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RAINFALL ESTIMATION BY RADAR FOR THE OTAKI CATCHMENT: THE OPERA PILOT STUDY

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ABSTRACT

The radar-based estimation of rainfall for operational hydrology has been attempted in many countries for decades. Obtaining data of an accuracy sufficient for practical routine application has proved difficult due to a variety of operational and more fundamental problems. The most significant of the latter are the occurrence of low-level growth of precipitation due to orographic effects and the existence of a strongly reflecting layer or 'bright band' due to the melting of snow to rain.

Two heavy rainfall events in the Otaki catchment near Wellington have been investigated using radar data obtained from the Met Service radar located at Outlook Hill. It was found that the radar data underestimated the surface rainfall rates by at least a factor of 2. The major cause of the error is shown to be consistent with an orographic enhancement of a factor of 4. The blocking of the radar beam by hilly terrain compounds the underestimation.

INTRODUCTION

For most hydrological activities, be it flash flood forecasting or general water resource planning, accurate estimates of areal rainfall are required. Rain gauges have traditionally been used for this purpose. However, the rain gauge gives point rather than areal estimates of rainfall and is subject to significant measurement errors. Rodda (1971), for example, found that local topography surrounding the gauge affects turbulence and rain-catch so that errors of up to 30% can occur in the estimates of annual rainfall. Many studies have been made concerning the errors that result from rain gauge catch (Sevruk, 1975) and of the representativeness of the rain gauge catch for the areas surrounding the gauge (Kitchen and Blackall, 1992). From experiments involving the collocation of rain gauges, it has been shown that, in most meteorological conditions, adjacent gauges will have catches that vary by less than 15%. However, there is greater difficulty in estimating the area over which a gauge may be said to be representative. This depends greatly on the wind speed and precipitation type, gauge location and exposure, and the topography both locally and over distances ranging from 10 to 100 km (Hutchinson, 1970; Zawadski et al., 1986; Austin, 1987).

The accuracy with which a gauge network can measure areal rainfall depends on the variability and intermittency of the rainfall event and the configuration of the gauge network. Seed and Austin (1990a) found that a regular grid of gauges gave somewhat less variable errors than a random array, but the total number of gauges was by far the most important determinant of the accuracy of areal rainfall estimates. The complexity or otherwise of the interpolation scheme had minor effects on areal rainfall estimates - given an intermittent field no interpolation scheme can guess the actual rainfall between gauges with any great skill. It seems to be necessary to have one gauge each "decorrelation length" of the rainfall pattern in order to achieve reasonable measurement accuracy. (The decorrelation length is the maximum distance between two gauges for which the correlation of the time series of rainfall rates remain significant). This means that, for convective rainfall, errors of less than 30% in areal rainfall require a gauge spacing of 10 km. In more stratiform rainfall events, 30 km or more may be adequate. Gauge data, then, do not represent the elusive "ground truth" and may not represent well the areal distribution of rainfall, especially during convective events.

RADAR RAINFALL ESTIMATION

An alternative to the rain-gauge-based techniques is, of course, to use meteorological radar data. This data has the immediate advantage of high spatial resolution (1 or 2 km) and can be operated at high time resolution (5 minutes or less), although, for a variety of reasons the New Zealand Meteorological Service radars (near Auckland, Wellington and Christchurch) are operated at 15-minute resolution. On the other hand radar data is subject to a variety of errors which limit the accuracy of the resulting areal rainfall data.

The statistical characteristics of radar observations of rainfall are markedly different to those of gauges (Zawadzki et al., 1986). Whereas gauges measure rainfall at a point, averaging in time, radar makes areal averages which represent a near instant in time. These differences lead to different perceptions of the structure of the true rainfall pattern. Thus radar will give significantly different frequencies-of-occurrence of high rainfall rates over its measurement cell, which is much bigger than a gauge and is, moreover, an increasing function of range. This can result in bias in the estimated statistics of rainfall events. The proper calibration against accumulated rainfall amounts, rather than comparing frequency statistics, can minimise the operational bias in the areal accumulation (Seed and Austin, 1990b).

One obvious source of radar error results from the radar beam intercepting the topography, particularly in mountainous terrain (Joss and Waldvogel, 1991). This leads to spurious ground clutter and shadow effects, as beyond these obstacles the radar beam will be partially blocked, reducing the returning signal.

A further result of the altitude of the radar beam (needed to clear mountains and due to the curvature of the earth) is that, if large changes in the structure of the precipitation pattern occur in the vertical, then large sampling errors can result. The two most significant effects result from the Bright Band (BB) and from orographic enhancement. Melting of snow to rain as it falls generates a layer of enhanced reflectivity known as the Bright Band. Orographic enhancement results from the sweeping out of enhanced liquid water content of the clouds produced by upward air motion as the air flows over hills. The latter effect is illustrated by the Bergeron "seeder-feeder" mechanism (Bergeron, 1965). These are both features of the vertical structure of reflectivity. This Vertical Profile of Reflectivity (VPR) is known to be highly variable on very short time scales (Fabry and Austin, 1991), although corrections for the storm-averaged structure can be made successfully (Gray, 1991a).

Attenuation of the radar beam by rainfall is another phenomenon that can result in the radar-based rainfall rates being in error. The effect is strongly wavelength dependent, with large attenuation at wavelengths as short as 3 cm and little at wavelengths in excess of 10 cm. At 5.6 cm, the wavelength of the radar used in this study, attenuation is a significant problem only at high rainfall rates and then only when this rainfall is widespread. Corrections can be made for this problem. The operational system uses a correction technique based on standard formulae (Battan, 1973). The attenuation is summed in range along each azimuth, beginning at 10 km, with the total correction being limited to 4.8 dB - the equivalent of doubling the rain rate. This limit is placed on the correction in order to prevent errors in the reflectivity generating unrealistic rain rates (see Hildebrand, 1977; Delrieu et al., 1992, based on previous work of Hitchfield and Bordon, 1954).

In addition to these problems, an empirical relationship is used to convert radar-measured reflectivity to rainfall rate. The formulation of this relationship makes assumptions about the raindrop size distribution and terminal velocity. Variations away from these assumed distributions can be quite large (Hosking and Stow, 1987), resulting in errors in radar-based rainfall estimates. In the early literature it was assumed that this microphysical problem dominated the errors, but more recently it has become apparent that the sampling and the vertical profile problems are much more significant (Collier, 1989; Joss and Waldvogel, 1989). Errors of less than 30% on hourly rainfall totals are thought to be possible within about 100 km of a radar if appropriate care is taken with the quality control and data analysis and provided the mountains are not too much of a dominating factor.

All these problems lead to radar estimates that are not generally in agreement with gauge data, and lead to bias and spread in the radar estimates of surface rainfall rate.

Adjustment of the radar data field is often performed in order to force the radar values above gauge sites to have similar values to those of the rain gauge data. This attempts to remove the bias and variability in the radar estimates, as perceived by the gauges. Although this process is not straightforward (Cain and Smith, 1976), improvement can be shown to result in well controlled situations (Collier et al., 1983). Ideally, the adjustment should take place after corrections for any known radar errors have been made since these may, for example, be a systematic function of range. Any correction process that results in the radar precipitation estimates having a smoother relationship with rain-gauge data will be more amenable to post-correction adjustment by rain-gauge data.

Careful electronic calibration of the radar is also necessary if accurate quantitative information is to be generated. Even in the best of situations, residual calibration errors of 1 dBZ are likely to remain, introducing errors of up to 15% in rainfall estimates. When careful and regular calibrations are not made, errors of 2-3 dBZ are likely, resulting in rainfall estimates with errors of up to 50% (Austin, 1987).

HYDROLOGICAL CONSIDERATIONS

The major motivation for this study is the estimation of hydrological outputs rather than just the accuracy of the rainfall rate estimates themselves. This suggests immediately that our major concern should be the estimation of reasonably heavy rainfall events since in very light rainfall most of the rain will not run off at all.

Verworn (1989) investigated the influence of the resolution of input rainfall information on hydrological model results by simulating the effects of spatial rainfall variability and rain gauge sampling on modelled peak flows. Results show that the variations in rainfall, when viewed by widely spaced rain gauges, could result in marked errors in the peak flows and runoff volumes. This led Verworn to conclude that "the spatial resolution of rain data input is of paramount importance to the accuracy of the calculated hydrograph". Verworn also found that rainfall is the most sensitive parameter in hydrological modelling, so that improvements here will yield the greatest improvements in accuracy. These results appear contrary to those of Cluckie et al. (1989), who showed that the precision of the rainfall input into a hydrological model could be reduced without significant degradation of the modelled forecasts. This work showed that when data, initially at 256 rainfall intensity levels, were gathered into only 8 levels, no significant difference resulted in the modelled hydrograph. Cluckie et al. suggest that the low-pass filter nature of hydrological flows effectively transforms a high-frequency process, rainfall, into a low-frequency one, runoff. However, Verworn indicates that, despite this low-pass filtering, errors in rainfall estimates are propagated and amplified by the rainfall-runoff transformation. These papers address different aspects of the problem. Verworn is primarily concerned with the rainfall totals being representative of the areal distribution of the rainfall, whereas Cluckie et al. is addressing the requirements on precision of these individual area totals. These results show that the radar, with its high spatial resolution, has great potential for initialising hydrological models, despite its problems with error sources. Yet care must be taken in the direct application of radar estimates to hydrological modelling.

Hydrological models contain parameters that can be "tuned" to generate the best correspondence between modelled and measured river response. Implicit in many of these parameters are corrections for the error characteristics of the rain gauge sampling and interpolation techniques. Radar data have significantly different error characteristics, and thus radar data must be used to retune the hydrological model (Cluckie et al., 1989).

In New Zealand most operationally-significant hydrological catchments are in hilly terrain. In order to investigate the severity of the problem outlined above for local operations, the Otaki Precipitation Estimation by RAdar (OPERA) project was initiated to investigate the application of radar-based techniques to the problems of measuring rainfall over complex terrain. This region is remote, with few routinely available observations and hills rising to 1500m. Thus radar observations could play an important role in determining accurate estimates of hydrological balances. It is also a region which is prone to flooding and regularly exhibits orographic enhancement of rainfall.

TABLE 1—Locations and elevations of tipping-bucket gauges in the Otaki region

Gauge	Latitude	Longitude	Height
Oriwa	40.751	175.349	1080 m
McIntosh	40.919	175.309	1100 m
Taungata	40.814	175.255	980 m
Warwicks	40.959	175.077	345 m

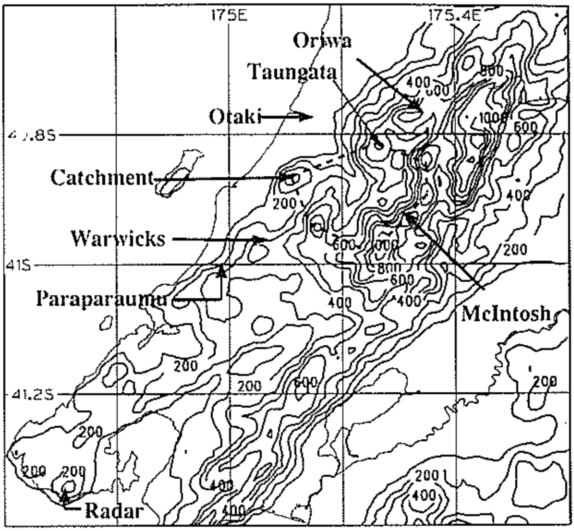


FIG. 1—The location of the rain-gauge, radar and balloon sites, in relation to the topography and catchment.

DATA AND CASE HISTORIES

The rain-gauge data used in this study were measured by recording tipping bucket gauges. The gauges have a bucket size of 0.5 mm and record the tip count for each 1-minute interval. These gauge data are then averaged over a 15-minute period, and recorded as rainfall rate. Four gauges were operational throughout this period, and their locations and heights are listed in Table 1 and displayed in Figure 1. The coarse intensity resolution, resulting from the large bucket size, leads to inaccuracies in their use in estimating 15-minute average rain rates (Gray, 1991b). This problem is more severe at low rain rates than at those rates encountered generally in the case studies investigated here. For rain rates of 5 mm.hr^{-1}, this quantisation error will lead to uncertainties of Order 1 mm.hr^{-1}, i.e., 20%. This will tend to lead to discrepancies between the time structure of the radar and rain gauge estimates but, for averages over entire rainfall events, no bias should result.

The radar provides information every 15 minutes, in polar co-ordinate form. These data have 2 km resolution in the range direction, with 0.86° resolution in azimuth. The radar has a beam width of 0.86°, and a wavelength of 5.6 cm. The location of the radar is displayed in Figure 1. The data used in this study come from the 5 lowest beam scans at elevations of 0.5°, 0.9°, 1.3°, 2.4° and 3.5°. Thus, we are comparing point, 15-minute time averages from the gauge with an instantaneous radar measurement corresponding to an area of a few square kilometres taken each 15 minutes. Substantial sampling differences are expected on this ground alone.

Constant Altitude Plan Position Indicator data (CAPPI) are also compared with the gauge data. CAPPI data are formed by interpolating between beam heights to give estimates of the rainfall at a constant altitude of 2 km. This is the form of radar rainfall estimate used operationally in the National Weather Forecasting Centre (NWFC). Corrections are also made for ground clutter and VPR-induced errors and the effects on the rainfall estimates monitored.

Two heavy rainfall events were investigated. Both events led to rapid river level rises and flooding. Figures 2 and 3 show the synoptic situations near the mid-point of each event. The first event, labelled Case 1, generated rain that fell for over 18 hours at average gauged rates of 7-11 mm.hr^{-1}. The rain during this event was influenced by the north-westerly wind flowing over the Tararua Ranges leading to orographic enhancement. Case 2 was more intense in nature, with 8 hours of rainfall at average rates of 13 mm.hr^{-1}. The second event appears similar to the first event with north-westerly winds at the onset. Figures 4 and 5 show the infra-red satellite images corresponding to times near that of Figures 2 and 3. These images also suggest that the rainfall structure may be similar for the two events. However the second event has lower humidity (75-80 % as opposed to 90-100 % for Case 1), is cooler (13-18°C as opposed to 18-20°C) and after only 4 hours of rain the southerly moves in, undercutting the warmer air. For these reasons the second event, though initially appearing similar, exhibited no orographic enhancement.

TABLE 2—Explained variances and raingauge-to-radar ratios for the storm of January 1992.

	RAW		CPVPR		CAPPI		CRVPR-CAPPI	
Gauge name	R^2-P	Ratio-B	R^2-P	Ratio	R^2-P	Ratio	R^2-P	Ratio
Oriwa	0.36-5	2.70-1	0.25-9	3.91-2	0.22-9	4.41	0.24-9	4.20
McIntosh	0.63-8	4.43-1	0.53-5	4.30-1	0.49-8	3.25	0.51-8	2.89
Taungata	0.44-7	2.37-1	0.40-5	2.77-3	0.33-5	2.72	0.40-5	2.55
Warwicks	0.53-1	1.87-1	0.37-2	2.09-2	0.26-2	2.47	0.30-2	2.38
Means	0.49	2.84	0.39	3.27	0.33	3.21	0.36	3.01
Std. Dev.	0.101	0.96	0.099	0.88	0.103	0.74	0.103	0.71

METHODOLOGY

The problem addressed in this paper is to assess the quality of the radar estimate, using rain gauge information, although the gauge information is not necessarily representative of ground truth. The method used is to linearly regress radar estimates of rainfall at locations near the gauges, against the rain gauge data. The results can then be compared by inspecting both the ratio of the means of the rain gauge and radar data, and from the explained variance, R^2. The ratio generates an estimate of the fractional bias in the radar rainfall estimation, while R^2 provides information on the similarity of the temporal structure of the two rainfall estimates. Linear regression was used as a tool for generating an estimate of the similarity of the time structures of the rain gauge and radar rainfall estimates. This is achieved by assuming the rain gauge and radar data are related by the linear model $Y = BX + C$ where Y is the estimate of the rain gauge observation, X the radar data and B and C are the model constants. The resulting explained variance provides information on the similarity of the time structures of the two estimates (rain gauge and radar). Bias must be eliminated independently.

TABLE 3—Explained variances and raingauge-to-radar ratios for the storm of February 1992.

Case 2	RAW		CPVPR		CAPPI		CRVPR-CAPPI	
Gauge name	R^2-P	Ratio-B	R^2-P	Ratio-B	R^2-P	Ratio	R^2-P	Ratio
Oriwa	0.41-9	2.63-1	0.53-1	2.53-3	0.38-7	2.31	0.50-7	2.84
McIntosh	0.51-2	2.13-2	0.52-5	2.40-4	0.35-6	1.85	0.43-2	2.26
Taungata	0.61-7	1.81-2	0.63-7	1.99-2	0.47-7	1.66	0.50-7	2.19
Warwicks	0.43-2	1.64-1	0.43-5	1.94-2	0.31-2	1.74	0.37-2	1.79
Means	0.49	2.05	0.53	2.21	0.38	1.89	0.45	2.27
Std. Dev.	0.079	0.38	0.071	0.25	0.059	0.25	0.054	0.37

Four different sets of radar data were used in the comparison (Tables 2 and 3). The first set is the unaltered radar data from the lowest 5 elevations (RAW). The next set is the same as the first, but with ground clutter returns removed and corrections made for the VPR- induced errors (designated CRVPR). The third set is a CAPPI scan at a nominal height of 2 km (CAPPI) - the data used operationally in the NWFC. The final set is the CAPPI data, with the clutter removed and corrections made for the VPR errors (CRVPR-CAPPI).

Tables 2 and 3 list the dependent sample explained variance (R^2), and the ratio of the means of the rain gauge data over the radar estimates (Ratio), for the radar estimates from each data set which showed the greatest similarity to the corresponding gauge. The estimate which will be chosen is not known *a priori.* This analysis procedure will highlight the factors which prevent one sample from being more representative than another. In the first two radar data sets, there are five elevations for correlation. For all the sets, there are nine adjacent radar range/azimuth samples to be compared, (the one over the gauge plus the eight surrounding ones): these are designated "bins". The "bin" (P) with the highest R^2 is listed alongside the R^2. The beam number (B) is indicated with the ratio for RAW and CRVPR.

Thus, by comparing the R^2 and bias ratios for these various data sets, the best radar rain estimator can be identified. In the tables, the means and standard deviations of these two measures are also listed. Biases in the ratio of radar data could be removed by adjustment with gauge data. The ideal radar data source would have little spread in the ratios for the different gauge sites. Thus, the data set which has the lowest standard deviation of the ratios should represent the best radar data source.

The VPR correction technique applied to the data (Gray, 1991a) parametrises the vertical profile with simple functions. These functions are a linear decrease of rain rate with height, superimposed with a function to represent the BB. The vertical profile is generated from the lowest 5 elevations, and represents the average vertical profile for the area out to a range of 120 km. The profile has been "normalised" with respect to the average rain rate encountered. The parametrisation function coefficients can also be interpreted as measurements of physical parameters. For example, the linear decrease of rain rate with height has a coefficient that represents the average, ground-level rain rate (normalised), and the "slope" represents the rate of decrease of rain rate with height.

Similarly, the BB function produces estimates of the amplitude of the melting layer effect (BB), and the BB height (ZBB). These physical parameters will be used to infer aspects of the vertical characteristics of the rain.

The ground clutter rejection algorithms attempt to detect and replace signals that have the characteristics of having intercepted either hills or ocean waves. These two forms of clutter are treated separately. Wave clutter is detected by the rapid decrease of its signal with height, whereas the hill clutter is detected by comparing the horizontal pattern of reflectivity with the patterns of the height and slope of the terrain. If there is a good correspondence, then it is likely that the hills in a region are reflecting the radar's beam, and the area is identified as clutter.

In conjunction with the above approach, the CAPPI data were summed over the entire event to present the radar estimate of the average rain rate across the radar domain. This will highlight some of the larger-scale problems of radar rainfall estimates.

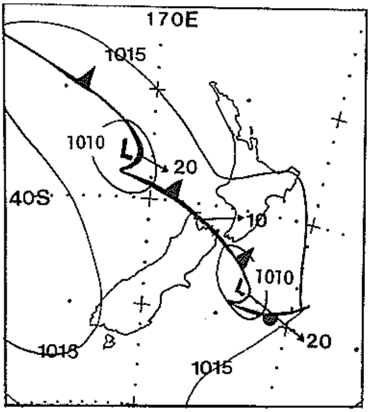


FIG. 2 — Synoptic situation for
12 GMT 25 Jan. 1992

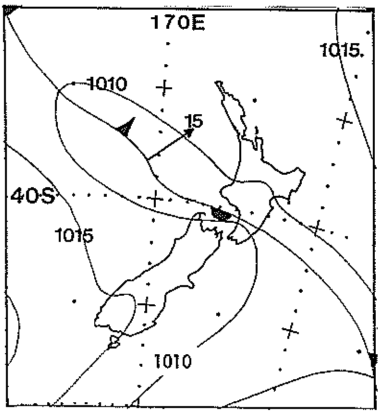


FIG. 3 — Synoptic situation for
06 GMT 5 Feb. 1992

RESULTS OF RADAR AND RAIN GAUGE COMPARISON
CASE 1

Table 2 shows the results of the comparison of the four forms of the radar data with the rain gauge data. Figures 6 and 7 show the coefficients from the parametrisation used in the VPR correction techniques.

From Table 2 it can be seen that the radar markedly under-estimated the amount of rain falling into the gauges, with ratios ranging from 1.9 to 4.4. The corrected data, CRVPR, have biases in excess of those of the raw data, as did both forms of CAPPI data. However, the raw data have the greatest spread in the ratios, with a standard deviation of 0.96, compared with 0.88, 0.74 and 0.71 for the other data forms respectively. However, many case studies would need to be investigated before one rainfall-estimation procedure could be said to be superior to another.

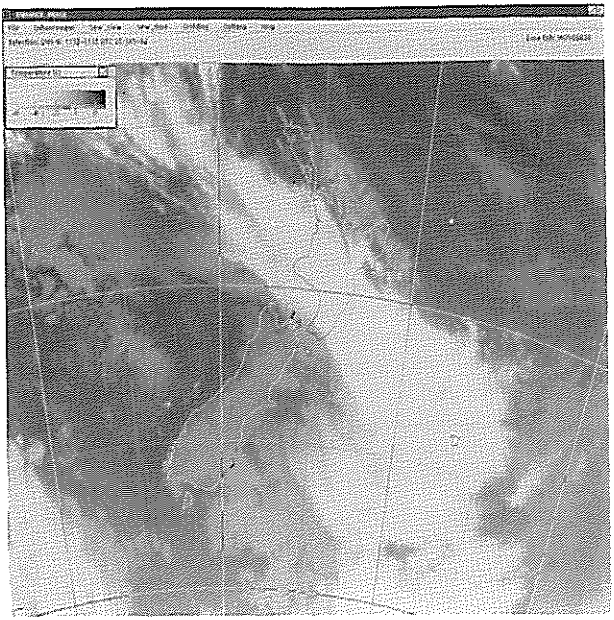
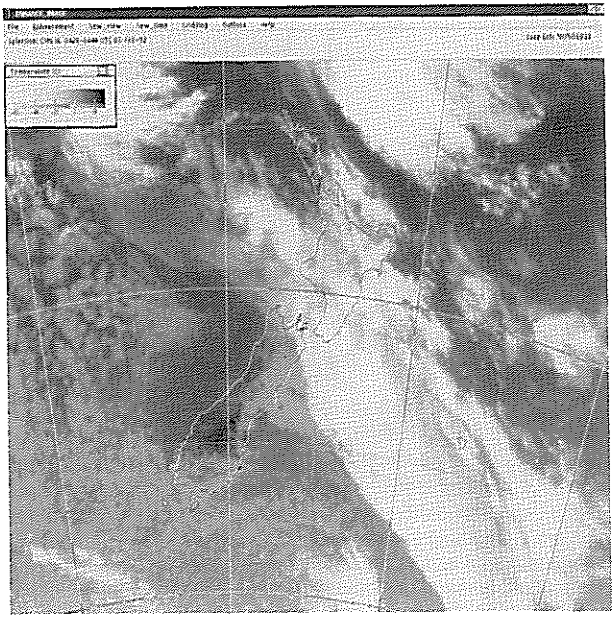


FIG. 4 — Satelite image for 12 GMT 25 Jan. 1992

FIG. 5 — Satelite image for 05 GMT 5 Feb. 1992



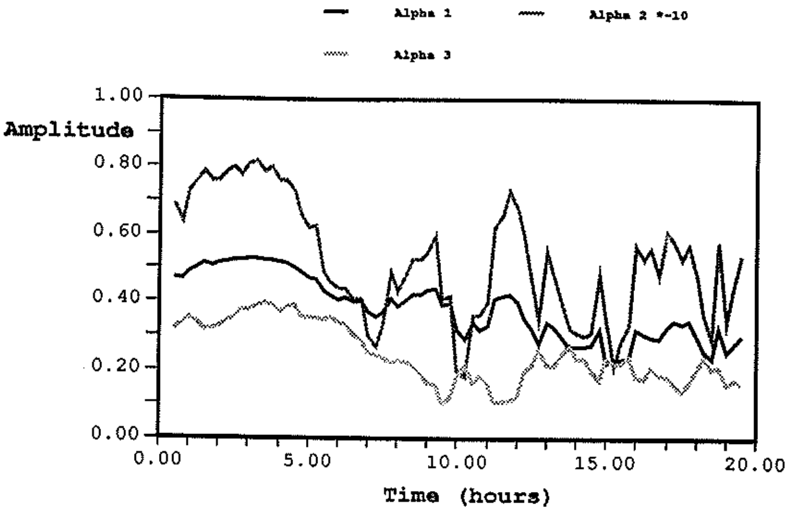


FIG. 6 — Time sequence plots for Case 1 on parametrisation coefficients Alpha 1, Alpha 2 and Alpha 3. Note that Alpha 2 is scaled by -10.

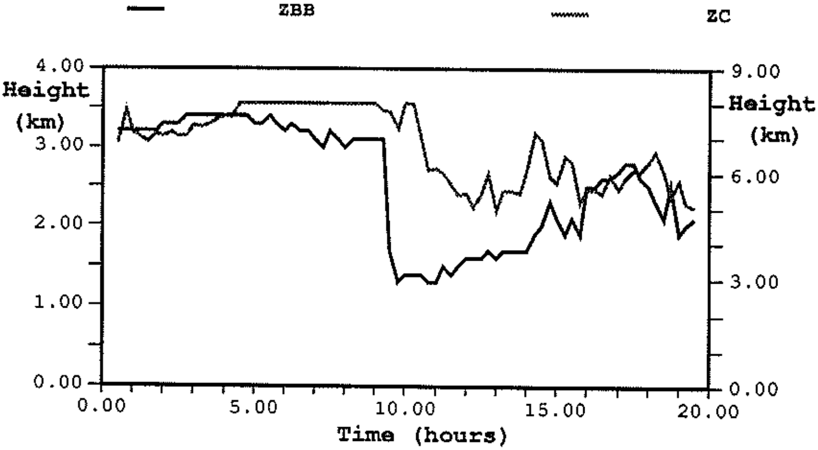


FIG. 7 — Time sequence plots for Case 1 on parametrisation coefficients ZBB and ZC.

Despite these being preliminary results, it is worth investigating the possible sources of the large bias, and to attempt to quantify these effects. As will be demonstrated later, this case study exhibits enhancement due to the flow over the hills. At the ranges considered in this study, the radar beam samples rain at altitudes between 0.75* and 1.80 km above mean sea level. Thus the radar may

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* This is for the Wellington radar, at a range of 60 km. The lowest figure is for an angle of 0°, the highest for 1.0°. These represent the half power angles of the radar beam.

fail to measure rain from the level at which this enhancement occurs. The correction for the domain-averaged VPR does not take into account this local low-level variation in the profile. This is shown in that the average ratio after VPR correction still indicates marked under-estimation.

Beam blocking is also a likely possibility. The region considered contains many mountains that may obscure part of the beam with the result that the returned signal would be significantly less than otherwise expected. Preliminary results indicate that the averaged topography lies below the 0.5° beam approach to 3 of the gauges, but intercepts the beam before the McIntosh gauge. This is reflected in the higher bias ratios for the McIntosh gauge. The CAPPI data for this gauge comparison performed better (ratio of 2.89 after correction) as less of the beam at 2 km was blocked by the hills.

However, when considering the maximum terrain heights, the 0° elevation barely skirts these peaks. For a 1° beam width at an elevation of 0.5°, approximately** 12% of the power will be at elevations of less then 0°. Presuming that this 12% is lost by blocking, adding 12% onto the power received would increase the average precipitation rate estimate of 5.0 mm.hr^{-1} to 5.4 mm.hr^{-1}. This, on its own, is too small an increase to explain the underestimation of the rainfall.

Orographic enhancement of the rain rate increases most in the lowest levels. When the lowest segments of the radar beam are blocked, this low-level enhancement will have no influence on the radar measurement. If, for example,

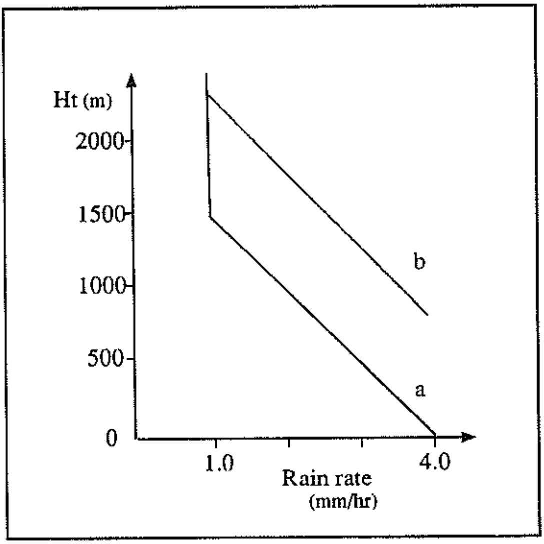

FIG. 8 — Model orographic profiles.

** By assuming a sin^{2}kx/(kx)^2 angular beam function, and that the radar has a half power width of 1°, it is possible to calculate the fraction of the beam that is at an angle of more than 0.5° (Brown et al., 1990) the rain rate increases linearly by a factor of 4 as it falls from 1500m aloft to the surface (Fig. 8a), the error in the radar rain rates can be estimated. If the Wellington radar had all the beam at angles below 0.0° blocked beyond a range of 75 km, it would measure only rainfall above a height of 750 m. Thus much of the enhancement cannot be observed at all. Assuming an angular beam distribution as above, the radar would observe a rate of 1.2 $mm.hr^{-1}$ instead of 4 $mm.hr^{-1}$. For an un-blocked beam a rate of 1.5 $mm.hr^{-1}$ would be inferred. If this orographic enhancement profile is "lifted" with the airflow over the hills (Fig. 8 b), then above a point of 870 m altitude, the radar will view a rain rate of 2.4 $mm.hr^{-1}$. Thus gauge to radar ratios of 1.65 to 2.75 may be expected for this region, under these idealised conditions.

This combination of beam blocking and overshoot of the orographic enhancement region is of the right order of magnitude to remove the bias in the rain gauge comparisons. However, more rigorous calculations are required for gauge-by-gauge corrections. These calculations would need to include the exact radar beam profile and estimates of the rain rate enhancement profile.

Drop-size distribution (DSD) problems may also have influenced the radar estimates. However, it is well known that the DSD can vary significantly (Hosking and Stow, 1987). The drop-size variations introduce errors by varying the relationship between reflectivity, Z, and rain rate R. Adjustment of the Z-R relationship could be made to account for the discrepancy between the rain gauge and radar data. For the large biases encountered here, this would result in values in the Z-R relationship well outside the expected range. Though the DSD variations may have had an influence on the radar estimates, their effects can be regarded as being second order when compared to those listed previously.

Average explained variances of around 40% match the experience of other operational experiments (Gray, 1991; Zawadzki et al., 1986; Harrold et al., 1974; Moore, 1989). The raw data exhibit the highest R^2, with the CRVPR correction decreasing the explained variance. However, the VPR correction improves the explained variance when applied to the CAPPI data.

For the RAW data set, the beam-1 data exhibit the greatest explained variance. For this example, the advantage of measuring close to the surface, and hence more of the low-level orographic enhancement, outweighs the disadvantages of beam blocking and ground clutter. However, when the broader scale features of the VPR are accounted for, data from beams 2 and 3 become better estimators of the surface rainfall.

The CAPPI data estimates are poor since they do not measure the rainfall as near the surface as possible. Measurements from near 2 km do remove contamination from ground clutter, but the local changes in reflectivity below this height reduce the information content in the observations. Correction for the general VPR improves both the explained variance and the mean bias of the CAPPI data.

Although the corrected data show only marginal improvement in terms of the explained variance and bias, there is a marked reduction in the standard deviation statistics. This implies that the spread in the mean errors has been decreased, and that the resulting data has a smoother relationship with the rain gauge data. This smooth field is more amenable to adjustment by rain gauge data (Collier et al.,1983; Gray 1991b).

Figures 6 and 7 contain estimates of the precipitation characteristics developed as part of the parametrisation of the vertical profile in the VPR correction algorithms. Alpha 1 represents the broad-scale average ground-level rain rate (normalised), Alpha 2 the decrease of rain rate with height (km^{-1}), Alpha 3 is the BB intensity (normalised), ZBB the BB height (km), and ZC is the echo top height estimate (km). These parameters provide estimates of the vertical structure of the precipitation. From these figures it can be seen that, for times up until 0900 GMT, the data exhibited only a slow change in echo top height (ZC) and BB height (ZBB). Alpha 2 and Alpha 3 show slow changes throughout this first section of the data. Alpha 2 increases from values of around -0.075 km^{-1} to values of -0.050 km^{-1}, indicating a general trend towards less rapid changes in reflectivity with height. Similarly the intensity of the BB decreases.

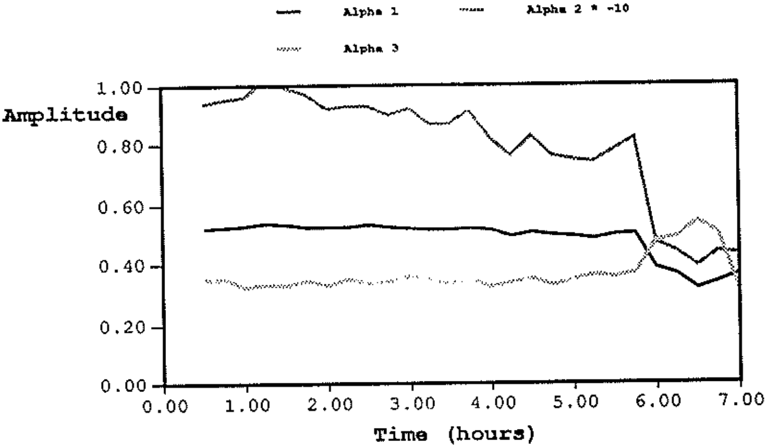

FIG. 9 — Time sequence plots for case 2 of parametrisation coefficients Alpha 1, Alpha 2 and Alpha 3. Note that Alpha 2 is scaled by -10

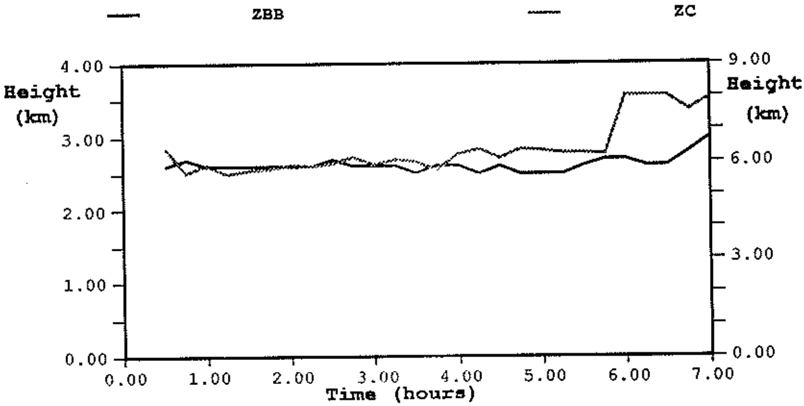

FIG. 10 — Time sequence plots for Case 2 of parametrisation coefficients of ZBB and ZC

Important changes in precipitation structure occurred at 0900 GMT. Within 30 minutes of that time the domain-averaged estimate of the precipitation depth decreased from over 8.0 km to 5.5 km. At the same time the apparent BB height decreased from 3.1 km to 1.4 km. From this point in time onwards, many of the parameters vary markedly. It appears that the synoptic scale support for the precipitation has moved in a NE direction (Fig. 4) and no longer lies over the catchment. The rain rate, as measured by the gauges, continued unabated and was maintained by a north-westerly flow which continued to form orographic rain. This continued until the southerly arrived, whereupon the rain stopped.

The utility of the coefficients in the parametrisation as estimators of physical characteristics of the rain needs further evaluation. For example ZBB appears to indicate a rapid change at 0900 GMT. Before the change, it indicated a melting level height of 3.1-3.4 km, which is consistent with temperature observations from Paraparaumu. After the change, its height was as low as 1.4 km. This would imply a rapid intrusion of very cold air into the radar domain. However, this appears inconsistent with other information. The temperature at Kaikoura (45.42S, 173.70E), in the coldest air, is 15°C with a dew point of 13°C (from Met Service NZ hourly observations). Even by assuming that this air was not unconditionally unstable, a melting level of 1.3 km is physically not possible. Indeed, BB heights below 2.4 km in this situation seem very unlikely. This very low BB height indicates that the use of the simple parametrisation scheme to provide estimates of these VPR attributes is not applicable in all cases. It appears that the parametrisation scheme has used the BB function to represent other aspects of the profile, in particular, it has accounted for changes in the rate of decrease of reflectivity with height. However, it is useful to have the extra degrees of freedom through having the BB function available to improve the fit of the measured profile by the parametrisation. If it was necessary for the BB function to represent only BB heights, then limits could be placed on ZBB, perhaps based on temperature soundings.

Coincident with these changes are the decrease in explained variance exhibited by the radar data, and the increase in the mean rain rate ratio. This is, again, consistent in that a decrease in the precipitation depth will decrease the quality of the radar estimate. For example, the explained variance of the CRVPR data around the Oriwa gauge, before the front, was 0.61, while after the transition it was 0.15. Similarly, ratios before average 2.11 for beam 1, while after, they are 4.83. This is in spite of the correction for average VPR induced errors. From viewing satellite imagery, it was apparent that by 0900 GMT, the deep precipitation had moved north-east over the domain and away from our area of interest. This left behind a shallow layer of moist north-westerly air, which enabled the rainfall to continue over the catchment, primarily due to orographic enhancement. It is apparent that the low-level flow was sufficient to keep significant amounts of rain falling into the gauges, while the decrease in the rain depth led the radar to further underestimate the rainfall. Further investigation of this change and the causes for the poor performance of the radar is needed to complete our understanding of this problem.

In future work balloon soundings of wind and temperature will play a vital role in determining the causes of the poor performance of the radar.

CASE 2

Table 3 contains the explained variances and bias ratios of the comparisons of the four radar data sets with the rain gauge measurements. Figures 9 and 10 show the coefficients of the parametrisations.

From Table 3 it is apparent that the radar data permitted significantly better estimates of the rainfall in this situation than was evident for Case 1. This is primarily a result of the lack of orographic enhancement. The correction algorithms have performed well here, increasing the average explained variance of both the raw and CAPPI data. Indeed the CRVPR data have the highest R^2, and the smallest standard deviations.

Here beams 2, 3, and 4 produced the best correlations, while beam 1 may have been contaminated with residual ground clutter and blocking. However, the mean ratios after correction are further from the ideal value of 1. Though this is disappointing, the primary aim of the correction for VPR errors is to reduce the standard deviation of the rain gauge to radar ratios, in order to provide a smoother field for post-correction adjustment by the rain gauge data.

Therefore more disappointing is the increase in the standard deviation when the CAPPI data are VPR corrected. The uncorrected CAPPI data perform well here, as the BB is found close to the CAPPI height of 2 km (Figs. 9 and 10). This would enhance the values observed, decreasing the mean ratio error. However, the correction for VPR induced error removes the BB enhancement and hence increases the mean ratio away from the ideal value of 1.0. From an inspection of the individual comparisons for Taungata, it is apparent that the RAW data at 4 of the 9 bins have ratios of less than 1, and probably have data that are contaminated with clutter. The correction algorithms alleviate these problems.

At greater elevations, especially when corrections are made for VPR-induced errors, the explained variance increases. This was not so in Case 1, as the low level enhancement played a significant role in determining the surface rain rate.

This weather system approached the Otaki basin from the south-west and showed little in the way of orographic modulation. The precipitation was remarkably stratiform in nature, with only small variations in BB height and precipitation depth (Figs. 9 and 10). There was, however, a general trend for the rain to become more convective as the event progressed. Early in the event the rate of decrease of rain rate with height was relatively rapid. Later on the structure was more vertical and hence convective. Estimates from the parametrisation algorithms show that Alpha 2, the rate of decrease of precipitation rate with height, goes from -0.09 km^{-1} early in the event to -0.04 km^{-1} towards the end (see Fig. 9). This is matched by an increase in the precipitation rate culminating in a heavy rainfall burst which abruptly ended the event.

RESULTS FROM STORM-AVERAGED RADAR RAINFALL RATES

Figures 11 and 12 show the Case 1, corrected, CAPPI data, for the entire radar domain. Figure 11 indicates the radar estimate in terms of reflectivity (as measured in dBZ), and Figure 12 shows the inferred rain rate ($mm.hr^{-1}$). Figures 13 and 14 are the same estimates for Case 2. From these plots various aspects of the average structure can be seen.

The area to the east of the North Island has significantly lower radar rainfall totals than for the surrounding areas. This may result from the blocking of the lower segment of the radar beam which would reduce the returned signal from this area. Also apparent from these images is the residual clutter over the Kaikoura Ranges (42°S, 173.5°E).

But of most interest is the area of higher rain rates over the Otaki catchment (40.8°S 175.2°E) for Case 1 which is not present in Case 2 (Figs. 13 and 14).

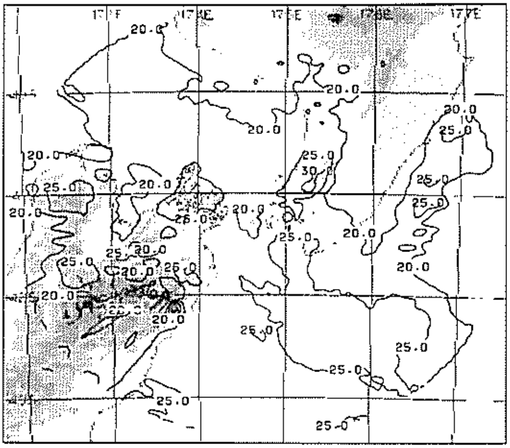

FIG. 11 — Plots of average reflectivity (dBZ) for Case 1.

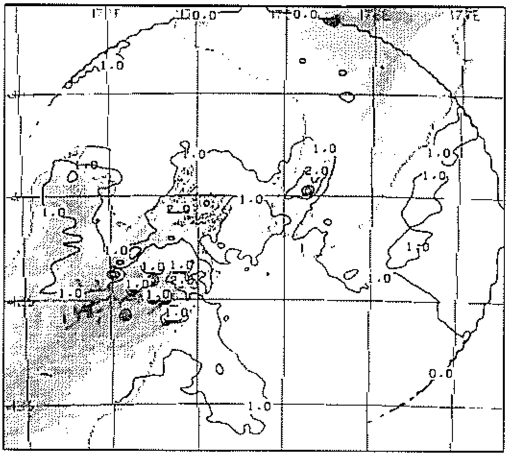

FIG. 12 — Plots of average rain rate (mm. hr^{-1}) for Case 1.

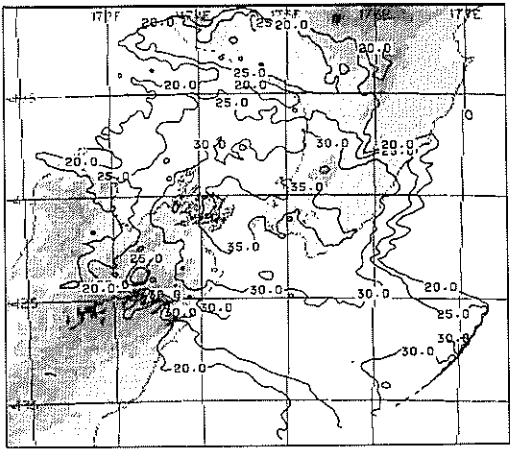

FIG. 13 — Plots of average reflectivity (dBZ) for Case 2.

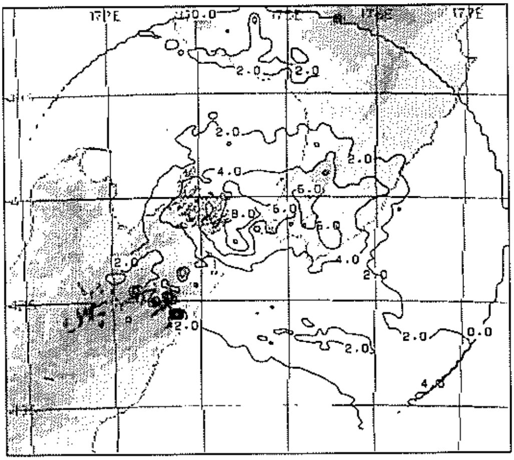

FIG. 14 — Plots of average rain rate (mm. hr^{-1}) for Case 2.

From Figures 11 and 12 an estimate of the degree of enhancement can be made. The average offshore rain rate adjacent to the catchment is around 1 mm.hr^{-1}, over the catchment values reach 2 mm.hr^{-1}. However, the radar estimates appear to underestimate the rainfall over the catchment, as has been discussed earlier. If the radar estimates the rainfall offshore accurately, but only sees half the enhancement onshore (i.e. a ratio of 2), then orographic enhancements of a factor of 4 could be expected. This is in keeping with the ideas stated earlier, as well as with model estimates (Grabowski, 1989) and anecdotal forecasting experience.

DISCUSSION AND CONCLUSIONS

In the introduction it was stated that radar has great potential application as input for hydrological modelling. This potential primarily lies in radar's ability to make real time observations of areal averages of precipitation. However, the realisation of this potential is hindered by the biases and variability induced by the various error sources.

From the research comparing radar and rain gauge observations, it has been shown that the ratios of the means of the rain gauge to radar estimates lie between 2 and 4. This was shown to be consistent with the radar failing to measure the low-level growth of precipitation that results from orographic enhancement. Attempts have been made to correct for this problem by modelling the enhancement as a linear increase in rain rate in the lowest 1500 m (Kitchen and Blackall, 1991). Application of these ideas to New Zealand conditions will require investigation of the detailed vertical structure of the enhancement, and its relationship to wind speed and direction, and humidity, in order to formulate a correction strategy tuned to the weather encountered here.

Also shown to be important is blocking of the lowest sections of the radar beam, leading to a further underestimation of the rainfall. Correction for this problem can be made. The current operational system has the facility to make corrections for beam blocking, assuming a constant rain rate with height. Future work will involve generating this information from detailed topographic maps.

Much of the recent radar research effort has been placed into attempting to quantify and correct for the errors induced by the vertical structure of precipitation (Gray, 1991; Fabry and Austin, 1991; Joss and Waldvogel, 1990). The approach of correction made on a single volume-scan basis was shown here to decrease the spread in the bias ratios, but made little impression on the absolute bias of the radar estimates. The biggest positive impacts of the correction method lie at ranges beyond about 120 km, where the radar beam climbs through the melting-level and samples rainfall in the regions of marked vertical structure. At the ranges used in this study, only the heights below 2 km are sampled by the radar's lowest beams. Profiles at this level are strongly influenced by the wind flow over the topography and this underlines the need for orographic enhancement corrections.

Ground clutter remains an important problem that is yet to be fully resolved. The clutter rejection techniques largely removes the wave clutter, but some residual hill clutter remains. Improvements in the hill clutter rejection algorithm are necessary in order that the radar data used for comparison can be said to truly represent the rainfall.

One factor needing further investigation is the effect of the attenuation correction scheme applied operationally. This correction begins at a range of 10 km and limits corrections to 4.8 dB. The rationale for these limits is to eliminate the possibility of the correction producing rain rates which are not physically realistic. However, in the light of the large residual bias, further consideration may need to be given to these limitations and the algorithms used, particularly the somewhat arbitrary choice of beginning at 10 km, as attenuation is a progressive phenomena with range, and the attenuation at close ranges may well induce marked errors at longer ranges.

If the hydrological potential of the Meteorological Service's radar is to be fully exploited, it would be desirable to reconsider the operational timing cycle of the radar scans to improve the time resolution of the radar data to 10 or even

5 minutes. Studies elsewhere have shown that poor time resolution has a major impact on accumulation accuracy (Fabry and Austin, 1991).

One of the primary aims of this study was to compare the various forms of radar data that are available. Though this is only the initial study, it was worth making an attempt to highlight the estimators that appear to have the greatest potential. From these results it can be seen that there are only slight differences in the quality of the radar estimates and that each data source was biased. A final assessment of which source is most reliable will require the investigation of many more case studies. This, along with the work on improving the correction techniques, will form the core of the main OPERA project.

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