being expected since a predominantly flat surface will tend to have a much larger back scatter cross-section at horizontal polarisation. However, an interesting feature is the band of somewhat lower values which appears to extend through this region and up towards one of the daughter cells. This has the appearance of a gust front boundary, and would therefore indicate a genuine meteorological feature which might otherwise go unnoticed. Thus in a qualitative sense, and in addition to identifying areas where attenuation may be a problem, measurements of Z_{DR} have a further potential role in supporting discrimination between meteorological and non-meteorological echoes. Although in both these cases Z_{DR} is providing information that supplements that provided by Z_H, it is equally true that exactly the same features will tend to interfere with attempting to use Z_{DR} quantitatively. Some adjustment is therefore required before attempting to derive information about variations in droplet size spectra and the consequent impact on derived rainfall intensities.

Clutter is perhaps the lesser of the two problems, since it will tend to occur in known locations, and a variety of techniques are available for identifying contamination. Accounting for the effects of differential attenuation is somewhat more problematic. Particularly in the case of a thunderstorm, with strong reflectivity gradients and the possible presence of hail, without any constraints a standard iterative correction is liable to become unstable, making the problem far worse. However, in this particular case it is possible to apply a constraint based on the expectation that Z_{DR} should tend towards zero in light rainfall. In other words, the cumulative adjustment made to Z_H and Z_V can be scaled so that Z_{DR} returns to zero on the far side of the storm. Naturally this relies on the storm being contained within the area for which coverage is available, something which is only broadly true in this instance.

Figure 3b shows the re-calculated values for Z_{DR}, together with the cumulative adjustment made to Z_H, based on a standard Z-K relationship derived for $\lambda = 5.6$ cm and assuming a water droplet temperature of 0^{o} C. This serves to highlight the areas where strong attenuation, and hence strong differential attenuation is likely to be occurring. It will be noted that no attempt is made to reflect the effect of variations on drop size distribution on the rate of attenuation, since this in itself would be prone to becoming numerically unstable, and instead a broad adjustment is sought which brings the values for Z_{DR} within reasonable limits rather than attempting to obtain precise values. The net effect is to strengthen Z_{DR} values in the vicinity of the main storm, and to remove the areas of negative Z_{DR} at longer range.

Values for D_o can then be obtained directly by reference to a pre-calculated look-up table based on the relationship shown previously in Fig 2b, with values for N_o subsequently being determined from Z_H assuming the appropriate exponential drop size distribution. However, it will be noted that in the vicinity of both the parent cell and one of the daughter cells there are instances of Z_{DR} values in excess of 4 dB, this suggesting a value for D_o larger than the maximum drop diameter of 8 mm. Since this occurs whether or not the correction for attenuation is applied, it appears to be a genuine feature and most probably represents a departure from the normal assumption underlying an exponential drop size distribution, namely that larger droplets will be present at lower concentrations. In areas with a strong updraft only the larger droplets will fall out, with the result that smaller droplets will be absent from a volume sampled within the rain shaft. The effect of this distortion is to enhance measured values for $Z_{DR,}$ giving an erroneous value for N_o if this is then calculated assuming that D_o and Z_H describe an exponential distribution.

Similarly, there are regions where Z_{DR} is very low, indicating a very small D_o. This is reasonable for lower rainfall intensities, but when combined with higher value for Z_H, the effect is for N_o to be many orders of magnitude larger than might be thought realistic. A situation in which this may occur would be in the presence of hail, where Z_{DR} would be expected to be much lower. Although hail particles do exhibit some variation in shape with size, and do have preferred fall modes, these are far less pronounced than for water droplets producing similar returns at horizontal and vertical polarisations.

The derived values for N_o and D_o shown in Fig. 3c have therefore been constrained to what are judged physically meaningful values, with D_o being allowed to vary between 1 and 4 mm, and with N_o being allowed to vary in the range 10^2 to 10^6. This results in images which are largely believable, though not necessarily accurate. For example, higher values for D_o are observed on the leading edge of the strongest daughter cell where a strong updraft might be expected, and this would be consistent with the observations of Schurr et al (2001) and others. Similarly, higher values for D_o are observed on the left flank of the parent cell, this again being consistent with where a strong updraft might be expected. However, the lower values for N_o to the rear of this feature could be taken as indicating a region of inflow, or could equally be explained as a consequence of Z_H not being adequately corrected for the effects of attenuation. Similarly, the higher values for N_o on the leading edge should probably be taken as indicating the presence of hail rather than large concentrations of smaller droplets.

This practical example serves to highlight both some of the strengths and the weaknesses of the Z_{DR} measurement, particularly with regard to use at shorter wavelengths. While it does clearly provide information which is potentially useful, it also a quantity which can potentially prove very misleading. Differential attenuation is clearly a problem, and while an adequate correction may or may not have been made in this specific example, it is much more difficult to postulate a similar scheme that would work reliably in an operational environment. However, even at longer wavelengths, it should be apparent that Z_{DR}, and particularly the derived quantities D_o and N_o are relatively sensitive to small errors which can arise from factors such as clutter contamination or noise. Without some form of constraint, these can swing wildly, giving very different estimates of rainfall intensity. In part, this stems from a reliance on the assumption of an exponential drop size distribution, with D_o and N_o inevitably being inter-dependent. With the possible exception of deep, tropical rainfall systems, this may not be a reasonable assumption in many of the situations where a discernible Z_{DR} signal is observed.

Differential Propagation Phase (K_{DP})

Liquid water has a refractive index which is higher than that of the surrounding air, with the effect that the speed of wave propagation will be slowed. Since larger rain droplets will tend to be more oblate, a wave which is horizontally polarized will in effect encounter more water content when traveling through rainfall, and hence experience a greater relative delay as compared with a wave which has a vertical polarization. There will thus be some difference in the phase measured at the two polarizations, and this relative difference will tend to increase in more intense rainfall as the average droplet size increases. This led Seliga and Bringi (1978) to propose the use of the Specific Differential Propagation Phase (K_{DP}) as a basis for estimating rainfall intensity.

In a practical sense, this technique has a number of attractive features not least of which is the fact that phase measurements will be virtually immune to the affects of attenuation, or to errors in the absolute calibration of the radar hardware. In addition, phase measurements will be unaffected by partial blockage of the radar beam, this allowing scans to be made at lower elevations so that the radar beam would be closer to the ground, and measured rainfall intensities would be closer to those at the surface.

Specific differential propagation phase (K_{DP}) is derived from measurements of the differential phase shift (Φ_{DP}) defined simply as

$$\Phi_{DP} = \Phi_H - \Phi_V \text{ deg} \qquad \textbf{(9)}$$

, where Φ_H and Φ_V represent the measured phase at horizontal and vertical polarisations respectively. Φ_{DP} is itself relatively insensitive to changes in drop size distribution, and tends to be a rather noisy. As a consequence, it is more convenient to refer to the rate of change over a distance r_1 to r_2, with the specific differential propagation phase (K_{DP}) being defined as

$$K_{DP} = [\Phi_{DP}(r_2) - \Phi_{DP}(r_1)] / 2(r_2 - r_1) \quad \text{deg km}^{-1} \qquad \textbf{(10)}$$

, and obtained using statistical approaches such as filtering or least squares. However, this requirement to spatially average represents an important practical limitation of the technique, since reliable estimates of precipitation can only be obtained in heavier rainfall and over larger areas.

Using the Prupacher and Pitter (1971) drop shape model and assuming a Marshal-Palmer drop size distribution, Sachidanada and Zrnic (1986) derived relationships between R and K_{DP} with the form

$$R = 37.1\ (K_{DP})^{0.866} \quad \text{mm hr}^{-1} \qquad \textbf{(11)}$$

for measurements at S-band. At C-band, Aydin and Giridhar (1991) give the equivalent relationship as being

$$R = 16.03\ (K_{DP})^{0.95} \quad \text{mm hr}^{-1} \qquad \textbf{(12)}$$

Gorgucci et al (2000) note that the relationship to rainfall intensity is almost linear and at S-band can be approximated as

$$R = 39.8\ K_{DP} \text{ mm hr}^{-1} \qquad \textbf{(13)}$$

, implying that a rainfall intensity of 10 mm hr^{-1} would result in a K_{DP} of only about 0.25 deg km^{-1}.

Analysis of S-POL data from the MAP campaign.

Normally, quantitative polarimetric methods would be examined in terms of the relative advantage over conventional, reflectivity only precipitation estimates. In other words, the assumption is that a particular technique will offer a more accurate estimate of rainfall intensity when compared with what is measured at the ground. In this case, the intention was to take a subtly different approach in examining what can be learnt about the scale of variations in the Z-R relations from a comparison of the rainfall estimates by different techniques. This makes sense if one remembers the difficulties in comparing any form of radar estimate, with so-called "ground truth" measurements from rain gauges or disdrometers. Quite simply, the sampling characteristics are so different that any smaller scale variations are liable to be lost in the noise. Comparing different radar based methods has the advantage that exactly the same atmospheric volume is being considered in each case.

The original intention, as described in the project workplan, had been to use data collected specifically for this purpose using the PODIRAD radar at DLR in southern Germany. Since this radar operates at C-band, any results might therefore be applicable to operational systems in Europe. However, prolonged delays in the refurbishment of the radar has meant that reliable data has not been available during the lifetime of the project. This together with doubts about the suitability of existing archived data sets, has meant it has instead been necessary to use data from the S-POL radar operated by NCAR, collected during the MAP campaign in Northern Italy during Autumn 1999. Although we are fortunate that this data was available, in terms of the work that had been planned there are two main disadvantages with using this data set. The first and most obvious is that the S-POL radar operates at S-band, and that while this has both advantages and disadvantages, any results would clearly be less relevant to the European situation. The second disadvantage is that the data available is limited to what was collected at the time, and to meet a different set of objectives.

Of the periods for which the S-POL radar was operating and during which there was significant precipitation in the vicinity, the following periods were selected for analysis.

- 17th September 1999 (IOP2A), Squall Line
- 20th September 1999 (IOP2B), Frontal
- 30th September 1999 (IOP4), Frontal
- 18th October 1999 (IOP7), Stratiform
- 21st October 1999 (IOP8), Stratiform
- 4th November 1999 (IOP14), Stratiform

The two cases where the heaviest precipitation occurred, and where the strongest effects might be apparent, were during thunderstorms associated with a squall line on the 17th September 1999, and during heavy rainfall associated with a frontal passage three days later on the 20th of September. igs.4 and 5 show examples of Z_H, Z_{DR} and K_{DP} recorded on each of these two dates. In both these cases, the radar was scanning a sector to the north west of the radar site where the terrain becomes mountainous. As a consequence the lowest elevation available is generally 2.5^o and even in these cases, there is clear evidence of partial blockages (e.g. at azimuths 288^o and 309^o) and some clutter contamination.

Examining these two cases, it will be immediately apparent that for S-band at least, K_{DP} is indeed both relatively insensitive and extremely noisy. There is some evidence of a signal emerging, most noticeably in the case on the 20th September where a band of very heavy rainfall (at azimuth 300^o, range 30 to 40km) is picked out by K_{DP} values in the range 0.5^o to 1^o. However, heavier but more localized rainfall that might be expected to be associated with the thunderstorm cells on the 17th September is only barely discernible as an enhancement to K_{DP}. In the case of the large cell at 65km range, 335^o azimuth this weak enhancement appears to occur predominantly to the rear of the areas where the strongest echoes are recorded by Z_H. A possible explanation would be that these echoes are themselves enhanced by the presence of hail to which K_{DP} should be relatively insensitive. This is certainly not a complete explanation since Z_{DR} does still appear to show at least some enhancement in the same region. Another featutre worth noting is the enhancement to K_{DP} values at ranges between 50 and 70km on the 20th of September. Although very patchy, there would be a strong suspicion that these higher values for K_{DP} are associated with the beam intercepting the melting layer.

To gain an appreciation of what this means in terms of estimated rainfall, a 64km square area was selected and divided into a grid at resolutions ranging from 1 to 16 km. These could represent river catchments or a model grid. The average rainfall intensity in each grid cell was then calculated based on Eqn.3 for Z_H and Eqn.11 for K_{DP}. Results are shown for each of the two events in Figures 6a and 6b respectively. Individual estimates of rainfall by each of the two methods and at different resolutions are shown by means of a scatter plots given as Figures 7a and 7b. Briefly, the assumption behind this method of analysis is that at some given resolution, the degree of averaging should be sufficient to bring the rainfall estimates obtained by the two methods more closely into agreement with the remaining differences being explained through variations in the drop size distribution. The resolution at which this occurs might reasonably be expected to show some dependence on the type of precipitation and the inherent variability of the underlying meteorological processes.

In practice, there is only very poor agreement between the rainfall estimates obtained by the two techniques, even as the resolution is reduced. Although, spatial filtering does improve the match between the two sets of estimates, scatter continues to dominate, and on the whole, this appears to be due to problems other than those which could be explained as drop spectra variations. It is worth noting that for the 17th Spetember event, there is noticeable trend suggesting that K_{DP} is systematically underestimating rainfall for higher values of Z_H. It is uncertain whether this should really be taken as indicating the presence of hail, and there appears to be some suggestion that possibly as a result of the way K_{DP} is processed, rainfall estimates obtained by this parameter may be being biased towards the rear of the main storm cell. For the event on the 20th September, K_{DP} again appears to be underestimating rainfall in the band close to the radar, but this is masked in the scatter plots by the effects of the radar beam intercepting the ice layer, which dominate over a much larger area.

Conclusions

The reason for comparing rainfall estimates at different resolutions is that not only is this more representative of what an end user might require, but also that this may serve to filter some of the noise observed in the K_{DP} estimates. Thus K_{DP} could potentially still provide a more accurate estimate of the average rainfall across an area (or equivalently. from a series of rapid scans averaged over a period of time). This necessarily implies some loss of temporal and spatial resolution, but this might in any case be required in order to provide a rainfall product suitable for a particular application. However, the results obtained are not very persuasive, and suggest that this is unlikely to be the case at the spatial and temporal resolutions which might typically be required by end users.

The original premise was to use the more accurate estimates of rainfall which, theoretically at least, could be obtained using a polarimetric radar, to identify variations in the quality of rainfall estimates obtained by conventional methods. This does not appear to be practicable, with the poor sensitivity and resultant noise in K_{DP} masking any meaningful information about variations in drop size distribution. Since the phenonema which is being examined will itself exhibit rapid spatial and temporal variations, application of spatial or temporal filtering does not appear to make the problem any more tractable. It should also be apparent that for a very weak signal, the precise way in which K_{DP} is being computed may have a significant effect.

Naturally, any conclusions need to be framed in the light of changes that were necessary to the work plan due to the unavailability of the POLDIRAD radar. In particular, it is important to note that K_{DP} is more sensitive to rainfall at C-band, and thus it might be easier to recover a meaningful signal. In addition, it should be borne in mind that since it has not been possible to collect data specifically for the purposes of the project, it has been impossible to attempt to improve the filtering of the K_{DP} estimates by, for example, rapid scanning and averaging in time. Both these factors would lead to some improvement, but whether this would be sufficient to give a meaningful result is clearly open to question. It is worth noting that from figures derived from other literature, Illingworth et al (2000) estimate that to achieve a rainfall estimate within 25% of the true value at 1.2 km spatial resolution, the rainfall intensity would need to be in excess of 66 mmhr^{-1} at S-band, as opposed to 43 mmhr^{-1} at C-band. Even if these figures are not optimistic, this still represents a very severe restriction on practical use even at the shorter wavelength.

In more general terms, it seems unlikely that, on its own, access to polarimetric information would lead directly lead to improved quantitative rainfall estimates in an operational environment. Propagation effects are generally small and noisy, while at the same time being prone to measurement problems in the same way that conventional reflectivity measurements are. In this sense, there is no net benefit being derived and from the results shown here, the precipitation estimates obtained by these methods are likely to be of lower quality than those obtained by conventional methods. In the case of K_{DP}, averaging may improve the estimate obtained, but the degree of averaging required and the loss of spatial and temporal resolution implied to routinely be able to reach the level of quality obtained by conventional methods, is likely to be unacceptable in most cases.

Where polarimetry may provide a benefit is in a secondary role, for detecting and delimiting quality problems affecting rainfall estimates obtained by conventional methods. Examples would include detecting the melting layer (as shown in Fig. 5), or at C-band, detecting stong attenuation (as shown in Fig. 3). It remains to be demonstated whether operationally robust methods for utilizing this information can be developed.

Acknowledgements

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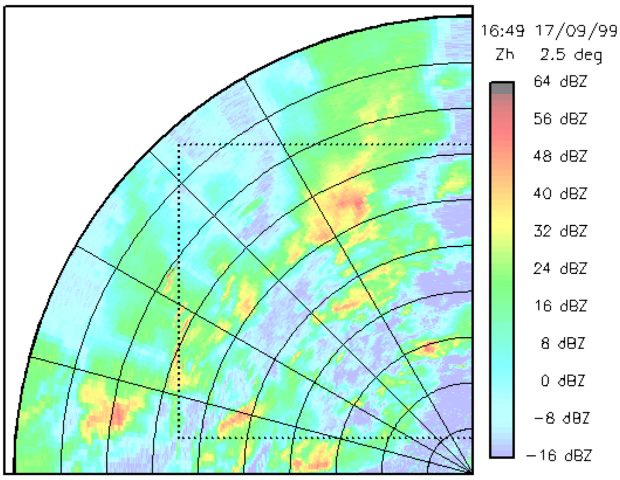


Fig. 4a) Reflectivity (Z_H), 17th September 1999

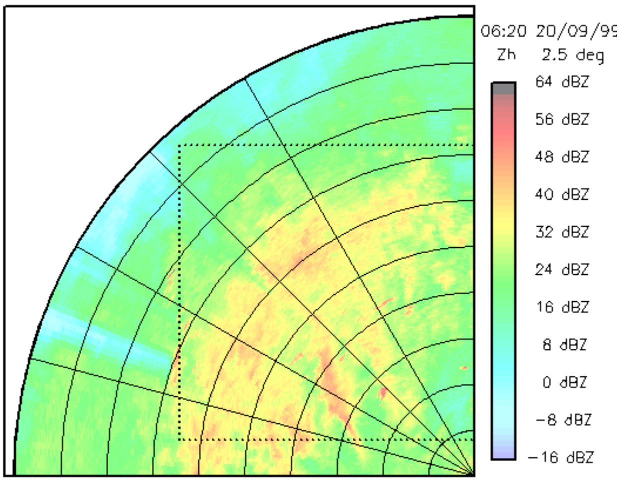


Fig. 5a) Reflectivity (Z_H), 20th September 1999

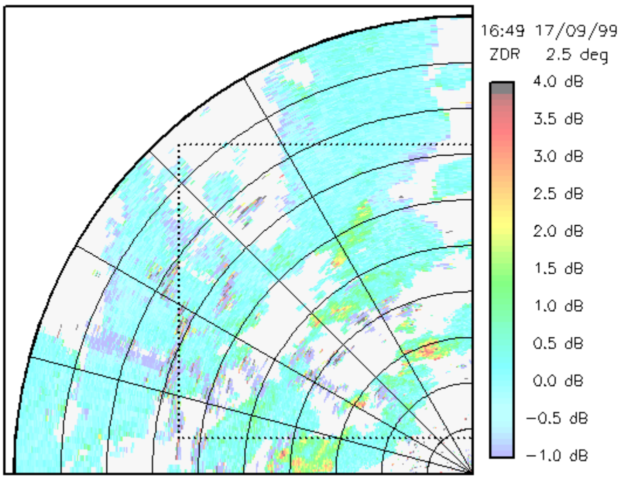


Fig. 4b) Differential Reflectivity (Z_{DR}), 17th September 1999

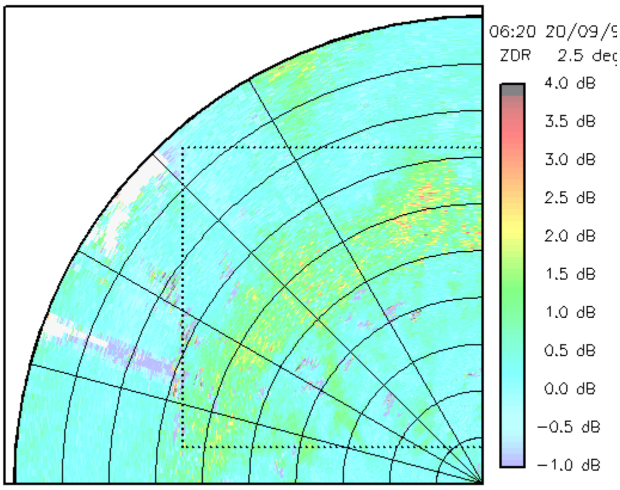


Fig. 5b) Differential Reflectivity (Z_{DR}), 20th September 1999

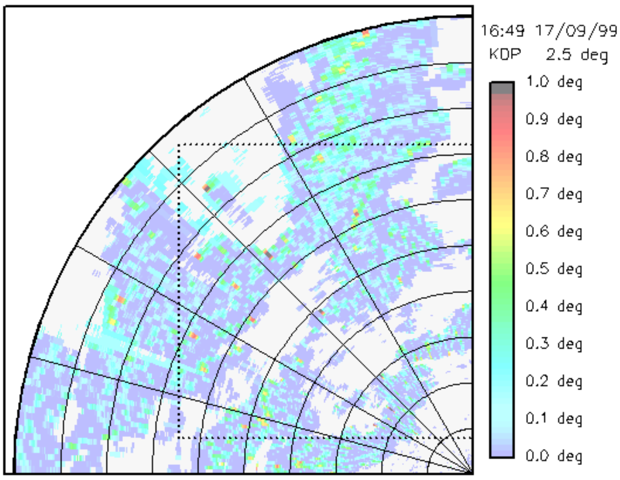


Fig 4c) Differential Phase (K_{DP}), 17th September 1999

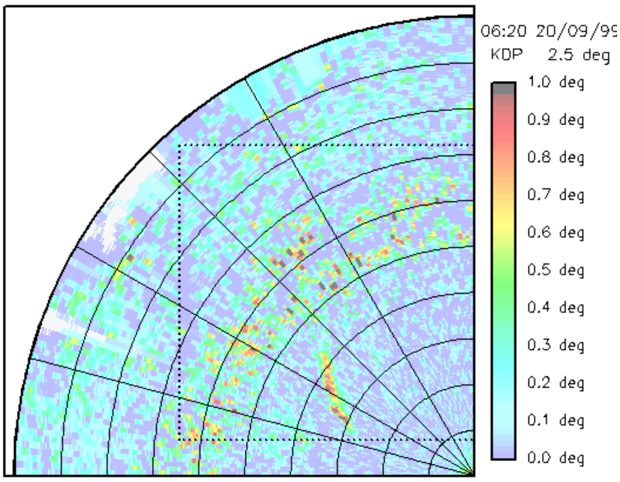


Fig 5c) Differential Phase (K_{DP}), 20th September 1999

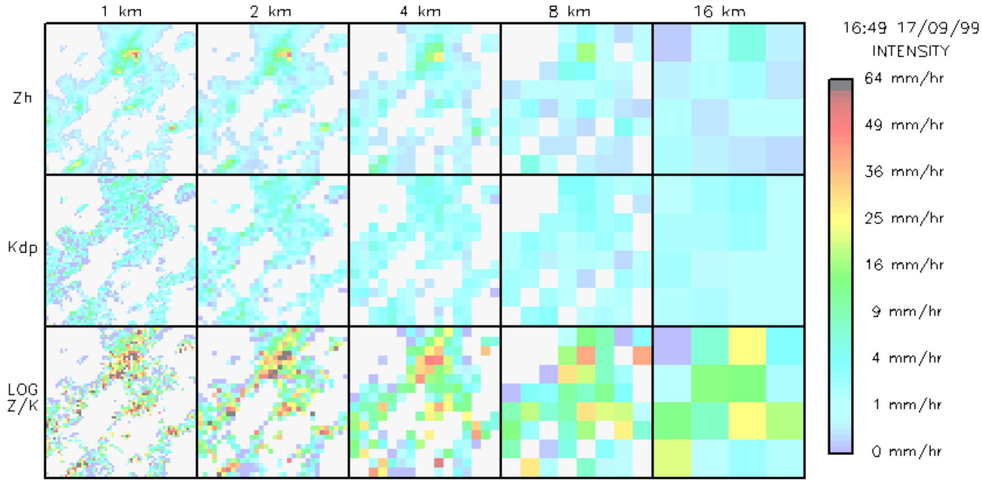
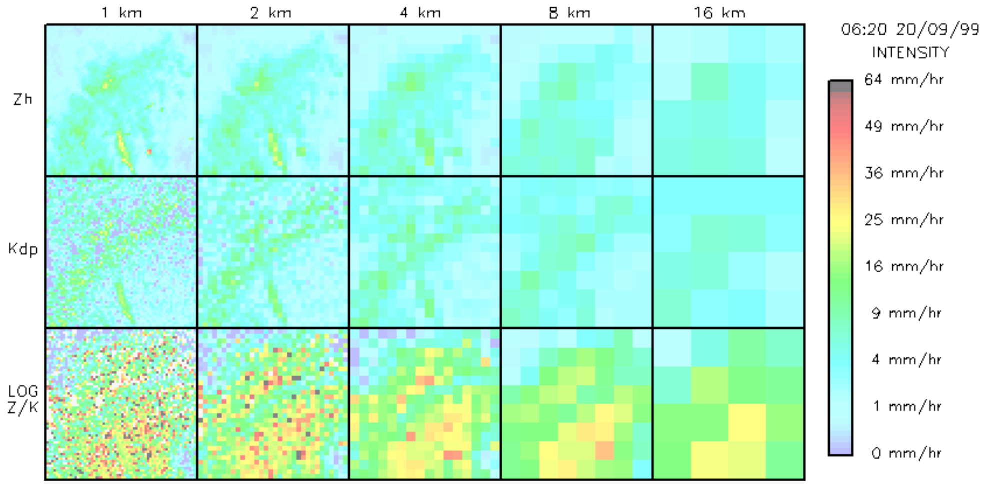
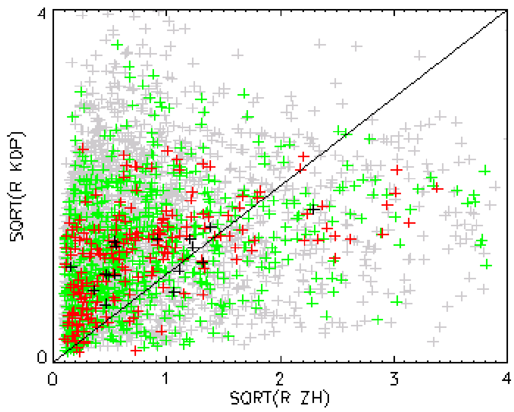


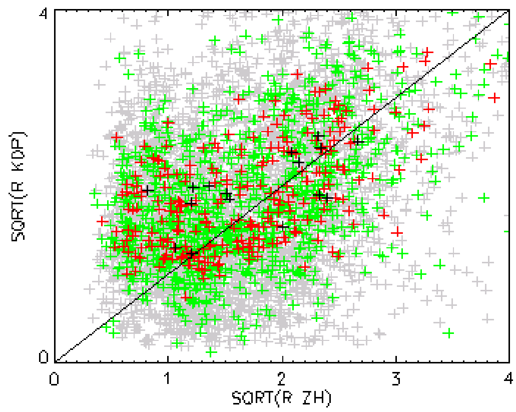
Fig 6a) 16:49 17th September 1999



b) 06:20 20th September 1999

Fig 6 : Rainfall estimates at different grid resolutions obtained from reflectivity (Z_H – top row) and from differential propagation phase (K_{DP} – middle row) for 16:49 on the 17th of September 1999. The total area represented is a 64km square and the bottom row shows the ratio between the two estimates expressed logarithmically so that the maximum differences (dark red or dark blue) occur where one quantity is ten times the other.





a) 16:49 17th September 1999　　b) 06:20 20th September 1999

Fig 7 : Rainfall estimates for Z_H plotted versus those for K_{DP}. The colours correspond to grid resolutions of 1km (grey), 2km (green), 4km (red) and 8km (black).