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METEOR. ASPECTS OF RADAR DEPOLARIZATION AT 16.5 GHz

METEOROLOGICAL ASPECTS OF RADAR DEPOLARIZATION

AT 16.5 GHz

By

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ABSTRACT

Radar polarization measurements of precipitation at 16.5 GHz are analyzed with reference to the ambient temperature profile of the atmosphere. The measured quantities are the reflectivity, the ratio of the main and cross-polarized signal powers, called the cancellation, and the correlation between these two signals.

Polarization observations of the bright band are discussed. It is suggested that correlation gives an indication of the relative number of water drops in the scattering ensemble. From the characteristics of correlation and cancellation observations it is inferred that small snowflakes melt first with the result that radar returns are dominated by large water-covered snow aggregates at levels just below the reflectivity peak in the bright band.

Correlation measurements are used to infer possible precipitation types at various altitudes in convective precipitation. A region of mixed-phase scatterers is postulated above the 0°C level, with ice-phase scatterers above and rain or rain-like particles below.

RÉSUMÉ

Des observations de précipitation, prises à l'aide d'un radar polarisateur à 16.5 GHz, sont analysées en rapport avec les données de température atmosphérique ambiante. Le coefficient de réflexion, le quotient des deux signaux polarisés, appelé le quotient de cancellation, et la corrélation entre ces signaux sont observés.

Les observations de polarisation dans la bande brillante sont discutées. Il est suggéré que l'analyse de corrélation indique le nombre relatif de gouttelettes d'eau dans le milieu diffusif. Des observations de corrélation et du quotient de cancellation suggèrent que les petits flocons de neige fondent en premier, occasionnant à ces niveaux un retour d'ondes largement dominé par de large masses de flocons imbibés d'eau liquide.

Des types de précipitation à différentes altitudes dans une cellule d'orage convective sont suggérés selon ces données. Une région de diffuseurs multi-phase est postulée au niveau immédiatement supérieur à celui de 0°C, ayant la phase d'eau solide au haut et la phase d'eau liquide au bas.

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TABLE OF CONTENTS

	<u>Page</u>
ABSTRACT	ii
ACKNOWLEDGEMENTS	iv
LIST OF TABLES	vi
LIST OF FIGURES	vii
I. INTRODUCTION	1
II. THE OTTAWA RADAR AND DATA FORMAT	
2.1 Radar Facilities	3
2.2 The Polarization Parameters	4
2.3 Data Selection	12
III. OBSERVATIONS IN STRATIFORM PRECIPITATION	
3.1 Mean Bright Band Profiles	13
3.2 10.4 cm Observations of the Bright Band	18
3.3 Other Bright Band Observations	20
3.4 Parameter Distributions in Stratiform Rain	22
IV. OBSERVATIONS IN CONVECTIVE PRECIPITATION	
4.1 Introduction	24
4.2 ORTT as a Function of Height in Convective Rain Data	24
4.3 The Low-ORTT Region	29
4.4 An Individual Elevation Scan	34
V. CONCLUSIONS AND DISCUSSION	
5.1 Conclusions	41
5.2 Experiments for Future Consideration	42
REFERENCES	44

LIST OF TABLES

Facing Page

- | | | |
|-----|--|----|
| 2.1 | Characteristics of the Ottawa radar. | 3 |
| 3.1 | The distribution of distances (d) between 0°C and the reflectivity peak in the bright band for 42 observations in stratiform rain. | 22 |

LIST OF FIGURES

	<u>On or Facing Page</u>
2.1 (a) Display of CAN for an elevation scan from 10 July 1972.	5
(b) Display of ORTT for an elevation scan from 10 July 1972.	6
(c) Display of Pz for an elevation scan from 10 July 1972.	7
(d) Display of ALD for an elevation scan from 10 July 1972.	8
3.1 (a) Display of CAN for an elevation scan through stratiform rain recorded 28 August 1971.	14
(b) Display of ORTT for an elevation scan through stratiform rain recorded 28 August 1971.	15
(c) Display of Pz for an elevation scan through stratiform rain recorded 28 August 1971.	16
3.2 Mean vertical profiles of Pz, CAN and ORTT for elevation scans through the bright band in stratiform precipitation recorded 9 December 1971.	17
3.3 Vertical profiles of Pz, CAN and ORTT ob- served in stratiform precipitation in Alberta at 10.4 cm 0028-0034 MDT 25 May 1972.	18
3.4 Mean vertical profiles of Pz, CAN and ORTT for elevation scans through the bright band in stratiform precipitation recorded 28 August 1971.	20
3.5 Mean vertical profiles of Pz, CAN and ORTT for elevation scans through the bright band in stratiform precipitation recorded 30 May 1972.	20
3.6 Mean vertical profiles of Pz, CAN and ORTT for elevation scans through the bright band in stratiform precipitation recorded 6 October 1971.	21

List of Figures (cont'd)

On or Facing Page

3.7	Theoretical calculations of cancellation and reflectivity for Marshall-Palmer rain.	21
3.8	The distribution of Pz, CAN and ORTT in rain below the bright band.	23
4.1 (a)	Display of CAN for an elevation scan through convective precipitation recorded 10 July, 1972.	25
(b)	Display of ORTT for an elevation scan through convective precipitation recorded 10 July 1972.	26
(c)	Display of Pz for an elevation scan through convective precipitation recorded 10 July 1972.	27
4.2	Vertical profile of ORTT for an elevation scan in convective rain recorded 4 September 1970.	28
4.3	The distribution of ORTT as a function of height in convective storm data.	29
4.4 (a)	Display of CAN for an elevation scan through convective precipitation recorded 14 August 1973.	30
(b)	Display of ORTT for an elevation scan through convective precipitation recorded 14 August 1973.	31
(c)	Display of Pz for an elevation scan through convective precipitation recorded 14 August 1973.	32
4.5	Scatter diagram of the height of the 40% contour above and below the low-ORTT region.	33
4.6 (a)	Display of CAN for an elevation scan through convective precipitation recorded 25 July 1972.	35
(b)	Display of ORTT for an elevation scan through convective precipitation recorded 25 July 1972.	36

List of Figures (cont'd)

On or Facing Page

- 4.6 (c) Display of Pz for an elevation scan through convective precipitation recorded 25 July 1972.

37

CHAPTER I

INTRODUCTION

Scientists at the National Research Council of Canada have developed two polarization diversity radars in use in Canada. These include a 10.4 cm radar at Penhold, Alberta, used primarily in the study of hail storms and a 1.82 cm radar at NRC in Ottawa used in the study of precipitation and its influence on the transmission of millimeter wavelength radio waves in the atmosphere.

The propagation and scattering theory developed by McCormick and presented by Barge (1972) and Humphries (1973), has been used in previous treatments of data from these radars. This theory shows that four independent measurable quantities may be used to describe the polarization characteristics of the signals returned from precipitation. These quantities represent the returned power, the cancellation ratio and the relative phase angle and correlation between the main and depolarized signals.

These parameters have been related to the observed precipitation in studies by Barge and Humphries from radar data collected in Alberta. Barge (1972) considered combinations of reflectivity and cancellation observed below the freezing level and related them to the occurrence of hail at the surface. He determined that the two parameters taken together could be used to indicate the occurrence of hail at the ground better than either of the quantities taken alone. Humphries (1973) examined the range dependency of the radar parameters (especially of the relative phase angle) and concluded that significant propagation effects can be observed at 2.88 GHz in rain as well as in hail.

Some observations from Ottawa have been related to meteorological parameters by McCormick and Hendry (1970, 1972) and by Hendry and McCormick (1971, 1974). These include the observation that correlation between the receiver channels is usually large and an increasing function of reflectivity in rain. They also report that relative phase angle measurements indicate that raindrops tend to fall with their major axes nearly horizontal.

RHI data have been collected in Ottawa since 1970. During this time observations have been concentrated on high reflectivity regions in precipitation. Consequently, these data are not comprehensive records of a given storm in space or time, nor do they represent all storms observable on a given day.

The purpose of this thesis was to determine some of the features of Ottawa data and to suggest how these might be explained in terms of meteorology. Efforts have been concentrated on relating the observed pattern of correlation values to the height of 0°C in the atmosphere. It appears from this study that correlation values may indicate information about the phase state of scatterers at various heights in the bright band and in convective showers.

Table 2.1. Characteristics of the Ottawa radar.

peak power	- 35 kw
pulse duration	- 0.4 μ sec
PRF	- 2000 sec^{-1}
frequency	- 16.5 GHz
beam width	- 0.89° (all planes)
antenna gain	- 43.0 dB (at transmitter terminals)
maximum elevation angle	- 35° (90° from June 1974)
azimuth sector	- 350° (approx.)

CHAPTER II

THE OTTAWA RADAR AND DATA FORMAT

In this chapter some of the characteristics of the Ottawa radar are described. The system and related theory have been described in greater detail by Hendry and McCormick (1968, 1971).

2.1 Radar Facilities

The Ottawa radar was designed and built by scientists at NRC to study propagation and scattering effects of precipitation at 16.5 GHz. The major characteristics of this radar are presented in Table 2.1.

The antenna is capable of transmitting any polarization and of resolving the returned signals into the transmitted polarization and the one orthogonal to it. The dual waveguide system employed with this antenna does not permit full 360° rotation in azimuth. Consequently, the radar system has been designed to enable the collection of data in an RHI mode.

The dual channel receiving system (one for each polarization) is equipped to measure four independent parameters related to the polarization characteristics of the returned signals. The four resulting video signals are sampled in twelve range gates after each transmitted pulse.

The 1 μ sec range gates are spaced 500 m apart, forming a sampling array with a total range extent of 5.5 km. This array is positioned at any desired range up to about 70 km to record precipitation echoes of interest. Data are recorded from the range gates while scanning the antenna in elevation, creating two-dimensional range-height records essentially the same as RHI's.

Samples of the four video signals in each of the twelve range gates are averaged by an array of 48 analogue integrators. The integration time constants are equal to 0.3 sec, corresponding to the time required for 600 samples (transmitted pulses). The integrator outputs are digitised and recorded at 1-second intervals during the observation period. Auxiliary information such as antenna position, date and time are included in the digital magnetic tape record.

Analysis of the raw data is executed by a conventional digital computer at NRC. The integrated video records are used to calculate a set of four independent radar parameters designated as Pz, CAN, ORTT and ALD, according to formulae given by McCormick (1968), McCormick and Hendry (1970, 1972) and Hendry and McCormick (1971). The computer prints a set of four range-height displays (one for each parameter) for every antenna elevation scan on record. A set of outputs, such as shown in Fig. 2.1, will be described hereafter as an "elevation scan".

The presentation of data in this computer-printout form results in printing 'errors' of two types. The finite number of character locations requires that the data positions be printed only approximately. These location errors are less than 167 m (1 line) and have not proven to be a serious problem. Secondly, the height axis is compressed slightly with respect to the range axis. This again is not a serious problem since exact comparison of distances between the horizontal and vertical has not been attempted here.

2.2 The Polarization Parameters

As noted earlier, four quantities are determined from the signals in each range gate. These quantities and the symbols used to identify them

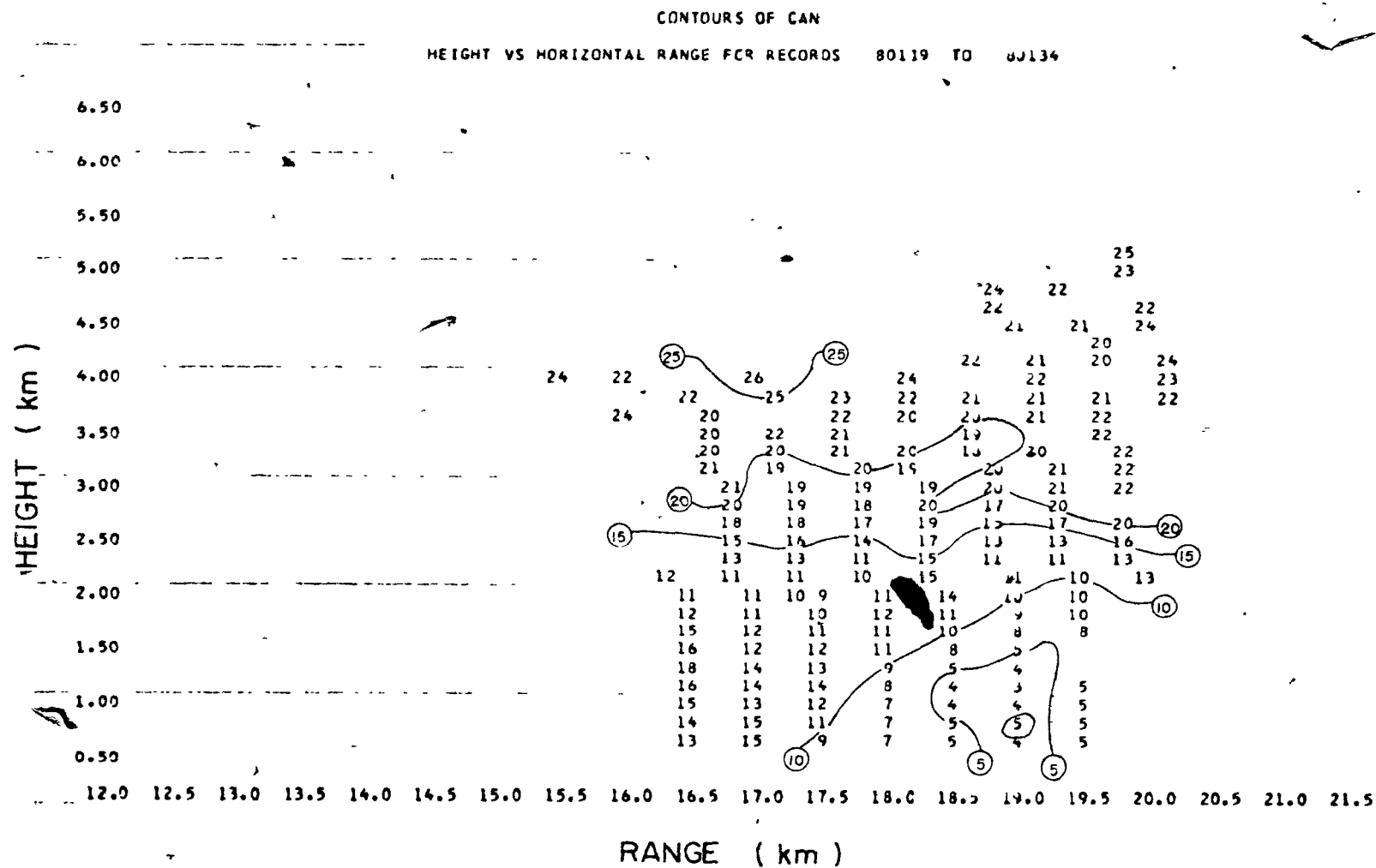


Fig. 2.1(a) Display of CAN for an elevation scan from 10 July 1972. CAN, the cancellation, represents the ratio of the powers in the main and depolarized receiver channels.

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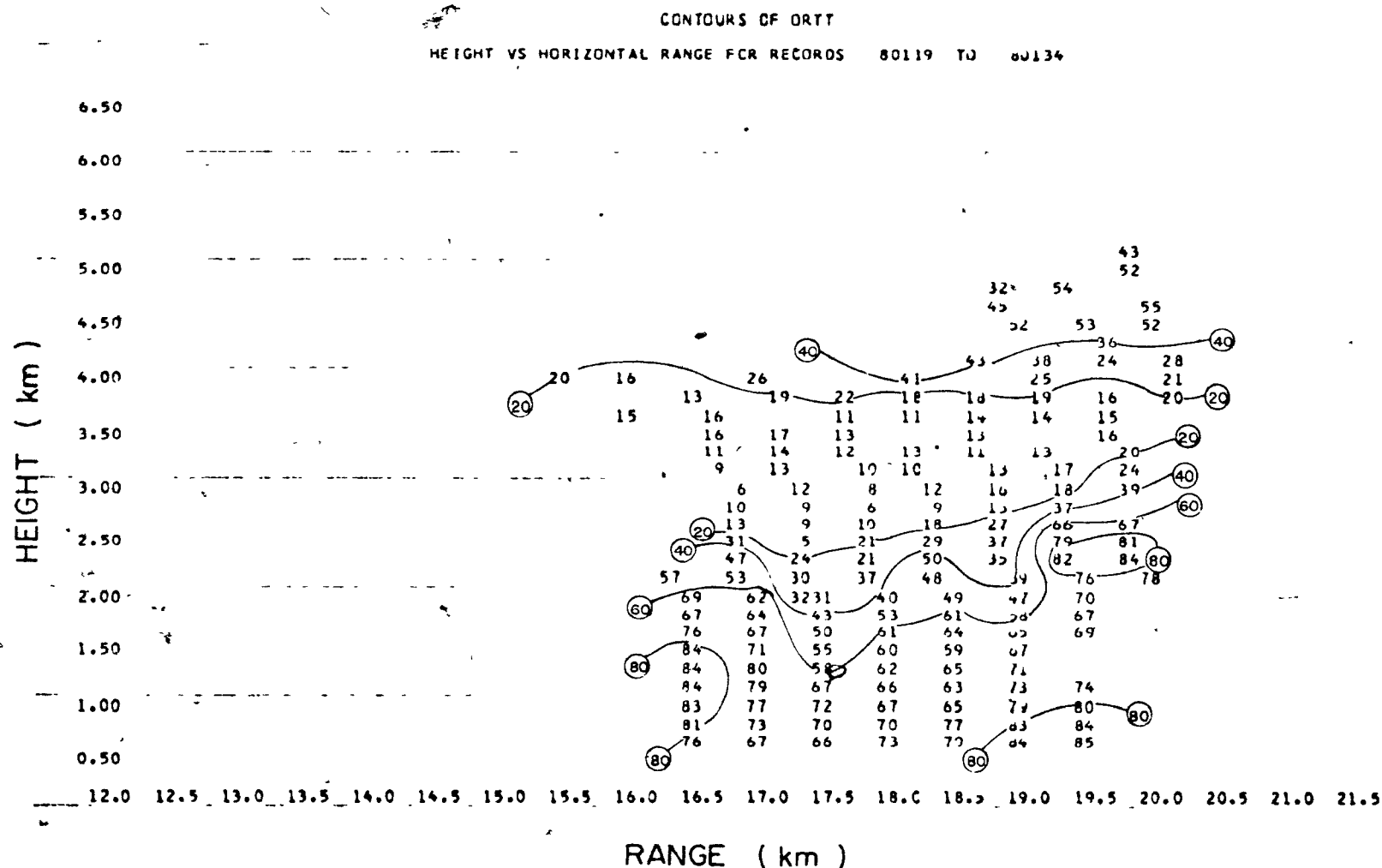


Fig. 2.1(b) Display of ORTT for an elevation scan from 10 July 1972. ORTT represents the correlation between the main and depolarized signals.

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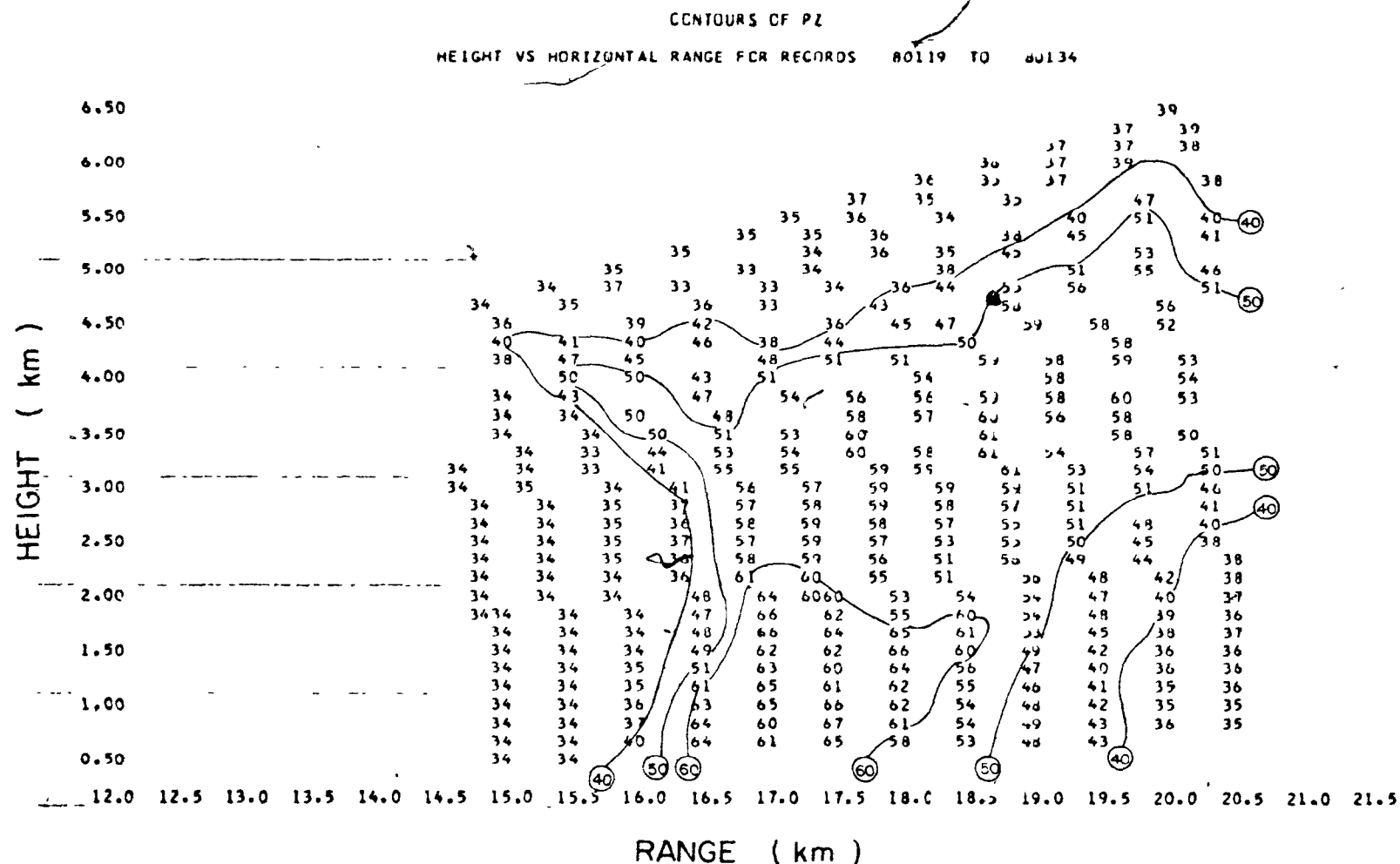


Fig. 2.1(c) Display of P_z for an elevation scan from 10 July 1972. P_z represents the range normalised backscattered power in the main channel.

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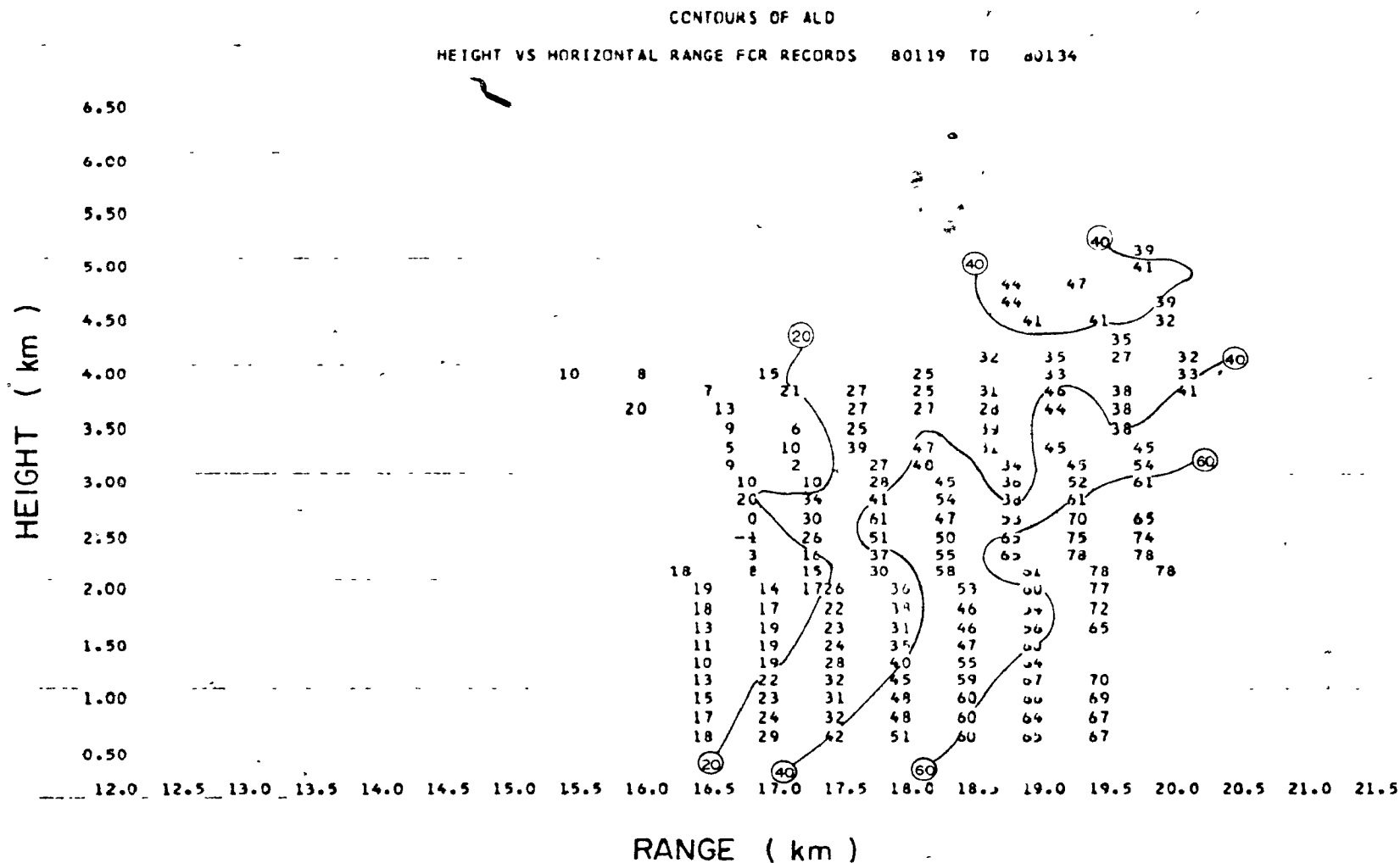


Fig. 2.1(d) Display of ALD for an elevation scan from 10 July 1972. ALD is a quantity derived from the average relative phase angle between the main and depolarised signals.

are as follows:

P_z , the range-normalized backscattered power in the main channel.

CAN, the cancellation, which represents the ratio of the powers in the main and depolarized channels.

ORTT, the correlation between the main and depolarized signals.

ALD, a quantity derived from the average relative phase angle between the main and depolarized signals.

These quantities are essentially the same as those described by Barge and Humphries although the terminology follows that of McCormick and Hendry. These radar parameters will now be discussed more specifically and in turn.

P_z

Hendry and McCormick (1971) indicated that P_z is related to the commonly used reflectivity factor Z_e by

$$P_z(\text{dB}) = 10 \log_{10} Z_e + 20 \text{ dB} \quad (2.1)$$

Of the four parameters, P_z is measured the most directly and therefore is subject to the smallest measurement error. However, large errors in P_z can occur from attenuation. Crane (1971) suggests, for example, that attenuation at 1.87 cm is greater than 1 dB/km at 20 mm/hr rainfall rate. Therefore, P_z measurements are not reliable when significant amounts of precipitation occur in the observation path.

CAN

This quantity, measured in decibels, may be defined by

$$\text{CAN}(\text{dB}) = 10 \log_{10} \left(\frac{P_2}{P_1} \right) \quad (2.2)$$

where P_2 is the power measured in the main channel and P_1 is the power measured in the cross polarized channel. This definition of cancellation is similar to the definition of circular depolarization ratio used by Barge and Humphries except for a difference of sign. Circular transmitted polarization was used exclusively for the observations considered here.

Calculations by Barge and later by Humphries based on measured droplet size distribution and size-shape relationship for oblate spheroids, have indicated that cancellation decreases with increasing rainfall rate. Cancellation values are observed to be greater than 25 dB in very light rain and snow and less than 20 dB in heavy rain (McCormick and Hendry, 1970).

Humphries (1973) has shown that differential propagation effects when present, tend to cause CAN values to decrease with range. Cancellation is often observed to be range dependent at 1.82 cm as shown in Fig. 2.1 at low elevation. It is clear that CAN values best represent the scatterers under clear path conditions.

ORTT

This parameter is derived from the correlation between the two returned signals. It is a measure of the tendency for the relative phase angle between the two receiver channels to remain constant over the integration time (0.3 sec). ORTT can assume any value between 0% and 100%. These limiting values occur when the relative phase angle varies widely and when it is constant, respectively.

Hendry and McCormick (1968) suggest that ORTT is a measure of the tendency for the observed precipitation to have a preferred orientation. In particular, for an idealized precipitation model in which all droplets

have the same size, shape and dielectric constant, ORTT represents the percentage of droplets having the same, fixed orientation angle in space. All remaining droplets are assumed to be randomly oriented.

Hendry and McCormick (1971) have determined characteristic ORTT values for various precipitation types. Values in rain below the melting layer are typically 60% or greater, and values in snow observed near the ground in winter are usually 40% or less. ORTT values around 20% are reported for observations in the bright band in stratiform rain. Humphries (1973) reports that correlation values are smaller in rain observed at 10.4 cm in Alberta than those at 1.82 cm in Ottawa.

Hendry and McCormick (1971) point out that ORTT indicates the relative number of scatterers with a preferred orientation only when propagation effects are negligibly small. For this reason correlation values best describe the scattering process under clear path conditions.

ALD

Whenever there is sufficient power in the cross polarised channel to be detectable and when the correlation is non zero, the relative phase angle between the receiver channels can be measured. With the Ottawa radar system, measurements of the relative phase angle are reliable whenever ORTT exceeds about 20%.

In the absence of propagation effects and non Rayleigh scattering effects, ALD represents the angle of preferred orientation of the scatterers. ALD has not been considered in the present analyses since it usually shows pronounced range dependency as in Fig. 2.1. The range-height display of ALD has been omitted from elevation scans presented in Chapters III and IV.

2.3 Data Selection

As noted, propagation effects at 1.82 cm result in pronounced range dependency of the radar parameters in moderate to heavy rain. In order that the data included here represent the scattering process as opposed to propagation effects, care was taken to exclude observations with significant amounts of precipitation in the path.

The exclusion of propagation effects is necessary since the radar parameters are interpreted here in terms of scattering theory. For elevation scans of convective rain, data points beyond the reflectivity maximum were disregarded. This procedure assured that at least those observations most strongly influenced by propagation effects were not included.

In stratiform rain propagation effects were not observed to be as important, probably because of the relatively low rainfall rates.

CHAPTER III

OBSERVATIONS IN STRATIFORM PRECIPITATION

In this chapter, observations of Pz, CAN and ORTT in stratiform precipitation are presented and discussed. The present data are compared with observations by others, with special emphasis on the characteristics of ORTT. The results are used in Chapter IV to help with the interpretation of convective rain data.

3.1 Mean Bright Band Profiles

Radar observations in stratiform precipitation presented by McCormick and Hendry (1972) and Hendry and McCormick (1974) show contours of Pz, CAN and ORTT that are virtually horizontal. Furthermore, the only large vertical gradients of these parameters occur near the melting layer.

An elevation scan through the bright band similar to those mentioned above, appears in Fig. 3.1. A high degree of horizontal uniformity is apparent. This suggests that horizontal averaging can be used to construct a representative vertical profile for each parameter. Mean bright-band profiles from four occasions were constructed as explained below.

The data from three adjacent range gates (as shown in Fig. 3.1) were averaged to determine a mean vertical profile for each elevation scan from a given day. The range gates were chosen to include data above the bright band to be as near to the radar as possible in order to minimize the influence of propagation effects. The mean profile determined from the three range gates was used with similar profiles from other elevation scans on a given day, to construct a single profile for each parameter showing the mean characteristics of the bright band for that day.

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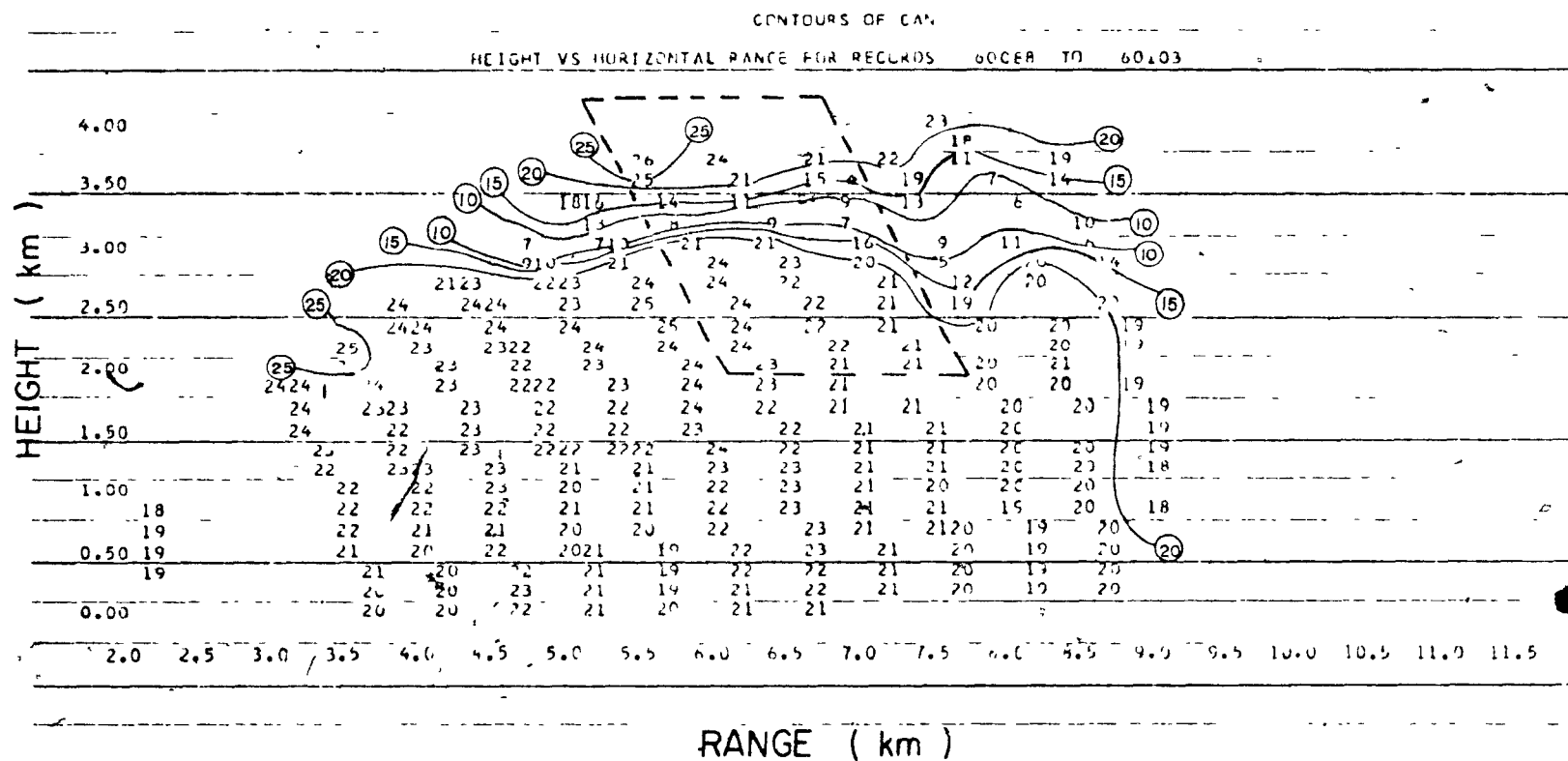


Fig. 3.1(a) Display of CAN for an elevation scan through stratiform rain recorded 28 August 1971. The enclosed region contains the data from this scan used in the construction of the mean vertical profile of the bright band for this day.

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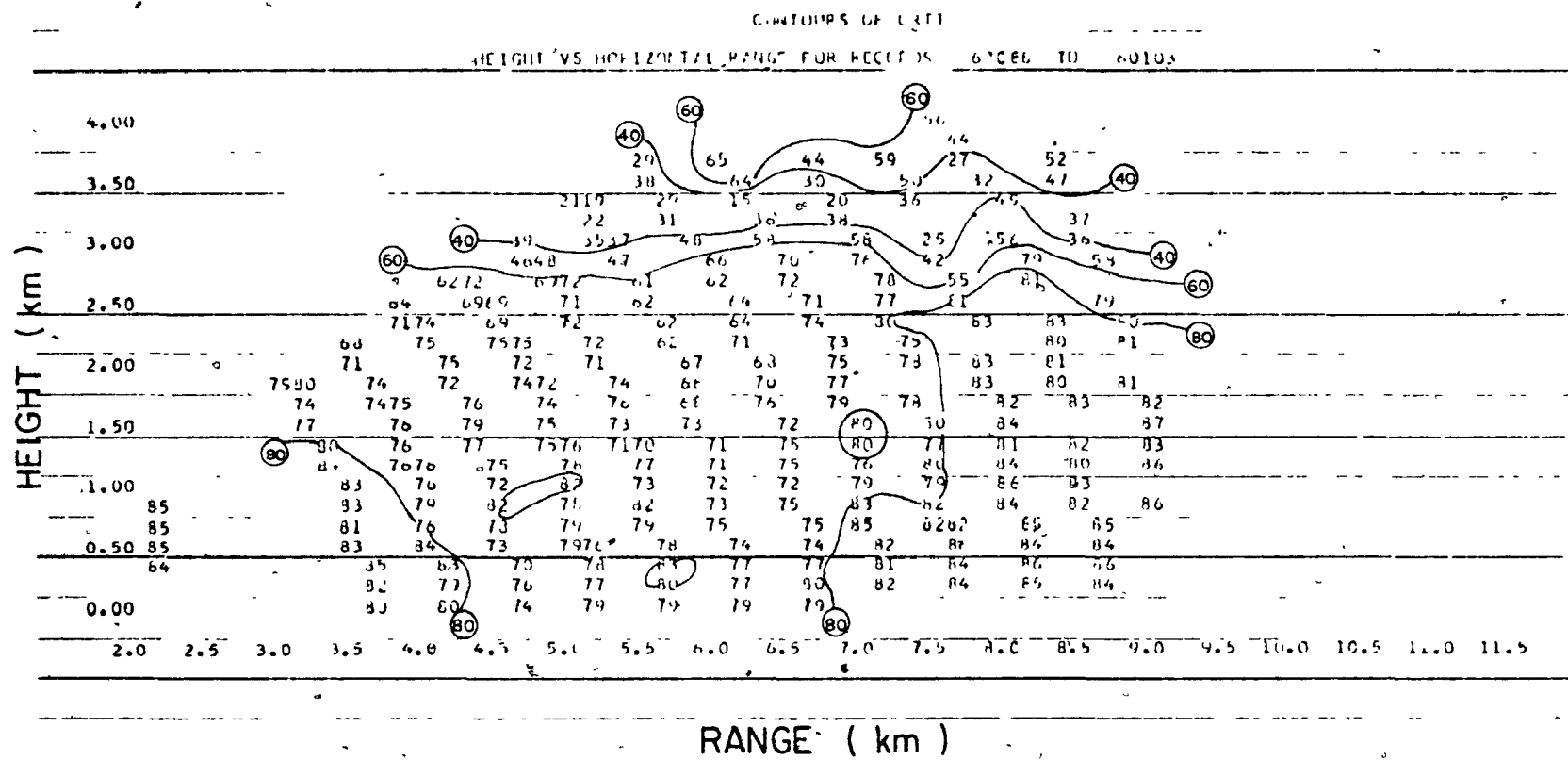


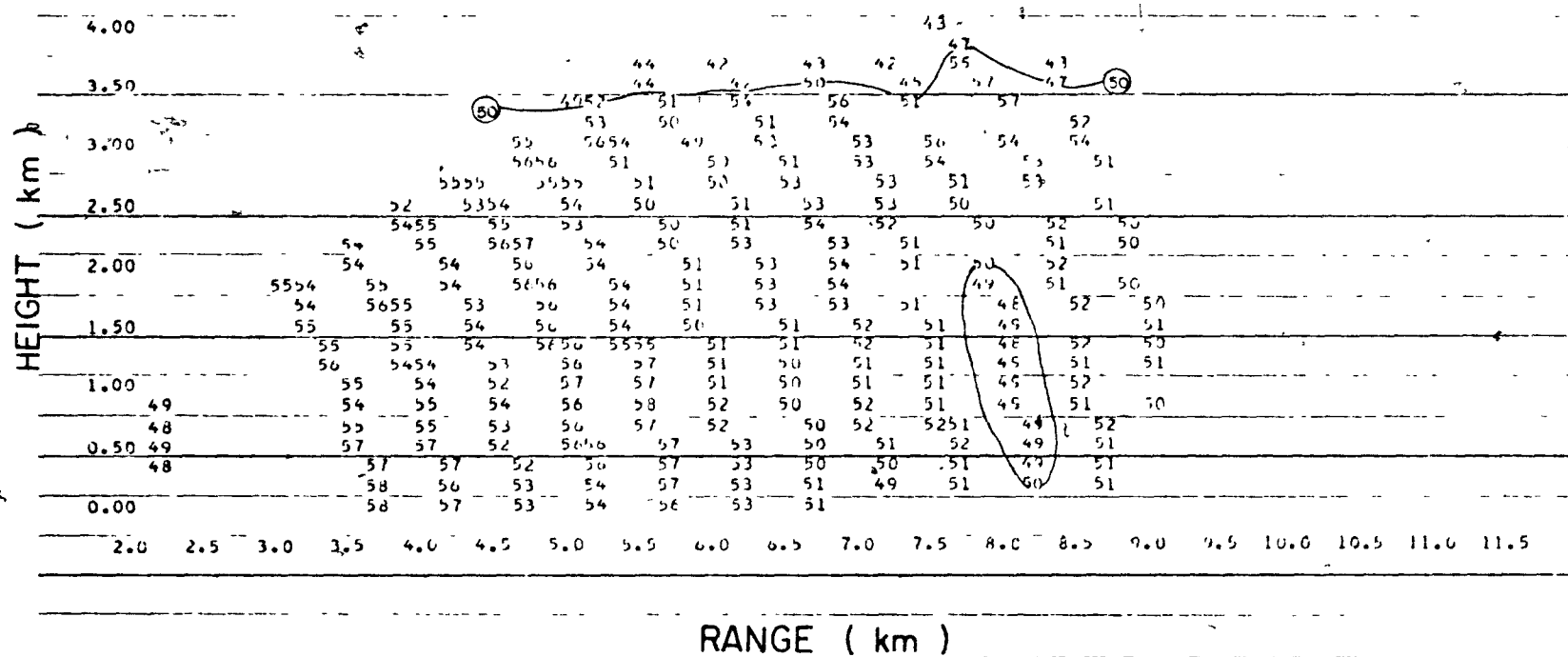
Fig. 3.1(b) Display of ORTT for an elevation scan through stratiform rain, recorded 28 August 1971.

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CONTOURS OF P_z

HEIGHT VS HORIZONTAL RANGE FOR RECORDS, 6000 TO 60103

Fig. 3.1(c) Display of P_z for an elevation scan through stratiform rain recorded 28 August 1971.

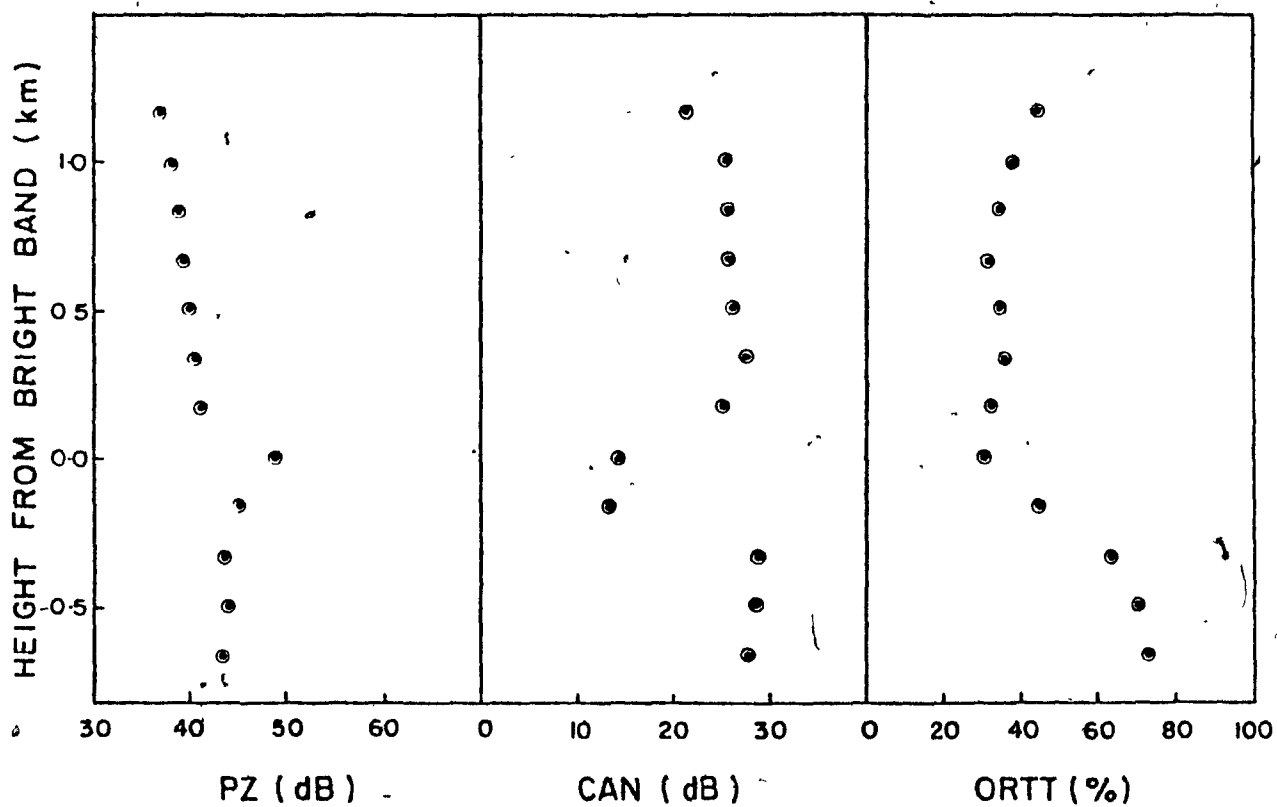


Fig. 3.2 Mean vertical profiles of P_z , CAN, and ORTT for elevation scans through the bright band in stratiform precipitation recorded 9 December 1971.

The mean vertical profiles of the bright band on 9 December 1971 appear in Fig. 3.2. This figure demonstrates many of the characteristics of the bright band discussed by Hendry and McCormick (1971, 1974) and by Humphries (1973). For example, the mean Pz profile shows the bright band reflectivity is only 5 dB greater than that of the rain. This value falls outside the 6 dB to 10 dB range quoted by Battan (1973) as characteristic of most observations. It might be expected that the averaging employed here would tend to smooth out the bright band maximum. However, it is apparent from Fig. 3.1 and from other stratiform rain elevation scans (McCormick and Hendry, 1972, Hendry and McCormick, 1974) that the weak bright band is a characteristic of the individual measurements. This peculiarity of the observations probably results from the short wavelength (1.82 cm) used at Ottawa, and consequent non-Rayleigh scattering.

A second characteristic of Fig. 3.2 which was previously observed by Hendry and McCormick (1974) is the well pronounced (15 dB) cancellation minimum in the melting layer. Fig. 3.2 indicates further, as do all available profiles, that the minimum CAN value occurs one height interval (167 m) below the reflectivity maximum as discussed by Humphries (1973) (see section 3.2).

The vertical profile of ORTT in Fig. 3.2 shows good agreement with values quoted by Hendry and McCormick (1971) for snow and rain. They report that snow is characterised by ORTT values less than 40%, as observed above the bright band in Fig. 3.2. Below the bright band, correlation values are greater than 60%, as suggested by Hendry and McCormick for rain.

The minimum correlation value in Fig. 3.2 and in all other bright band examples (Figs. 3.4, 3.5 and 3.6 in Section 3.3), occurs at the same

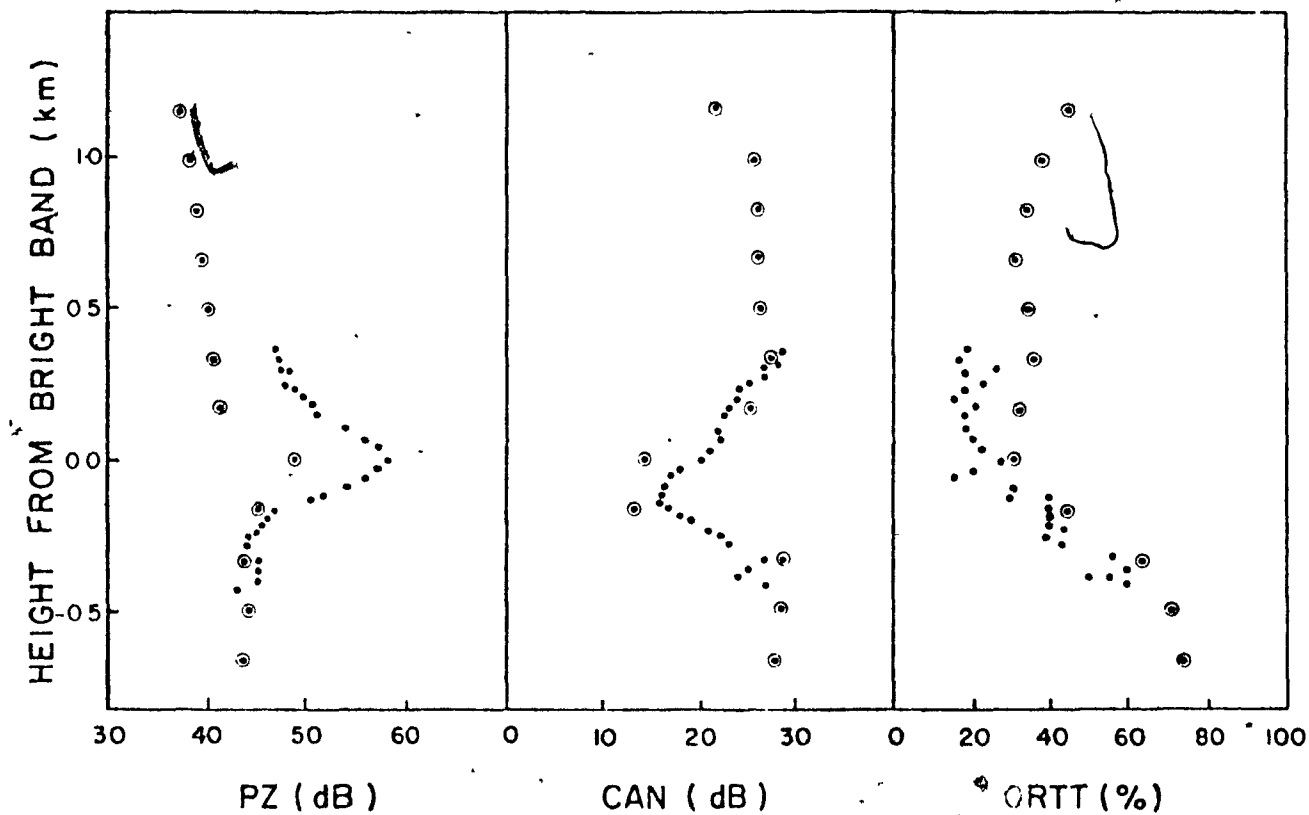


Fig. 5.3 Vertical profiles of Pz, CAN and ORTT observed in stratiform precipitation in Alberta at 10.4 cm 0028 - 0034 MDT 25 May 1972 (plain points). The Ottawa observations (circled points) are from 9 December 1971 as in Fig. 3.2. The Alberta data are adapted from Humphries (1973).

level as the reflectivity peak. The values at this minimum (near 20% in Figs. 3.4, 3.5 and 3.6) are similar to those reported by Hendry and McCormick (1971) for observations in the bright band.

3.2 10.4 cm Observations of the Bright Band

Humphries (1973) presented three unaveraged vertical profiles of the bright band measured in Alberta at 10.4 cm. A representative profile from his observations and the data from Fig. 3.2 are shown together in Fig. 3.3 for comparison.

The profile sets differ in two ways: the reflectivity peak is more pronounced and the correlation values are lower in the Alberta data than in the Ottawa data. As mentioned, the bright band is weak in all observations at Ottawa at 1.82 cm. This may account for the large difference in reflectivity between the 10.4 cm profile and 1.82 cm profile at the reflectivity maximum. Also, the tendency for the correlation values to be smaller in the Alberta observations is consistent with the observation by Humphries (1973) that in general, rain observations at 10.4 cm in Alberta show less correlation than those at 1.82 cm in Ottawa.

These differences aside, the agreement of the two data sets is quite good. The cancellation profiles are similar in value everywhere and both demonstrate the tendency for the minimum value to occur just below the reflectivity maximum. The correlation profiles are similar as well. They show nearly the same increase from small values at the reflectivity peak to larger values below in the rain. This rapid increase of ORTT values is confined to a 500 m interval immediately below the reflectivity maximum in both profiles.

Humphries (1973) speculated about the processes occurring in the bright band. He suggested that the cancellation minimum occurs below the reflectivity peak because of the oscillation of large drops prior to breakup. Equally plausible, perhaps, is the following argument which does not depend upon droplet break up, but which considers the behavior of correlation as well as cancellation below the reflectivity maximum in the bright band.

Brazier-Smith and Stromberg (1972) studied the shape and fall speed of large water drops (diameter 4-5 mm). They observed that initial oscillations of the drops, associated with their release in the wind tunnel, died out after a free fall of less than 4 m. Their photograph shows that when oscillations cease, the drops are distorted and falling with a flat side down. This suggests that a raindrop in the bright band develops a strongly preferred orientation almost as soon as the ice in that particle melts completely.

The ORTT minimum at the bright band indicates that the scatterers there show less tendency to have a preferred orientation than do the snowflakes above. This may be a scattering effect associated with the change in dielectric constant which occurs when the surfaces of the snow aggregates become water covered at the bright band. Another possibility is that the scatterers begin to tumble because of changes in density and shape that result when melting occurs.

Assuming that the mixed-phase scatterers return only weakly correlated signals and that the pure-water scatterers show a preferred orientation as soon as they are created, ORTT indicates the relative percentage of scatterers that are liquid at various heights below the reflectivity maximum in the bright band. The higher the ORTT value the higher the

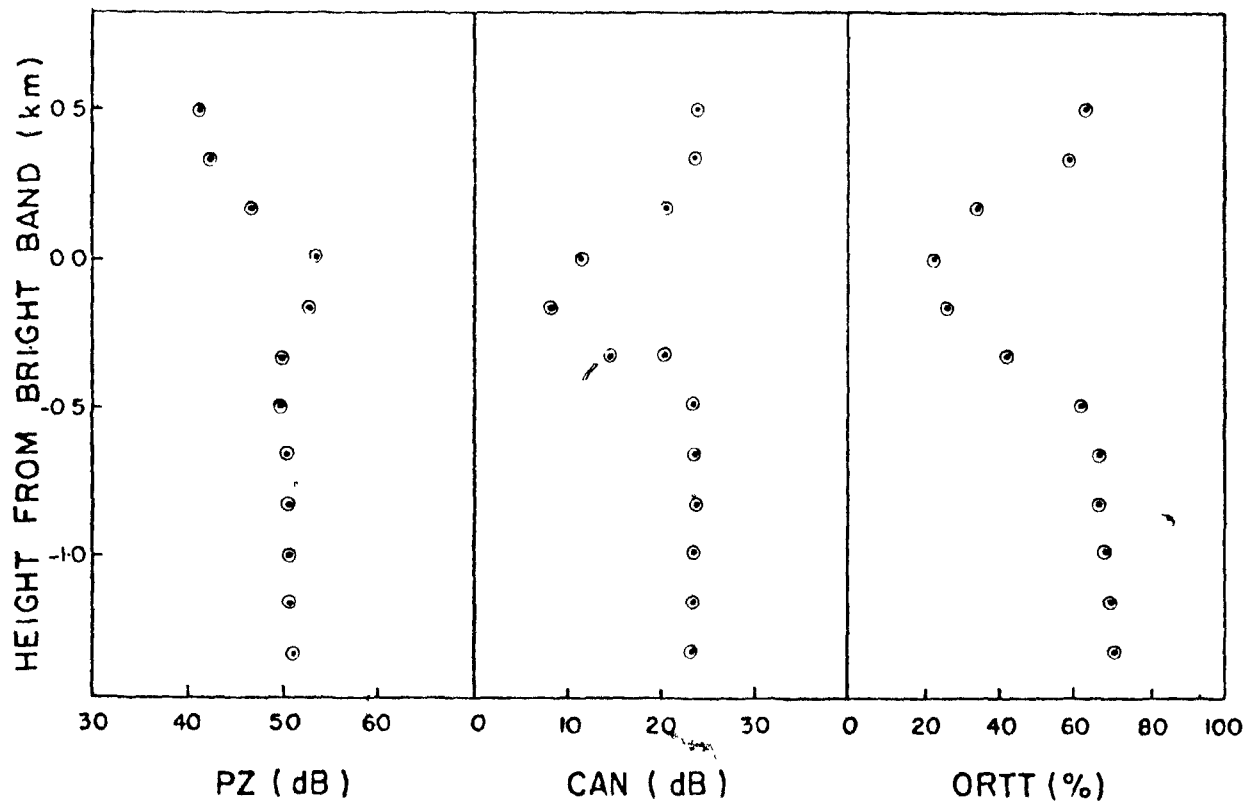


Fig. 3.4 Mean vertical profiles of P_z , CAN and ORTT for elevation scans through the bright band in stratiform precipitation recorded 28 August 1971.

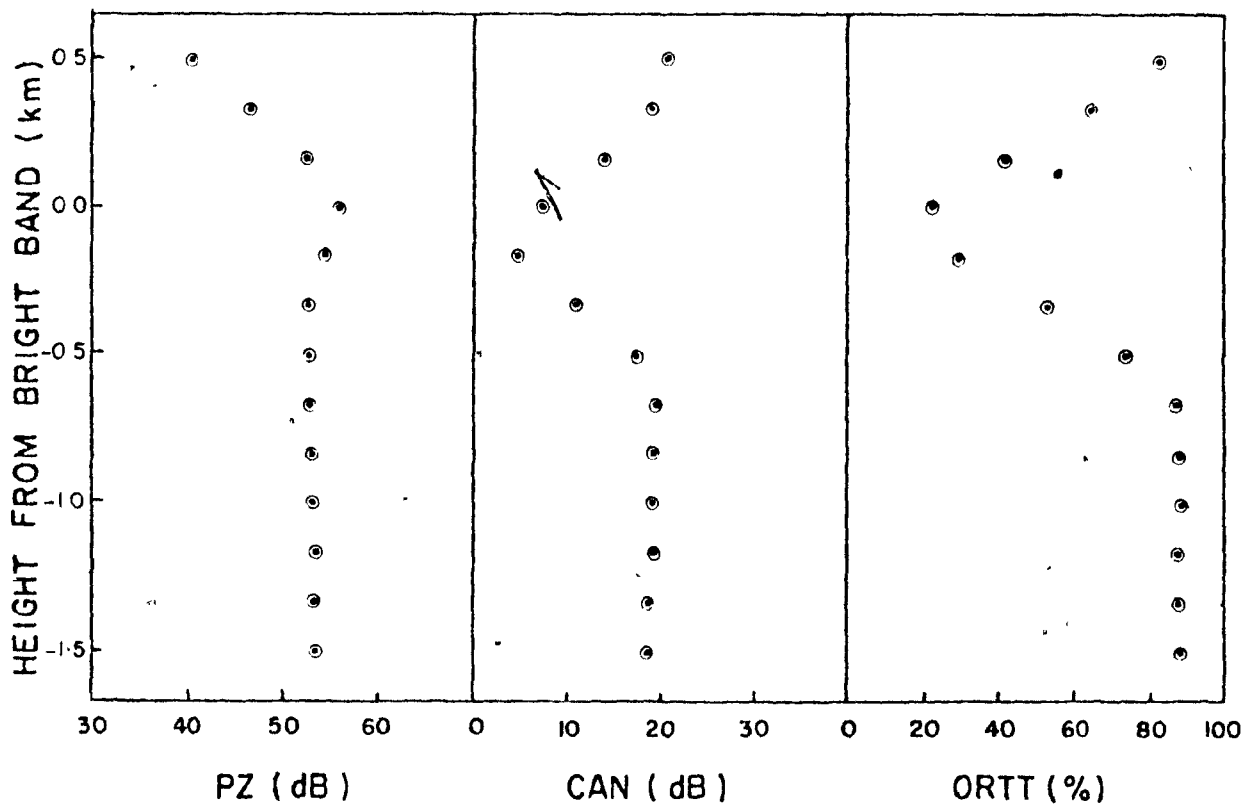


Fig. 3.5 Mean vertical profiles of P_z , CAN and ORTT for elevation scans through the bright band in stratiform precipitation recorded 30 May 1972.

percentage of rain droplets at that level. About 500 m below the reflectivity peak all the mixed-phase scatterers seem to have melted since ORTT becomes constant ($>60\%$) with height. This latter feature is shown more clearly in Figs. 3.4, 3.5 and 3.6 presented in the next section.

The occurrence of the cancellation minimum below the level of the reflectivity maximum would be explained as follows. Some of the water-covered snowflakes, probably the smallest ones, have been converted to raindrops by the time they fall to level of the cancellation minimum. These small scatterers have a smaller cross-section and lower number density (due to acceleration) in raindrop form than they did as snowflakes. As a result, the large scatterers dominate the returned signals more heavily at this level than at the reflectivity peak. Since the large water-covered snow aggregates probably have a more irregular shape than the smaller ones, the cancellation minimum occurs below the reflectivity peak where the large scatterers dominate most heavily.

3.3 Other Bright Band Observations

Several characteristics of stratiform rain data have been determined from examples in which observations of rain below the bright band extend over two or three kilometers in elevation. Three such examples are presented in Figs. 3.4, 3.5 and 3.6. These profiles differ from Fig. 3.2, in two ways: the reflectivity in the rain is greater than observed in Fig. 3.2, and ORTT values above the bright band exceed 40%, the value expected for snow (Hendry and McCormick, 1971). These two differences may be related. It is possible that ORTT values above the bright band are large because of propagation effects which occurred in the rain (see McCormick et al, 1972). The larger reflectivity in the rain suggests that

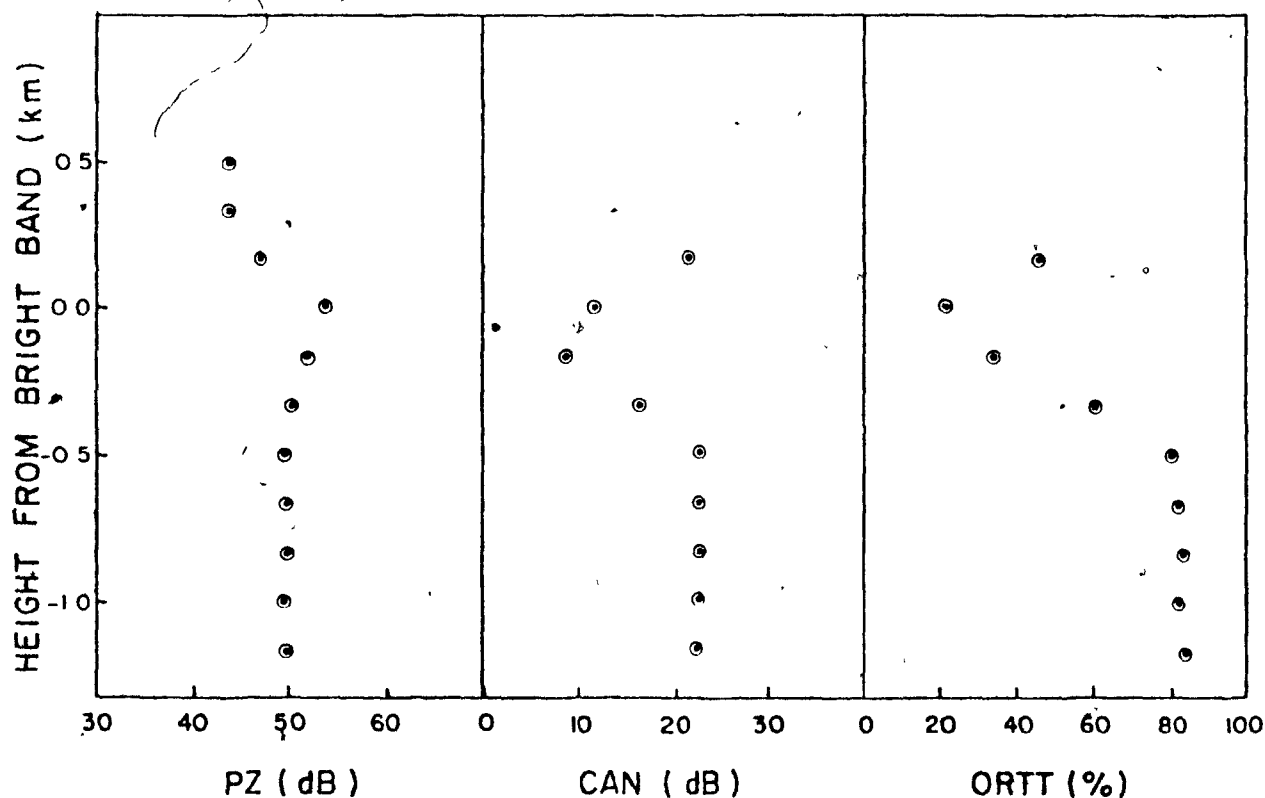


Fig. 3.6 Mean vertical profiles of P_z , CAN and ORTT for elevation scans through the bright band in stratiform precipitation recorded 6 October 1971.

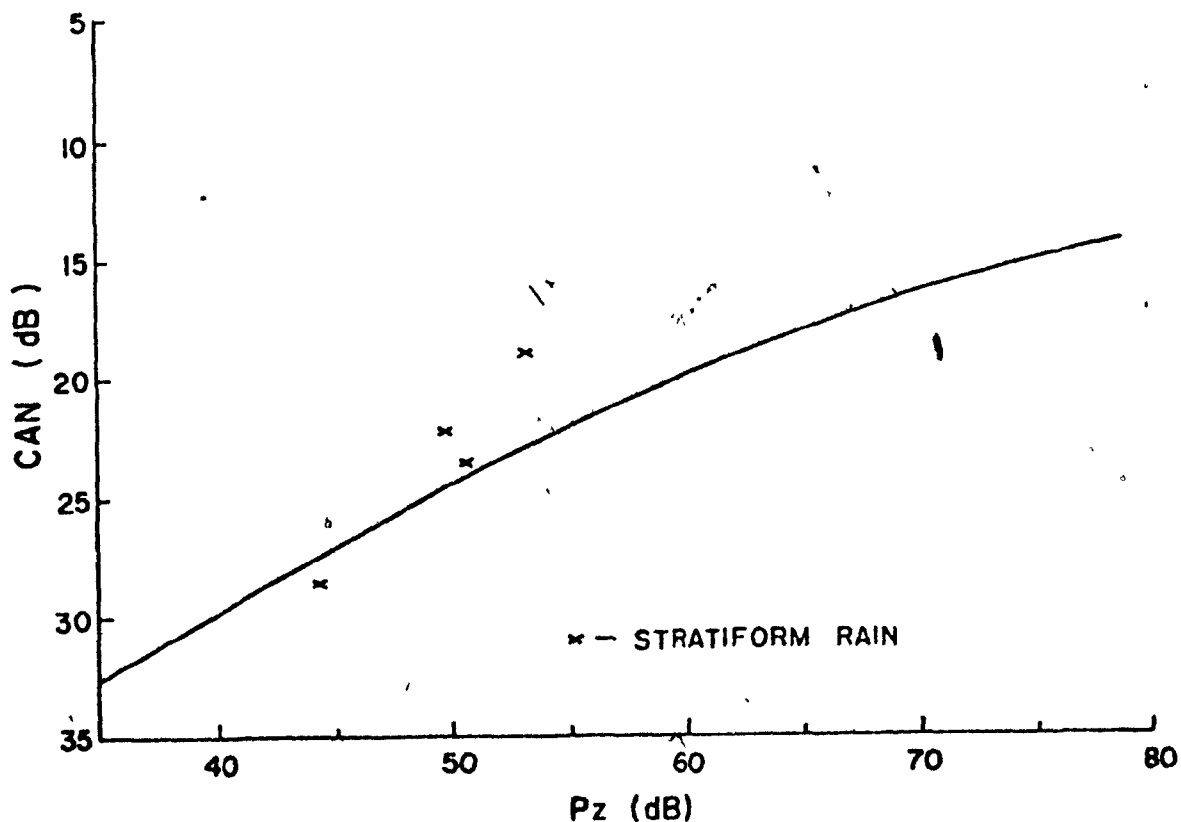


Fig. 3.7 Theoretical calculations of cancellation and reflectivity for Marshall-Palmer rain. The curve is adapted from Humphries (1973). The points correspond to an average P_z -CAN value from the rain region in Figs. 3.2, 3.4, 3.5 and 3.6.

such propagation effects are more important in these examples than in Fig. 3.2.

Humphries (1973) determined a theoretical relationship between reflectivity and cancellation based on the Marshall and Palmer (1948) raindrop size distribution and the Pruppacher and Pitter (1971) size-shape relationship for raindrops. This curve, presented in Fig. 3.7, shows that cancellation is a decreasing function of reflectivity for Marshall-Palmer rain. Humphries asserted that propagation effects should tend to make observations lie above his calculated values. An average P_z -CAN value for the rain region in each of the stratiform rain examples studied, appears in Fig. 3.7. These observations lie quite close to the line, suggesting that propagation effects were not dominant in the rain region of stratiform rain examples. Propagation effects, apparently incurred in the rain, appear to dominate correlation values above the bright band, as noted above. This probably occurs since accumulated propagation effects would become relatively more important in the snow region where cancellation values are large.

Well below the bright band, in Figs. 3.4, 3.5 and 3.6, P_z and ORTT increase and CAN decreases toward the ground. This pattern agrees with the observation by McCormick and Hendry (1972) that ORTT is an increasing function of P_z and CAN is a decreasing function of P_z . The profiles of P_z suggest that the mean drop size increases toward the ground. The ORTT and CAN profiles show that the larger drops near the ground tend to be more deformed and have a greater tendency to fall with a preferred orientation than do the smaller drops above.

The bright band profiles of Figs. 3.4, 3.5 and 3.6 were constructed from data observed in 3 range gates in each of 17 elevation scans.

Table 3.1. The distribution of distances (d) between
0°C and the reflectivity peak in the bright
band for 42 observations in stratiform rain.

DISTANCE BELOW 0°C(m)	NUMBER
d < 250	6
250 < d < 500	11
500 < d < 750	25

In these 42 profiles, the reflectivity maximum occurred below the height of 0°C as measured at Maniwaki¹ at 1200Z on those days. A tabulation of the distances that the reflectivity peak occurred below the 0°C isotherm is presented in Table 3.1. Thirty three of these observations lie between 240 m and 540 m. Battan (1973) indicates that most observers report values between 100 m and 400 m. This difference suggests that the height errors caused by applying atmospheric data from Maniwaki are of the order of 200 m.

3.4 Parameter Distributions in Stratiform Rain

It was indicated in Section 3.1 that the radar parameters show very little horizontal variation in stratiform rain. To verify this, the distributions of Pz, CAN and ORTT were constructed for various heights from data recorded 28 August 1971. The distributions of these parameters for a 500 m interval in the rain below the bright band are presented in Fig. 3.8. The vertical profiles for this day (Fig. 3.4) show that the mean parameter values vary only slightly in any 500 m interval in the rain region. As a result, vertical variations in the 500 m interval do not widen the distributions significantly.

The very narrow distributions in Fig. 3.8 show that horizontal variations of Pz, CAN and ORTT were indeed small and that values remained nearly constant over the observation period.

1

Maniwaki, a radiosonde station located 110 km NNW of Ottawa.

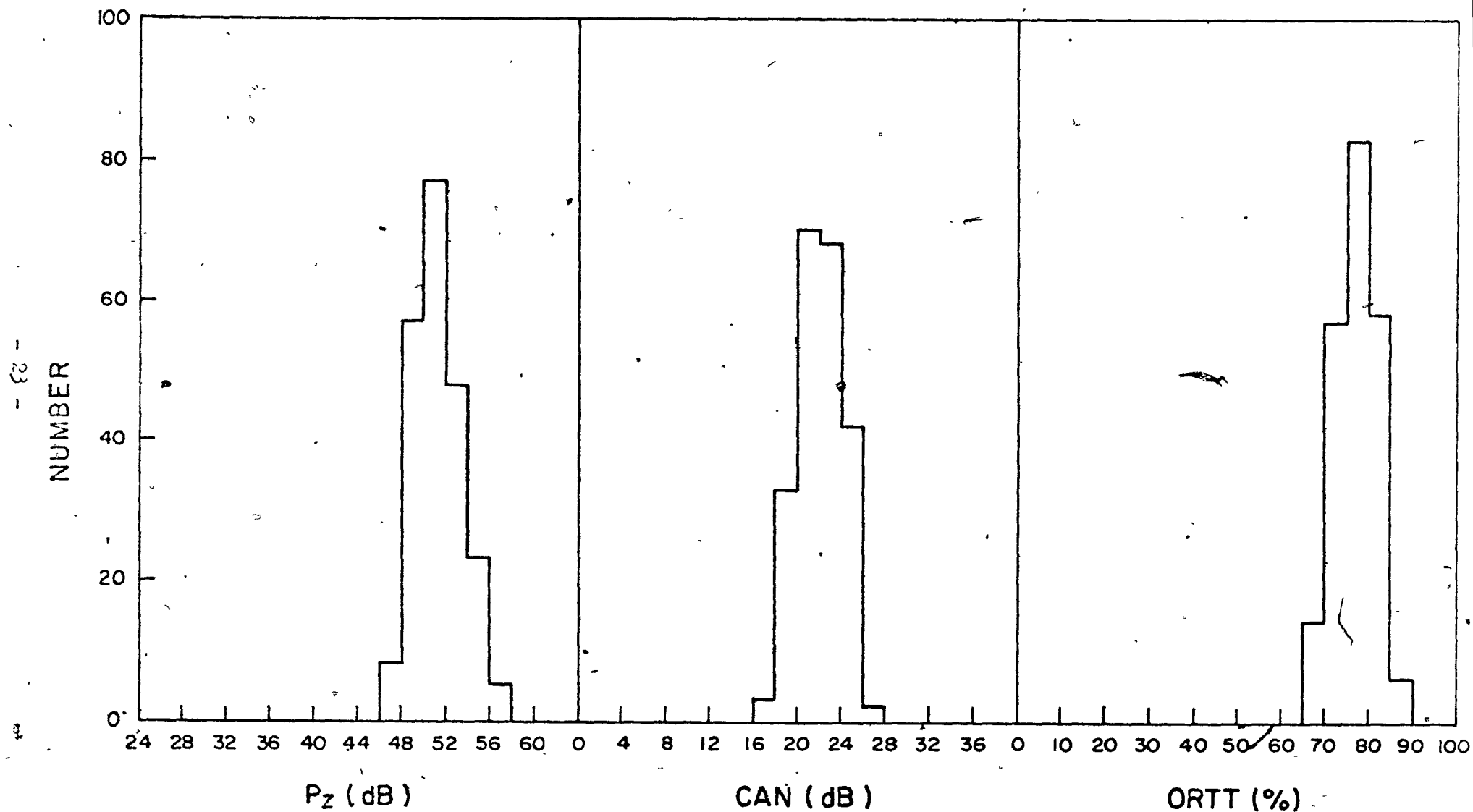


Fig. 3.8 The distribution of P_z , CAN and ORTT in rain below the bright band. The histograms represent observations at elevations between 1.0 km and 1.5 km for data recorded 28 August 1971.

CHAPTER IV

OBSERVATIONS IN CONVECTIVE PRECIPITATION

4.1 Introduction

In comparison with bright band data, elevation scans through convective rain (see Fig. 4.1) show relatively large horizontal variations of reflectivity and cancellation. On the other hand, correlation is nearly a function of height alone. Small values of correlation tend to occur at mid-levels with larger values above and below. This structure of ORTT is similar to the pattern observed in stratiform rain.

In stratiform rain, parameter values consistent with snow and rain were observed above and below the bright band respectively (Fig. 3.2). For the convective rain elevation scan in Fig. 4.1, the large ORTT values (>60%) below 3.0 km suggest the presence of rain. Above 4.5 km the ORTT values near 40% must represent some form of precipitation other than snow since the associated CAN values (approximately 20 dB) are lower than those reported for snow (McCormick and Hendry, 1970).

This pattern of ORTT values is different from stratiform rain observations in a second way: the ORTT minimum in Fig. 4.1 occurs above the height of 0°C, whereas in stratiform rain it occurred at the reflectivity maximum, below the 0°C level. The proximity of the low ORTT region to 0°C suggests that low correlation values in convective storms may indicate the presence of mixed-phase scatterers as they do in stratiform rain.

4.2 ORTT as a function of Height in Convective Rain Data

The extent to which ORTT is a function of height in an individual elevation scan is demonstrated in Fig. 4.2 from Hendry and McCormick

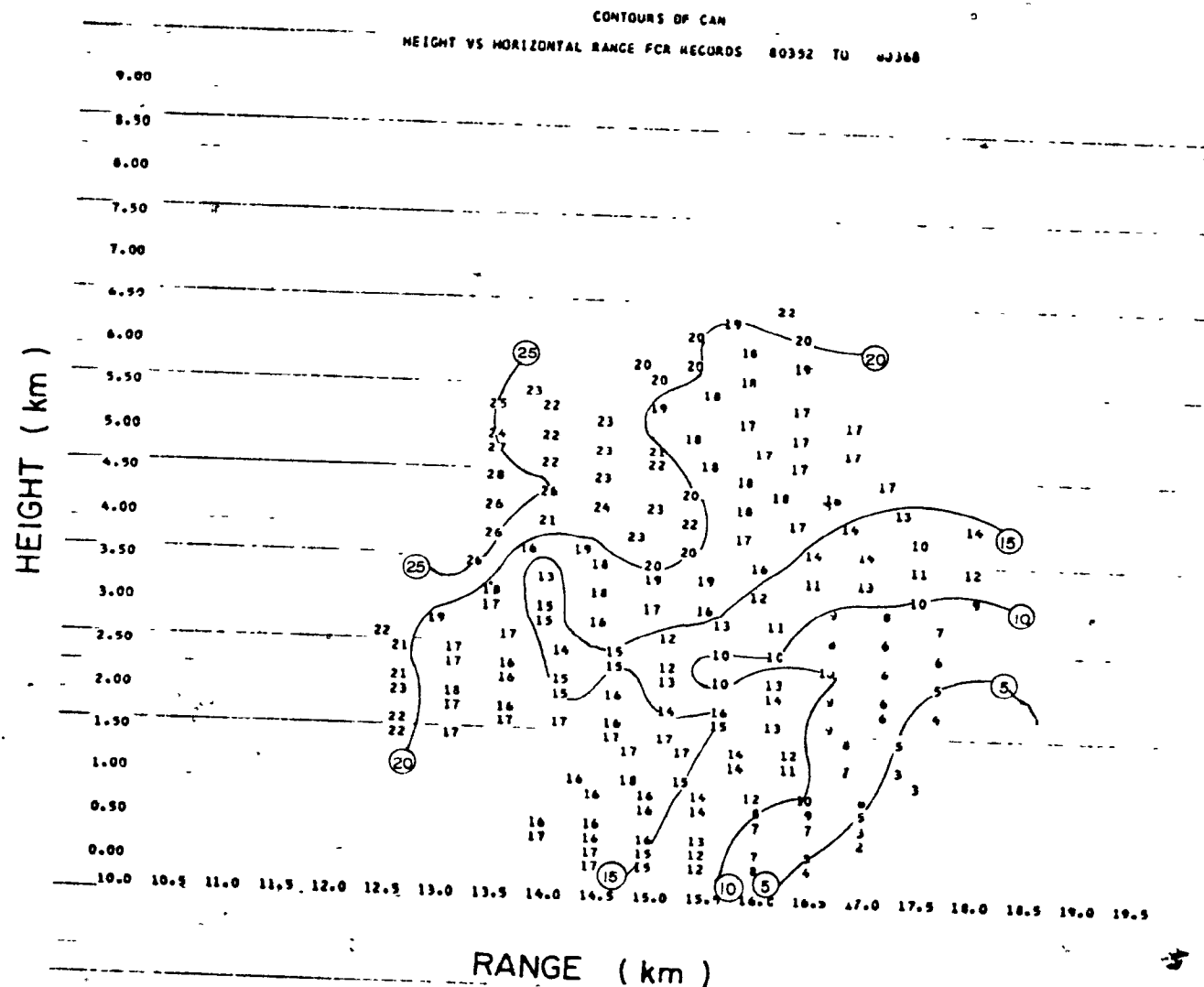


Fig. 4.1(a) Display of CAN for an elevation scan through convective precipitation recorded 10 July 1972. The height of the 0°C isotherm at Maniwaki was 2.8 km.

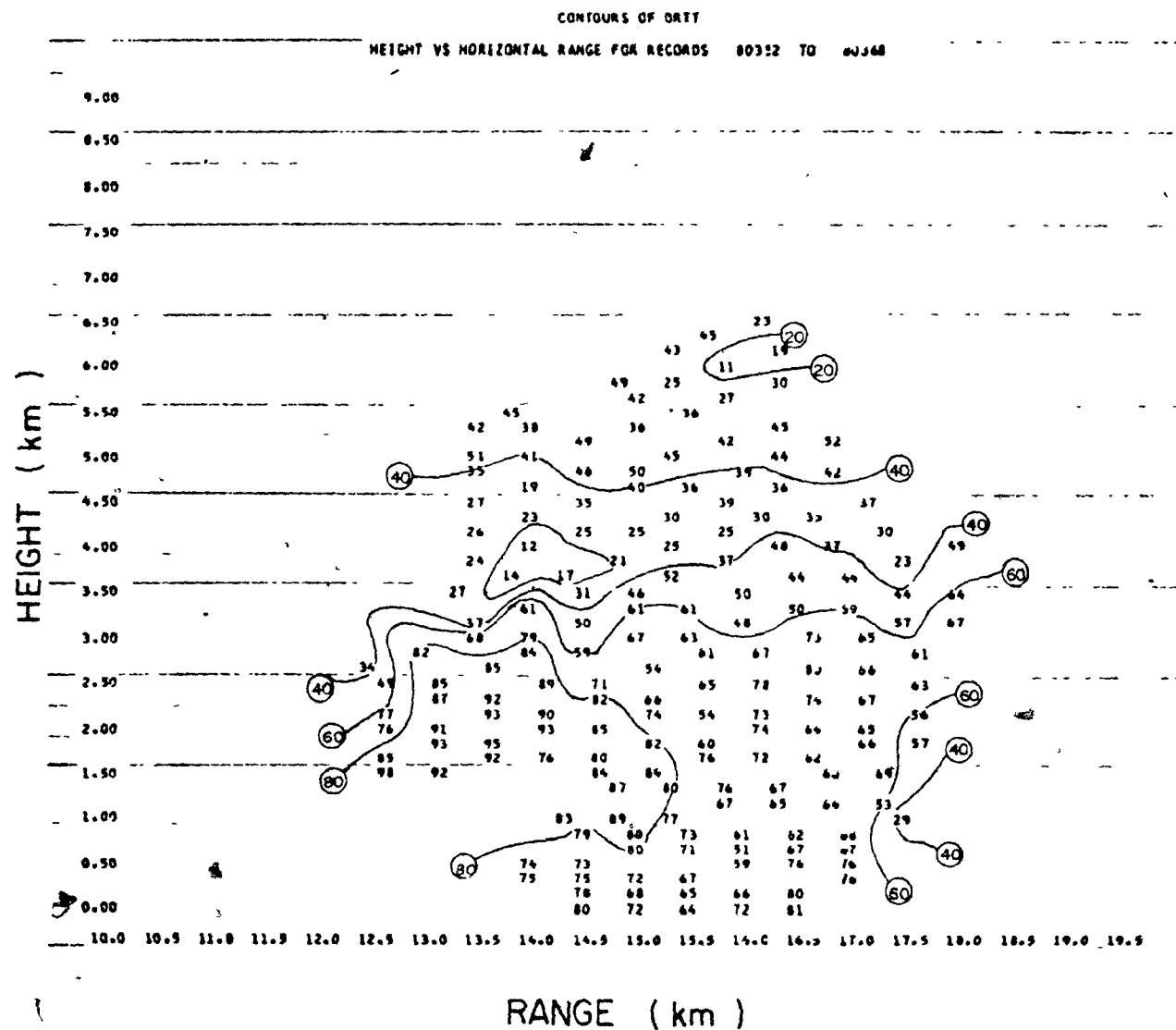


Fig. 4.1(b) Display of ORTT for an elevation scan through convective precipitation recorded 10 July 1972.

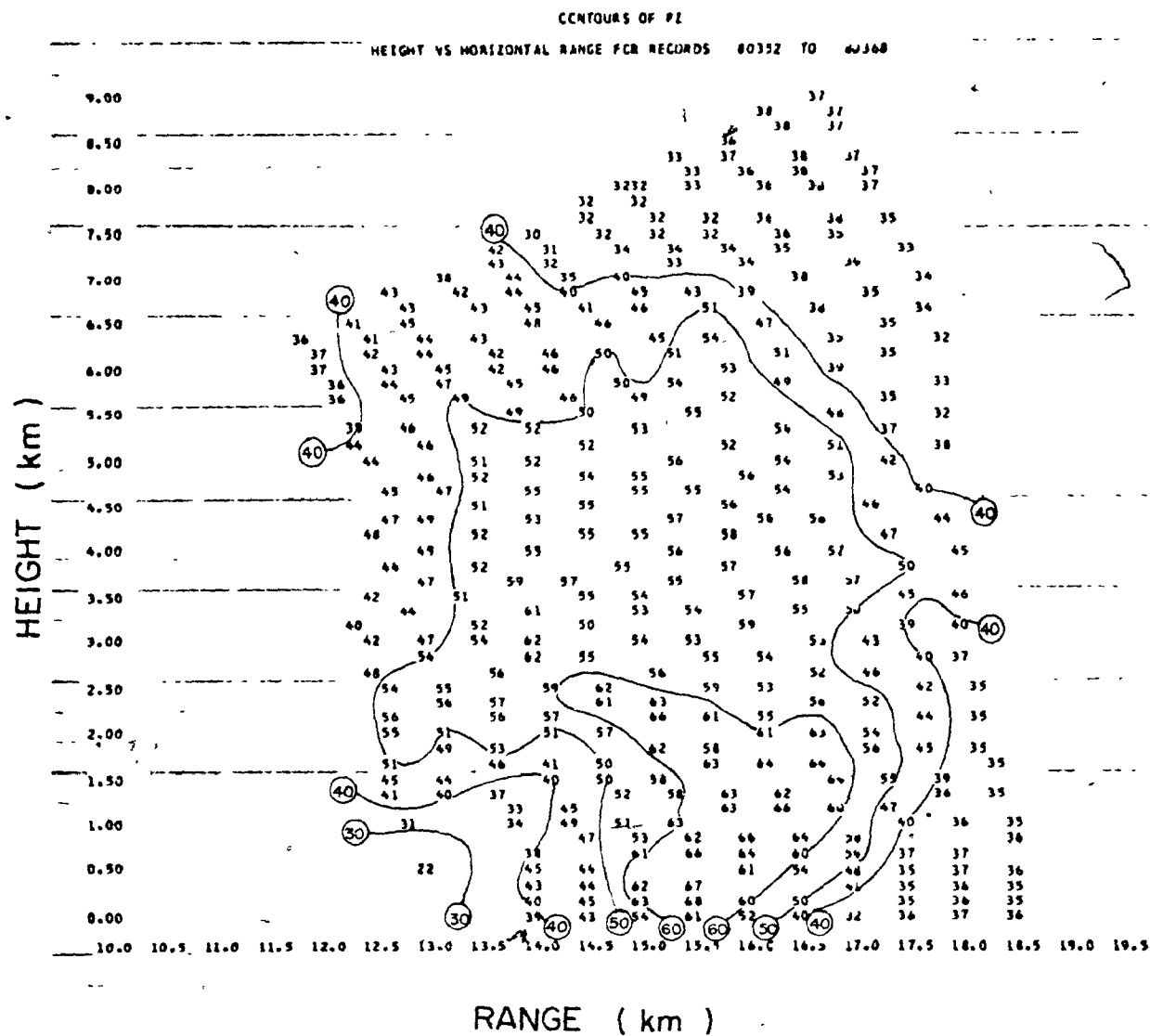


Fig. 4.1(c) Display of P_2 for an elevation scan through convective precipitation recorded 10 July 1972.

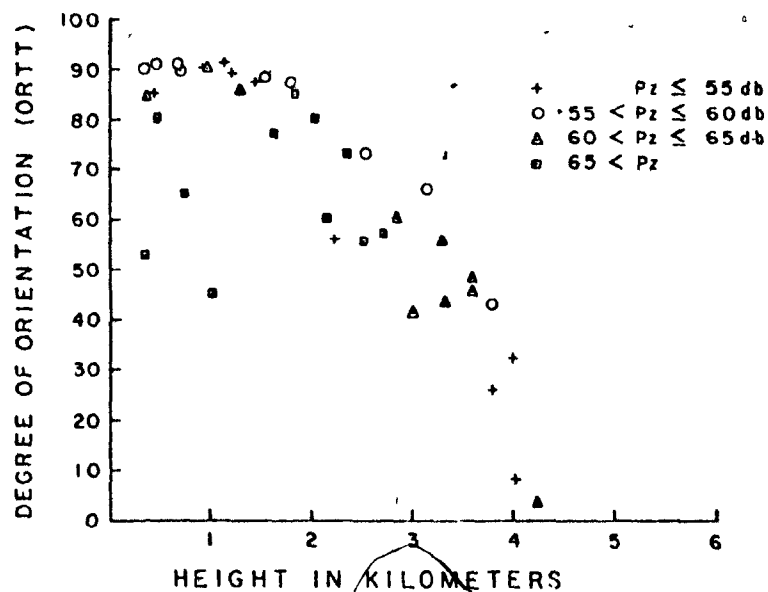


Fig. 4.2 Vertical profile of ORTT for an elevation scan in convective rain recorded 4 September 1970. This figure is reproduced from Hendry and McCormick (1971).

(1971). In this figure correlation values greater than 80% predominate below 2.0 km. Above 2.0 km ORTT values decrease with height becoming less than 10% at 4.0 km. Fig. 4.1 demonstrates the tendency for ORTT values to increase again at high elevations. In elevation scans with large vertical extent (such as Fig. 4.1) low ORTT values appear in a band at mid-levels.

Fig. 4.2 shows that the vertical gradient of ORTT below the band of low values extends from 2.0 km to 4.0 km. This height interval is much wider than 500 m ORTT gradient zone observed below the reflectivity peak in the bright band (see Section 3.2). A relatively wide ORTT gradient zone such as this, is common to most elevation scans from convective rain.

When the data from a given day are considered as a whole, a similar pattern of ORTT values is apparent. The height dependency of correlation values was determined for four occasions on which eight or more convective rain elevation scans were recorded. For a given day, a histogram of correlation values was constructed for each 500 m height interval between 1.0 km and the maximum height of observations on that day. Observations at greater range than the reflectivity maximum were disregarded in an attempt to reduce contamination of the results by propagation effects. The histograms for each of the four occasions show a similar height dependency of correlation values. The ORTT distributions representing data in thirty elevation scans recorded 10 July 1972 appear in Fig. 4.3.

The distributions show that ORTT values greater than 60% predominate near the ground and the values less than 30% predominate at mid-levels (4.0 km in this case). The relative increase in the number of large ORTT values in the distributions above 4.0 km indicates that the low ORTT values at 4.0 appear as a minimum in high elevation scans. This low-ORTT region existed well above 2.8 km, the height of 0°C measured at Maniwaki

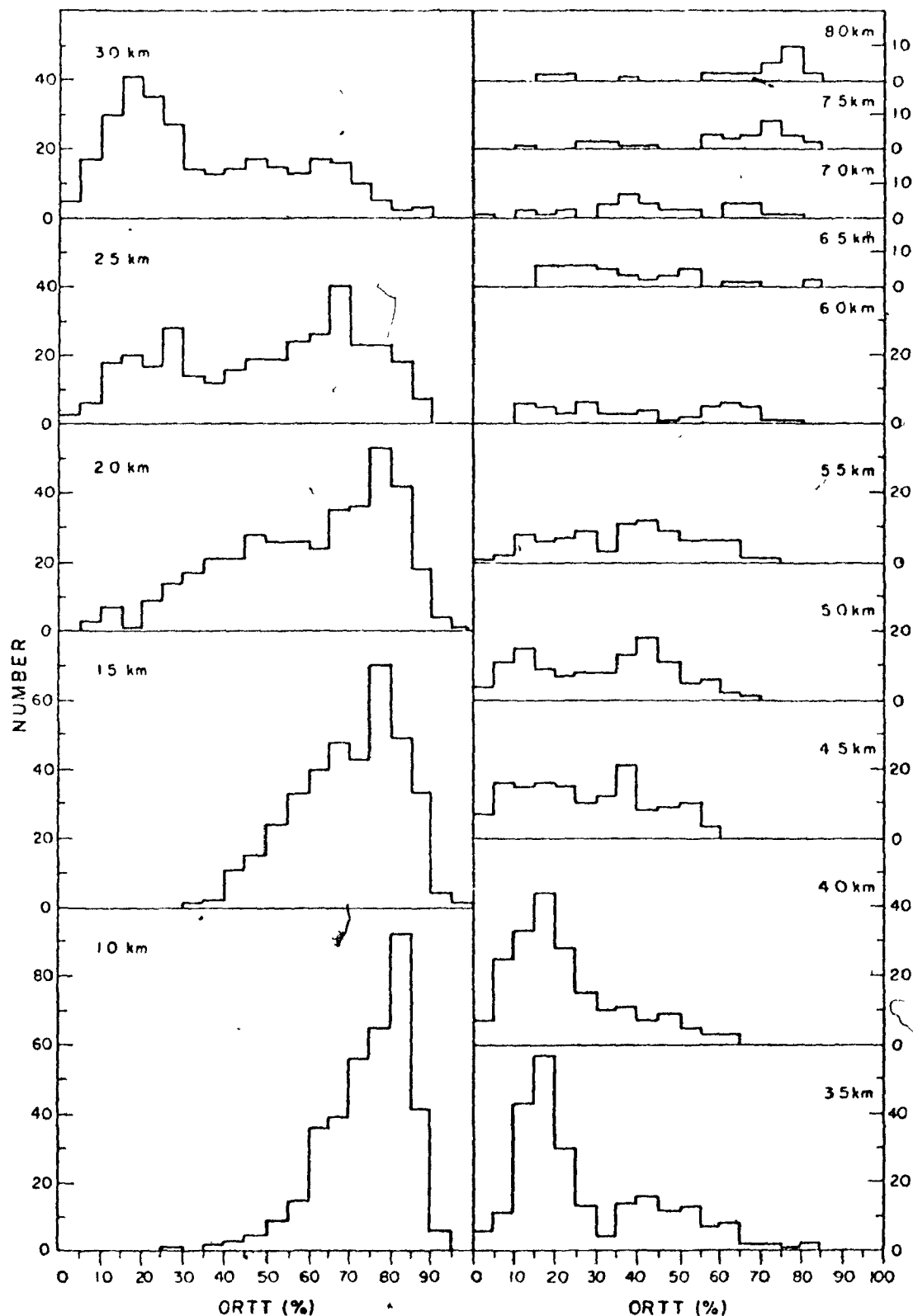


Fig. 4.3 The distribution of ORTT as a function of height in convective storm data. The histograms represent ORTT values observed 10 July 1972 in 500 m height intervals between 1.0 km and 8.0 km. Observations beyond the maximum in reflectivity have not been included. The height of the 0°C isotherm at Maniwaki was 2.8 km.

on this day.

It is clear that the ORTT distributions at 7.5 and 8.0 km are heavily weighted toward elevation scans from strongly developed storms since these are the only scans with data points printed at such high elevations. Some values greater than 70% appear there. It is only near the top of the tallest elevation scans, however, that such large values of ORTT are observed at elevations above the ORTT minimum.

4.3 The Low-ORTT Region

On two occasions in August 1973 elevation scans were recorded in rapid succession at fixed azimuth and range. These elevation scans constitute a time history of the precipitation which existed in the storm above a fixed line at the surface. The records also contain spatial information due to the movement of the storm above the line of observation.

On 14 August more than one hundred elevation scans were recorded in eight groups, from a storm system which passed south of Ottawa. Within each of the eight groups, elevation scans were recorded at about 15-second intervals until precipitation echoes at that azimuth and range died out. After recording ceased, the antenna was repositioned to record data from a high reflectivity region near the location of the preceding observations. Weather radar records from Ste. Anne de Bellevue indicate that the storm system diameter was about 45 km and that it moved due eastward between 1840 EST and 1930 EST (the observation period). An estimate from these records suggests that the storm moved about $1/4$ beamwidth between successive elevation scans.

An example of the elevation scans recorded on this day appears in Fig. 4.4. No suggestion is made that elevation scans within a group trace

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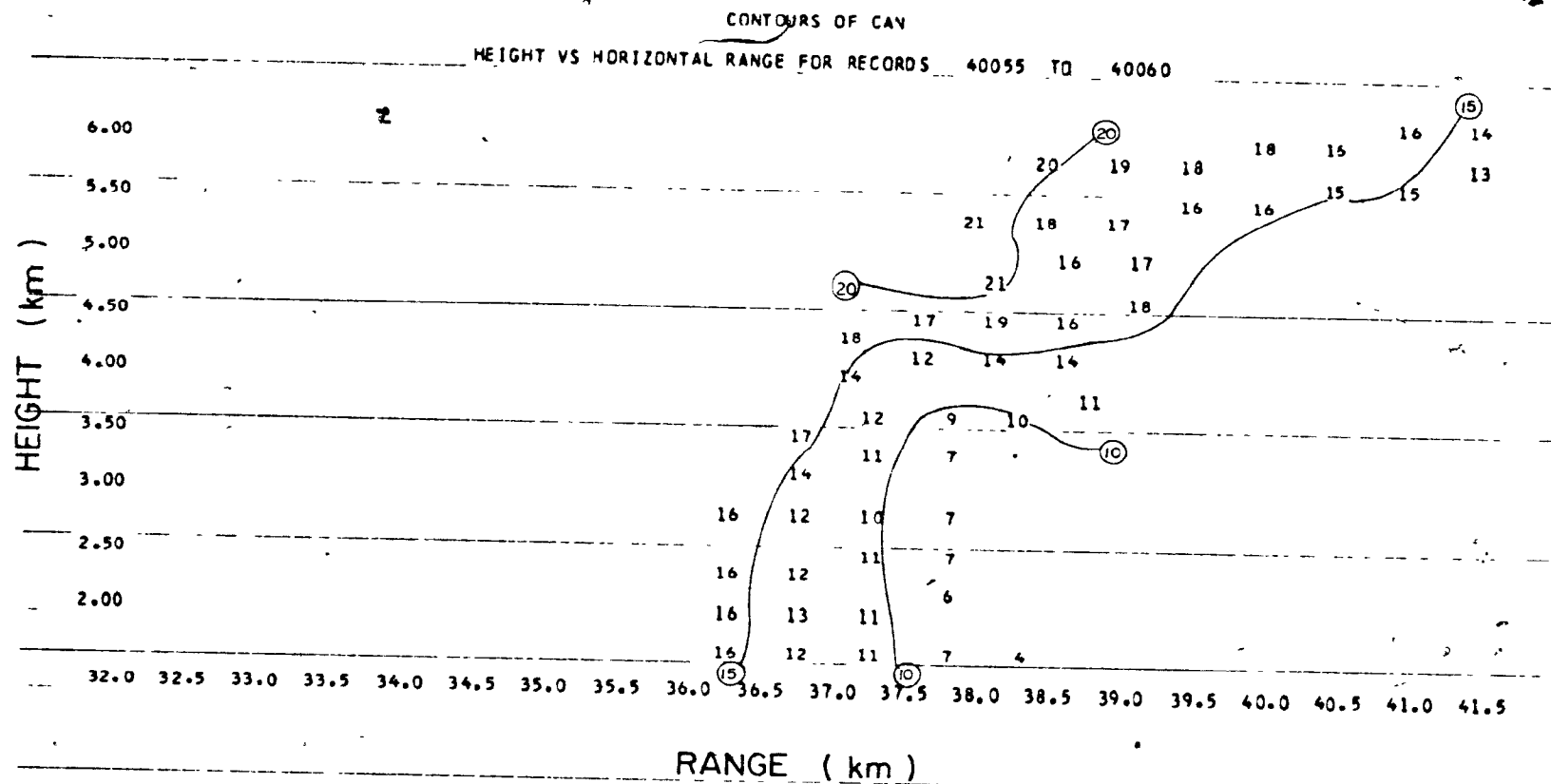


Fig. 4.4(a) Display of CAN for an elevation scan through convective precipitation recorded 14 August 1973. The height of the 0°C isotherm at Maniwaki was 3.1 km.

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DATE= 14 08 73 RUN NO. 2

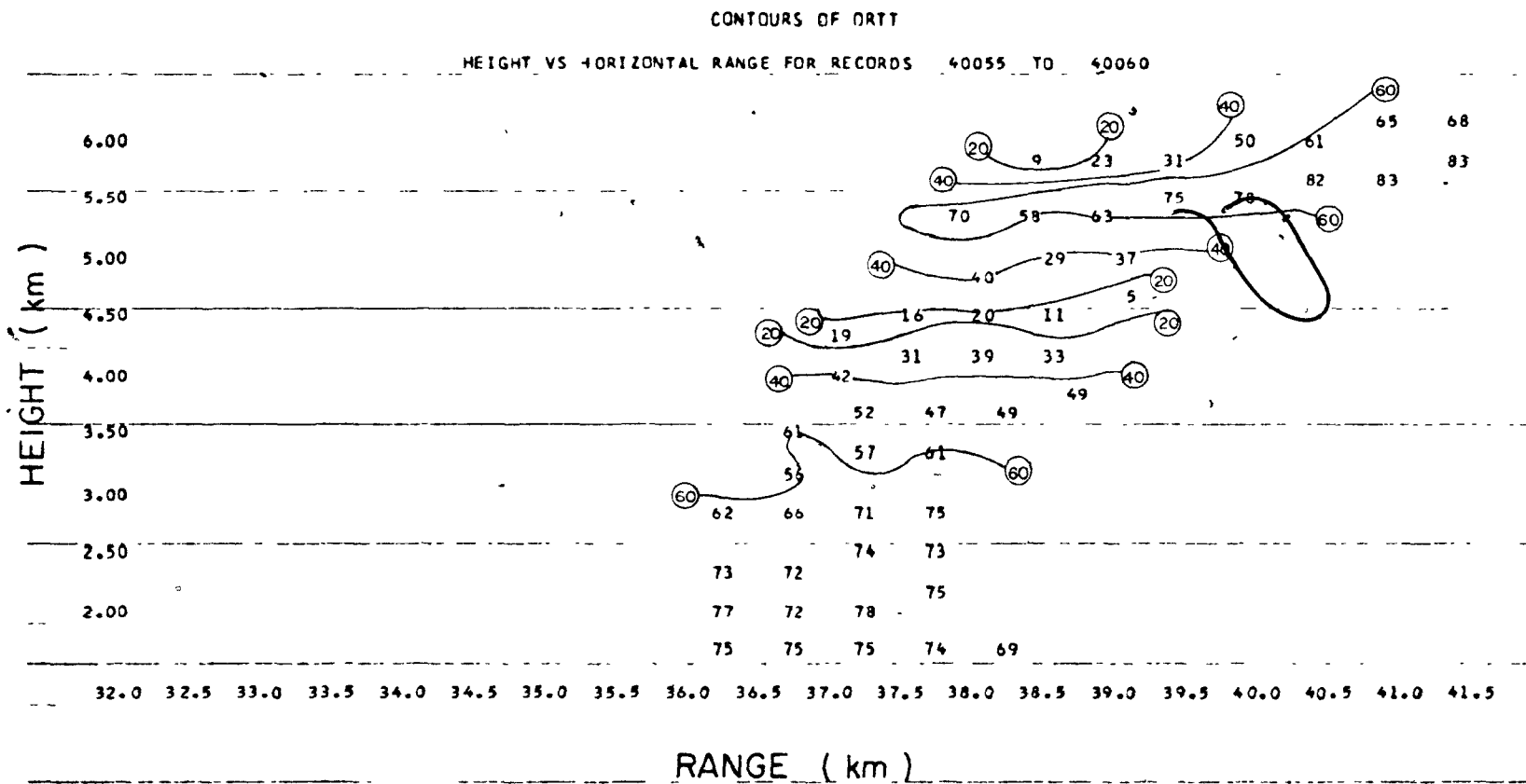


Fig. 4.4(b) Display of ORTT for an elevation scan through convective precipitation recorded 14 August 1973.

PEFQ02 UR180U

DATE= 14 08 73 RUN NO. 2

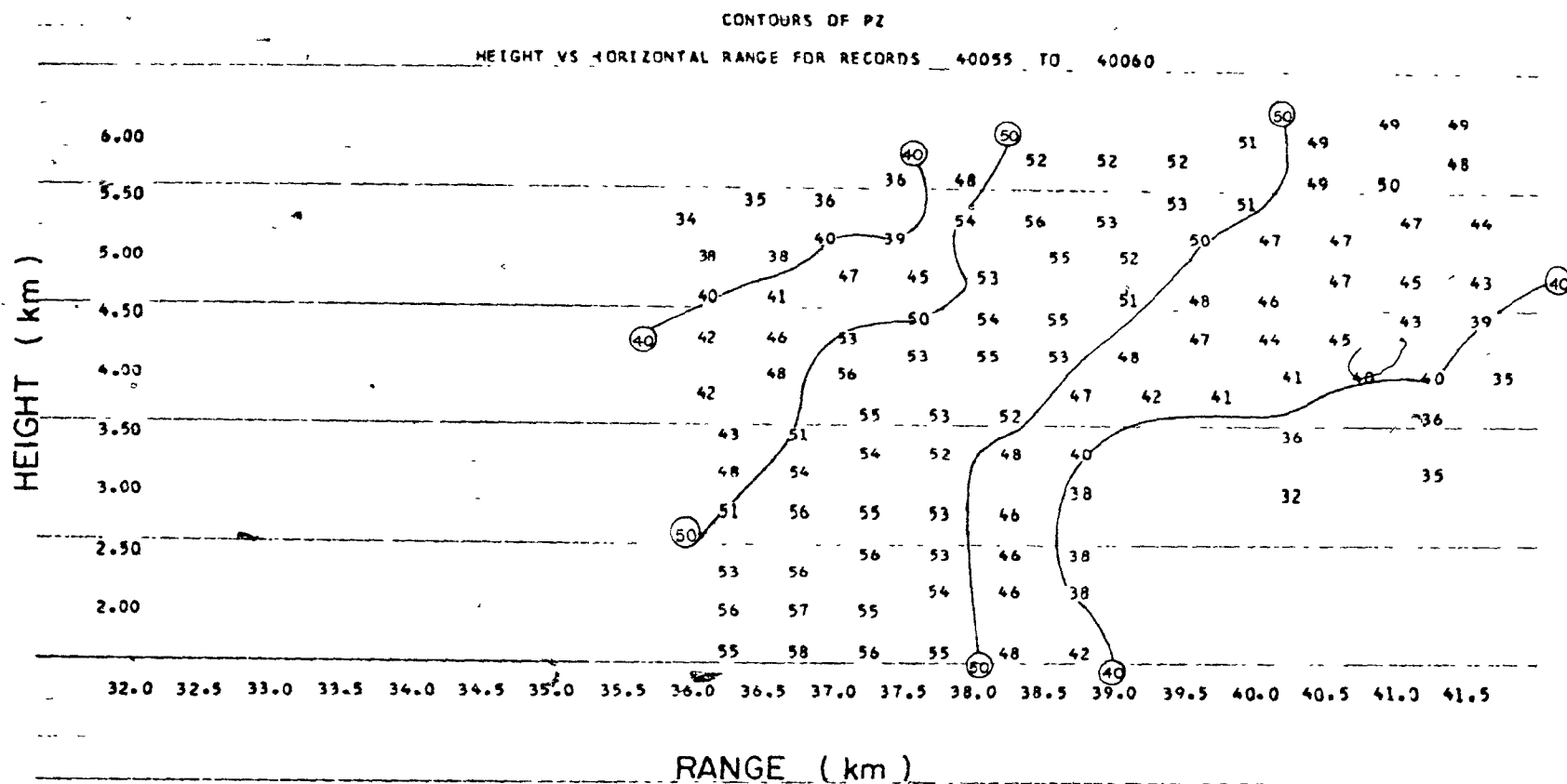


Fig. 4.4(c) Display of P_Z for an elevation scan through convective precipitation recorded 14 August 1973.

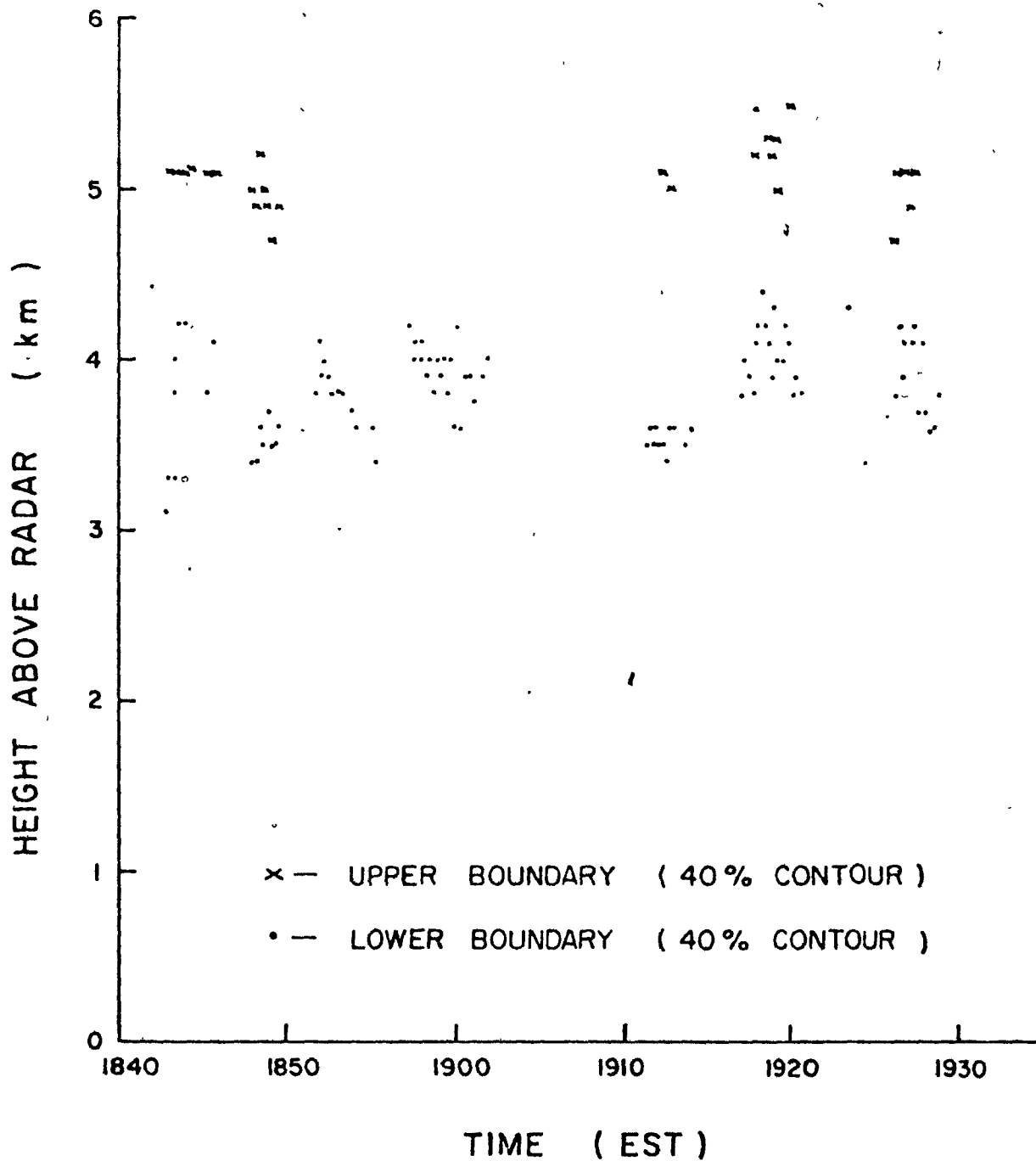


Fig. 4.5 Scatter diagram of the height of the 40% contour above and below to low-ORTT region. Points represent elevation scans recorded in a 50 min interval 14 August 1973. The height of the 0°C isotherm at Maniwaki was 3.1 km.

the history of a given precipitation parcel or that the eight groups are records of a unique portion of the storm system. Rather, the elevation scans included in the subsequent analysis represent a series of measurements, closely spaced in time, from a variety of locations within a single storm system.

All elevation scans recorded on this day demonstrate large values of correlation near the ground and a region of smaller values above, forming a pattern of the type described in Section 4.2. Those elevation scans which include data points above 4.5 km indicate that a low-ORTT region occurred in a narrow zone centered near 4.5 km as shown in Fig. 4.4.

In order to determine the time dependency of the height and vertical extent of this low-ORTT region, the heights of the 40% contour below and above (when available) the region were plotted against time as shown in Fig. 4.5. While the 40% contour is a rather arbitrary criterion to use to locate the boundary of the low-ORTT region, it is suitable for determining the gross characteristics of the region.

It is clear from Fig. 4.5 that the center of the low-ORTT region remained near 4.5 km throughout the observation period. Neither the upper nor lower boundary indicates a long term height trend in spite of the scan-to-scan variations which are present. The persistence of the low-ORTT region at a given height throughout these observations is consistent with a temperature dependency of ORTT, since atmospheric temperatures are known to be more or less constant over a time interval of this length (50 min).

It is also apparent from Fig. 4.5 that the general pattern of ORTT values is virtually the same in all elevation scans recorded on that day. The scans, however, cover a wide variety of locations within the storm system, as mentioned. Such uniformity throughout the storm system might be expected if the correlation were a function of temperature.

4.4 An Individual Elevation Scan

In view of the characteristics of the low-ORTT region determined so far, some speculation about the nature of the scattering medium in various parts of an individual elevation scan can be made. An elevation scan through a relatively small convective storm on 25 July 1972 appears in Fig. 4.6.

Most of the region below 2.5 km in this figure is characterized by correlation values greater than 60% and cancellation values less than 15 dB. The large ORTT values show that rain predominates in this region and the low cancellation values indicate that many large (greatly deformed) raindrops are present.

McCormick and Hendry (1972) indicated that low cancellation values (typical of moderate to heavy rain) are usually observed in low reflectivity regions near the edge of convective cells. This phenomenon is apparent in the 9.0 km range gate at low elevation in Fig. 4.6. Humphries (1973) cautioned that low cancellation values, such as these, could be caused by propagation effects or by the presence of hail. In this example, however, propagation effects are probably small or negligible due to the short range. The high ORTT values suggest that rain predominates there, not hail. The observed reflectivity-cancellation values ($P_z < 40$ dB, $CAN < 15$ dB) lie well above the line calculated by Humphries for Marshall-Palmer rain (see Fig. 3.7). It appears, therefore, that the precipitation in the 9.0 km range gate was rain with many more large drops than predicted by the Marshall-Palmer raindrop size distribution for this reflectivity.

In the 9.0 km range gate at higher elevation there are three

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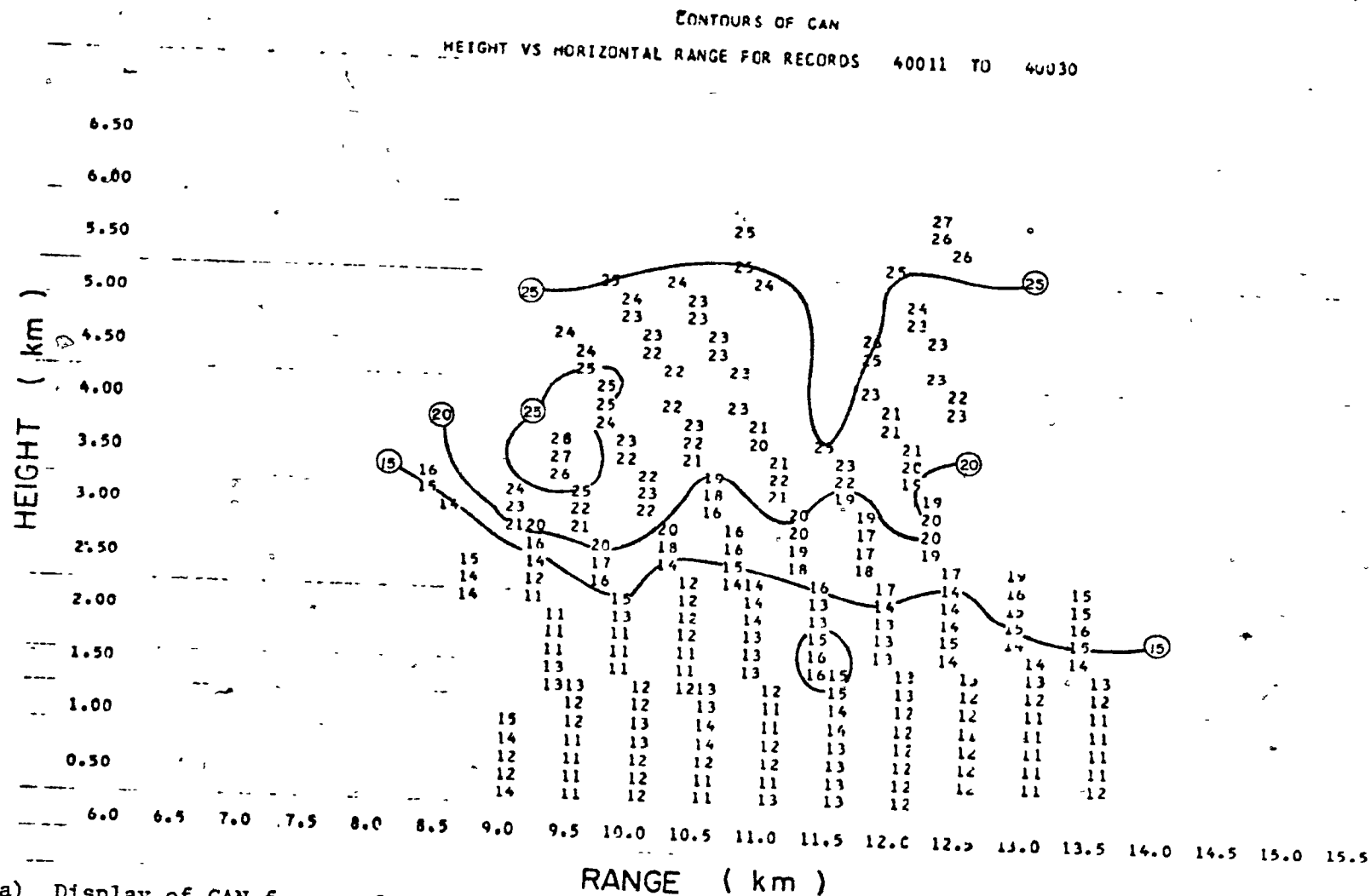


Fig. 4.6(a) Display of CAN for an elevation scan through convective precipitation recorded 25 July 1972.
The height of the 0°C isotherm at Maniwaki was 2.8 km.

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CONTOURS OF ORTT.

HEIGHT VS HORIZONTAL RANGE FOR RECORDS 40011 TO 40030

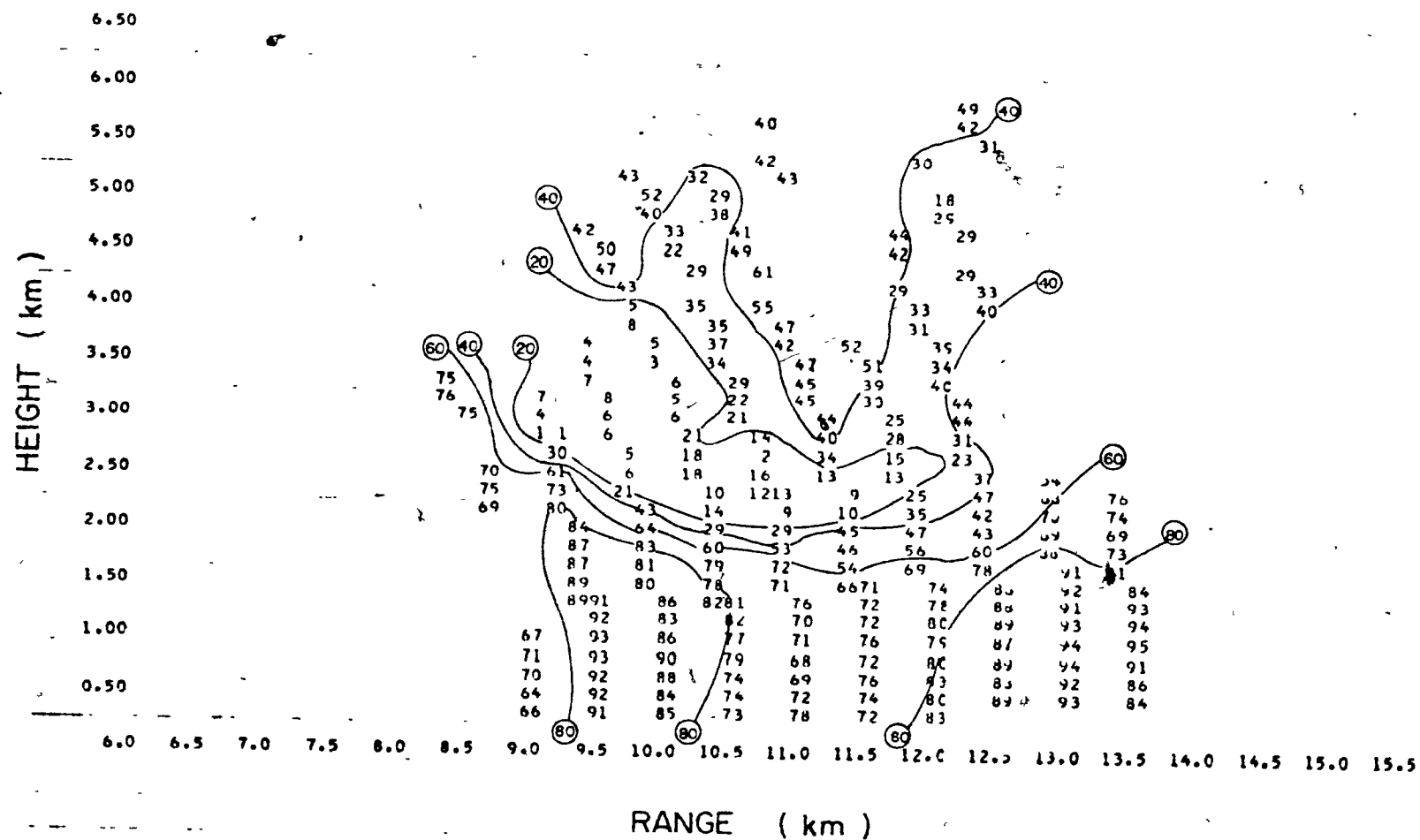


Fig. 4.6(b) Display of ORTT for an elevation scan through convective precipitation recorded 25 July 1972.

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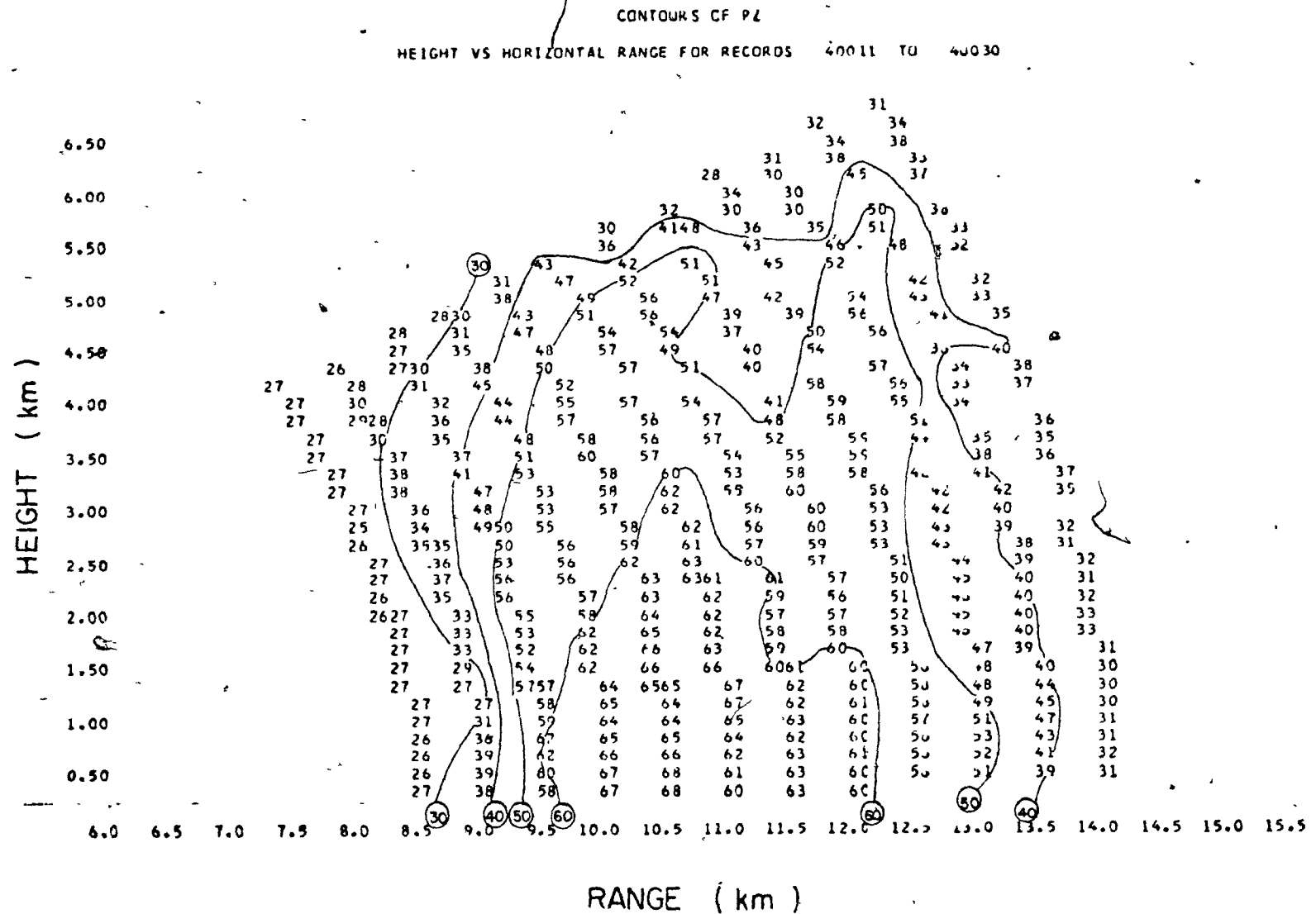


Fig. 4.6(c) Display of P_z for an elevation scan through convective precipitation recorded 25 July 1972.

data points which have large correlation values and low cancellation values as described above. These values suggest the presence of rain yet the data points lie above 2.8 km the height of 0°C on this day. It would appear that rain from below was carried up to this elevation by an updraught and remained in liquid form until observed by radar.

Above 2.8 km cancellation values increase with height becoming greater than 25 dB near the maximum elevation. The values of cancellation in this entire region are somewhat lower than those observed in snow (McCormick and Hendry, 1970). Some Pz values in this region are greater than 55 dB which is 15 dB greater than the observations in snow above the bright band in Fig. 3.2. Atmospheric temperatures above 5 km were colder than -15°C on this day so one would expect to find frozen precipitation there. It appears, however, that this precipitation is some form other than snow. It is possible that graupel or small hail dominate the radar returns from this region.

The ORTT display in Fig. 4.6 indicates that the percentage of particles with a preferred orientation in the region near the top of the echo lies between 30% and 50%. Between this and the rain region there exists a zone of very low correlation, with some values less than 10%. The low-ORTT region in this example extends below 2.8 km, whereas, in previous examples the low ORTT region occurred well above the height of 0°C . The extension below the level of 0°C occurs in the same range gates as the zone of high reflectivity. The high reflectivity may indicate the presence of a downdraught region which would be expected to lower the 0°C isotherm in that part of the storm. This may account for the occurrence of part of the low-ORTT region below the level of 0°C as measured at Maniwaki.

The gradient of correlation values below the low-ORTT region is confined to a 500 m interval, as observed in stratiform rain. Such a narrow ORTT gradient zone, below the 0°C level, is consistent with melting, as deduced in Chapter III. Fig. 4.6 shows, however, that low correlation values also occur well above the 0°C level. Consequently, melting alone cannot explain the occurrence of low ORTT values in convective rain.

The following model based on mixed-phase scatterers can explain the observed characteristics of the low ORTT region in convective rain. If one postulates that water-covered graupel or water-covered hail is characterised by low ORTT values either because of the lack of a preferred orientation or because of a peculiar scattering effect associated with the water coating, then the low ORTT region merely indicates the presence of these scatterers.

The 30% to 50% ORTT values near the top of Fig. 4.6 characterise the hail or graupel when dry and the low ORTT region above 0°C indicates the appearance of a water coating due to collisions with supercooled water droplets. The surface of the ice-phase scatterers could become wet at elevations well above the height of 0°C if the collisions with droplets were frequent enough. This water surface could then be maintained over the substantial height interval associated with the low-ORTT region so long as water were supplied by supercooled droplets. Below 2.8 km in Fig. 4.6, the rapid increase in ORTT values suggests that the hail or graupel particles were small enough to melt at roughly the same rate as the snow particles in stratiform precipitation when falling into air with temperatures above 0°C.

A slight variation of this mechanism can explain the wider ORTT gradient zone observed in most convective examples (see Figs. 2.1,

4.1, 4.4). The graupel or hail becomes water coated well above the height of 0°C and results in the low ORTT values observed at those levels. If the water is collected at a rate faster than it can be frozen then the water coating becomes thicker. When the water mass begins to be comparable with the ice mass in a given particle, it seems likely that the particle would begin to behave as a raindrop with the higher associated ORTT value. This may account for the gradual increase in ORTT values which begins above the height of 0°C in Fig. 4.4 and the other elevation scans from that day. When particles of this type fall into air warmer than 0°C no rapid increase in ORTT values occurs since they behave essentially as raindrops even before melting begins.

In elevation scans from tall convective storms presented by McCormick and Hendry (1972) and Hendry and McCormick (1974), low values of correlation extend to high elevations. They suggest that such values probably indicate the presence of large hail. These scatterers are likely to have wet surfaces since hail is thought to have a water surface when growing. It is possible that large hail, wet or dry, has low associated ORTT values due to tumbling.

CHAPTER V

CONCLUSIONS AND DISCUSSION

5.1 Conclusions

In stratiform rain data a correlation minimum was observed at the level of the reflectivity maximum in the bright band. The rapid increase in ORTT values just below this level coincides with the conversion of water-covered snowflakes into raindrops. It appears that the increase in ORTT values is effected by the increase in the relative number of raindrops as the snow melts. To that extent ORTT indicates the relative number of raindrops (with respect to water-covered snowflakes) in the scattering ensemble at levels below the reflectivity peak in the bright band.

The cancellation minimum was observed below the height of the reflectivity peak in all bright band examples studied. This may be associated with the conversion of the small snowflakes into raindrops immediately below the reflectivity peak. If so, the low cancellation values observed there characterise the large water-covered snow aggregates which remain.

In convective rain data a region of low ORTT values exists at midlevels with higher values above and below. High correlation values characteristic of rain occur near the ground. The moderate ORTT values at high elevations appear to characterise ice-phase scatterers since these probably dominate at the cold ambient temperatures which prevail at these levels. The low-ORTT region, centered above the height of 0°C , is evident in virtually all convective rain data. This region appears to be a temperature dependent phenomenon since it was observed near the same height in

a variety of locations within the storm on 14 August 1973. Furthermore, the height did not appear to change during the 50 min of observations. The occurrence of most low correlation values just above the height of 0°C suggests that these values indicate the presence of water-covered ice particles. Ice particles falling from above would grow by accreting super-cooled cloud droplets. As the particles approach the 0°C level there would be an increasing tendency for the accreted liquid to remain unfrozen; the precipitation might therefore resemble that in the bright band of stratiform rain. The large increase in correlation values which occurs near 0°C suggests that the scatterers soon accumulate enough liquid to resemble raindrops very closely, though they may still have ice in their centers. At temperatures warmer than 0°C , melting completes the conversion from ice to water but the particles have already been established essentially as raindrops.

5.2 Experiments for Future Consideration

The following experiments may serve to clarify the interpretation of some of the polarization characteristics of precipitation.

The Alberta radar could be equipped to record correlation data much more extensively than attempted in the past. The PPI correlation display described by Hendry and Allan (1973) could be routinely photographed as has been done for reflectivity and cancellation. This would produce a record of ORTT in three spatial dimensions and time. Such data, which are presently unavailable, would be invaluable in the study of correlation in convective storms.

Additional information about the processes which occur in the bright band could be determined by using Doppler techniques. In observations

of the bright band, discussed in Chapter III, the cancellation minimum was observed below the height of the reflectivity peak. A comparison of the Doppler spectra in the two receiver channels could determine which scatterers cause the cancellation minimum. At this level, a shift toward higher velocities in the cross-polarized channel would indicate that large droplets are present. A shift toward lower velocities would indicate that the water-covered snow aggregates return the power in the cross polarized channel.

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