P WAVE INVESTIGATIONS OF THE EARTH'S STRUCTURE
USING THE WARRAMUNGA SEISMIC ARRAY

A thesis submitted for
the degree of
DOCTOR OF PHILOSOPHY
in the
Australian National University
by
Cedric Wright
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STATEMENT

This thesis is an account of a research project undertaken while the writer was a full time research student in the Department of Geophysics and Geochemistry at the Australian National University during the period July 1966 to July 1970.

Many of the mathematical techniques described in Chapters 2 to 4 have been taken from published material referred to in the text. In certain cases, however, these techniques have been modified and extended by the writer. The interpretive work of Chapter 3 and Chapters 5 to 8 is entirely the writer's. A part of Chapter 2 and the whole of Chapter 9 are an account of a short investigation performed in collaboration with Dr. K.J. Muirhead, though the geophysical interpretation of Chapter 9 is my own. All array records used in the preparation of the dT/dΔ data were read and interpreted by the writer.

The work described is therefore entirely my own, except where due acknowledgements are made in the text. This thesis has never been submitted to another university or similar institution.

Cedric Wright,
Canberra,
July 1970.
I am indebted to my supervisor, Dr. J.R. Cleary, for advice and encouragement at all stages of the project, to Mr. H.A. Doyle and Dr. R. Underwood for valuable discussions, and to Dr. K.J. Muirhead for his collaboration in part of the work and for supplying one of the computer programmes (ZCW101D1). I also wish to thank Dr. E.W. Carpenter and Mr. B.S. Gopalakrishnan for providing the least-squares azimuth and dT/d\Delta programme (ZCW101A5).

During the four years I spent as a full time research student I was in receipt of a Commonwealth post-graduate scholarship for the first three and a half years, and an Australian National University post-graduate scholarship for the final six months.
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CHAPTER 1

INTRODUCTION

1.1 THE HISTORY AND DEVELOPMENT OF SEISMIC ARRAYS

The four seismic arrays of the United Kingdom Atomic Energy Authority (U.K.A.E.A.) type were designed and put into operation as part of a research programme into the detection and recognition of underground nuclear explosions. Their history and development have been described in detail in a report entitled "The Detection and Recognition of Underground Explosions" published in 1965 by the U.K.A.E.A. Towards the end of 1958 investigations of the means of detecting possible violations of an agreement to end nuclear tests indicated that the use of the 'first motion' of P waves as a major criterion in the identification of underground explosions was unsatisfactory. Consequently research into seismological methods for detecting underground nuclear explosions was started at the Atomic Weapons Research Establishment at Aldermaston in England in 1959. Both British research at Aldermaston and American research in the Advanced Research Projects Agency Vela Uniform Program was initially directed towards the improvement of the signal to noise ratio of distant seismic events. This gave rise to the concept of using arrays of sensors similar in principle to those used in radio astronomy, radar and underwater acoustics. The central theme in this thesis is the application of the British-designed medium aperture arrays to the investigation of the structure of the earth's interior. Two important American arrays are also described briefly to facilitate subsequent comparison of results and to illustrate the differences in the basic techniques used to elucidate similar problems. In May 1962 an array nine kilometres long was set up at Eskdalemuir in southern Scotland; the small size of the array was due to the emphasis at that time on detection in the 'first zone' (i.e. at distances of between 0° and 10° from the source). By December 1963 an array consisting of two perpendicular lines of ten seismometers spaced at 2.5 km intervals had begun operation near Yellowknife just to the north of the Great
Slave Lake in Canada. In 1965 a small array similar in size to the Eskdalemuir array (EKA), consisting of two nearly perpendicular lines of five seismometers, was installed at Gauribidanur, Mysore State, in southern India, and the Warramunga array (WRA), similar to the Yellowknife array (YKA), was installed near Tennant Creek in the Northern Territory of Australia. The Gauribidanur array (GBA) was subsequently extended so that it now consists of twenty instruments and is similar to WRA. Another array of the U.K.A.E.A. type is at present being constructed in Brazil by the United Kingdom Institute of Geological Sciences. Each of the U.K.A.E.A. arrays use short-period vertical component Willmore Mk II seismometers.

The Tonto Forest Seismological Observatory (TFSO) was installed as part of the Vela Uniform project, and began operation in 1963. It is located in central Arizona, and originally contained a circular array 3 km in diameter and two perpendicular lines about 10 km long, each containing eleven short-period vertical component instruments of the Johnson-Matheson type. During 1965 the array was temporarily extended by the addition of eight long range seismic measurement (LRSM) mobile vans; the NW-SE and NE-SW arms of the extended array were 325 km and 285 km long respectively. TFSO has been briefly described by Niazi and Anderson (1965) and by Johnson (1967). Progress and developments made under the sponsorship of the U.S. Vela Uniform Program and at Aldermaston gave rise to the concept of a large aperture seismic array (LASA). Such an array was constructed in Montana, U.S.A., and began full time operation in September 1965. It consists of 525 Hall Sears HS-10-1 short-period vertical component seismometers distributed in 21 clusters with an overall aperture of 200 km. Each subarray is 7 km in diameter, and each of the 25 seismometers is placed in a borehole 200 feet deep. Digital telemetry to the array centre is used rather than the f.m. telemetry used in the U.K.A.E.A. arrays. A plan of the array has been published by Frosch and Green (1966), Chinnery and Toksöz (1967) and by Greenfield and Sheppard (1969).
Figure 1.1 Noise spectra at WRA, uncorrected for seismometer response.
1.2 THE WARRAMUNGA SEISMIC ARRAY AND ITS APPLICATIONS

The Warramunga seismic array was installed by the U.K.A.E.A. near Tennant Creek in the Northern Territory of Australia, and began operation in October 1965; it is classed as a medium aperture array, and is now operated by the Australian National University. The array is situated on granite outcrops of the lower Proterozoic Warramunga Geosyncline, and is about 500 km from the nearest part of the Australian coast. The noise level is low though micro-seismic activity is substantially increased by low pressure atmospheric systems near the coast. Typical noise spectra on 'quiet' and 'noisy' days are shown in Figure 1.1. One stringent requirement of an array site is that the geology should be reasonably homogeneous and uncomplicated by major elastic discontinuities. It is difficult to say whether this criterion is satisfied at WRA since very little is known about the local geology. However, the results presented later do throw some light on the problem, and suggest that the choice of array site may have been an unfortunate one.

The array itself consists of twenty short-period vertical component Willmore Mk II seismometers working with a natural period of one second and a damping factor of 0.6, arranged in two lines of ten, 22.5 km long and approximately at right angles to each other as shown in Figure 1.2. Each seismometer is operated at a displacement magnification of about 250,000. The geographic coordinates of the individual seismometers and the cartesian coordinates relative to the point of intersection of the two arms of the array are listed in Table 1.1. Signals from each seismometer are telemetered to a central recording station and recorded simultaneously on 24-track magnetic tape together with a digital time code consisting of a continuous train of one second, ten second and one minute pulses. 5-bit, 6-bit and 5-bit binary codes provide the hour, minute and day respectively. The precision timing system in current use at all of the U.K.A.E.A. arrays except YKA, and the f.m. telemetry system have been described in detail by Truscott (1964) and Keen, Montgomery, Mowat, Mullard and Platt (1965). To economise in tape usage a small array or cluster
Figure 1.2 The Warramunga Seismic Array.
consisting of two lines about 2.5 km long, parallel to the main arms of the array and each containing five vertical component seismometers, was installed in 1967; this cluster discriminates against random noise by direct summation, and triggers the tape recording system only when a correlator output exceeds a specified trigger level. The mode of operation is similar to that of the 24-element cluster in use at YKA (see Whiteway, 1965). A computer system for automatic digital processing of the magnetic tape records from WRA has been developed at the Australian National University and described by Muirhead (1968b). Of more importance to this thesis, analogue outputs of all seismometers can be filtered using a number of different passbands between 0 and 12 cps and transferred on to paper using a sixteen-channel pen recorder.

An important application of an array of short-period vertical component seismometers is the determination of a velocity-depth distribution for P waves throughout the whole of the earth's mantle. Before the invention of seismic arrays the travel times of P and S phases provided the most useful data for investigating the structure of the earth's interior. The Herglotz-Wiechert method used in obtaining a velocity distribution from travel time data requires knowledge of the first derivative of the travel time curve, $dT/d\Delta$, as a function of distance; consequently $dT/d\Delta$ was obtained by differentiating a travel time curve derived by smoothing observations made at a large number of different stations. This method has in the past yielded velocity profiles that are in reasonable agreement except in the upper 1000 km of the mantle. An array can be used to measure $dT/d\Delta$ directly, and therefore should be capable of supplying more detail about the structure of the mantle than the conventional travel time method, provided the relative onset times of events at the array elements can be measured with high precision, and that the structure of the crust and upper mantle in the vicinity of the array is fairly homogeneous. All the same there is still considerable arbitrariness involved in interpreting array data.

The inverse process of using $dT/d\Delta$ and azimuth measurements, together with data from refraction experiments, to infer the structure of the crust
Table 1.1 Coordinates of the Individual Seismometers of the Warramunga Seismic Array

<table>
<thead>
<tr>
<th>Seismometer</th>
<th>Latitude E</th>
<th>Longitude S</th>
<th>Cartesian Coordinates X, km</th>
<th>Cartesian Coordinates Y, km</th>
<th>Elevation, feet</th>
</tr>
</thead>
<tbody>
<tr>
<td>B1</td>
<td>134.34776</td>
<td>19.96097</td>
<td>-0.310</td>
<td>-1.476</td>
<td>1294.85</td>
</tr>
<tr>
<td>B2</td>
<td>134.35250</td>
<td>19.94417</td>
<td>0.183</td>
<td>0.373</td>
<td>1284.27</td>
</tr>
<tr>
<td>B3</td>
<td>134.35681</td>
<td>19.92453</td>
<td>0.638</td>
<td>2.558</td>
<td>1265.89</td>
</tr>
<tr>
<td>B4</td>
<td>134.36050</td>
<td>19.90497</td>
<td>1.025</td>
<td>4.724</td>
<td>1252.65</td>
</tr>
<tr>
<td>B5</td>
<td>134.36954</td>
<td>19.87971</td>
<td>1.762</td>
<td>7.519</td>
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<tr>
<td>B6</td>
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<td>19.85645</td>
<td>1.863</td>
<td>10.095</td>
<td>1200.07</td>
</tr>
<tr>
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<td>19.84227</td>
<td>3.150</td>
<td>11.662</td>
<td>1268.09</td>
</tr>
<tr>
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<td>19.81541</td>
<td>3.291</td>
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<tr>
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<tr>
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<td>19.76862</td>
<td>4.554</td>
<td>19.816</td>
<td>1188.50</td>
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<tr>
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<td>134.34085</td>
<td>19.94411</td>
<td>-1.033</td>
<td>0.391</td>
<td>1281.07</td>
</tr>
<tr>
<td>R2</td>
<td>134.36555</td>
<td>19.94890</td>
<td>1.552</td>
<td>-0.140</td>
<td>1265.19</td>
</tr>
<tr>
<td>R3</td>
<td>134.38831</td>
<td>19.95000</td>
<td>3.934</td>
<td>-0.264</td>
<td>1236.49</td>
</tr>
<tr>
<td>R4</td>
<td>134.40802</td>
<td>19.95181</td>
<td>5.998</td>
<td>-0.465</td>
<td>1228.55</td>
</tr>
<tr>
<td>R5*</td>
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<td>19.95949</td>
<td>8.052</td>
<td>-1.314</td>
<td>1217.83</td>
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<tr>
<td>R6</td>
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<td>19.95525</td>
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<tr>
<td>R7</td>
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<td>19.95667</td>
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<td>19.96109</td>
<td>20.016</td>
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<td>1161.54</td>
</tr>
</tbody>
</table>

* moved to new site 18 September, 1968.
R2 and R6 have subsequently been moved to new sites.

The geographic coordinates of the origin of the cartesian coordinate system are 19.94777°S, 134.35081°E.
and upper mantle in the vicinity of an array is also possible if \( \frac{dT}{d\Delta} \) values derived from one of the established travel time curves are assumed to be correct. The signal enhancing ability of an array can be used to study later phases and other coherent signals following P; this is not restricted to waves travelling in the crust and mantle, and is a very powerful tool in investigating core structure.

1.3 PREVIOUS WORK ON THE STRUCTURE OF THE EARTH'S INTERIOR USING SEISMIC ARRAYS

In view of the extensive progress made in array seismology since I started working on WRA data, it is pertinent at this stage to trace the history of array investigations with special emphasis on the structure of the earth's mantle. Array studies of P wave travel time gradients were pioneered by Niazi and Anderson (1965) who investigated the upper mantle structure below western North America using P arrivals from 70 events at distances between 10° and 30° recorded at the original 10 km Tonto Forest array. Niazi (1966) considered the effect of a dipping M-discontinuity on \( \frac{dT}{d\Delta} \) and azimuth measured at an array, and showed with reference to data recorded at TFSO at distances between 32.5° and 80.1° how the details of an underlying structure can be inferred from measurements of both \( \frac{dT}{d\Delta} \) and azimuth. A rather similar type of study was made by Otsuka (1966a, b) using events recorded at the Central California Seismographic Array.

When I started work on WRA data towards the end of 1966, the only velocity distribution obtained from array data was that of Niazi and Anderson; no array results for the lower mantle had yet been published. Kanamori (1967) made measurements of apparent velocities for about 50 earthquakes recorded at the Wakayama Microearthquake Observatories in Japan over the distance range 5° to 55°. Johnson (1967) published a velocity structure for P waves in the uppermost 750 km of the mantle derived from \( \frac{dT}{d\Delta} \) data from 52 earthquakes and explosions in the distance range 0° to 30° recorded at the extended TFSO. Chinnery and Toksöz (1967) gave the first set of P wave velocities for the lower mantle derived from array data. They used 167 events in the distance range
27° to 90° recorded at the LASA in Montana. Greenfield and Sheppard (1969) measured the travel time gradients for epicentres to the northwest and southeast of the LASA, and proposed a model for the crust beneath the array which explains the station residuals, and accounts for the major differences between the measured and the Jeffreys and Bullen (1958) \( \frac{dT}{d\Delta} \) values. Chinnery (1969) presented observations of \( \frac{dT}{d\Delta} \) on P wave arrivals at the LASA in Montana for about 400 events over the distance range 25° to 98° in a northwesterly azimuth from the array. Johnson (1969), using the extended TFSO, continued his earlier work to the base of the mantle by measuring \( \frac{dT}{d\Delta} \) for direct P from 212 earthquakes in the distance range 30° to 100°.

Preliminary work on the effects of local structure on the measurement of \( \frac{dT}{d\Delta} \) at WRA was published by Cleary, Wright and Muirhead (1968). I (Wright, 1968) found evidence for a low velocity layer at a depth close to 800km using \( \frac{dT}{d\Delta} \) measurements at WRA from 30 earthquakes in the Mariana Islands and surrounding regions. Gopalakrishnan (1969) made measurements of \( \frac{dT}{d\Delta} \) for P arrivals from 88 events recorded at GBA over the distance range 25° to 95°, but used too little data to enable a really detailed velocity model to be deduced. A \( \frac{dT}{d\Delta} \) curve for the distance range 30° to 104° using data from all four U.K.A.E.A. arrays was obtained by Corbishley (1970). His method of analysis of the relative arrival times at the arrays involves attempting to estimate \( \frac{dT}{d\Delta} \) and the crustal effects beneath each array simultaneously. In November 1969, I (Wright, 1970) submitted for publication a discussion of the preliminary \( \frac{dT}{d\Delta} \) curve described in Chapters 4 and 8. Since that time the WRA data for the lower mantle have been amplified, and a refined \( \frac{dT}{d\Delta} \) curve has been derived.

The signal enhancing ability of arrays has not yet been extensively adapted to studies of the structure of the earth's interior. Various applications have been described by Agger and Carpenter (1964), Birtill and Whiteway (1965), Hannon and Kovach (1966), Key (1967), Whitham and Weichert (1968) and Weichert and Whitham (1969). Nevertheless much valuable information
stored in array tape records still remains virtually unexploited. Consequently, in collaboration with Dr. K.J. Muirhead, I have attempted to show how some of this array data can be used to study the structure of the earth's core, as well as the crust and upper mantle, employing both visual matching and array phasing techniques (see Wright and Muirhead, 1969).

1.4 AIM AND SCOPE OF THESIS IN THE LIGHT OF PREVIOUS WORK

The major aim of this study has been to derive a P wave velocity distribution for the whole of the earth's mantle, using travel time and travel time gradient measurements of events recorded at WRA. The scope of the thesis is perhaps best conveyed by considering the basic problem that the writer has tried to solve, which may be formulated as follows: how can a seismic array of about 20 km in aperture be used to investigate the structure of the earth's interior, and how detailed is the information that it is capable of yielding? The work described subsequently is an attempt to demonstrate that a medium aperture array such as WRA is just as useful in revealing the fine details of mantle structure as a large aperture device such as the LASA in Montana or the extended TFSO, even though the problems associated with local structure may be serious; the approach has been to use the P wave velocity models for the earth's mantle derived from dT/dA measurements at WRA to try and resolve some of the current ambiguities and controversies concerning the existence of velocity anomalies in the mantle. It is stressed that the data would be better if a more satisfactory array site had been chosen. The greatest difficulty in this work is in ascertaining the relative extent to which the fine detail in the dT/dA data is due to lateral variations in the structure of the crust and upper mantle in the vicinity of the array, to real anomalous regions at great depth and to rapid lateral changes in mantle structure close to the source. Unfortunately the only seismic refraction survey of the Tennant Creek area (Underwood, 1967) was insufficiently exhaustive to reveal detailed crustal structure. In an effort to overcome some of the ambiguities in interpretation, two different approaches to the problem of the interpretation of
the $dT/dA$ data have been tried.

As well as the $dT/dA$ studies the signal enhancing ability of the array has been exploited in an investigation of later P phases and coherent signals following the first arrival generated by a large nuclear explosion in Novaya Zemlya. Since very little work has been done in applying phasing techniques to the study of the earth's interior, some ideas and preliminary results have been advanced, first to show that information of very high quality and of considerable value can be obtained from the U.K.A.E.A. arrays, and secondly to illustrate the applications and potentialities of the array processing scheme developed at the Australian National University and described by Muirhead (1968b).

To clarify the division of Chapters 5 to 7, the mantle has been divided into two major regions: the upper mantle and the lower mantle. Using the notation invented by Bullen (1963, pp. 222–223), the upper mantle and lower mantle are defined as comprising the B and C regions and the D region respectively. Following Anderson (1967a) the boundary between the C and D regions has been taken at a depth close to 700 km. The thesis has been logically split into three main sections as outlined below.

(a) **Measurements and Techniques**

**Measurement of Travel Time Gradients and Azimuths of P Wave Arrivals** (Chapter 2).

Since the seismometer outputs are usually coherent across a 20 km array, visual matching techniques can be used to obtain accurate relative onset times at the individual seismometers. The $dT/d\Delta$ (or slowness) and azimuth values have been calculated by means of a modified version of a least squares programme designed by Dr. E.W. Carpenter and written by Mr. B.S. Gopalakrishnan. Alternative means of $dT/d\Delta$ and azimuth determination have been examined, and a careful study of the sources of error, both random and systematic, has been included. For $dT/d\Delta$ studies, events showing reasonably clear P onsets have been selected from a wide range of distances and azimuths,
and have been classified into a number of different groups depending on the focal depth and the quality of the recorded information.

The Structure beneath the Array (Chapter 3).

In the early stages of this project it was important to find out how serious were the effects of structure in the vicinity of the array on the measurement of $dT/d\Delta$ and azimuth. For events at teleseismic distances ($\Delta > 28^\circ$) the azimuth values often differed considerably from the true azimuths. Further the $dT/d\Delta$ values often differed significantly from those predicted from the Jeffreys-Bullen (J-B) Tables (1958) or the Seismological Tables for P Phases (1968). In addition the errors in the measurements of onset times, in epicentre determinations or in the travel time tables themselves can only account for anomalies in both $dT/d\Delta$ and azimuth an order of magnitude less than those observed. Niazi (1966) considered the effect of a dipping M-discontinuity on measurements of slowness and azimuth. Using the theory outlined in Niazi's paper, a computer programme has been written that will work out the effect on $dT/d\Delta$ and azimuth of any combination of dipping interfaces for a seismic event whose azimuth and distance are accurately known. From the results of the seismic experiment WRAMP, Underwood (1967) inferred the presence of a near-surface dipping structure. Although P arrivals from some azimuths do show azimuth and $dT/d\Delta$ anomalies consistent with a structure similar to that derived by Underwood, the overall picture is one of great complexity. Since the simple approach of trying to fit one or more dipping interfaces beneath the array was not successful, an attempt to derive a 'structure surface' that will explain the complex pattern of slowness and azimuth anomalies has been made. The main purpose in deriving models of the local structure is to enable corrections to be applied to $dT/d\Delta$ values before inverting the data to obtain a velocity distribution. The station correction has been calculated from the travel time residuals for teleseisms to try and detect a correlation with the $dT/d\Delta$ and azimuth anomalies. Finally attempts have been made to give a meaningful physical explanation of the complex pattern of $dT/d\Delta$ and azimuth anomalies.
The Construction of a $dT/d\Delta$ Curve and a Velocity Model (Chapter 4).

Having obtained the raw $dT/d\Delta$ data corrected for structure, the next step is to reduce the random error as much as possible without losing useful information. Arnold (1968) drew attention to the Method of Summary Values invented by Jeffreys (1937, 1961, pp. 223–227), and suggested that it has advantages over a polynomial regression; this technique has been used to smooth the $dT/d\Delta$ data in preference to the standard technique of fitting a polynomial or a series of polynomials by least squares. Owing to the extreme complexity of the local structure at WRA, initially a purely empirical approach to the problem of correcting the $dT/d\Delta$ values beyond $27.9^\circ$ has been adopted. This simple approach has subsequently been refined making use of much of the work on the local structure discussed in Chapter 3. The computational methods used in inverting body wave data and in investigating ray paths for regions of anomalous velocity change have been briefly described.

(b) Results and Interpretation

Upper Mantle Structure. (Chapter 5).

During recent years a considerable body of evidence has been published indicating the presence of two regions in the upper mantle, at depths close to 400 km and 650 km, where the velocity gradients for P waves are high. These regions have now been identified for the mantle below the Australasian region. Slowness measurements of first and second P onsets from earthquakes in Celebes, the Moluccas, Halmahera and surrounding regions have been compared with similar measurements over a similar distance range from events in Eastern New Guinea, New Britain and surrounding regions. Two upper mantle velocity models have been derived by the Herglotz-Wiechert method. Regional differences and rapid lateral variations in upper mantle structure have been considered.
The Lower Mantle, Part 1: the Region between 700 km and 1000 km 
($28^\circ < \Delta \leq 42^\circ$). (Chapter 6).

There is evidence of considerable complexity in the structure of the mantle between about 700 km and 1000 km. It is certain that there is an anomalous region close to 800 km, and current explanations of its nature are in conflict. Exhaustive $dT/d\Delta$ measurements from earthquakes of the Caroline and Mariana Islands and surrounding regions have been compared with similar data from the Solomon, Santa Cruz, New Hebrides and Loyalty Islands and surrounding regions, and also with less copious data from other areas. Using simple ray theory $dT/d\Delta$ curves for different types of velocity anomaly that might explain some of the $dT/d\Delta$ observations have been presented. Finally regional differences in structure and petrological implications have been discussed.

The Lower Mantle, Part 2: 1000 km to 2890 km ($42^\circ < \Delta \leq 99^\circ$). (Chapter 7).

The basic slowness data come from earthquakes and explosions of five main regions: (i) the East China Sea, Taiwan, the Ryukyu Islands and surrounding regions, (ii) the Bonin Islands, the Japan region, the Kurile Islands, Kamchatka and surrounding regions, (iii) the Aleutian Islands and Alaska regions, (iv) the Fiji, Tonga and Kermadec Islands region and (v) the Asian continent and surrounding regions. The aim has been to determine whether the data support the recent evidence for anomalous velocity gradients at depths close to 1000, 1200, 1600, 1900 and 2300 km, and whether any regional differences in structure at these depths can be detected.

P Wave Velocities in the Mantle below 700 km. (Chapter 8).

The smoothed $dT/d\Delta$ curves presented in Chapter 4 have been inverted by the Herglotz-Wiechert method to obtain two P velocity distributions for the lower mantle. A third velocity model and two regional $dT/d\Delta$ curves are also included. The scatter of the $dT/d\Delta$ data has been examined in relation to the quality of the recorded information and the focal depths of the events used. Finally my results have been compared with recent travel time and amplitude
studies as well as other array investigations. I have attempted to show that when my work is taken in conjunction with other array studies, a fairly coherent picture of lower mantle structure emerges.

(c) Further Applications of the Array. (Chapter 9).

The applications of the signal enhancing ability of the array to the study of the structure of the earth's interior have been investigated in collaboration with Dr. K. J. Muirhead. A large Soviet nuclear explosion detonated on Novaya Zemlya, at a distance of 106.0°, was well recorded by the array and a number of later phases as well as P were visible. Measurements of apparent velocity and azimuth have been used as an aid in identifying the phases PKiKP, PP, PcPPcP, PKKP, PcPPKP and PKPPKP in the explosion record. Similar measurements have been made for coherent signals not corresponding to any of the conventional phases; these signals include precursors to PP, and possible explanations of their origin have been presented.

Conclusions. (Chapter 10).

The results of this study have been summarised, and the relationship of this work to other branches of geophysics and geochemistry has been evaluated. A number of suggestions for future investigations in array seismology have been put forward.

The computer programmes used throughout this thesis, together with the relevant references and theory, have been incorporated in a separately bound collection of appendices. These appendices also contain details of all the seismic events used and lists of the basic dT/dΔ and azimuth data. In the literature citations at the end of the thesis I have chosen the form recommended in the pamphlet 'JGR Style - A Guide for Authors' (1968) published by the American Geophysical Union. This list of references also includes all those citations occurring in the appendices.
CHAPTER 2

MEASUREMENT OF TRAVEL-TIME GRADIENTS AND AZIMUTHS OF P WAVE ARRIVALS

2.1 INTRODUCTION.

With an array of short-period vertical component seismometers the travel-time gradient or slowness, \( \frac{dT}{d\Delta} \), which is required as a function of distance in deriving a velocity distribution for P waves throughout the earth's mantle, can be measured directly. Both \( \frac{dT}{d\Delta} \) and the azimuth of arrival at an array of P wave onsets can be estimated from a set of arrival times at the individual seismometers. For an array to be a useful tool in investigating mantle structure it is necessary to keep the standard error on any \( \frac{dT}{d\Delta} \) value below 0.2 sec/deg. To achieve this sort of precision it is essential to be able to measure onset times at the seismometers of a medium aperture array with a standard deviation of between 0.01 and 0.03 seconds. Since the seismometer outputs are usually coherent across such an array, visual matching techniques can be used to obtain relative onset times of the specified accuracy. The slowness and azimuth values have been calculated by means of a least-squares computer programme designed by Dr. E.W. Carpenter and written by Mr. B.S. Gopalakrishnan. A subroutine has been added to this programme that calculates the standard errors in \( \frac{dT}{d\Delta} \) and azimuth using the formulae given by Kelly (1964). It was also important to determine whether the least-squares technique was the best available for deriving precise values of \( \frac{dT}{d\Delta} \) and azimuth. Other means of \( \frac{dT}{d\Delta} \) and azimuth determination include a correlation method described by Birtill and Whiteway (1965), and a Fourier transform method described by Shima, McCamy and Meyer (1964). With the help of computer programmes developed and described by Muirhead (1968b), these techniques have been compared with the least-squares method.

It is important to consider sources of systematic error in the \( \frac{dT}{d\Delta} \) and azimuth measurements. By far the most serious effect on \( \frac{dT}{d\Delta} \) and
Figure 2.1 Diagram to illustrate a least-squares method of estimating $dT/d\Delta$ and azimuth.
azimuth is produced by the structure of the crust and possibly the upper mantle beneath the array, but discussion of this problem has been deferred until Chapter 3. It was also necessary to find out how serious were the effects of the different instrumental constants of the individual seismometers on the measured values of $dT/d\Delta$ and azimuth. This problem has been examined in the light of a study by Muirhead (1968b) of phase shifts caused by variations in instrument characteristics. Another small source of systematic error in the $dT/d\Delta$ measurements is the ellipticity of the earth, and this has also been examined quantitatively. Finally the criteria used to select events suitable for $dT/d\Delta$ measurements have been stated, and the events themselves have been classified into a number of different groups depending on the focal depth and the quality of the recorded information.

2.2. THE MEASUREMENT OF $dT/d\Delta$ AND AZIMUTH OF P WAVES.

(i) Fitting a $dT/d\Delta$ and Azimuth Value through a Set of Onset Times by the Method of Least Squares.

This least-squares method has been described by Kelly (1964) and Otsuka (1966a), and is summarised below. Select an arbitrary origin $O$, in practice the point of intersection of the two arms of WRA, and take a cartesian coordinate system with the $y$ and $x$ axes pointing north and east respectively, as shown in Figure 2.1. For an array the size of WRA any curvature of the earth's surface can be neglected. Consider a seismometer at $S_i$ distance $R_i$ from $O$, and let $\angle S_i O y$ be $\theta_i$. Now suppose a P wave arrival crosses the array from azimuth $\phi$ and with apparent surface velocity $V$. It is assumed that the wavefront is plane; this assumption is approximately true for events at distances greater than 10° provided the structure of the crust and the upper mantle in the vicinity of the array is reasonably simple. At this stage it is worth pointing out that in the literature some authors use apparent velocities, $d\Delta/dT$, in km/sec and others the reciprocal velocity or slowness, $dT/d\Delta$, in sec/deg. As far as deriving a velocity model for the earth's mantle is concerned it is more logical to use $dT/d\Delta$ which will generally be used throughout
this thesis.

Let the arrival time at $S_i$ be $T_i$ subject to an error $\epsilon_i$. Arrival
time at $O$ is $T_o$. Then

$$T_o = T_i + \frac{R_i \cos (\phi - \theta_i)}{V} + \epsilon_i$$  \hspace{1cm} (2-1)$$

Hence

$$\sum_{i=1}^{N} \epsilon_i^2 = \sum_{i=1}^{N} (T_o - T_i - [x_i P + y_i Q])^2$$

$$= E(P, Q, T_o)$$ \hspace{1cm} (2-2)$$

where $P = \sin \phi / V$ and $Q = \cos \phi / V$. $N$ is the number of seismometers.

To obtain best estimates of the three unknowns $P$, $Q$ and $T_o$, the usual least
squares condition that $E(P, Q, T_o)$ is a minimum is used: i.e. $\frac{\delta E}{\delta P} = 0$. In this way three normal equations are obtained as follows:

$$\sum_{i=1}^{N} x_i (T_o - T_i - [x_i P + y_i Q]) = 0 \hspace{1cm} (2-3)$$

$$\sum_{i=1}^{N} y_i (T_o - T_i - [x_i P + y_i Q]) = 0 \hspace{1cm} (2-4)$$

$$\sum_{i=1}^{N} (T_o - T_i - [x_i P + y_i Q]) = 0 \hspace{1cm} (2-5)$$

Expressions (2-3) to (2-5) can conveniently be rewritten in the following form:

$$[XX] P + [XY] Q + [XT] - [X] T_o = 0 \hspace{1cm} (2-3a)$$

$$[XY] P + [YY] Q + [YT] - [Y] T_o = 0 \hspace{1cm} (2-4a)$$

$$[Y] P + [X] Q + [T] - N T_o = 0 \hspace{1cm} (2-5a)$$

where $[XY]$ denotes the summation $\sum_{i=1}^{N} x_i y_i$.

These three normal equations are solved to give best estimates of $P$, $Q$ and $T_o$. 
Then
\[ V = (P^2 + Q^2)^{-\frac{1}{2}} \quad (2-6) \]
\[ \frac{dT}{d\Delta} = r_0 (P^2 + Q^2)^{\frac{1}{2}} \quad (2-6a) \]

where \( r_0 \) is the mean radius of the earth.

\[ \phi = \tan^{-1} \left( \frac{P}{Q} \right) \quad (2-7) \]

A computer programme for determining \( V \), \( \frac{dT}{d\Delta} \) and \( \phi \), using the method described above, was designed by Dr. E.W. Carpenter and written by Mr. B.S. Gopalakrishnan. In his description of this technique Otsuka simplifies the calculation by choosing the origin of the coordinate system to be the centroid of the array seismometer points, when from (2-5a)

\[ T_o = \frac{1}{N} \left[ T \right] \quad (2-8) \]

This simplification has not been used for WRA data for the following reason: the number of seismometers working satisfactorily varies from day to day so that the centroid would vary from one event to another; this is undesirable as calculations of the seismometer residuals with respect to the true azimuth and the expected value of \( \frac{dT}{d\Delta} \) derived from an established travel-time curve would be meaningless. For events recorded at WRA it is useful to be able to compare the residuals from one event to another so that a fixed origin is necessary.

**Random Errors in \( \frac{dT}{d\Delta} \) and Azimuth.** Suppose the errors in the onset time measurements, \( \epsilon_i \), are independent Gaussian variables, and that each \( \epsilon_i \) has mean zero and variance \( \sigma^2 \). It is not suggested that this is a realistic model of the actual errors, but if will at least help to indicate the sensitivity of the results to measurement errors. On this model Kelly (1964) showed that the root mean square errors in \( V \), \( \frac{dT}{d\Delta} \) and \( \phi \) are given by

\[ \left\langle \frac{\delta V}{V} \right\rangle_{\text{r.m.s.}} = \left\langle \frac{\delta \left( \frac{dT}{d\Delta} \right)}{\text{r.m.s.}} \right\rangle \]

\[ = \frac{\sigma V}{\sqrt{N D}} \left[ \text{Var} x \cos^2 \phi - 2 \text{Cov}(x,y) \sin \phi \cos \phi + \text{Var} y \sin^2 \phi \right]^{\frac{1}{2}} \quad (2-9) \]
where $\langle \delta \phi \rangle_{\text{r.m.s.}}$ is in radians and

$$D = \text{Var} x \text{Var} y - [\text{Cov}(x, y)]^2$$

(2-11)

Also

$$\text{Var} x = \frac{1}{N} \sum_{i=1}^{N} (x_i - \bar{x})^2$$

(2-12a)

$$\text{Var} y = \frac{1}{N} \sum_{i=1}^{N} (y_i - \bar{y})^2$$

(2-12b)

$$\text{Cov}(x, y) = \frac{1}{N} \sum_{i=1}^{N} (x_i - \bar{x})(y_i - \bar{y})$$

(2-12c)

$$\bar{x} = \frac{1}{N} \sum_{i=1}^{N} x_i$$

(2-13a)

$$\bar{y} = \frac{1}{N} \sum_{i=1}^{N} y_i$$

(2-13b)

Kelly and Otsuka both discussed the application of formulae (2-9) and (2-10) to specific arrays. It is, of course, important to devise a satisfactory way of estimating $\sigma$. One approach to this problem is to take a number of events occurring in a small region of the earth and to estimate a random reading error in the manner described by Corbishley (1970). Four important considerations have led the writer to approach this problem in a different way. If a plane wave front crosses the array the error in each onset time depends on the following factors: (i) the waveform of the event - low frequency onsets give less clearly defined peaks and cross-over points (or zeros); (ii) the signal to noise ratio - random bursts of noise can cause spurious changes of waveform from one seismometer to another when the signal to noise ratio is only moderately large; (iii) the instrumental constants of the seismometers and (iv) rapid variations
STANDARD DEVIATION ON EACH ONSET TIME = 0.015 SECONDS  
APPARENT VELOCITY = 20.0 km/sec  
$\frac{dT}{d\Delta}$ = 5.560 sec/deg  

Figure 2.2 Errors in apparent velocity, $dT/d\Delta$ and azimuth for the complete array.
of local structure - the assumption that a plane wave front crosses the array is a better approximation for some azimuths and distances than for others. The last two factors are strictly speaking sources of systematic error, but have been included as they do have an important bearing on the precision of the dT/dΔ and azimuth determinations. Consequently it seems preferable to calculate a value of σ for each set of relative onset times. This is done by assuming that the residuals have zero mean and variance \( \sigma^2 \) characteristic of that particular set of arrival times. Each value of the residual \( \epsilon_i \) is calculated once \( P \), \( Q \) and \( T_0 \) have been estimated, and hence

\[
\sigma^2 = \frac{\sum_{i=1}^{N} \epsilon_i^2}{N - 3}
\]

(2-14)

The writer has added a subroutine to the basic computer programme of Carpenter and Gopalakrishnan that will calculate the standard errors in \( \phi \) and dT/dΔ using expressions (2-9) and (2-10). It is stressed that the estimate of \( \sigma \) contains not only random reading errors, but also systematic errors due to possible differences in instrumental constants of the individual seismometers and to curvature of the allegedly plane wave front. The programme (ZCW101A5) also calculates the apparent velocity or dT/dΔ assuming the azimuth calculated from the U.S.C.G.S. epicentre determination is correct. Two sets of time residuals are calculated: (i) for the least-squares azimuth and dT/dΔ fit, and (ii) taking a dT/dΔ value from an established travel-time curve (e.g. Jeffreys and Bullen, 1958) and the U.S.C.G.S. azimuth.

The theoretical errors in dT/dΔ, apparent velocity and azimuth were calculated using computer programme ZCW101A7 and plotted graphically in Figures 2.2 and 2.3 as a function of azimuth; in each case an apparent velocity of 20.0 km/sec (i.e. dT/dΔ = 5.560 sec/deg) was used, and the standard deviation on a single onset time was taken as 0.015 seconds. In Figure 2.2 the errors have been plotted for the complete array and in Figure 2.3 for the array with seismometers R7-R10 removed. Note the marked asymmetry in the azimuth and dT/dΔ errors in Figure 2.2, even though the two lines of the
Figure 2.3 Errors in apparent velocity, dT/dΔ and azimuth for the array without R7 - R10.
array are approximately of equal length and almost at right angles; note also
that the azimuths corresponding to maximum error in $dT/d\Delta$ also correspond
to minimum error in azimuth and vice versa. Another important result is that
the errors in $dT/d\Delta$ for a fixed standard deviation and array configuration are
independent of $dT/d\Delta$. Let $A = \text{Var } x, \ B = \text{Var } y$ and $H = \text{Cov } (x, y)$. Then (2-9) becomes

$$
\left( \frac{\delta (dT/d\Delta)}{dT/d\Delta} \right)^2 = R^2 = K^2 [A \cos^2 \phi - H \sin 2 \phi + B \sin^2 \phi] \quad (2-15)
$$

where $K = \frac{\sigma V}{\sqrt{N \cdot D}}$. The problem is to find the directions in which the errors
are maximum or minimum. Differentiating (2-15)

$$
2R \frac{dR}{d\phi} = K^2 [(B - A) \sin 2 \phi - 2H \cos 2 \phi] \quad (2-15a)
$$

Since $R > 0$ for all $\phi$, $dR/d\phi = 0$ when

$$
tan 2 \phi = \frac{2H}{B-A} \quad (2-16)
$$

As $\tan (\pi + x) = \tan x$, there will in general be four solutions of (2-16) between
0 and $2\pi$. Using (2-16) greatest precision in azimuth and least precision in
$dT/d\Delta$ for the complete array are obtained when $\phi = 52^\circ 33' \text{ and } 232^\circ 33'$. In
addition least precision in azimuth and greatest precision in $dT/d\Delta$ occur
when $\phi = 142^\circ 33'$ and $322^\circ 33'$. The maximum error divided by the minimum
error for either $dT/d\Delta$ or azimuth is 1.45. For Figure 2.3 greatest precision
in azimuth and least precision in $dT/d\Delta$ occur when $\phi = 88^\circ 33'$ and $268^\circ 33'$. Similarly least precision in azimuth and greatest precision in $dT/d\Delta$ occur when
$\phi = 178^\circ 33'$ and $358^\circ 33'$. In this case the ratio of the maximum error to the
minimum error for either $dT/d\Delta$ or azimuth is 2.28. These simple calculations
demonstrate the strong dependence of the precision of $dT/d\Delta$ measurements on
azimuth and on the number of seismometers operating.
(ii) Correlation and Other Techniques.

A correlation method of azimuth and \( \frac{dT}{d\Delta} \) determination has been described by Birtill and Whiteway (1965). The main application of this technique is in picking out small coherent signals partly masked by noise, and its uses will be explored in detail in Chapter 9. The method may be summarised briefly as follows. For each line the appropriate time delays corresponding to tuning to a particular azimuth and \( \frac{dT}{d\Delta} \) are inserted for each seismometer, and a summed output for each line is determined. Then if \( E_R \) and \( E_B \) are the normalised output amplitudes of the summed red and blue lines respectively, the cross correlation integral assuming a sinusoidal signal of angular frequency \( \omega \), (see Birtill and Whiteway, 1965), is given by

\[
\phi_{RB}(t_1) = \frac{1}{T_1} \int_{t_1-T_1}^{t_1} E_R(t) E_B(t) \cos \omega t \cos (\omega t + \gamma_R - \gamma_B) \, dt \quad (2-17)
\]

taking a square window of integration of length \( T_1 \). \( \gamma_R \) and \( \gamma_B \) are the phase angles of the vectors representing the summed outputs of the red and blue lines respectively. Provided \( \omega T_1 \gg 1 \) (in practice about 10 to 40) \( \phi_{RB} \approx \frac{1}{2} E_R E_B \cos (\gamma_R - \gamma_B) \quad (2-18) \)

Thus, for a continuous sinusoidal signal, \( \phi_{RB}(t_1) \) is independent of \( T_1 \) if \( T_1 \) is sufficiently large. The correlator output \( \phi_{RB} \) presented as a function of azimuth and \( \frac{dT}{d\Delta} \) will give a maximum value at the signal azimuth and velocity. This assumption of a continuous sinusoidal signal is at best only a very rough approximation to the truth. A computer programme to calculate apparent velocity and azimuth automatically on the IBM 360/50 has been written and described by Muirhead (1968b). The seismometer outputs are digitised at about 25 samples per second, and the array is tuned to a wide range of velocities and azimuths to find an approximate maximum value of \( \phi_{RB} \); then a finer set of correlations is made over a limited range of velocity and azimuth.
Table 2.1. Correlation and Least Squares Measurements of Azimuth, Apparent Velocity and dT/d\(\Delta\) for the P Onset of the Novaya Zemlya Nuclear Explosion of October 27, 1966.

### Least Squares Measurements

<table>
<thead>
<tr>
<th>Apparent Velocity (km/sec)</th>
<th>dT/d(\Delta) sec/deg</th>
<th>Measured Azimuth deg</th>
<th>Variance, (\text{sec}^2 x 10^{-3})</th>
<th>No. of Seismometers Used</th>
<th>Matching Technique Used</th>
</tr>
</thead>
<tbody>
<tr>
<td>22.14 ± 0.20</td>
<td>5.023 ± 0.045</td>
<td>344.4 ± 0.7</td>
<td>0.159</td>
<td>20</td>
<td>First zero</td>
</tr>
<tr>
<td>22.64 ± 0.30</td>
<td>4.912 ± 0.065</td>
<td>344.2 ± 0.9</td>
<td>0.292</td>
<td>18</td>
<td>First peak</td>
</tr>
<tr>
<td>22.04 ± 0.24</td>
<td>5.044 ± 0.056</td>
<td>345.1 ± 0.8</td>
<td>0.241</td>
<td>19</td>
<td>Trace</td>
</tr>
<tr>
<td>22.30 ± 0.27</td>
<td>4.985 ± 0.060</td>
<td>345.2 ± 0.9</td>
<td>0.287</td>
<td>20</td>
<td>First zero</td>
</tr>
<tr>
<td>22.46 ± 0.46</td>
<td>4.951 ± 0.101</td>
<td>343.5 ± 1.3</td>
<td>0.417</td>
<td>15</td>
<td>Second zero</td>
</tr>
</tbody>
</table>

### Correlation Measurements

<table>
<thead>
<tr>
<th>Apparent Velocity km/sec</th>
<th>dT/d(\Delta) sec/deg</th>
<th>Measured Azimuth, deg</th>
<th>Length of Window of Integration, sec</th>
<th>Time for Start of Correlation, (GMT)</th>
</tr>
</thead>
<tbody>
<tr>
<td>21.0</td>
<td>5.30</td>
<td>344.2</td>
<td>1.0</td>
<td>6-12-12.0</td>
</tr>
<tr>
<td>20.6</td>
<td>5.40</td>
<td>347.8</td>
<td>3.0</td>
<td>6-12-11.0</td>
</tr>
</tbody>
</table>
values to give the maximum value of $\phi_{RB}$ and hence the azimuth and $d\Delta/dT$. Birtill and Whiteway also considered taking a simple vector sum of the seismometer outputs with the appropriate time delays inserted for tuning to a specific apparent velocity and azimuth; they showed that the sum squared response for an L-shaped array is inferior to the correlator response.

The correlation technique was compared with the least-squares procedure for several events, and the former was shown to give insufficiently high precision to be useful in the work on mantle P wave velocities. To illustrate this, some selected correlation and least-squares apparent velocity and azimuth measurements for the P onset of the Novaya Zemlya nuclear explosion of October 27, 1966, have been included in Table 2.1 (see also Tables 9.1 and 9.2). A new correlation technique that appears to give more reliable azimuth and $d\Delta/dT$ measurements has been described by Muirhead (1968a).

A Fourier transform technique of $d\Delta/dT$ and azimuth determination has been described by Shima et al. (1964); by measuring the differences in phase angle of Fourier spectral densities across a fixed array 1.25 km in length, they determined apparent velocities at a number of different frequencies. A modified version of their method was devised for automatic determination of $d\Delta/dT$ and azimuth on the IBM 360/50 computer by Muirhead (1968b), and I compared the results obtained from Muirhead's programme with the results obtained from the same events by fitting a velocity and azimuth through visually measured onset times by least squares. The results using the Fourier transform technique were not encouraging, and the method was consequently abandoned.

2.3 THE MEASUREMENT OF RELATIVE ONSET TIMES.

In order to determine $dT/d\Delta$ with a precision similar to that obtained from a large aperture array, such as the LASA in Montana or the extended TFSO, onset times must be measured with a standard deviation of between 0.01 and 0.03 seconds. It is virtually impossible to measure P onset
Figure 2.4 Matching technique for determining relative arrival times at each seismometer. From the record at pit R4(a), a family of curves with different amplitudes is constructed (b), and matched with the records at the other pits (c), (d).
times with such an accuracy. However, if the P waves recorded at each seismometer are matched with each other, the measurement of relative P times to the required accuracy is quite feasible. Accurate measurement of the relative arrival times of surface waves by matching waveforms has been a common technique in the calculation of the phase velocities of surface waves across tripartite arrays since the method was first described by Evernden (1953). Matching P waves having periods of about one second to an accuracy approaching 0.01 seconds is no more difficult than matching surface waves with periods of 20 seconds and greater to an accuracy of about 0.2 seconds, provided a sufficiently expanded time scale is used. For this array study a manual technique illustrated by the Longshot records in Figure 2.4 has been used. The procedure is as follows:

1. For a particular event each seismometer output is band-pass filtered between 0.4 and 2.0 cps and transferred on to graph paper at a speed of about 40 mm/second using a 16-channel pen recorder.

2. In order to compensate for variations in amplitude between records, the P waves from one record are used to construct, on transparent paper, a family of curves having a range of amplitudes.

3. Using this paper as an overlay, the curves are matched with the P waves recorded at each seismometer. The portion of the record most suitable for accurate matching is the section between the first peak and the first trough of the P wave train. In this manner a set of relative onset times is obtained.

4. The position of the first peak (or trough), second peak (or trough), or first zero of each seismometer output is measured relative to a fixed reference time, thus giving other sets of relative onset times.

So for each event, by matching different parts of the P wave train, a number of different sets of relative onset times can be obtained. In practice two or three sets of onset times for each event were derived, depending on
whether the P onset was only fairly clear or exceptionally good; each set of
arrival times was regarded as independent, and a value of dT/dΔ and azimuth
was fitted through each set using programme ZCW101A5. In this way a total
of about 700 values of dT/dΔ and azimuth for first arrivals from 292 events
over the distance range 13.1° to 106.0° has been obtained; this covers the
region of the mantle from a depth of about 100 km to the mantle-core boundary.
Over some distance ranges, particularly between 17° and 25° which covers a
region of the upper mantle in which there are two important discontinuities,
dT/dΔ and azimuth measurements have been made for second arrivals. All
measured values of dT/dΔ and azimuth, together with relevant information
on the events themselves, have been listed in Tables 1 to 3 of Appendix 2.
If, in any dT/dΔ and azimuth determination, a seismometer residual exceeded
0.04 seconds, the corresponding onset time was discarded and new dT/dΔ
and azimuth values were calculated.

The least-squares method has been adapted for automatic dT/dΔ
and azimuth determination on the IBM 360/50 using digitised seismometer
outputs. The procedure used was to locate the first zero or cross-over point
beyond the first peak whose magnitude exceeded a certain specified value. The
programme used for this was ZCW101A9. Relative onset times were meas­
ured automatically in this fashion for the first two cross-over points of the
P onset of the Novaya Zemlya nuclear explosion of October 27, 1966; the
results are listed in Table 2.1 for comparison with the azimuth and dT/dΔ
values derived from visual measurements of relative onset times. For events
with large, clear P onsets the results are most satisfactory even with only
25 samples per second, and this automatic measurement of relative arrival
times is well worth exploring further in the future.

2.4 SOURCES OF SYSTEMATIC ERROR IN dT/dΔ AND AZIMUTH MEASURE­
MENTS.

Large systematic errors in dT/dΔ and azimuth values due to the
structure of the crust and upper mantle in the vicinity of the array are known
to exist, and the whole of Chapter 3 is devoted to a discussion of the effects of local structure on measurements of $dT/d\Delta$ and azimuth. However, in this section only the problem of whether errors in these parameters due to differences in the instrumental constants of the individual seismometers and to the ellipticity of the earth will be considered.

(i) **The Effect of Differences in Seismometer Constants.**

Muirhead (1968b) investigated the problem of phase shifts due to small differences in seismometer constants in a medium aperture array. He showed that for an instrument with a natural frequency of 0.9 cps recording a sinusoidal motion of period one second, the first zero or cross-over point of the recorded wave would be 0.04 seconds later than if the instrument had a natural frequency of 1.1 cps; he also showed that the effect of differences in damping constants of the seismometers was of second order and could be neglected. The seismometer calibration scheme in operation for U.K.A.E.A. arrays has been described by Keen et al. (1965). The natural frequency of each seismometer can be obtained from the damped oscillations of the seismometer magnets recorded regularly on the WRA tapes as part of the calibration routine. Such a calibration trace is given by the expression

$$F(t) = A_o \exp(-\lambda \omega t) \cos(\sqrt{1-\lambda^2} \omega t + \epsilon)$$  \hspace{1cm} (2-19)

where $\lambda$ is the damping constant, $\omega$ the natural angular frequency of oscillation of the seismometer magnet and $t$ denotes time; $A_o$ and $\epsilon$ are constants. This is damped harmonic motion of period

$$T = \frac{2\pi}{(\omega \sqrt{1-\lambda^2})}$$  \hspace{1cm} (2-20)

Though the oscillations of any calibration trace can be seen to die away, for R1 on 27.8.67, $\lambda$ was found to be $(1.69 \pm 0.13) \times 10^{-2}$. Therefore the error introduced in $T$ by neglecting the $\sqrt{1-\lambda^2}$ term in (2-20) is only about 1 part in 6000 so that the measured period is effectively the natural period of the seismometer magnet. The damping constants for all other free period traces
Table 2.2. Measurements of Natural Frequencies of Seismometer Magnets and Numerical Corrections for P Arrivals of Different Frequencies.

<table>
<thead>
<tr>
<th>Seismometer</th>
<th>27/10/66</th>
<th>26/12/66</th>
<th>14/8/67</th>
<th>27/8/67</th>
<th>7/10/67</th>
</tr>
</thead>
<tbody>
<tr>
<td>B1</td>
<td>1.071 ± 0.001</td>
<td>1.070 ± 0.001</td>
<td>1.078 ± 0.001</td>
<td>1.074 ± 0.001</td>
<td></td>
</tr>
<tr>
<td>B2</td>
<td>1.085 ± 0.001</td>
<td>1.087 ± 0.001</td>
<td>1.003 ± 0.002</td>
<td>1.012 ± 0.001</td>
<td></td>
</tr>
<tr>
<td>B3</td>
<td>1.032 ± 0.001</td>
<td>1.032 ± 0.001</td>
<td>1.099 ± 0.001</td>
<td>1.096 ± 0.001</td>
<td></td>
</tr>
<tr>
<td>B4</td>
<td>1.092 ± 0.001</td>
<td>1.093 ± 0.001</td>
<td>1.026 ± 0.001</td>
<td>1.027 ± 0.001</td>
<td></td>
</tr>
<tr>
<td>B5</td>
<td>1.019 ± 0.001</td>
<td>1.018 ± 0.001</td>
<td>0.960 ± 0.001</td>
<td>0.960 ± 0.001</td>
<td></td>
</tr>
<tr>
<td>B6</td>
<td>1.167 ± 0.001</td>
<td>1.027 ± 0.001</td>
<td>1.018 ± 0.001</td>
<td>1.019 ± 0.001</td>
<td></td>
</tr>
<tr>
<td>B7</td>
<td>1.059 ± 0.001</td>
<td>1.056 ± 0.001</td>
<td>0.973 ± 0.001</td>
<td>0.972 ± 0.001</td>
<td></td>
</tr>
<tr>
<td>B8</td>
<td>1.064 ± 0.001</td>
<td>0.966 ± 0.003</td>
<td>0.969 ± 0.001</td>
<td>0.968 ± 0.001</td>
<td></td>
</tr>
<tr>
<td>B9</td>
<td>1.060 ± 0.001</td>
<td>1.058 ± 0.001</td>
<td>0.985 ± 0.001</td>
<td>0.983 ± 0.001</td>
<td></td>
</tr>
<tr>
<td>R1</td>
<td>1.066 ± 0.001</td>
<td>1.058 ± 0.001</td>
<td>1.073 ± 0.001</td>
<td>1.073 ± 0.001</td>
<td></td>
</tr>
<tr>
<td>R2</td>
<td>1.057 ± 0.001</td>
<td>1.041 ± 0.001</td>
<td>0.943 ± 0.001</td>
<td>0.943 ± 0.001</td>
<td></td>
</tr>
<tr>
<td>R3</td>
<td>1.041 ± 0.001</td>
<td>1.045 ± 0.001</td>
<td>1.053 ± 0.001</td>
<td>1.049 ± 0.001</td>
<td></td>
</tr>
<tr>
<td>R4</td>
<td>1.037 ± 0.001</td>
<td>0.999 ± 0.001</td>
<td>0.997 ± 0.001</td>
<td>0.997 ± 0.001</td>
<td></td>
</tr>
<tr>
<td>R5</td>
<td>1.055 ± 0.001</td>
<td>1.054 ± 0.001</td>
<td>1.058 ± 0.001</td>
<td>1.056 ± 0.001</td>
<td></td>
</tr>
<tr>
<td>R6</td>
<td>1.144 ± 0.001</td>
<td>0.951 ± 0.001</td>
<td>0.950 ± 0.001</td>
<td>0.950 ± 0.001</td>
<td></td>
</tr>
<tr>
<td>R7</td>
<td>1.135 ± 0.001</td>
<td>1.135 ± 0.001</td>
<td>1.145 ± 0.003</td>
<td>1.142 ± 0.002</td>
<td></td>
</tr>
<tr>
<td>R8</td>
<td>1.002 ± 0.001</td>
<td>1.016 ± 0.002</td>
<td>0.986 ± 0.001</td>
<td>0.985 ± 0.002</td>
<td></td>
</tr>
<tr>
<td>R9</td>
<td>1.142 ± 0.001</td>
<td>0.976 ± 0.001</td>
<td>0.980 ± 0.001</td>
<td>0.983 ± 0.002</td>
<td></td>
</tr>
<tr>
<td>R10</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
Table 2.2 (Contd)

$\frac{dt}{dT}$ - The Rate of Change of Time Shift $t$ with Natural Period of Seismometer $T$. (Derived from Figure 24 of Muirhead (1968b)).

<table>
<thead>
<tr>
<th>Frequency of P Arrival, $\text{cps}$</th>
<th>Phase $\pi$</th>
<th>Phase $2\pi$</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.5</td>
<td>0.112</td>
<td>0.165</td>
</tr>
<tr>
<td>1.25</td>
<td>0.150</td>
<td>0.234</td>
</tr>
<tr>
<td>1.0</td>
<td>0.187</td>
<td>0.323</td>
</tr>
<tr>
<td>0.75</td>
<td>0.253</td>
<td></td>
</tr>
</tbody>
</table>

* In the Tennant Creek log sheets there is no record of when the seismometers were changed. Between 27/10/66 and 26/12/66, B6 was obviously changed. During the wet season between 26/12/66 and 27/8/67 probably ten seismometers were replaced. During the period 27/8/67 to 7/10/67 it also seems likely that B3 was changed, though fortunately the alteration in natural frequency is small.
Table 2.3. Azimuth and $dT/d\Delta$ Values Corrected for the Natural Periods of the Array Seismometers.

<table>
<thead>
<tr>
<th>Event No.</th>
<th>Date</th>
<th>Uncorrected Values</th>
<th>Corrected Values</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>Apparent Velocity</td>
<td>Variance $\sigma^2$</td>
</tr>
<tr>
<td></td>
<td></td>
<td>km/sec</td>
<td>sec/$d\Delta$</td>
</tr>
<tr>
<td></td>
<td></td>
<td>sec/deg</td>
<td>deg</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>3</td>
<td>14/9/67</td>
<td>11.92$^{+0.10}_{-0.10}$</td>
<td>9.332$^{+0.078}_{-0.078}$</td>
</tr>
<tr>
<td>20</td>
<td>17/10/66</td>
<td>12.42$^{+0.13}_{-0.13}$</td>
<td>8.950$^{+0.097}_{-0.097}$</td>
</tr>
<tr>
<td>26</td>
<td>20/10/66</td>
<td>12.50$^{+0.06}_{-0.06}$</td>
<td>8.894$^{+0.046}_{-0.046}$</td>
</tr>
<tr>
<td>81</td>
<td>2/1/67</td>
<td>12.44$^{+0.12}_{-0.12}$</td>
<td>8.938$^{+0.089}_{-0.089}$</td>
</tr>
<tr>
<td>86</td>
<td>2/10/67</td>
<td>13.50$^{+0.10}_{-0.10}$</td>
<td>8.238$^{+0.064}_{-0.064}$</td>
</tr>
<tr>
<td>120</td>
<td>8/11/67</td>
<td>14.93$^{+0.18}_{-0.18}$</td>
<td>7.448$^{+0.092}_{-0.092}$</td>
</tr>
<tr>
<td>207</td>
<td>14/9/67</td>
<td>21.76$^{+0.23}_{-0.23}$</td>
<td>5.111$^{+0.055}_{-0.055}$</td>
</tr>
</tbody>
</table>
are clearly very close to $1.7 \times 10^{-2}$. The natural periods of the seismometers were measured on four days, and the results are summarised in Table 2.2.

This disparity in the natural periods of the instruments and its possible effect on the measured onset times was realised early in 1967, but owing to the large systematic errors found in $dT/dA$ and azimuth measurements due to local structure, it was decided not to pursue the problem quantitatively until a way of correcting adequately for local structure had been found. When a complete $dT/dA$ curve had been obtained by the method described and successfully applied in Chapter 4, it was then pertinent to find out whether it was possible to reduce the scatter in the data and some of the systematic errors by correcting the relative onset times to allow for the differences in the natural periods of the seismometers. It is possible in principle to remove errors due to this cause by using the following simple technique. For a particular event the period of the first cycle of one seismometer output is measured. Then a correction to each relative onset time measured either by matching peaks (or troughs) or cross-over points is obtained from Table 2.2 prepared from the graphs of Figure 24 of Muirhead (1968b), and revised values of $dT/dA$ and azimuth are calculated from the new set of onset times. However, for the majority of events the first half second only of the P arrival has been matched, and the frequency was generally between 1.0 and 2.0 cps so that the corrections are negligible. 17 events occurring between 1.8.67 and 31.10.67 and between 1.10.66 and 31.1.67 were examined, and of these it was found that for ten of them the corrections for the natural periods of the seismometers were negligible for any of the sets of onset times; the results of applying corrections to seven events are summarised in Table 2.3. It is evident that this approach does not give a noticeable improvement in the precision of $dT/dA$ and azimuth measurements, though it is worth remarking that in each case the value of $dT/dA$ is increased by about half its standard error; a slight improvement in precision was obtained for the four events occurring in the latter half of 1967, and a slight reduction was found for the other three. As expected the variations in natural
period across either arm of the array are not systematic, so that any resulting systematic errors in $dT/d\Delta$ and azimuth would be small. It is thus concluded that the effect of the differences in the natural periods of the seismometer magnets on the onset times is, in almost all instances, too small to produce appreciable scatter or systematic errors in values of $dT/d\Delta$ and azimuth. Nevertheless it is important that the natural periods of the seismometers at WRA should be kept very close to one second. The heights of the individual seismometers are listed in Table 1.1. The difference in elevation of the highest and lowest seismometers of the array is about 130 ft (40 m.). The systematic difference in arrival time for an event at a distance of $30^\circ$ that would be caused by this height difference is $0.005$ seconds, assuming a P wave velocity of 6.0 km/sec at the surface. The errors introduced by neglecting the differences in altitudes of the seismometers are therefore negligible.

(ii) The Effect of the Earth's Ellipticity.

Another source of systematic error in $dT/d\Delta$ measurements is the ellipticity of the earth. According to Bullen (1937, 1963, pp. 179-181) the ellipticity correction to the travel time $\delta T$ is given approximately by

$$\delta T = f(\Delta) (h_0 + h_1)$$

(2-21)

where $h_0$ and $h_1$ are the differences between the actual and the mean radii of the earth at the epicentre and observing station, and are listed on p. 162 of Bullen (1937); $f(\Delta)$ is given on p. 323 of Bullen (1938a). Since $\delta(dT/d\Delta)$ is required, the midpoint of each of Bullen’s latitude intervals corresponding to a particular $h$ has been found. Then, using these latitudes, an interpolation polynomial has been fitted through the points between $5.0^\circ$ and $84.3^\circ$ by the method of divided differences; $h$ was tabulated as a function of latitude at intervals of $0.2^\circ$. A similar procedure was performed for $f(\Delta)$ which Bullen gave at intervals of $10^\circ$; $f(\Delta)$ was tabulated at intervals of one degree. $\delta T$ was calculated for a number of imaginary epicentres at intervals of $5^\circ$ from $5^\circ$ to $100^\circ$ and at azimuths of $0^\circ$, $30^\circ$, $60^\circ$, $90^\circ$, $120^\circ$, $150^\circ$ and $180^\circ$ from
Table 2.4. Event Classification Scheme.

<table>
<thead>
<tr>
<th>Class</th>
<th>Quality of WRA Record</th>
<th>No. of Stations Used in Epicentre Determination</th>
<th>No. of Seismometers Used</th>
<th>Focal Depth, h</th>
</tr>
</thead>
<tbody>
<tr>
<td>A α</td>
<td>Very Good</td>
<td>At least 36</td>
<td>At least 18</td>
<td>Shallow</td>
</tr>
<tr>
<td></td>
<td>&quot; &quot;</td>
<td>&quot; &quot;</td>
<td>&quot; &quot;</td>
<td>&gt; 130 km</td>
</tr>
<tr>
<td>B β</td>
<td>Good</td>
<td>At least 20</td>
<td>16 or more</td>
<td>Shallow</td>
</tr>
<tr>
<td></td>
<td>&quot; &quot;</td>
<td>&quot; &quot;</td>
<td>&quot; &quot;</td>
<td>&gt; 130 km</td>
</tr>
<tr>
<td>C γ</td>
<td>Moderate to Good</td>
<td>At least 12</td>
<td>12 or more</td>
<td>Shallow</td>
</tr>
<tr>
<td></td>
<td>&quot; &quot;</td>
<td>&quot; &quot;</td>
<td>&quot; &quot;</td>
<td>&gt; 130 km</td>
</tr>
<tr>
<td>D δ</td>
<td>Poor to Moderate</td>
<td>-</td>
<td>10 or more</td>
<td>Shallow</td>
</tr>
<tr>
<td></td>
<td>&quot; &quot;</td>
<td>&quot; &quot;</td>
<td>&quot; &quot;</td>
<td>&gt; 130 km</td>
</tr>
</tbody>
</table>

Each division has been split into three further categories as follows:

- C+ an event close to being placed in class B.
- C typical event of class C.
- C− an event close to being placed in class D.
WRA with the aid of computer programme ZCW101B2. Finally approximate values of $\frac{dT}{d\Delta}$ were determined by differentiating the tables of $\delta T$. The reason for interpolating Bullen's tables was to enable values of $\delta T$ to be calculated that were sufficiently precise to facilitate this numerical differentiation. The results are displayed in Table 1 of Appendix 3, and will be discussed further in Chapter 4. At this stage it suffices to comment that for an array such as WRA at a latitude of about 20° the maximum error that can be introduced into a portion of a $dT/d\Delta$ curve due to ellipticity is $\sim 0.024$ sec/deg; the effect of the earth's ellipticity is therefore quite significant.

2.5 THE SELECTION OF EVENTS FOR AZIMUTH AND $dT/d\Delta$ MEASUREMENTS, AND ERRORS IN EPICENTRE DETERMINATIONS.

In this section the criteria used to select events for azimuth and $dT/d\Delta$ measurements are examined. Events recorded at WRA were inspected on a helicorder output of a single seismometer channel, and if the P onset seemed clear and the signal to noise ratio was fairly large, the event was transferred from magnetic tape on to paper in the manner described earlier, provided that at least five seismometers in each line were working. Thus only the visual characteristics of the events recorded at the array were used in selecting data; no reference to U.S.C.G.S. magnitude determinations was made. The azimuth, distance, focal depth and the expected arrival time derived from the J-B tables, together with the number of stations used in the epicentre determination of each event, were obtained from the 'Gedess' bulletins prepared monthly from the U.S.C.G.S. Preliminary Determination of Epicentre (P.D.E.) Cards and supplied by the U.K.A.E.A. The records used have been divided into four main classes as explained in Table 2.4. Each of these classes has been subdivided into three further categories on a more qualitative basis. The class allotted to each event is listed in Table 1 of Appendix 2. This classification scheme has been devised to indicate the reliability of the data obtained from any event, and its significance and usefulness will be considered in Chapter 8. Prior to May 1966, the azimuths in the 'Gedess' bulletins were given only to the nearest degree. As
Figure 2.5 Section through earth showing a ray path from an earthquake to a recording station.
azimuths were required to the nearest tenth of a degree, it was necessary to calculate the azimuths of all events occurring before May 1966 using computer programme ZCW 101A1.

A section through the earth showing a ray path of parameter p is shown in Figure 2.5. For an earthquake occurring at depth h and epicentral distance Δ, the value of dT/dΔ measured at an array will correspond to an adjusted distance Δ₀, where Δ₀ is measured from the point where the extension of the ray path beyond the focus reaches the surface. For array measurements of dT/dΔ it is preferable to adjust all distance values to correspond to Δ₀, and this procedure has been adopted for all earthquakes used in this thesis. At distances greater than 28° events at any focal depth were used, but at shorter distances earthquakes at depths of less than 130 km were preferred. The reason for this is that at large distances the uncertainty in the distance correction for different upper mantle models is small, but becomes larger as the distance decreases. In the early stages of this work values of Δ₀ were calculated using an assumed structure of the crust and upper mantle for shallow events or using the extended distance tables of Hodgson and Storey (1953) for earthquakes occurring at depths greater than 60 km. After the publication of the 1968 Seismological Tables for P Phases, a computer programme (ZCW 101 C2) was written which enables adjusted distances to be calculated for ray paths of specified parameter travelling through any upper mantle model. Adjusted distances were subsequently calculated for the upper mantle model used in preparing the 1968 Seismological Tables for P Phases; they are listed in Table 2 of Appendix 2.

It is important to point out that all epicentres used were the preliminary U.S.C.G.S. determinations obtained using the J-B travel-time tables. Recent travel-time studies by Carder, Gordon and Jordan (1966), Cleary and Hales (1966) and Lilwall and Douglas (1970), as well as those of Herrin, Tucker, Taggart, Gordon and Lobdell (1968), have revealed systematically shorter travel times at teleseismic distances than those of Jeffreys and Bullen. These new travel-time curves do, however, show systematic
Figure 2.6 Diagram showing geometry of array, true epicentre and U.S.C.G.S. epicentre.
differences among themselves, and this problem will be examined again in
Chapter 4. The important fact is that significant systematic errors in epicentre
and focal depth determinations almost certainly exist, and very little can be
done about this until the travel time controversy has been resolved. As far as
random variations in epicentre locations for a particular region are concerned,
the epicentres themselves are in all probability accurate to 50 km or about
0.5°; this is not terribly serious at teleseismic distances, but is a problem
in upper mantle work. Superposed on these random errors will be some
sort of regional effect or bias. The effect of errors in epicentre locations on
dT/dΔ measurements is to increase the scatter of the data especially in those
parts of the dT/d Δ curve where d^2T/d Δ^2 is large. The effect on azimuth
can be calculated by applying the sine theorem of spherical trigonometry (see
Otsuka, 1966a) to ΔSEE' of Figure 2.6.

\[
\sin d \phi = \sin D \sin \theta / \sin \Delta
\]

(2-22)

\(d \phi\) is maximum when \(\theta = \pm 90^\circ\).

Hence

\[
\sin |d \phi|_{\text{max}} = \sin D / \sin \Delta
\]

(2-23)

Taking \(D = 0.5^\circ\), \(|d \phi|_{\text{max}} = 1.9^\circ\, 1.0^\circ\) and \(0.5^\circ\) at distances of 15°, 30°
and 80° respectively. Errors in focal depth determinations, particularly
for deep focus earthquakes, may well produce as much scatter in the \(dT/d \Delta\)
data as epicentre mislocations. As an example, for an event at a distance of
40.0° and a focal depth of 400 km an error in the depth determination of 50 km
will produce a corresponding error in the adjusted distance of 0.4°.
CHAPTER 3

THE STRUCTURE BENEATH THE ARRAY

3.1 INTRODUCTION.

In the early stages of this project it was important to find out how serious were the effects of structure in the vicinity of the array on the measurements of dT/d\(\Delta\) and azimuth. For events at teleseismic distances (\(\Delta > 28^0\)) the measured azimuths and dT/d\(\Delta\) values often differed significantly from the true azimuths and the values of dT/d\(\Delta\) derived from either the Jeffreys-Bullen (J-B) Tables (1958) or the new Seismological Tables for P Phases (1968). In addition the errors in the measurements of onset times, in epicentre determinations or in the travel-time tables themselves could only account for anomalies in both dT/d\(\Delta\) and azimuth an order of magnitude less than those observed. Niazi (1966) considered the effect of a dipping M-discontinuity on measurements of slowness and azimuth. Using the theory outlined in Niazi's paper, two computer programmes have been written that will work out the effect on dT/d\(\Delta\) and azimuth of any combination of dipping interfaces for a seismic event whose azimuth and distance are accurately known. From the results of the seismic experiment WRAMP, Underwood (1967) inferred the presence of a near-surface dipping structure. As more dT/d\(\Delta\) and azimuth measurements accumulated, it became evident that the structure beneath the array could not be approximated by a single plane dipping interface. Although P arrivals from some azimuths do show azimuth and dT/d\(\Delta\) anomalies consistent with a structure similar to that derived by Underwood, the overall picture is one of great complexity.

The main purpose in deriving models of the local structure is to enable corrections to be applied to dT/d\(\Delta\) values before inverting the data to obtain a velocity distribution. However, if the structure is of great complexity, this is no longer feasible, and an empirical method of correcting dT/d\(\Delta\) measurements must be used. Nevertheless, the anomalies in dT/d\(\Delta\) and
azimuth still provide valuable information on the structure beneath the array. In recent years several travel-time studies have shown that the travel-time residuals at a single station are azimuthally dependent (Ritsema, 1959, Cleary and Hales, 1966, Otsuka, 1966a, b, Bolt and Nuttli, 1966 and Herrin and Taggart, 1968). This suggests that a search for azimuthal variations of the travel-time residuals for teleseisms at a medium aperture array, such as WRA, might well permit some limitations to be placed on the depth and cause of the \( \frac{dT}{d\Delta} \) and azimuth anomalies. Since the simple approach of trying to fit one or more dipping interfaces beneath the array has not been entirely successful, an attempt to derive a 'structure surface' that will produce the observed azimuth and \( \frac{dT}{d\Delta} \) anomalies has been made. Curvature of the wave front or marked changes in wave form of the P arrivals from one seismometer to another seem to originate from specific regions below the array. Consequently a method that enables a map of these regions to be drawn has been devised and tentatively applied.

3.2 ANOMALIES IN \( \frac{dT}{d\Delta} \) AND AZIMUTH MEASUREMENTS.

There is evidence from nuclear explosion studies that the earth's mantle becomes relatively homogeneous between 700 and 800 km. In view of this, P arrivals from events at distances greater than 30° are often more suitable for deriving reliable \( \frac{dT}{d\Delta} \) and azimuth values owing to their comparatively simple wave forms. For this reason P arrivals from events between 30° and 100° were studied first. The array itself was put into operation just in time to record the nuclear explosion Longshot. When the array records of this explosion were analysed towards the end of 1966, they provided a measurement of \( \frac{dT}{d\Delta} \) which was about 11 per cent higher than that predicted by the Jeffreys-Bullen travel-time curve. At that time travel-time studies by Cleary and Hales (1966) and by Carder et al. (1966) strongly suggested that the J-B value for \( \frac{dT}{d\Delta} \) was in error by less than 1 per cent at a distance of 81.2°. It seemed likely that the large systematic error in \( \frac{dT}{d\Delta} \) was due to crustal structure beneath the array, and consequently a first approximation to this
Figure 3.1 The Fiji Islands Region earthquake of 9.10.66 showing two distinct arrivals. (i) pass band 2.0 to 4.0 cps. (Note that the first phase is almost completely damped out at B8 to B10. First detectable motion is shown by the small markers.)
Figure 3.2 The Fiji Islands Region earthquake of 9.10.66 showing two distinct arrivals. (ii) pass band 0.4 to 2.0 cps. (Note that the output of R1 is faulty.)
structure was obtained using data from Longshot and some selected earthquakes. The results of this work have been described in one of the supporting papers, Cleary et al. (1968).

It soon became evident that appreciable systematic errors in both $dT/d\Delta$ and azimuth occurred for P arrivals from events at nearly all azimuths and distances. Early in 1967, I noticed that for events from the Fiji and Tonga Islands regions, at distances between $40^\circ$ and $50^\circ$ and at azimuths between $93^\circ$ and $101^\circ$, $dT/d\Delta$ fluctuated in a completely unpredictable way by as much as 1.7 sec/deg. Further, the azimuth anomaly varied only slightly, and there were marked systematic changes of wave form of the P onset across both lines of the array; also for some events the P arrival was sharp at R1 to R4 and across the blue line, but was both strongly damped and severely distorted at seismometers R7 to R10. Soon after I had drawn attention to this remarkable effect, Dr. Muirhead of the Engineering Physics Department discovered an interesting event at an azimuth of $98.3^\circ$, an epicentral distance of $46.5^\circ$ and a focal depth of 220 km; it showed two distinct onsets that became progressively further apart across the red line but not across the blue line, and is shown in Figures 3.1 and 3.2. This event is No. 160 of Appendix 2, and event 163 and possibly 162 show these two separate phases. These two phases are only seen clearly when the records are band-pass filtered between 0.4 and 4.0 cps or preferably between 2.0 and 4.0 cps. An interpretation of this phenomenon will be presented in section 3.6.

The changes of wave form, but not the separation into two phases, were subsequently found for events from New Zealand at a distance of about $41^\circ$ and an azimuth of about $125^\circ$, from regions between azimuths of $290^\circ$ and $320^\circ$ at distances of less than $55^\circ$, and also from the New Hebrides and Loyalty Islands regions at azimuths between $88^\circ$ and $100^\circ$ and distances of about $35^\circ$. For these three azimuth ranges the irregularities are associated largely, but not entirely, with the red line of the array. Moreover, earthquakes of the Japan region, at azimuths between $0^\circ$ and $10^\circ$, show totally different azimuth anomalies from events between $10^\circ$ and $30^\circ$. Also three events from the West
Table 3.1. Azimuth and dT/dΔ Measurements, and Seismometer Residuals Relative to the Origin of the Array. (Calculated Using the Cleary and Hales Apparent Velocity and the Station to Epicentre Azimuth for 4 West Caroline Islands Earthquakes.)

<table>
<thead>
<tr>
<th>Event</th>
<th>Corrected Distance, deg</th>
<th>Azimuth, deg</th>
<th>Cleary and Hales Apparent Velocity, km/sec</th>
</tr>
</thead>
<tbody>
<tr>
<td>39</td>
<td>31.8</td>
<td>9.9</td>
<td>12.31</td>
</tr>
<tr>
<td>42</td>
<td>32.7</td>
<td>11.8</td>
<td>12.48</td>
</tr>
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</table>

<table>
<thead>
<tr>
<th>Seismometer</th>
<th>Peak A First Zero Trace</th>
<th>Peak A First Zero Trace</th>
</tr>
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<tbody>
<tr>
<td>B1</td>
<td>0.018 0.002 0.008 0.007</td>
<td>0.004 0.005</td>
</tr>
<tr>
<td>B2</td>
<td>-0.001 0.004 0.006 -</td>
<td>-</td>
</tr>
<tr>
<td>B3</td>
<td>-0.019 -0.024 -0.029 0.006</td>
<td>0.004 -0.007</td>
</tr>
<tr>
<td>B4</td>
<td>-0.027 -0.026 -0.024 -0.050</td>
<td>-0.052 -0.057</td>
</tr>
<tr>
<td>B5</td>
<td>-0.030 -0.012 -0.022 -0.051</td>
<td>-0.049 -0.040</td>
</tr>
<tr>
<td>B6</td>
<td>-0.068 -0.073 -0.083 -0.064</td>
<td>-0.076</td>
</tr>
<tr>
<td>B7</td>
<td>-0.058 -0.048 -0.053 -0.091</td>
<td>-0.079 -0.079</td>
</tr>
<tr>
<td>B8</td>
<td>-0.064 -0.056 -0.062 -0.120</td>
<td>-0.116 -0.107</td>
</tr>
<tr>
<td>B9</td>
<td>-0.052 -0.022 -0.027 -0.099</td>
<td>-0.100 -0.104</td>
</tr>
<tr>
<td>B10</td>
<td>-0.055 -0.060 -0.075 -0.109</td>
<td>-0.100 -0.101</td>
</tr>
<tr>
<td>R1</td>
<td>-0.009 -0.009 -0.012 -0.015</td>
<td>-0.025 -0.017</td>
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<tr>
<td>R2</td>
<td>0.031 0.048 0.046 0.004</td>
<td>0.026 0.022</td>
</tr>
<tr>
<td>R3</td>
<td>- - - 0.010 0.011 0.009</td>
<td>-</td>
</tr>
<tr>
<td>R4</td>
<td>0.023 0.030 0.026 0.038</td>
<td>0.038 0.033</td>
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<tr>
<td>R5</td>
<td>0.041 0.051 0.045 0.046</td>
<td>0.042 0.048</td>
</tr>
<tr>
<td>R6</td>
<td>0.028 0.046 0.035 0.057</td>
<td>0.062 0.060</td>
</tr>
<tr>
<td>R7</td>
<td>0.057 0.077 0.072 0.042</td>
<td>0.035 0.042</td>
</tr>
<tr>
<td>R8</td>
<td>0.099 0.119 0.119 0.031</td>
<td>0.034 0.038</td>
</tr>
<tr>
<td>R9</td>
<td>0.084 0.092 0.079 0.051</td>
<td>0.044 0.047</td>
</tr>
<tr>
<td>R10</td>
<td>0.074 0.097 0.084 0.060</td>
<td>0.062 0.060</td>
</tr>
</tbody>
</table>

Azimuth, deg: 6.6±0.4 5.9±0.5 6.2±0.5 9.1±0.4 9.2±0.4 9.1±0.4
Apparent Velocity, km/sec: 11.75±11.79±11.72±11.52±11.58±11.58±
0.08 0.10 0.11 0.08 0.09 0.08
### Table 3.1 (Contd)

<table>
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<th>Corrected Distance, deg</th>
<th>Event 41</th>
<th></th>
<th>Event 40</th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>32.7</td>
<td></td>
<td>32.6</td>
<td></td>
</tr>
<tr>
<td>Azimuth, deg</td>
<td>11.6</td>
<td></td>
<td>12.6</td>
<td></td>
</tr>
<tr>
<td>Cleary and Hales Apparent Velocity, km/sec</td>
<td>12.48</td>
<td></td>
<td>12.46</td>
<td></td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Residual for Each Matching Technique, sec</th>
</tr>
</thead>
<tbody>
<tr>
<td>Seismometer</td>
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<tr>
<td>-------------</td>
</tr>
<tr>
<td>B1</td>
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<td>B3</td>
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<td>B8</td>
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</tr>
<tr>
<td>R3</td>
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</tr>
<tr>
<td>R5</td>
</tr>
<tr>
<td>R6</td>
</tr>
<tr>
<td>R7</td>
</tr>
<tr>
<td>R8</td>
</tr>
<tr>
<td>R9</td>
</tr>
<tr>
<td>R10</td>
</tr>
</tbody>
</table>

| Azimuth, deg | 10.7±0.5 10.6±0.5 10.5±0.5 13.6±0.5 13.9±0.6 13.6±0.5 |
| Apparent Velocity, km/sec | 11.76± 11.93 ± 11.78 ± 11.77 ± 11.80 ± 11.81 ± |
|                | 0.11     | 0.12     | 0.11     | 0.10     | 0.12     | 0.10     |
Caroline Islands region close to 10° show similar dT/dΔ anomalies, but the azimuth anomaly changes by 5° over an azimuth range of 2.7°; relative to the origin or cross-over point of the array this corresponds to a change of seismometer residual at R10 of 0.15 seconds. This striking set of observations has been summarised in Table 3.1. Note the systematic change in residuals across the red line, particularly from R7 to R10, as the azimuth changes from 9.9° to 12.6°. It is also worth remarking that for events at azimuths between 0° and 10° and over the distance range 48° to 65°, the azimuth anomaly is changed dramatically by removing R7 to R10 without altering dT/dΔ by any significant amount. A similar effect is observed for events at distances between 30° and 83° at azimuths between 290° and 320°, but both dT/dΔ and azimuth are affected. For both of these azimuth ranges the assumption that a plane wave crosses the array is no longer a good approximation, and the data show once again that much of the trouble is connected with the red line of the array. Thus, at this stage, it was apparent that some kind of irregularity in local structure was causing complexity in P arrivals and curvature of the wave front for several distinct azimuth ranges. Some pertinent observations to illustrate these points have been displayed in Table 3.2.

3.3 PRELIMINARY WORK ON LOCAL STRUCTURE.

If a thin planar beam of seismic rays is incident on a dipping interface between a half space and an overlying surface layer, the rays are deflected out of the vertical plane of incidence of azimuth ϕ relative to the direction of strike, and arrive at the surface from a new azimuth ϕ′ and with a new apparent velocity. Niazi (1966) showed that the direction cosines of the beam reaching the surface (l, m, n) are given by

\[ l = -(mb + nc) / a \]  
\[ m = (\cos r' - n \cos \alpha) / \sin \alpha \]  
\[ n = \cos r' \cos \alpha \pm \sin \alpha \sin r' \]

where (a, b, c) are the direction cosines of the normal to the apparent plane
Table 3.2: Ranges of Azimuths and Distances for which Changes of Waveform across the Array or Curvature of the Wave Front Are Observed

<table>
<thead>
<tr>
<th>Event Numbers</th>
<th>Distance Range, deg</th>
<th>Azimuth, deg</th>
<th>Features Observed</th>
</tr>
</thead>
<tbody>
<tr>
<td>159-169</td>
<td>47.5-50.1</td>
<td>93.0-100.9</td>
<td></td>
</tr>
<tr>
<td>212-213</td>
<td>40.7-41.6</td>
<td>124.7-129.0</td>
<td>Slight change of waveform across red line.</td>
</tr>
<tr>
<td>170-189</td>
<td>27.9-52.9</td>
<td>291.1-307.6</td>
<td>Changes of waveform across both lines.</td>
</tr>
<tr>
<td>190-191</td>
<td>58.9-90.9</td>
<td>314.3-315.5</td>
<td>Curvature of waveform across red line.</td>
</tr>
<tr>
<td>192-195</td>
<td>70.1-70.2</td>
<td>297.7-298.0</td>
<td>Changes of waveform across both lines.</td>
</tr>
<tr>
<td>202-205</td>
<td>80.0-80.5</td>
<td>306.5-306.7</td>
<td>Curvature of waveform across red line.</td>
</tr>
<tr>
<td>28, 31, 34-37</td>
<td>33.6-33.7</td>
<td>88.5-100.8</td>
<td></td>
</tr>
<tr>
<td>110-126</td>
<td>51.9-59.5</td>
<td>3.5-7.8</td>
<td></td>
</tr>
</tbody>
</table>

A number of events for other azimuth and distance ranges also show evidence for late arrivals in the middle of the red line.
of incidence, $\alpha$ is the angle of dip and $\gamma'$ is the angle of refraction. Let $i$ be the angle the incident beam in the lower medium makes with the vertical, and let $v_1$ and $v_2$ be the wave velocities in the upper and lower medium respectively (see Figure 1 of Niazi, 1966). Then according to Niazi

\begin{align}
a &= (\tan \alpha \cot i - \sin \phi) / R \quad (3-4) \\
b &= \cos \phi / R \quad (3-5) \\
c &= -\tan \alpha \cos \phi / R \quad (3-6)
\end{align}

where

\begin{align}
R &= \left[ 1 + \tan^2 \alpha \cot^2 i + \tan^2 \alpha \cos^2 \phi - 2 \sin \phi \tan \alpha \cot i \right]^{\frac{1}{2}} \quad (3-7)
\end{align}

and

\begin{align}
\sin \gamma' &= v_1 \left[ 1 - (\sin i \sin \phi \sin \alpha + \cos i \cos \alpha)^2 \right]^{\frac{1}{2}} / v_2 \quad (3-8)
\end{align}

The apparent velocity $V'$ and azimuth $\phi'$ of the refracted beam can then be determined from the expressions

\begin{align}
V' &= v_1 / \sqrt{1 - n^2} \quad (3-9) \\
\phi' &= \tan^{-1} (m/1) \quad (3-10)
\end{align}

This basic theory has been used in computer programmes **ZCW101A4** and **ZCW101A8**. It is worth noting that Niazi states that only the minus sign in (3-3) leads to the correct numerical results; this is wrong: the plus sign gives the correct results. Niazi also published 18 tables giving corrections to $dT/d\Delta$ and azimuth determinations for a series of dip angles and velocity contrasts. For the work on the 'structure surface' described in section 3.5, Niazi's tables were found to be insufficiently detailed. Consequently programme **ZCW101A4** was designed to enable a much more detailed set of tables to be prepared. Programme **ZCW101A8** calculates the effect of one or more dipping interfaces on the P arrival from a specific event.

Having identified the azimuth ranges in which irregularities in local structure are serious, it was then important to find out whether it was possible to derive a crustal model, consisting of a single dipping interface, which would partly explain the anomalies in $dT/d\Delta$ and azimuth, and give first order corrections to $dT/d\Delta$ and azimuth for the remaining azimuth ranges.
An alternative approach to the problem of correcting \( dT/d\Delta \) for local structure would be to derive a mathematical function to represent each seismometer residual. For \( P \) arrivals from most events the assumption that a plane wave front crosses the array is completely justifiable. The standard deviation \( \sigma_i \) on a single onset time due to random reading errors is in the range 0.010 to 0.015 seconds. This gives an indication of the 'internal' consistency of the onset times. When, however, the seismometer residuals themselves are used (see equation 2.14), the standard deviation \( \sigma_e \) is in the range 0.010 to 0.030 seconds, thus giving an estimate of the 'external' consistency of the data; the higher values of \( \sigma_e \) frequently occur for events in the distance and azimuth ranges of Table 3.2, showing that the plane wave front model is usually adequate for other azimuth ranges. So fitting a plane dipping interface below the array can be regarded as equivalent to, and in many ways preferable to, the alternative method of determining empirically a correction for each seismometer.

Underwood, Elliston and Mathews (1968) have described a seismic experiment which involved firing an explosive charge in a mine close to the array. From an analysis of arrival times from this shot, Underwood (1967) has inferred the presence of a dipping interface between metasediments and basement below the array, which he estimates to dip 5° in a direction 200°. The \( P \) wave velocities in the upper and lower media were estimated to be 5.42 and 6.10 km/sec respectively, and the depth of the layer beneath the cross-over point of the array was estimated to be 1 km.

In working out a crustal structure consisting of one or more dipping interfaces, using a single teleseism or several teleseisems at similar azimuths, accurate knowledge of both the azimuths and the expected values of \( dT/d\Delta \) of the \( P \) waves crossing the array are required. For the majority of earthquakes the azimuth is known accurately, but the \( dT/d\Delta \) values, determined by differentiating the J-B or any other travel-time tables are not, owing to the presence of errors in the tables. Hence the procedure adopted in the early stages of this work was to select a realistic model of the crust
Figure 3.3 Anomalies in dT/dΔ and azimuth for a plane dipping interface. Azimuths are measured from the dip direction. 
\( v_1 = 5.74 \text{ km/sec}, \ v_2 = 8.20 \text{ km/sec}, \ \alpha = 7.0^\circ \).
and to adjust the dip angle and dip direction, keeping the P wave velocities constant, until the apparent velocity and azimuth calculated using the structure agreed with the measured values. Obviously an infinite number of models would fit the data, and any structure with a specific velocity contrast could be replaced by an equivalent structure having a different velocity contrast and dip angle.

Jeffreys (1962b) realised from a study of Pacific nuclear explosions that the J-B travel times for P required small modifications. At this stage it suffices to comment that more recent travel-time studies give consistently shorter travel times than J-B at distances greater than 24°, but show small systematic errors among themselves. Consequently dT/dΔ is not yet known sufficiently accurately over any distance range to enable part of a dT/dΔ curve to be used to calibrate an array.

The effect of a plane dipping interface is to introduce approximately sinusoidal variations in azimuth and dT/dΔ as a function of azimuth that are 90° out of phase. This phenomenon is illustrated in Figure 3.3 for three different dT/dΔ values. In the dip direction the dT/dΔ anomaly is least, and the azimuth anomaly changes from negative to positive. An important result is that the effect of several plane dipping interfaces on dT/dΔ and azimuth measurements is indistinguishable from that of a single interface. The effects of a structure consisting of a single plane dipping interface can be separated from errors in a given dT/dΔ curve if data from earthquakes occurring at similar distances, but at opposite azimuths, are used (Niazi, 1966). The distribution of regions of high seismicity with respect to WRA is such, however, that approximately equidistant events at opposite azimuths are rare occurrences. One such occurrence has been used to determine two possible structures. The first is a simple structure involving a single dipping interface between sediments and basement beneath the array. The second structure involves two dipping interfaces; a dipping M-discontinuity has been placed below the superficial structure derived by Underwood (1967).
Table 3.3. Structures Determined for the Warramunga Array.

<table>
<thead>
<tr>
<th>Structure</th>
<th>Dip Direction, deg</th>
<th>P Wave Velocity</th>
</tr>
</thead>
<tbody>
<tr>
<td>1 (a) *</td>
<td>One dipping interface using Sacks velocity</td>
<td>179.0</td>
</tr>
<tr>
<td>1 (b) *</td>
<td>One dipping interface using J-B velocity</td>
<td>175.0</td>
</tr>
<tr>
<td>2 (a) *</td>
<td>Two dipping interfaces using Sacks velocity</td>
<td>205.5</td>
</tr>
<tr>
<td></td>
<td></td>
<td>162.0</td>
</tr>
<tr>
<td>2 (b) *</td>
<td>Two dipping interfaces using J-B velocity</td>
<td>205.5</td>
</tr>
<tr>
<td></td>
<td></td>
<td>162.0</td>
</tr>
<tr>
<td>3 ≠</td>
<td>One dipping interface</td>
<td>235.0</td>
</tr>
<tr>
<td>4 ≠</td>
<td>Two dipping interfaces</td>
<td>205.5</td>
</tr>
<tr>
<td></td>
<td></td>
<td>243.0</td>
</tr>
<tr>
<td>5 ≠</td>
<td>One dipping interface</td>
<td>215.0</td>
</tr>
</tbody>
</table>

* Structures determined from the Novaya Zemlya nuclear explosion.

≠ Structures determined from three Aleutian Islands earthquakes, the nuclear explosion Longshot, and one South of Africa earthquake.

≠ Structure determined from seven Mariana Islands region earthquakes.
These structures were determined using three earthquakes from the Aleutian Islands and the nuclear explosion Longshot, corresponding to a distance range of 79.4° to 84.7° and an azimuth range of 20° to 30°, together with one earthquake south of Africa at a distance of 84°0° and an azimuth of 215.7° (events 143, 146, 147, 149 and 214 of Appendix 2). These structures are listed in Table 3.3. The 'reversal' technique gives only an approximate correction for the local structure since the P waves from the South of Africa event are affected by a slightly different structure from that affecting the arrivals from the Aleutian Islands.

It has already been pointed out that measurements of dT/dΔ and azimuth for a wide range of azimuths and distances have revealed that the crustal structure can change quite sharply over a small azimuth range. In particular, it has been discovered that the structure affecting P waves arriving between 340° and 10° in azimuth is quite different from that between 20° and 30°. Moreover, for some azimuths the structure appears to be different for P waves arriving from different distances. Consequently the P onset from the Novaya Zemlya nuclear explosion of October 27, 1966, has been used to determine two possible types of structure, involving a single dipping interface and two dipping interfaces respectively, which it was hoped would give reasonably good seismometer corrections in the azimuth range 335° to 355° (see Wright and Muirhead, 1969).

WRA Structure from the Novaya Zemlya Explosion. The initial P onset of the explosion was sharp, and a large quantity of energy was present in the first 3 seconds of the record. Figure 3.4 shows the traces of P on fifteen seismometers. It is not clear whether P is diffracted or direct. Recent work on P in the shadow zone has been published by Sacks (1966, 1967). Ergin (1967) has suggested that a reduction in the velocity of P is required at the base of the mantle, on evidence from PKP_2 travel times near 180° and beyond; this reduction would cause direct P to propagate up to 130° and beyond.
Figure 3.4 Traces of P on fifteen seismometers for the Novaya Zemlya nuclear explosion of October 27, 1966.
To determine a reliable structure it is essential that the apparent velocity, \( \frac{d\Delta}{dT} \), of diffracted or direct P at 106.0° is known to within 0.2 km/sec. The J-B P curve straightens at about 90°, and has very little curvature from there on; \( \frac{d\Delta}{dT} \) for \( \Delta > 100° \) is 25.3 km/sec. Sacks (1967) has, however, given preliminary travel times of diffracted P to 167° and has given \( \frac{d\Delta}{dT} \) as 24.55 ± 0.08 km/sec. This value has been taken as the best available for determining a crustal structure. From their travel-time curve for P, Cleary and Hales (1966) give an apparent velocity of 24.72 km/sec at 95°, and Carder et al. (1966) give a value of 24.44 km/sec for the distance range 94° to 100°. From the new Seismological Tables for P Phases (1968) the apparent velocity is 24.38 km/sec for P beyond 97.5°. Structures have been determined assuming that the apparent velocity given by Sacks is correct; for comparison similar structures are given assuming the J-B value is correct. The results are listed in Table 3.3, and have been obtained using an apparent velocity of 22.11 km/sec and an azimuth of 344.5°; these values have been obtained by averaging the three independent least-squares determinations for P from the nuclear explosion.

The extent to which the computed structures are applicable to other events was originally investigated using seven earthquakes occurring at similar azimuths to the Novaya Zemlya explosion, but at different distances; the results have been discussed in the supporting paper by Wright and Muirhead, (1969). Since this paper was published more data have been obtained for the azimuth range 335° to 355°; consequently the problem of correcting \( \frac{dT}{d\Delta} \) measurements within this azimuth range will be examined in sections 4.4, 6.6 and 7.2.

From the work on Aleutian Islands events and the Novaya Zemlya explosion, structures have been determined for correcting \( \frac{dT}{d\Delta} \) for the azimuth ranges 335° to 355° and 20° to 30°. The crustal model derived by Cleary et al. (1968), to give corrections to \( \frac{dT}{d\Delta} \) and azimuth for Aleutian Islands events at distances close to 80° and at azimuths between 20° and 30°,
gave corrections to \( \frac{dT}{d\Delta} \) that were too small, and azimuth corrections that were too large, when Mariana Islands earthquakes in the azimuth range 13° to 25° were used. For this reason another model involving a single dipping interface within the crust, to give approximate corrections to \( \frac{dT}{d\Delta} \), was derived in the following way: seven clear records of shallow events were selected, corresponding to distances of between 36.0° and 43.8°; the effects on apparent velocities and azimuths of a number of different structures with velocity contrasts of 0.7 were investigated. The best structure was taken to be that which gave corrected azimuths closest to the true azimuth, and apparent velocities closest to the values obtained by Cleary and Hales (1966) for this distance range. The results of Cleary and Hales were chosen because at the time this work was completed neither the 1968 Seismological Tables for P Phases nor the revised travel times of Hales, Cleary and Roberts (1968) had been published. The calculated structure differs from the model of Cleary et al. (1968) by 20° in dip direction and 0.5° in dip angle, and is given in Table 3.3 (see also Wright, 1968). Thus, in deriving this model of the crust, it has been assumed that the form of the \( \frac{dT}{d\Delta} \) curve beyond 36° was in good agreement with that of Cleary and Hales. The usefulness of the three sets of structures of Table 3.3 will be considered again in Chapter 4 in the light of more recent data.

Fairborn (1966) has shown that with certain approximations the effect of horizontal P velocity gradients on \( \frac{dT}{d\Delta} \) and azimuth determinations cannot be distinguished from the effect of plane dipping interfaces. Johnson (1967) reached a similar conclusion. It is implicitly assumed in this work that horizontal gradients in the elastic parameters, apart from those required by a dipping interface, are negligible, and the problem of anisotropy will not be considered again. However, it is stressed that anisotropy could well provide an alternative explanation of, or contribute to, the \( \frac{dT}{d\Delta} \) and azimuth anomalies. In this preliminary study it has been shown that the structure beneath the array is too complex to be approximated by a single dipping interface. The crustal models of Table 3.3 may well be useful for a
Figure 3.5 Azimuth anomalies as a function of azimuth for first arrivals from teleseisms.
Figure 3.6  Azimuth anomalies for the azimuth range 344° to 32°. (Δ > 27°).
Figure 3.7 Azimuth anomalies for the azimuth range $72^\circ$ to $120^\circ$. ($\Delta > 27^\circ$).
limited range of azimuths, but their applicability is restricted. The results do suggest a general south to southwest trend of the dipping structures. To illustrate the complexity of the structure beneath WRA, the azimuth anomalies have been plotted in Figures 3.5 to 3.7.

It is important to realise that the models of the crust presented in Table 3.3 are merely formal solutions derived to correct dT/dΔ and azimuth determinations, and do not necessarily give a realistic indication of the true nature of the crust beneath the array; these models were all derived using programme ZCW101A8. At this stage it is relevant to comment briefly on this point and to examine whether other observational evidence can help to convert a formal solution into a plausible model. Nuttli and Whitmore (1961, 1962) measured apparent angles of incidence of P waves recorded on seismograms from the Galitzin-Wilip instruments at Florissant. The 'half periods' of these waves varied from 1.5 to 3.5 seconds, and the data indicated that the velocity of the P waves at the earth's surface was approximately 8 km/sec. This suggests that the P waves did not 'see' or were not affected by the earth's crust, even though the crustal thickness below the station was known to be approximately 35 km. In this case the wavelengths of the P waves concerned were between 24 and 60 km. Nuttli and Whitmore (1962) pointed out that incorrect instrument calibration could account for their results, but this seemed less likely when similar results were obtained for other stations and from studies of the polarization angle of S waves. Now at Tennant Creek the periods of the P arrivals are generally between 0.7 and 1.0 seconds which correspond to wavelengths within the crust of about 4 to 6 km. Thus it seems probable that structure within say 2 km of the surface will not significantly affect the measured values of dT/dΔ and azimuth at WRA. It is sufficient to remark that it does seem a little unrealistic to place dipping interfaces within a few kilometres of the surface, and it is necessary to try and find a technique which will give some idea of the depth of the dipping interface, if indeed this is the cause of the trouble.
3.4 TRAVEL-TIME RESIDUALS AND THEIR INTERPRETATION.

Valuable information on structure beneath the array might be provided by the travel-time residuals for teleseisms measured at the origin of the array. First it is important to consider the definition of a station correction. If an average travel-time table is used to predict the arrival time at a particular station, a set of corrections must be added to the tabulated travel times. Thus, ignoring the source correction, the predicted arrival time at a station is given by (Herrin and Taggart, 1968).

\[ t_i = t_o + t(\Delta_i) + C_i \]  

(3-11)

where \( t_o \) is the origin time, \( t(\Delta_i) \) is the travel time given in the tables and \( C_i \) is the station correction.

It has been known for some time that station corrections vary considerably with station to source azimuth (see Ritsema, 1959, Cleary and Hales, 1966, Otsuka, 1966a, b, Bolt and Nuttli, 1966 and Herrin, 1966). This led Herrin and Taggart (1968) to assume a dependence on azimuth \( \Phi_{ij} \) of the form

\[ C_{ij} = A_i + B_i \sin (\Phi_{ij} + E_i) \]  

(3-12)

where \( C_{ij} \) is the correction to be added to the tabulated travel time for the distance from source \( j \) to station \( i \), and \( A_i \), \( B_i \) and \( E_i \) are constants to be determined by the usual least-squares procedure. The objective in this section is to derive the station correction for WRA, to examine using simple models of the crust (thus for simplicity neglecting any upper mantle effects) whether expression (3-12) for the station correction is realistic, and to ascertain whether there is a correlation between the station correction and the \( dT/d\Delta \) and azimuth anomalies.

These considerations give rise to the following idea: consider a recording station \( A \) above a crust of a certain thickness, say 30 km, with a fixed upper mantle velocity \( v_1 \) and crustal velocity \( v_2 \) for which the station correction for a specific travel-time curve is by definition zero; then consider
a station B situated on a crust of different thickness and with a dipping M-
discontinuity. Two questions arise: how does the station correction vary with
crustal thickness, and is this correction distance dependent? It seems clear
that for a plane dipping M-discontinuity, an approximately sinusoidal variation
in station correction will occur, which will increase in amplitude as the depth
to the interface is increased. Thus if anomalies in dT/dΔ and azimuth are
used to calculate a dip angle and dip direction, it might be possible to detect
a periodic variation in the travel-time residuals and hence derive an approxi­
mate depth to the dipping interface. In Appendix 4 the variation of station
correction with azimuth for a plane dipping interface has been investigated,
and some numerical results using both the basic theory and the actual travel­
time residuals for teleseisms recorded at WRA have been presented. These
travel-time residuals did reveal a small periodic variation in station correction.
However, if a plane dipping interface was responsible for the results, the dip
direction indicated by the phase angle E₁ in equations (3-12) and (A4.13) is
not correlated with the predominant structure suggested by the dT/dΔ and
azimuth anomalies. Therefore the residuals have failed to reveal any signifi­
cant azimuthal variation due to local structure.

3.5 AN ATTEMPT TO DERIVE A 'STRUCTURE SURFACE'.

Since the structure of the crust beneath the array is not simple,
the logical extension of the idea of fitting a plane dipping interface is to try
and derive a smooth surface that will explain the complex pattern of dT/dΔ
and azimuth anomalies. To facilitate this process, it is first essential to
prepare a detailed set of tables giving dT/dΔ and azimuth anomalies as a
function of both dT/dΔ and the azimuth from the dip direction, for a number
of different plane dipping structures.

A dipping interface between crust and mantle was selected with
a velocity contrast of 0.7 (i.e. v₂ = 8.20 km/sec, v₁ = 5.74 km/sec). Then
since evidence has shown that the predominant structure beneath the array
has a dip direction close to 220°, the tables were prepared as described below
using programme ZCW101A4. Taking a dip direction of 215° and a fixed dip angle, the azimuth was varied at intervals of 20° from 0° to 360°. For each azimuth the following quantities were calculated for expected apparent velocities of 11, 13, 15, 17, 19, 21, 23, 25, 59 and 100 km/sec: the azimuth anomaly, the measured values of dT/dA and apparent velocity, and the ratio measured velocity/expected velocity. Tables were prepared for dip angles of 0.5° to 8.0° in intervals of 0.5°, for dip angles of 9° to 13° in intervals of 1° and for 13°, 15°, 17°, 20° and 25°. Similar tables were prepared for dip directions of 220°, 225° and 230°, thus giving a complete set of tables of azimuth and dT/dA anomalies at azimuth intervals of 5° for each structure. Since the problem is symmetrical with respect to the dip direction, it is necessary only to take azimuths from 0° to 180° from the dip direction. However, since the sign of the azimuth anomaly changes in either side of the up dip or down dip directions, the tables were prepared for all azimuths to save time and thought when using them.

The set of tables of dT/dA and azimuth anomalies were used to work out a dip direction and dip angle for every dT/dA and azimuth determination beyond 28°. These dip vectors were then divided into 48 groups according to distance and azimuth and whether the P arrivals appeared to be affected by approximately the same structure. For each group it was assumed that the sample vectors were randomly distributed about some mean value. The problem of the statistics of a spherical distribution is commonly encountered in palaeomagnetism, and has been extensively discussed in the literature by Fisher (1953), Watson (1956) and McElhinny (1964). Suppose each direction is given unit weight. For this case Fisher (1953) has suggested that these directions will be distributed over the unit sphere with a probability density \( P \) given by

\[
P = \frac{x \exp(x \cos \psi)}{4\pi \sinh x}
\]  

(3-13)

where \( \psi \) is the angle between the direction of a sample and the direction in
in which the probability density is maximum; $\chi$ is the precision parameter.

Then the best estimate of the mean direction is given by the vector sum of the dip vectors whose direction cosines are $(l_i, m_i, n_i)$.

Let

$$R^2 = (\sum l_i)^2 + (\sum m_i)^2 + (\sum n_i)^2$$  (3-14)

Then the mean dip direction and dip angle $L$ and $A$ are given by

$$L = \tan^{-1} \left( \frac{\sum m_i}{\sum l_i} \right)$$  (3-15)

and

$$A = \sin^{-1} \left( \frac{\sum n_i}{R} \right)$$  (3-16)

The best estimate $K$ of the precision parameter $\chi$ is for $K > 3$ given by

$$K = \frac{N - 1}{N - R}$$  (3-17)

where $N$ is the number of sample vectors.

An expression for calculating the precision of the mean direction for a particular probability level can be found in Fisher (1953, p. 303) and McElhinny (1964).

Unfortunately this statistical theory is not strictly applicable to the problem of averaging the dip vectors. The reason is that the dip angle itself is known with greater precision than the dip direction, so that the probability distribution is anisotropic. Thus a correct statistical model would be one intermediate between a spherical distribution and a circular distribution. Also the data should be weighted. Using computer programme ZCW 101D3 the procedure outlined below has been followed. For each of the 48 groups of dip vectors the measured azimuth and apparent velocity have been used to define an average $dT/d\Delta$ and azimuth vector by calculating the angle of incidence of the P waves at the surface, assuming a velocity of 5.74 km/sec, and then averaging the direction cosines of the resulting $dT/d\Delta$ vectors. The resultant vector was then used to find the point on a dipping interface from which a ray reaching the origin along this resultant direction appeared to have originated; the dipping interface was arbitrarily selected having $h = 30$ km, $\lambda = 220^\circ$ and
### Table 3.4. Averaged Dip Vectors and Their Position on a Plane Dipping M-Discontinuity

(h = 30 km, \( \lambda = 220^\circ \), \( \alpha = 7^\circ \)).

<table>
<thead>
<tr>
<th>Mean Azimuth, deg</th>
<th>Parameters of Dip Vectors</th>
<th>Fisher's 'K'</th>
<th>Resultant Dip, deg</th>
<th>Resultant Azimuth, deg</th>
<th>No. of Points Used</th>
<th>No. of Different Events</th>
<th>Distance Range, deg</th>
<th>Measured Azimuth Range, deg</th>
<th>Coordinates of Projection on to M-Discontinuity, km</th>
</tr>
</thead>
<tbody>
<tr>
<td>4.1</td>
<td>3.4</td>
<td>141.5</td>
<td>47.6</td>
<td>32</td>
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* Equation (3.17) is only valid for K > 3.

‡ Only second arrivals were used.

§ PKP phase used.
Figure 3.8 Diagram showing average dip vectors projected back from the origin of the array on to a dipping M-discontinuity. (Lengths of dip vectors denote dip angle. Tails of arrows correspond to points on dipping plane.)
and $\alpha = 7^\circ$, where $h$, $\lambda$ and $\alpha$ are defined in Appendix 4. Each vector was weighted according to the reciprocal of the square of the r.m.s. error on the corresponding $dT/d\Delta$ measurement. The dip vectors were summed with the weighting factors $w_i$ normalised so that the sum of the weights was equal to the number of observations used.

$$L = \tan^{-1} \frac{\sum w_i m_i}{\sum w_i l_i}$$

(3-15a)

$$A = \sin^{-1} \left[ \frac{\sum w_i n_i}{\left\{ (\sum w_i l_i)^2 + (\sum w_i m_i)^2 + (\sum w_i n_i)^2 \right\}^{\frac{1}{2}}} \right]$$

(3-16a)

$$\sum w_i = N$$

(3-18)

The parameter $K$ of equation (3-17) is of some value in indicating how the sample vectors are clustered about the resultant, and has therefore been calculated. However, no estimate of the error in the direction of the resultant vector has been made owing to the anisotropy of the probability distribution. As far as errors are concerned, perhaps it would be preferable to treat the dip directions as a case of the von Mises distribution of vectors on a circle and consider the dip angles separately; the theory of the circular normal distribution has been described by Gumbel, Greenwood and Durand (1953), Greenwood and Durand (1955) and Vistelius (1966, pp. 54–64).

The results of these calculations on the dip vectors are listed in Table 3.4, and details of the raw data used are supplied in Appendix 5. The dip vectors on an assumed dipping M-discontinuity have been plotted in Figure 3.8 to see if some definite 'structural surface' can be derived; the arrows point in the averaged dip direction, and the length of each arrow is proportional to the dip angle. The diagram shows that it would be futile to try and work out numerically a detailed structure surface. There is, however, a slight predominance of southwesterly trending vectors indicating an average dip of between $6^\circ$ and $10^\circ$. To the northwest of the array there is a tendency to have large dip
Table 3.5. $dT/d\Delta$ and Azimuth Measurements, and the Separation of the Two Phases for the Fiji Islands Event of 9 October, 1966.

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<th>Measured Azimuth, deg.</th>
<th>Measured $dT/d\Delta$, sec/deg</th>
<th>Apparent Velocity, km/sec</th>
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<th>Depth at which Second Phase is Produced, km</th>
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<th>Difference in Arrival Times of Phases 1 and 2, sec</th>
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<td>18.2</td>
<td>R8</td>
<td>0.83</td>
<td>34.1</td>
</tr>
<tr>
<td>B9</td>
<td>0.45</td>
<td>18.5</td>
<td>R9</td>
<td>0.83</td>
<td>34.3</td>
</tr>
<tr>
<td>B10</td>
<td>-</td>
<td>-</td>
<td>R10</td>
<td>0.94</td>
<td>38.6</td>
</tr>
</tbody>
</table>
angles, and dip directions with a southeasterly trend. A line of demarcation has been sketched across the northwestern portion of the diagram. The dip vectors at azimuths between 290° and 330° support some kind of basin structure northwest of the array. The abrupt change in structure close to 10° in azimuth shows up clearly. The southeasterly trend in the dip vectors is also present to the southeast of the origin of the array. The structure indicated by events to the east of the array is known to be highly complex and needs no further comment, except to point out that there is a group of four vectors dipping northwest between azimuths of 75° and 95°. As a final comment it is sufficient to remark that Figure 3.8 illustrates very well the extreme complexity of the structure in the vicinity of the array and the great difficulty of trying to find a crustal model to account for the anomalies in $dT/d\Delta$ and azimuth.

3.6 THE REGIONS OF SHARPLY CHANGING STRUCTURE AND SOME POSSIBLE EXPLANATIONS.

The Fiji Islands event of 9 October 1966 had two distinct arrivals at WRA separated by less than one second, and is shown in Figures 3.1 and 3.2. The $dT/d\Delta$ and azimuth measurements are listed in Table 3.5 together with the measured separation of the two phases at all seismometers; note the curious break between R5 and R6 — a feature that seems to be closely connected with the curvature of the wave front across the red line of the array for P arrivals from many azimuths. Possible causes of these two phases will now be examined. Because of its high apparent velocity it is unlikely that the second phase is due to structure in the Fiji Islands region or near the deepest point of the ray path. Thus it is most probable that near-surface structure in the vicinity of the array is responsible.

To start with suppose that both phases are produced by refraction at a plane dipping interface. This gives rise to a simple idea. Since both phases can be used to give precise $dT/d\Delta$ and azimuth values across most of the array (the break between R5 and R6 does not show up so well when
Figure 3.9  (a) An illustration of the method of projecting the array back on to two intersecting dipping planes which might produce two P phases.
(b) An illustration of the postulated splitting of the wave front to enable an approximate depth of its cause to be calculated.
the peaks of the second phase are matched), it is reasonable to suppose that if the P velocities above each plane are the same, the array itself can be 'projected' back on to each plane. If it is postulated that these two projections must not overlap, then a minimum depth of the point of intersection can be calculated. This is illustrated in Figure 3.9 for the simple two dimensional case, where AB represents the red line of the array. The first and second phases arrive at the surface at angles $\theta_1$ and $\theta_2$. On this simple geometrical argument the minimum depth of the intersection of the two planes is given by $d$ in Figure 3.9a.

AB = 20 km. Hence

$$d \geq \frac{20}{(\tan \theta_1 - \tan \theta_2)} \quad (3-19)$$

Using a crustal velocity of 6 km/sec and the apparent velocities of the two phases of Figure 3.1, $d \geq 106$ km, which is not realistic. If the velocity used is increased to 8 km/sec $d$ is decreased slightly. The difference in travel time between the paths DB and CB will give the minimum travel-time difference between the arrivals of the two phases. This difference is given by

$$T \geq \frac{d \sec \theta_2 \left[ 1 - \cos (\theta_1 - \theta_2) \right]}{v_1} \quad (3-20)$$

If $v_1 = 8.0$ km/sec, $T \geq 1.75$ seconds at R10. In actual fact $T = 0.94$ seconds. This argument shows that the two phases cannot possibly be due merely to refraction at different parts of the same discontinuity.

Alternatively the difference in arrival times of the phases can be used to calculate an approximate depth $d$ to the structure causing the anomalies. The basic assumption is that the first phase is refracted at a dipping interface and that the second results from the splitting of the wave front due to diffraction or scattering as shown in Figure 3.9. On this simple assumption and with $v_1 = 6.0$ km/sec

$$d = 41.4 T \quad (3-21)$$

where $T$ is once again the difference in arrival times of the two phases. Since
no adequate diffraction theory has yet been devised it is not possible to say whether the concentration of energy in a certain direction is very likely. It seems plausible that two structures less than a wavelength apart are responsible for the two phases and for the other phenomena listed in Table 3.2. Thus the high apparent velocity second phase would show up more clearly for events with considerable energy at high frequencies. d has been calculated for each seismometer of the array and the results are displayed in Table 3.5. The significant fact is that for the whole of the blue line and half of the red line d ≈ 20 km; but for R6 to R10 d > 30 km. Hence it is tentatively suggested that fairly deep in the crust there is a source of diffracted arrivals or a cause of multiple arrivals, and this irregularity in structure is responsible for all of the curious effects noted in Table 3.2. The change in the value of T between R5 and R6 make it reasonable to speculate, as an alternative to having two closely spaced structures, that faulting deep in the crust may be largely responsible for the irregular pattern of dT/dA and azimuth anomalies.

Another possibility is that the second phase is SV. Reflection and transmission coefficients for a plane longitudinal wave incident on a plane boundary between two media have been published by Koefoed (1962) and McCamy, Meyer and Smith (1962). A small P wave followed by a large S wave can only be produced when the P wave is incident in the medium of lower P wave velocity, and further, the velocity ratio across the interface has to be small to enable significant refracted S energy to be produced for any angle of incidence (see Figure 10 of McCamy et al., 1962). In the case when P is incident in the higher velocity medium, S amplitudes comparable with P amplitudes can only be produced when the velocity ratio is high ( > 2) and for relatively large angles of incidence, i.e. 40° - 75° (see Figure 22 of McCamy et al., 1962). It is therefore unlikely that the second phase is SV. It is worth remarking that the errors in the diagrams in the paper by McCamy et al. due to an incorrect sign convention, which have been pointed out by Singh, Ben-Menahem and Shimsoni (1970), do not affect the results for incident P waves.
Another possible explanation of the two phases is Bragg reflection. This phenomenon is believed to play a significant role in reflection seismology (Grant and West, 1965, pp. 88-90), but is less likely to be an important effect in the first second or so after the P arrivals from teleseisms owing to the longer wavelengths involved.

Refraction, Reflection, Diffraction or Scattering? An Idea.

Since the evidence now available strongly suggests that the major cause of the anomalies in dT/dΔ and azimuth is structure near the base of the crust, it is possible to some extent to determine the coordinates of the region in which the anomalies originate. The idea is as follows: consider a dipping M-discontinuity of perpendicular distance h from the origin of the array. Take the mean dip as 8.0° in a direction 220°. The whole array can then be 'projected' back on to the dipping plane by tracing back the ray paths that correspond to the measured azimuth and apparent velocity. Suppose the events are divided into four groups as follows:

(a) curious changes of wave form across the array or two distinct arrivals: i.e. strong splitting of the wave front.

(b) no changes of wave form but evidence for different structures affecting arrivals at R1 to R5 and R6 to R10: i.e. marked curvature of the wave front.

(c) clear wave form and good least-squares fit to onset times assuming a plane wave front - dT/dΔ and azimuth anomalies indicate refraction by the predominating structure.

(d) as for (c) except that the dT/dΔ and azimuth anomalies indicate refraction by a structure different from the predominating one.

For each of these groups 'zones' on the dipping interface can be defined which act either as poor or efficient generators of diffracted waves or multiple arrivals. But in order to define these zones it is essential to start with some mathematical theory.
Figure 3.10 The geometry of the points on a plane dipping interface from which a beam of seismic rays crossing an array originates.
The Points on a Plane Dipping Interface from which a Beam of Seismic Rays Crossing an Array Originates.

Referring to Figure 3.10, consider an array, origin O, and let the x, y and z axes be oriented east, north and vertically downwards respectively. Suppose a parallel beam of seismic rays emanates from a plane dipping interface; whether this beam results from refraction, diffraction or multiple reflections is of no concern at this stage. A ray travelling from Q reaches the origin O, and another from Q' arrives at the surface at seismometer B. The problem is as follows: given the dip direction and dip angle of the plane, and its perpendicular distance from O, what are the coordinates, both cartesian and spherical polar, of Q and Q', if the angle of incidence and the azimuths of the waves reaching O and B are known?

\[
\overrightarrow{OB} = (a \sin \beta, a \cos \beta, 0)
\]

\[
\overrightarrow{R} = (R \sin \theta \sin \phi, R \sin \theta \cos \phi, R \cos \theta)
\]

\[
\overrightarrow{N} = (-h \sin \alpha \sin \lambda, -h \sin \alpha \cos \lambda, h \cos \alpha) - \text{perpendicular from O on to plane}
\]

\[
\overrightarrow{R'} = (r \sin \theta_1 \sin \phi_1, r \sin \theta_1 \cos \phi_1, r \cos \theta_1)
\]

\[
\overrightarrow{PQ} = \overrightarrow{R} - \overrightarrow{N}.
\]

But \(\overrightarrow{PQ} \cdot \overrightarrow{n} = 0\), where \(\overrightarrow{n}\) is a unit vector in the direction of \(\overrightarrow{N}\). Using this scalar product condition it is found that

\[
R = h / [\cos \theta \cos \alpha - \sin \theta \sin \alpha \cos (\phi - \lambda)] \quad (3-22)
\]

Let \(|\overrightarrow{x}| = x\), and its direction cosines are \((\sin \theta \sin \phi, \sin \theta \cos \phi, \cos \theta)\).

Since \(\overrightarrow{R'} = \overrightarrow{OB} + \overrightarrow{BQ'}\)

\[
a \sin \beta + x \sin \theta \sin \phi = r \sin \theta_1 \sin \phi_1 \quad (3-23)
\]

\[
a \cos \beta + x \sin \theta \cos \phi = r \sin \theta_1 \cos \phi_1 \quad (3-24)
\]

\[
x \cos \theta = r \cos \theta_1 \quad (3-25)
\]
There are three equations but four unknowns, $x$, $r$, $\theta_1$ and $\phi_1$. The aim is to derive expressions for $\theta_1$ and $\phi_1$.

Now \( \overrightarrow{PQ'} = \overrightarrow{R'} - \overrightarrow{N} \) and \( \overrightarrow{PQ'} \cdot \mathbf{n} = 0 \):

\[
\begin{align*}
\text{hence } r &= \frac{h}{\cos \theta_1 \cos \alpha - \sin \theta_1 \sin \alpha \cos (\phi_1 - \lambda)} \quad \text{(3-26)}
\end{align*}
\]

This is the fourth equation. $x$, $r$ and $a$ could be eliminated from (3-23 - 3.25) algebraically to give $\theta_1$ in terms of $\phi_1$ or vice versa. However, this process is simplified by noting that $\overrightarrow{R}$, $\overrightarrow{R'}$ and $\overrightarrow{OB}$ are coplanar so that $(\overrightarrow{R} \times \overrightarrow{OB}) \cdot \overrightarrow{R'} = 0$. Using this condition it is found that

\[
\tan \theta_1 \sin (\phi_1 - \beta) = \tan \theta \sin (\phi - \beta) \quad \text{(3-27)}
\]

However, this simple expression cannot be used in practice as it will only work when $\phi \neq \beta$, and trouble with numerical analysis arises when $\phi - \beta$ is small. So separate expressions for $\theta_1$ and $\phi_1$ have to be derived.

If $x$ is eliminated from (3-23 - 3-25) and $a/h$ is replaced by $Z$, using (3-26),

\[
\begin{align*}
\tan \theta \sin \phi &= \tan \theta_1 \sin \phi_1 - Z \sin \beta \left[ \cos \alpha - \tan \theta_1 \sin \alpha \cos (\phi_1 - \lambda) \right] \quad \text{(3-28)} \\
\tan \theta \cos \phi &= \tan \theta_1 \cos \phi_1 - Z \cos \beta \left[ \cos \alpha - \tan \theta_1 \sin \alpha \cos (\phi_1 - \lambda) \right] \quad \text{(3-29)}
\end{align*}
\]

Substituting for $\tan \theta_1$ using (3-27) and solving for $\phi_1$ the following result is obtained:

\[
\begin{align*}
\tan \phi_1 &= \frac{\sin \phi + Z \left[ \cos \alpha \sin \beta \cot \theta - \sin \alpha \cos \lambda \sin (\beta - \phi) \right]}{\cos \phi + Z \left[ \cos \alpha \cos \beta \cot \theta + \sin \alpha \sin \lambda \sin (\beta - \phi) \right]} \quad \text{(3-30)}
\end{align*}
\]

Suppose $\beta = \phi$. In this case $\phi = \phi_1$.

Then (3-23) and (3-24) become

\[
\begin{align*}
a + x \sin \theta &= r \sin \theta_1 \quad \text{(3-31)}
\end{align*}
\]

Using (3-31), (3-25) and (3-26) with $\phi = \phi_1$, it is found that

\[
\begin{align*}
\tan \theta_1 &= \frac{\tan \theta + Z \cos \alpha}{1 + Z \sin \alpha \cos (\phi - \lambda)} \quad \text{(3-32)}
\end{align*}
\]
Using (3-28) and (3-29)

\[
(tan \theta \sin \phi + Z \cos \alpha \sin \beta) \cot \theta_1 = \sin \phi_1 \left(1 + Z \sin \alpha \sin \beta \sin \lambda\right) + \cos \phi_1 \left(Z \sin \alpha \sin \beta \cos \lambda\right) 
\]  

(3-33)

\[
(tan \theta \cos \phi + Z \cos \alpha \cos \beta) \cot \theta_1 = \sin \phi_1 \left(Z \sin \alpha \cos \beta \sin \lambda\right) + \cos \phi_1 \left(1 + Z \sin \alpha \cos \beta \cos \lambda\right) 
\]  

(3-34)

To simplify subsequent algebra (3-33) and (3-34) can be written in the form

\[
A \sin \phi_1 + B \cos \phi_1 = X \]  

(3-35)

\[
C \sin \phi_1 + D \cos \phi_1 = Y \]  

(3-36)

Eliminating \( \phi_1 \)

\[
(BC - AD)^2 = (XC - YA)^2 + (YB - XD)^2 
\]  

(3-37)

Substituting in (3-37) where

\[
A = 1 + Z \sin \alpha \sin \beta \sin \lambda \]  

(3-38)

\[
B = Z \sin \alpha \sin \beta \sin \lambda \]  

(3-39)

\[
C = Z \sin \alpha \cos \beta \sin \lambda \]  

(3-40)

\[
D = 1 + Z \sin \alpha \cos \beta \cos \lambda \]  

(3-41)

\[
X = \cot \theta_1 \left(tan \theta \sin \phi + Z \cos \alpha \sin \beta\right) 
\]  

(3-42)

\[
Y = \cot \theta_1 \left(tan \theta \cos \phi + Z \cos \alpha \cos \beta\right) 
\]  

(3-43)

it is found that

\[
[\tan^2 \theta \left(1 + 2Z \sin \alpha \sin (\phi - \beta) \left(\cos \lambda - \sin \lambda\right) + Z^2 \sin^2 \alpha \sin^2 (\phi - \beta)\right) + Z^2 \sin^2 \alpha \sin^2 (\phi - \beta) \sin (\beta - \lambda) + Z^2 \cos^2 \alpha]^{\frac{1}{2}} 
\]

\[
\tan \theta_1 = \frac{1 + Z \sin \alpha \cos (\beta - \lambda)}{1 + Z \sin \alpha \cos (\beta - \lambda)} 
\]  

(3-44)

When \( \beta = \phi \) this expression for \( \tan \theta_1 \) reduces to (3-32) as expected. Programme ZCW101D2 enables \( \theta_1 \) and \( \phi_1 \) and hence \( r \) and the cartesian coordinates of \( Q' \) to be calculated if \( B \) is any seismometer of the array, using the apparent velocity and azimuth of the ray \( \overrightarrow{X} \).
Figure 3.11 Plan of the deep crust beneath WRA divided into zones according to the nature of the P arrivals and the dT/dΔ and azimuth anomalies. ($\lambda = 220^\circ$, $\alpha = 8^\circ$, $h = 30$ km). For explanation of (a), (b), (c) and (d) see opposite page.
Using this mathematical theory the region near the base of the crust has been divided into the four zones (a) - (d) listed earlier. These zones have been plotted out in Figure 3.11 for h = 30 km, and give some idea of the geographic extent of the sources of the distorted P arrivals. This figure could have been plotted simply by projecting the origin of the array on to the plane and then graphically drawing the two projected arms parallel to the two arms of the array; the reason is that the length and orientation of the arms projected on to an M-discontinuity dipping at $\theta^0$ does not differ significantly from the length and orientation of the real arms of the array. The exact mathematical theory was derived because it is envisaged that it could be used as the basis of a method of projected planes for correcting $dT/d\Delta$ and azimuth measurements and for elucidating the local structure. This method involves projecting a $dT/d\Delta$ vector from any seismometer back on to any inclined planar discontinuity within the crust, the inclination of such a plane being determined for a specific region of the crust by the $dT/d\Delta$ and azimuth anomalies themselves. Deep reflection studies would then be able to establish or deny the reality of these discontinuities. Thus the method would aim to correlate reflecting and refracting or diffracting horizons.

In preparing Figure 3.11, R1, R10, B1 and B10 were projected on to the plane for one or two measured $dT/d\Delta$ and azimuth values for each of the 48 groups used in determining the dip vectors of Figure 3.10. The preliminary test of this technique was only moderately successful, though the southeast trend of the region producing the anomalies may be of outstanding significance. This region is parallel to the Quartz Hill and Rocky Range Fault Zone to the north of the array (Underwood, 1967, Figure 11a), and once again it is tempting to speculate that faulting may be part of the cause of the observed anomalies. This approach, which is essentially a simplified version of the more general method of projected planes will eventually be applied for $h = 10$ km and $h = 20$ km to see if an improvement can be obtained.
3.7 DISCUSSION.

The observational evidence presented in this chapter does strongly suggest that the major causes of the anomalies in $dT/d\Delta$ and azimuth are irregularities in structure at or near the base of the crust. Deeper structure may indeed be partly responsible, but structure at depths of less than a few kilometres is unlikely to be of any importance owing to the phenomenon first observed by Nuttli and Whitmore for structure less than one wavelength below the surface. The next important step is to find out whether the interpretation of the data is reasonable, by examining some other recent investigations of crustal structure in different parts of the world. The evidence in favour of structural complexity deep in the crust is corroborated by Mack (1969) who investigated short-period P wave signal variations at the LASA in Montana. He showed that the P wave signal recorded at a particular subarray consists of a series of closely spaced signals; further, the interference of these multiple arrivals with one another causes the recorded signal variations between subarrays. He assumed that the signal recorded at each subarray consists of a series of primary signals $s(t)$, arriving at slightly different times, which could be decomposed into two or three primary signals commencing within the first second of the recorded signal and having amplitudes of the same order of magnitude as the first arrival. Recent deep reflection studies have suggested considerable complexity in structure near the base of the crust (Dohr and Fuchs, 1967, Junger, 1951 and Clowes, Kanasewich and Cumming, 1968). The detailed reflection survey in Alberta of Clowes et al. showed a seismic cross section with a structural relief of 8 km over a distance of 25 km with individual dips as great as $20^\circ$; this causes a significant, but not great, decrease in Bouguer gravity (see Figure 9 of Clowes et al). Evidence for a dipping M-discontinuity beneath array sites has been published by Niazi (1968), Otsuka (1966a, b) and Kanestråm (1969). Thus at present the explanation of the anomalies in $dT/d\Delta$ and azimuth, in spite of its obvious limitations, is at least in broad general agreement with results obtained for other parts of the
world.

No adequate gravity data for the array site have yet been obtained, and one of the most important tasks to be performed in the near future is a detailed gravity survey. On the seismological side an extensive refraction and reflection survey would certainly provide valuable $P_n$ velocities, a check on the average dip of the $M$-discontinuity and a determination of the depths to the reflecting layers. Weichert and Whitham (1969) have described a crustal seismic experiment used for calibration of the Yellowknife array. They found that the Mohorovicic discontinuity is horizontal under the array within the resolution of the experiment, and that the crust in the immediate vicinity of the array is very uniform. At distances of a few tens of kilometres, however, the crust was found to be more complex, but the effect of these inhomogeneities on teleseismic signals was believed to be negligible. This work gives support for the view that the structure beneath WRA is indeed complex in comparison with that below YKA.
CHAPTER 4

THE CONSTRUCTION OF A $dT/d\Delta$ CURVE AND A VELOCITY MODEL

4.1 INTRODUCTION.

In Chapter 3 it was demonstrated that the complex pattern of $dT/d\Delta$ and azimuth anomalies cannot be explained by any relatively simple crustal model. However, to construct a single $dT/d\Delta$ curve for the lower mantle from the raw $dT/d\Delta$ data, it is necessary to be able to correct for irregularities in structure in the vicinity of the array. Owing to the complexity of this local structure, it seems reasonable to try and accomplish this by using travel-time data. First a purely empirical approach to this problem has been adopted. Secondly, with the addition of extra data, a more elaborate semi-empirical technique has been devised that takes account of the variations in local structure and includes ellipticity corrections. Unfortunately the fine details present in the $dT/d\Delta$ data, even for a narrow azimuth range, may not be due to velocity anomalies at great depths; irregularities in both the structure of the crust and the upper mantle in the vicinity of an array, and possibly even lateral variations in mantle structure near an earthquake focus, can introduce complexity into a $dT/d\Delta$ curve. (Velocity anomalies are defined as regions in which the velocities change abruptly or in which the velocity gradients differ significantly from those expected for a chemically homogeneous earth devoid of phase changes.) In correcting for systematic errors in $dT/d\Delta$ these fine details are preserved as far as possible, and their significance is discussed in Chapters 6 to 8 together with some possible velocity models. The principal aim in this chapter, therefore, is to present the mathematical techniques which enable the $dT/d\Delta$ data to elucidate the structure of the earth's mantle.

After correcting the $dT/d\Delta$ measurements for local structure, the next step is to reduce the random errors as much as possible without losing useful information. This problem has generally been solved by fitting a
polynomial or a series of polynomials through the data by the method of least squares. Arnold (1968), however, drew attention to the Method of Summary Values invented by Jeffreys, and suggested that it has advantages over a polynomial regression; the two versions of this technique have been used in the smoothing of the dT/dΔ data. Finally the mathematical methods used in studying seismic ray theory and in inverting the dT/dΔ data to obtain a velocity distribution are discussed.

4.2 METHODS OF SMOOTHING OBSERVED DATA.

Consider a series of measured values of dT/dΔ through which a smooth curve is to be fitted. Estimates of the r.m.s. errors in dT/dΔ are available, but owing to epicentre mislocations and errors in focal depth determinations there are also errors in the values of the corrected distance Δ. Techniques have been devised for fitting a straight line through a set of observational points when there are errors in both the x and y coordinates (see for example York, 1966), but no similar techniques exist for fitting a smooth curve. Further, it is difficult to obtain reliable estimates of the errors in Δ; the data supplied by the U.S.C.G.S. on their preliminary epicentre determinations enable estimates of the precision of each epicentre to be made, but systematic errors will be caused by errors in the focal depth determinations, by errors in the J-B tables and by bias produced in the source region of the earthquake itself. Consequently the techniques available that assume errors in one coordinate only must be used, and any errors in the corrected distances are neglected.

The dT/dΔ data, even when corrected as far as possible for local structure, contain random errors and also systematic errors due to irregularities in structure near the array site or in the source region; in smoothing, the objective is to reduce the random part as much as possible and to estimate the amount by which the random error is reduced. The relative merits of three important smoothing techniques have been discussed briefly by
Arnold (1968). In the first of these one twelfth of the fourth difference is subtracted from, or one quarter of the second difference is added to, each observed value except the two at each end (Jeffreys, 1934, 1936). The second method is to fit a simple function, such as a polynomial or a series of polynomials, to the data by least squares. The third technique, which does appear to be preferable to the other two, is the Method of Summary Values devised by Jeffreys (1937, 1961, pp. 223-227); there are two versions of this technique, and the mathematical details of the more elaborate of these is given only in Jeffreys (1937). Since the two versions of this method have been used in all of the smoothing of the dT/dΔ data and are not widely known, they are outlined tersely below.

Consider a number of observed values of the function p (Δ) for a certain range of the argument Δ in which there is little curvature. If a straight line is fitted to these points by least squares, any two points (p₁, Δ₁) and (p₂, Δ₂) on this line can be taken to represent the data, but in general the errors will not be independent. However, Jeffreys showed that a straight line can be fitted through the data in such a way that the errors at two points (p₁, Δ₁) and (p₂, Δ₂) are independent of each other and of a possible quadratic term; p₁ and p₂ are called summary values and represent p (Δ) in the range of the argument given. Smoothing is therefore performed in the following way. The data are divided into several ranges of the argument such that the curvature in each range is small; the two summary points for each range are calculated and the whole table is interpolated to any required tabular interval by the method of divided differences (see Jeffreys and Jeffreys, 1956, pp. 261-265). This first technique will subsequently be referred to as the Method of Summary Values A. To determine whether adequate smoothing has taken place, a χ² test is performed on the residuals. If χ² is too large, important features of the data are being smoothed out, and more ranges should be used; on the other hand, if χ² is too small, the amount of smoothing is insufficient, and fewer ranges are required. If v is the number of degrees of freedom, χ² should lie in the interval
$v \pm \sqrt{2}v$ if the error is wholly random.

The summary values principle can also be used to fit a quadratic so that the errors at three points are independent of one another and of a possible cubic term (version B). This means that if the data are capable of estimating a curvature but possibly not a cubic term, three values of the argument can be chosen such that the errors at these points will be independent of each other and of the cubic term. Thus, in a table of observed values, such as a series of measurements of $dT/d\Delta$, covering a long interval in which the structure of the curve may be complicated, the values can be divided into a number of ranges, and each range of data can be specified adequately by two or three isolated values called summary points; then interpolation will give satisfactory smoothed values. Computer programmes ZCW101C1 and ZCW101C4 enable observed data to be smoothed by versions A and B respectively of the Method of Summary Values. The main disadvantage of this method in comparison with a conventional polynomial regression is that the smoothing has to be applied by trial and error until an acceptable solution is found.

Arnold (1968) has applied version A to the raw travel times of Herrin et al. (1968) that form the basis of the 1968 Seismological Tables for P Phases. Early in 1969 I decided to check computer programme ZCW101C1 using the raw data and the smoothed travel-time curve supplied in Arnold's paper. The process of checking this programme led to the discovery of some numerical errors in Arnold's computations which were also found independently by Dr. D.J. Corbishley of the U.K.A.E.A. Since the errors are serious it was considered worthwhile to send a letter to the Editor of the Bulletin of the Seismological Society of America drawing attention to these mistakes. The main points of our letter (Wright and Corbishley, 1970) are presented and discussed in Appendix 6.
4.3 A $dT/\Delta$ CURVE FOR THE LOWER MANTLE.

(i) Preliminary Method.

Owing to the complexity of the crust beneath the array, it is evident that an empirical method of correcting the measured values of $dT/\Delta$ must be devised. At WRA the azimuth and $dT/\Delta$ anomalies are for most azimuths slowly changing functions of azimuth and of the angle of incidence of the P waves at the array site. Recent travel-time studies (Cleary and Hales, 1966, Carder et al., 1966, Herrin et al., 1968, Lilwall and Douglas, 1970) have shown marked deviations from the J-B times without enabling distinct anomalous regions of the lower mantle to be identified. Cleary and Muirhead (1969) and Muirhead and Cleary (1969) have shown that the major discrepancies between these recent studies are systematic and are approximately straight lines; further, travel times from four nuclear explosions suggest that the travel times of Cleary and Hales, modified in a later paper by Hales et al. (1968), are basically correct. This means that at distances greater than $25^\circ$ these recent travel-time studies differ by a constant $dT/\Delta$. Thus the shapes of $dT/\Delta$ curves derived from any of these studies are essentially the same. This suggests that a method combining travel-time data and array $dT/\Delta$ measurements might be the answer to the difficulties in interpreting array data. Since the events used in obtaining the $dT/\Delta$ data are frequently fairly small, travel-time data from these events themselves would not be particularly useful. Now since none of these recent travel-time investigations has shown any distinct anomalous regions in the lower mantle, it seems reasonable to suppose that any irregularities in lower mantle velocities will cause perturbations of one of these modern travel-time curves no larger than the random errors in the travel times themselves. Therefore, by constraining the $dT/\Delta$ data to an acceptable travel-time curve, the effect on the P wave arrivals of the local structure can be largely eliminated.

The method used is as follows. 535 values of $dT/\Delta$ (listed in Table 2 of Appendix 2) from 192 events over the distance range $28^\circ$ to $99^\circ$ have
<table>
<thead>
<tr>
<th>Group</th>
<th>Event Numbers</th>
<th>Number of Points</th>
<th>Distance (a) and Azimuth (b) Range, (deg)</th>
<th>Points Rejected</th>
<th>Total $\chi^2$</th>
<th>Number of Degrees of Freedom</th>
<th>Correction, (sec/deg)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1A</td>
<td>1-22, 24-28</td>
<td>72</td>
<td>(a) 27.9 - 34.9</td>
<td>9A, 12B, 18B</td>
<td>66.49</td>
<td>66</td>
<td>-0.185 ± 0.038</td>
</tr>
<tr>
<td></td>
<td>31, 34, 35</td>
<td></td>
<td>(b) 73.0 - 94.6</td>
<td>22C, 28A</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1B</td>
<td>29-30, 32-34A, C, 36-37</td>
<td>19</td>
<td>(a) 34.1 - 35.7</td>
<td>-</td>
<td>18.35</td>
<td>17</td>
<td>0.152 ± 0.090</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>(b) 93.8 - 100.8</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>2A</td>
<td>39-64</td>
<td>79</td>
<td>(a) 31.8 - 37.5</td>
<td>44B</td>
<td>82.09</td>
<td>71</td>
<td>-0.935 ± 0.035</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>(b) 9.9 - 25.6</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>2B</td>
<td>65-80</td>
<td>36</td>
<td>(a) 39.2 - 46.4</td>
<td>65B, 66A, 72B, C, 74A, B, 78A</td>
<td>40.50*</td>
<td>32</td>
<td>-0.829 ± 0.035</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>(b) 10.5 - 19.7</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>3A</td>
<td>81-92</td>
<td>35</td>
<td>(a) 41.7 - 53.3</td>
<td>-</td>
<td>28.26</td>
<td>29</td>
<td>-0.672 ± 0.035</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>(b) 342.9 - 357.8</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>3B</td>
<td>93-95</td>
<td>7</td>
<td>(a) 73.5 - 77.7</td>
<td>-</td>
<td>5.97</td>
<td>5</td>
<td>-0.710 ± 0.043</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>(b) 338.5 - 352.4</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>3C</td>
<td>97-104</td>
<td>18</td>
<td>(a) 30.1 - 39.1</td>
<td>99A, B, C</td>
<td>12.91</td>
<td>14</td>
<td>-0.643 ± 0.045</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>(b) 333.0 - 344.5</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>4A</td>
<td>106-125</td>
<td>55</td>
<td>(a) 47.9 - 59.5</td>
<td>-</td>
<td>36.19$^{*}$</td>
<td>51</td>
<td>-0.334 ± 0.028</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>(b) 3.5 - 9.7</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>5A</td>
<td>126-142</td>
<td>44</td>
<td>(a) 65.4 - 78.2</td>
<td>128A, 132A, 134C, 139A</td>
<td>46.67</td>
<td>40</td>
<td>-0.307 ± 0.023</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>(b) 5.3 - 16.8</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>6A</td>
<td>143-152</td>
<td>25</td>
<td>(a) 79.4 - 98.8</td>
<td>150A</td>
<td>25.79</td>
<td>21</td>
<td>-0.745 ± 0.046</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>(b) 21.4 - 33.4</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>7A</td>
<td>153-159</td>
<td>17</td>
<td>(a) 43.4 - 47.5</td>
<td>155A, B, C</td>
<td>20.00</td>
<td>15</td>
<td>-0.052 ± 0.041</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>(b) 100.9 - 119.0</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
Table 4.1 (Contd.)

<table>
<thead>
<tr>
<th>Group</th>
<th>Event Numbers</th>
<th>Number of Points</th>
<th>Distance (a) and Azimuth (b) Range, (deg)</th>
<th>Points Rejected</th>
<th>Total $\chi^2$</th>
<th>Number of Degrees of Freedom</th>
<th>Correction, (sec/deg)</th>
</tr>
</thead>
<tbody>
<tr>
<td>7B$\dagger$</td>
<td>161-163, 165, 167A, C</td>
<td>11</td>
<td>(a) 48.1 - 49.4 (b) 96.3 - 99.8</td>
<td>165, A, C 167A</td>
<td>9.87</td>
<td>9</td>
<td>-0.829 ± 0.088</td>
</tr>
<tr>
<td>7C$\dagger$</td>
<td>164, 166, 167B, 168-169</td>
<td>8</td>
<td>(a) 49.1 - 50.1 (b) 93.0 - 96.8</td>
<td>168</td>
<td>6.57</td>
<td>6</td>
<td>0.404 ± 0.127</td>
</tr>
<tr>
<td>8A</td>
<td>170-181</td>
<td>27</td>
<td>(a) 27.9 - 40.6 (b) 291.1 - 296.6</td>
<td>-</td>
<td>28.87</td>
<td>23</td>
<td>-1.249 ± 0.051</td>
</tr>
<tr>
<td>8B</td>
<td>182, 185, 188-191</td>
<td>16</td>
<td>(a) 47.0 - 60.5 (b) 299.0 - 315.5</td>
<td>-</td>
<td>11.91</td>
<td>12</td>
<td>-0.634 ± 0.067</td>
</tr>
<tr>
<td>8C</td>
<td>192-200</td>
<td>25</td>
<td>(a) 70.1 - 71.4 (b) 297.7 - 312.5</td>
<td>-</td>
<td>15.03$\neq$</td>
<td>23</td>
<td>0.23 ± 0.079</td>
</tr>
<tr>
<td>8D</td>
<td>201, 206-207</td>
<td>10</td>
<td>(a) 80.0 - 85.7 (b) 325.0 - 327.3</td>
<td>-</td>
<td>9.61</td>
<td>8</td>
<td>-0.159 ± 0.043</td>
</tr>
<tr>
<td>8E</td>
<td>202-205, 208-209</td>
<td>16</td>
<td>(a) 80.3 - 90.9 (b) 283.4 - 306.8</td>
<td>208C</td>
<td>10.67</td>
<td>14</td>
<td>0.571 ± 0.039</td>
</tr>
<tr>
<td>9</td>
<td>211, 212A, C, 213A</td>
<td>6</td>
<td>(a) 37.6 - 41.8 (b) 124.7 - 147.7</td>
<td>-</td>
<td>2.07</td>
<td>4</td>
<td>0.444 ± 0.044</td>
</tr>
</tbody>
</table>

$\dagger$ data not used in preparing an overall $dT/d\Delta$ curve.

$\neq$ slightly undersmoothed owing to underestimation of errors.

* denotes a situation in which $\chi^2$ was just outside the range $u \pm \sqrt{2u}$, but the solution was allowed to stand.

$u$ is the number of degrees of freedom.
been divided into seventeen groups, each corresponding to a narrow range of azimuths and a distance range of about 10°; these groups are summarised in Table 4.1, and the azimuth ranges are seen to vary from 2.3° to 23.4° and the distance ranges from 1.3° to 19.4°. Thus some of the groups do cover an inordinately large azimuth range or an undesirably small distance range; unfortunately this is inescapable where the data from a particular region of the earth are sparse. In preparing an overall dT/dΔ curve no data from the Fiji and Tonga Islands region have been used, apart from events 158 and 159 of Appendix 2, owing to the complexity of the local structure affecting the P arrivals between azimuths of 90° and 100°. However, the results obtained from this region have been listed in Table 4.1 (groups 7B and 7C). A smooth curve was fitted through each group by the Method of Summary Values A, and a \( \chi^2 \) test was used to check the smoothing procedure. Each value of dT/dΔ was treated as an independent datum, and was weighted according to the reciprocal of the square of its standard error. The two or three dT/dΔ determinations for a particular event are not strictly independent, but it is justifiable to treat them as such. If, for example, the source were in a region of anomalous upper mantle velocities, each point would contain a bias characteristic of the source region; a similar bias could conceivably be introduced by structure near the recording station. The reason for taking more than one measurement of dT/dΔ and azimuth for each event is to compensate to some extent for the changes in waveform from one seismometer to another; these changes might be due to multiple arrivals caused by structure at the deepest point of the ray path rather than by local structure. In this study events of variable quality and with different numbers of seismometers working have been used, so that equal weight should not be given to each datum. Bias for a particular event and the errors in distance make it likely that the errors in dT/dΔ are underestimated. Consequently, if a \( \chi^2 \) test is applied in the smoothing process, the data will clearly be undersmoothed, and too much detail will be retained in the smooth curve. Since, at this stage, the data are smoothed merely to determine a correction, this does
not matter. If an acceptable value of $\chi^2$ could not be obtained it was assumed that some points contained bias, and the points giving the largest values of $\chi^2$ were rejected and the data smoothed again until an acceptable total $\chi^2$ was obtained. In practice it was found that the number of points that had to be rejected was small (see Table 4.1).

The area under each smooth curve was evaluated and subtracted from the travel-time difference between the end points of the range of integration derived from the 1968 Seismological Tables for P Phases. This enabled a correction to be applied to every value of $dT/d\Delta$ in a particular group. This method merely involves changing the base line of each set of $dT/d\Delta$ measurements, and is similar to using a single plane dipping interface to correct each group of data. Having devised a method for correcting the $dT/d\Delta$ values, it is now important to estimate the error on a correction term. According to Arnold (1968) the summary points computed in smoothing the data are approximately independent observations, and a standard error for each summary point can be obtained. If the coordinates of each summary point are $(p_i \pm \alpha_i, \Delta_i)$, $i = 1, 2 \cdots N$, where $N$ is the number of summary points and $\alpha_i$ is the standard error, an estimate of the error in the area under the curve can be obtained by integrating the curve interpolated between the points $(p_i + \alpha_i, \Delta_i)$ or $(p_i - \alpha_i, \Delta_i)$. To find the time differences between the end points of the range of integration, Table 2 of Arnold (1968) has been used even though it has already been shown that it is undersmoothed; the errors so-introduced are insignificant. Because the exact nature of the errors in the smoothing of the 1968 Seismological Tables for P Phases is not yet known, the correctly smoothed table of Appendix 6 has not been used. If the value of the distance was not integral, the error in the travel time at any distance has been taken as the maximum value of the errors at the two nearest integral values of $\Delta$. Thus for $\Delta = 34.3^0$, the standard error is taken as 0.10 seconds.

Consider a single group of data. Let the errors in the travel times $T_1$ and $T_2$ at the end points of the range of integration, $\Delta_1$ and $\Delta_2$, be
\( \sigma_1 \) and \( \sigma_2 \) respectively, and let the area under the \( \frac{dT}{dA} \) curve be \( A - \sigma_3 \).

Hence the correction is given by

\[
C = \frac{(T_2 + \sigma_2) - (T_1 + \sigma_1) - (A + \sigma_3)}{\Delta_2 - \Delta_1}
\]

\[
= \frac{(T_2 - T_1 - A) + \sqrt{\sigma_1^2 + \sigma_2^2 + \sigma_3^2}}{\Delta_2 - \Delta_1}
\]

This correction \( C \) is then added to each value of \( \frac{dT}{dA} \) in the group, and the error estimate in \( \frac{dT}{dA} \) is appropriately revised. The error in the correction \( C \) will be smaller for groups having both a large distance range and an abundance of points. Therefore data in those azimuth ranges where there is a large number of events distributed over a considerable distance range and clearly affected by a similar local structure, will tend to be given more weight relative to values of \( \frac{dT}{dA} \) in regions where the data are sparse, or in azimuth ranges in which the local structure changes relatively rapidly. Although there are probably systematic errors in the 1968 Seismological Tables for P Phases (Cleary and Muirhead, 1969, and Muirhead and Cleary, 1969), they were used in preference to the surface focus travel times of Hales et al. (1968) because standard errors in the travel times were supplied; a knowledge of these errors is essential if a statistical significance test is to be applied in the smoothing of the corrected data. In addition much more data were used than in the original work of Cleary and Hales (1966), so that they are potentially capable of giving a more reliable travel-time curve once the systematic errors have been removed.

Finally a smooth curve was fitted through the corrected values of \( \frac{dT}{dA} \) by the Method of Summary Values A, and this is shown in Figure 8.1. These corrected values are also listed in Table 2 of Appendix 2. Only 492 of the 535 points were used in deriving the smooth curve, and a \( \chi^2 \) test gave 487.5 on 470 degrees of freedom. The remaining 43 points were rejected.
### Table 4.2  List of Summary Points of Preliminary dT/dΔ Curve

<table>
<thead>
<tr>
<th>Range, (deg)</th>
<th>First Summary Point</th>
<th>Second Summary Point</th>
<th>Total ( \chi^2 )</th>
<th>No. of Degrees of Freedom</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>( \frac{dT}{d\Delta} ) (sec/deg)</td>
<td>( \Delta ) (deg)</td>
<td>( \frac{dT}{d\Delta} ) (sec/deg)</td>
<td>( \Delta ) (deg)</td>
</tr>
<tr>
<td>27.9 - 30.9</td>
<td>9.084 ± 0.025</td>
<td>28.106</td>
<td>8.845 ± 0.032</td>
<td>30.372</td>
</tr>
<tr>
<td>31.3 - 34.5</td>
<td>8.716 ± 0.012</td>
<td>31.960</td>
<td>8.666 ± 0.013</td>
<td>33.634</td>
</tr>
<tr>
<td>34.6 - 39.2</td>
<td>8.657 ± 0.014</td>
<td>35.316</td>
<td>8.556 ± 0.021</td>
<td>37.775</td>
</tr>
<tr>
<td>39.6 - 44.4</td>
<td>8.342 ± 0.016</td>
<td>40.632</td>
<td>8.066 ± 0.019</td>
<td>43.397</td>
</tr>
<tr>
<td>44.8 - 49.0</td>
<td>7.895 ± 0.021</td>
<td>45.334</td>
<td>7.541 ± 0.019</td>
<td>48.433</td>
</tr>
<tr>
<td>49.9 - 54.7</td>
<td>7.464 ± 0.019</td>
<td>50.736</td>
<td>7.317 ± 0.018</td>
<td>53.374</td>
</tr>
<tr>
<td>55.4 - 65.4</td>
<td>7.088 ± 0.022</td>
<td>56.221</td>
<td>6.506 ± 0.047</td>
<td>62.898</td>
</tr>
<tr>
<td>65.4 - 72.1</td>
<td>6.387 ± 0.021</td>
<td>66.344</td>
<td>6.054 ± 0.015</td>
<td>71.197</td>
</tr>
<tr>
<td>73.4 - 81.2</td>
<td>5.864 ± 0.016</td>
<td>74.412</td>
<td>5.441 ± 0.017</td>
<td>79.914</td>
</tr>
<tr>
<td>81.2 - 87.8</td>
<td>5.103 ± 0.042</td>
<td>82.868</td>
<td>4.906 ± 0.025</td>
<td>86.169</td>
</tr>
<tr>
<td>93.7 - 98.8</td>
<td>4.678 ± 0.054</td>
<td>93.700</td>
<td>4.596 ± 0.080</td>
<td>98.800</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Total</td>
<td>489.18</td>
</tr>
</tbody>
</table>
because they contained bias probably due to either station or source effects. The 22 summary points of the smooth curve are listed in Table 4.2; these points have been interpolated at 0.1° intervals using fifth divided differences from 27.9° to 86.2°, and using only fourth divided differences beyond 86.2° to avoid reading too much detail into a portion of the curve where the data are sparse.

It is worth mentioning that care has to be taken that this method of analysis does not introduce spurious discontinuities or remove real anomalies between the separate groups of data, owing to the inaccuracy of the assumption that the correction to dT/dA is a constant for each group. To investigate this possibility the data of group 1A have been divided into two groups and a separate correction derived for each. The results are summarised in Table 4.3, and indicate that the assumption of a constant correction to dT/dA is an excellent approximation. The groups in similar azimuth ranges and adjacent distance ranges for which the corrections differ substantially are 2A and 2B, and also 8A, 8B and 8E. In groups 8A, 8B and 8E the density of the data is low and there is a large distance gap between the separate groups. It is therefore important to point out that since there is a large difference in the corrections for each group, which becomes more positive as the distance increases, the data of each group are tilted; ideally, therefore, distance dependent corrections are required. This tilt has little effect on the overall dT/dA curve as the amount of data involved is small. For groups 2A and 2B the corrections to dT/dA are -0.935 sec/deg and -0.829 sec/deg respectively. The effect is to make the dT/dA curve between 37° and 42° a little flatter than it should be, possibly obliterating a rapid decrease in dT/dA within this distance range. The method of correcting the dT/dA data does also result in a large amount of scatter beyond 80°. The overall smooth curve of Figure 8.1 is the smoothest possible curve that will fit the data, so that any first or second order discontinuities in the lower mantle will tend to be smeared out. A detailed discussion of the dT/dA profiles and possible interpretations will be presented in Chapters 6 to 8.
Table 4.3  Empirical Corrections to $dT/d\Delta$ with Data Redistributed

<table>
<thead>
<tr>
<th>Group</th>
<th>Event Numbers</th>
<th>Number of Points</th>
<th>Distance (a) and Azimuth (b) Range, (deg)</th>
<th>Points Rejected</th>
<th>Total $\chi^2$</th>
<th>Number of Degrees of Freedom</th>
<th>Correction, (sec/deg)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1C</td>
<td>1-8, 10, 12</td>
<td>28</td>
<td>(a) 27.9 - 31.8</td>
<td>-</td>
<td>19.75</td>
<td>24</td>
<td>-0.186</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>(b) 73.0 - 86.0</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1D</td>
<td>9, 11, 13-22,</td>
<td>46</td>
<td>(a) 31.5 - 34.9</td>
<td>19A, 22C,</td>
<td>47.48</td>
<td>42</td>
<td>-0.186</td>
</tr>
<tr>
<td></td>
<td>24-28, 31,</td>
<td></td>
<td>(b) 78.2 - 94.6</td>
<td>28A</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>34B, 35</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
4.4 A dT/dΔ Curve for the Lower Mantle.

(ii) Refined Method.

There are several refinements that can be introduced to improve the technique for correcting dT/dΔ described in the previous section. First the raw dT/dΔ data used in preparing the final dT/dΔ curve were those of Table 3 of Appendix 2. Thus a reduction in the errors and in the scatter of the data has been sought by calculating each dT/dΔ and azimuth value with all onset times giving a residual greater than 0.04 seconds removed. In section 4.3 the ellipticity of the earth was not considered. In Chapter 2 it was shown that the ellipticity correction can be as great as -0.024 sec/deg. However, this is irrelevant if dT/dΔ for a range of azimuths and distances is corrected by a constant amount. Of particular interest is the variation of the ellipticity correction with distance for a specific azimuth profile. For the groups of Table 4.1 the maximum difference in ellipticity correction between two events is 0.01 sec/deg. Therefore the errors introduced by neglecting the ellipticity correction are indeed very small. In applying the ellipticity corrections the effect of abnormal focal depth has been ignored. In view of the comments of Bullen (1939) this seems quite justifiable. If the P arrivals at the array cross a dipping interface with a significant velocity contrast, the correction to dT/dΔ becomes slightly distance dependent and fairly strongly azimuth dependent, even for an azimuth range of only 10° to 20°. This effect should really be allowed for. Since the preliminary method was tested and successfully applied in August 1969, 63 additional dT/dΔ measurements from 32 more events have been obtained, so that in the final analysis there are 616 dT/dΔ values from 224 events. These extra events are numbers 261-292 of Appendix 2.

The most important piece of information that was not used in section 4.3 is the measured azimuth of arrival of each event. This suggests that the vector summation method described in section 3.5 could be applied to the dT/dΔ and azimuth data if split into fewer groups than in Table 3.4.
### Table 4.4 List of Structures Used to Correct the dT/dΔ Measurements

<table>
<thead>
<tr>
<th>Structure Group</th>
<th>Event Numbers</th>
<th>Number of Points</th>
<th>Distance Range (deg)</th>
<th>Azimuth Ranges, (deg)</th>
<th>Fisher's 'K'</th>
<th>Dip Direction (deg)</th>
<th>Dip Angle (deg)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>1-22, 24-28, 31, 34B, 35, 38</td>
<td>79</td>
<td>27.9 - 40.9</td>
<td>73.0 - 94.6</td>
<td>9.96</td>
<td>314.4</td>
<td>3.84</td>
</tr>
<tr>
<td>2*</td>
<td>29-30, 32-33, 34A, C, 36-37</td>
<td>19</td>
<td>34.1 - 35.7</td>
<td>93.8 - 100.8</td>
<td>2.83</td>
<td>129.3</td>
<td>3.75</td>
</tr>
<tr>
<td>3</td>
<td>40, 43-80A, C, 261-262, 106-107</td>
<td>125</td>
<td>32.6 - 47.9</td>
<td>8.6 - 25.6</td>
<td>42.40</td>
<td>221.0</td>
<td>9.59</td>
</tr>
<tr>
<td>4</td>
<td>81-92, 263-267</td>
<td>45</td>
<td>41.7 - 53.3</td>
<td>342.9 - 357.8</td>
<td>22.34</td>
<td>158.3</td>
<td>6.59</td>
</tr>
<tr>
<td>5</td>
<td>274-275, 93-95</td>
<td>11</td>
<td>64.5 - 77.7</td>
<td>338.5 - 358.5</td>
<td>28.56</td>
<td>177.3</td>
<td>7.48</td>
</tr>
<tr>
<td>6*</td>
<td>39, 41-42, 97, 99, 102-104</td>
<td>21</td>
<td>30.1 - 39.1</td>
<td>9.9 - 11.8</td>
<td>61.30</td>
<td>165.1</td>
<td>7.45</td>
</tr>
<tr>
<td>7*</td>
<td>98, 100-101, 210-211</td>
<td>14</td>
<td>31.3 - 37.3</td>
<td>333.0 - 338.5</td>
<td>61.81</td>
<td>198.7</td>
<td>7.13</td>
</tr>
<tr>
<td>8</td>
<td>80B, 108-112, 114, 116-125, 269-273</td>
<td>56</td>
<td>46.4 - 63.9</td>
<td>4.8 - 10.5</td>
<td>15.88</td>
<td>156.0</td>
<td>3.43</td>
</tr>
<tr>
<td>9</td>
<td>126-142, 144, 276-277</td>
<td>54</td>
<td>65.4 - 80.1</td>
<td>5.3 - 16.8</td>
<td>33.64</td>
<td>228.5</td>
<td>3.95</td>
</tr>
<tr>
<td>10</td>
<td>143, 145-152, 279-284</td>
<td>35</td>
<td>74.2 - 98.8</td>
<td>21.4 - 33.4</td>
<td>59.39</td>
<td>241.0</td>
<td>8.84</td>
</tr>
<tr>
<td>Structure Group</td>
<td>Number of Points</td>
<td>Distance Range (deg)</td>
<td>Azimuth Ranges (deg)</td>
<td>Fisher's K'</td>
<td>Dip Angle (deg)</td>
<td>Dip Direction (deg)</td>
<td></td>
</tr>
<tr>
<td>-----------------</td>
<td>-----------------</td>
<td>----------------------</td>
<td>----------------------</td>
<td>-------------</td>
<td>----------------</td>
<td>---------------------</td>
<td></td>
</tr>
<tr>
<td>11</td>
<td>21</td>
<td>43.4 - 47.5</td>
<td>100.9 - 119.5</td>
<td>23.02</td>
<td>2.95</td>
<td>201.3</td>
<td></td>
</tr>
<tr>
<td>12</td>
<td>29</td>
<td>27.9 - 44.6</td>
<td>291.1 - 297.5</td>
<td>32.15</td>
<td>11.39</td>
<td>124.4</td>
<td></td>
</tr>
<tr>
<td>13</td>
<td>27</td>
<td>47.0 - 70.2</td>
<td>297.7 - 314.3</td>
<td>26.26</td>
<td>7.73</td>
<td>183.3</td>
<td></td>
</tr>
<tr>
<td>14</td>
<td>57</td>
<td>71.2 - 90.9</td>
<td>283.4 - 327.3</td>
<td>89.50</td>
<td>9.45</td>
<td>233.7</td>
<td></td>
</tr>
</tbody>
</table>

* Denotes structures derived from $dI/d\Lambda$ measurements from more than one azimuth range.

This structure has not been used to correct the data of group 2 owing to the small value of $K'$. (43 Rejected)
Consequently the data have been divided into 15 groups by visual examination of the structures of Appendix 5. It is thus postulated that each $dT/d\Delta$ value within one of these structure groups results from refraction by the same local structure, and a vector sum of the corresponding dip vectors will give a good approximation to this structure. Generally the data within a particular structure group have been taken from fairly narrow ranges of azimuths and distances; however, in three cases it was thought justifiable to consider $P$ arrivals from different regions but from similar distances to be affected by the same local structure. So by using the dip vectors of Appendix 5, an average dip vector for each structure group has been calculated using programme ZCW101D3; this vector was used to correct the $dT/d\Delta$ values within the group with the aid of tables of $dT/d\Delta$ and azimuth anomalies prepared using programme ZCW101A4. These special tables are listed in Appendix 7. The structure groups differ slightly from the groups of Table 4.1 and are shown in Table 4.4. Ellipticity corrections have been applied to the data of those groups labelled with the suffix E; for the remaining groups the variation of ellipticity correction with distance or azimuth was insignificant. Any random errors in the structure corrections have been neglected.

The $dT/d\Delta$ values corrected to first order for local structure using both the 1968 Seismological Tables for $P$ Phases and the azimuth anomalies, but still possibly containing systematic error, have been corrected again by the method described in section 4.3. It is necessary to constrain each profile to a travel-time curve since the structures can introduce bias if there is a large number of observations close to an anomalous region of the $dT/d\Delta$ curve. For some structure groups the $dT/d\Delta$ values have been combined to give a profile covering a large distance interval. 11 such profiles have been used, 8 of which can be classed as regional and the remainder contain data from events occurring in two or more regions of the earth. In smoothing the data to calculate the correction, the Method of Summary Values B was used for profile 2R, and version A was used for the rest. The results, which are essentially a refined version of Table 4.1, are displayed in Table 4.5.
### Table 4.5  \( \frac{dT}{dA} \) Profiles Corrected by the Method of Summary Values

<table>
<thead>
<tr>
<th>Profile</th>
<th>Event Numbers</th>
<th>Structure Groups Used</th>
<th>Number of Points</th>
<th>Distance (a) and Azimuth (b) Range (deg)</th>
<th>Points Rejected</th>
<th>Total ( \chi^2 )</th>
<th>Number of Degrees of Freedom</th>
<th>Correction, ( \text{sec/deg} )</th>
</tr>
</thead>
<tbody>
<tr>
<td>1R*</td>
<td>1-37</td>
<td>1, 2</td>
<td>90</td>
<td>(a) 27.9 - 35.7</td>
<td>12B, 17A, 18B, 22C, 28A, 36A</td>
<td>92.12</td>
<td>82</td>
<td>0.087 ± 0.035</td>
</tr>
<tr>
<td>2R</td>
<td>40, 43-80, 262, 106-107</td>
<td>3, 8E</td>
<td>107</td>
<td>(a) 32.6 - 47.9</td>
<td>61A, B, C, 65A, B, 66A, 68A, 72B, C, 74A, B, 78A, 262A, B, 80B</td>
<td>117.12</td>
<td>98</td>
<td>0.123 ± 0.027</td>
</tr>
<tr>
<td>3R</td>
<td>81-92, 263-267</td>
<td>4E</td>
<td>43</td>
<td>(a) 41.7 - 53.3</td>
<td>84A, B</td>
<td>33.63</td>
<td>37</td>
<td>0.085 ± 0.024</td>
</tr>
<tr>
<td>4</td>
<td>39, 41-42, 97-104, 210-211</td>
<td>6E, 7E</td>
<td>31</td>
<td>(a) 30.1 - 39.1</td>
<td>210A, B, 42A, 100B</td>
<td>28.37</td>
<td>27</td>
<td>-0.002 ± 0.030</td>
</tr>
<tr>
<td>5R</td>
<td>108-142, 144, 269-273, 276-277</td>
<td>8E, 9E</td>
<td>109</td>
<td>(a) 50.8 - 80.1</td>
<td>111B, 128A, 136A, 139A, 141A</td>
<td>104.69</td>
<td>97</td>
<td>0.029 ± 0.028</td>
</tr>
<tr>
<td>6R</td>
<td>143, 145-152, 280-284</td>
<td>10E</td>
<td>25</td>
<td>(a) 79.4 - 93.7, 98.5</td>
<td>149B, C, 150A, B, 282A</td>
<td>27.49</td>
<td>21</td>
<td>-0.039 ± 0.024</td>
</tr>
<tr>
<td>7</td>
<td>93-95, 274-275, 5E, 13E, 279, 192-200, 14E, 202A, 203A, 204A</td>
<td>41</td>
<td>(a) 64.5 - 80.4</td>
<td>201A</td>
<td>38.91</td>
<td>37</td>
<td>0.043 ± 0.023</td>
<td></td>
</tr>
</tbody>
</table>
Table 4.5 (Contd)

<table>
<thead>
<tr>
<th>Profile</th>
<th>Event Numbers</th>
<th>Structure Groups Used</th>
<th>Number of Points</th>
<th>Distance (a) and Azimuth (b) Range (deg)</th>
<th>Points Rejected</th>
<th>Total $\chi^2$</th>
<th>Number of Degrees of Freedom</th>
<th>Correction, (sec/deg)</th>
</tr>
</thead>
<tbody>
<tr>
<td>8R</td>
<td>38, 153-159, 161-163, 165, 167</td>
<td>11, 1</td>
<td>28</td>
<td>(a) 40.9 - 49.4</td>
<td>155A, B, C, 156D, 158B</td>
<td>26.18</td>
<td>24</td>
<td>- 0.014 ± 0.033</td>
</tr>
<tr>
<td>9R</td>
<td>170-175, 177-179, 181, 268</td>
<td>12E</td>
<td>26</td>
<td>(a) 27.9 - 44.6</td>
<td>175A, B</td>
<td>29.19</td>
<td>22</td>
<td>0.059 ± 0.033</td>
</tr>
<tr>
<td>10R</td>
<td>182, 185, 188-191</td>
<td>13E</td>
<td>15</td>
<td>(a) 47.0 - 60.5</td>
<td>188A</td>
<td>16.86</td>
<td>11</td>
<td>- 0.101 ± 0.059</td>
</tr>
<tr>
<td>11</td>
<td>201 B, C, - 204B, C, 205-209, 285-292</td>
<td>14E</td>
<td>31</td>
<td>(a) 80.0 - 90.9</td>
<td>286A, B, 201C, 203B, 204C, 205B</td>
<td>22.12</td>
<td>25</td>
<td>0.101 ± 0.038</td>
</tr>
</tbody>
</table>

* R denotes a regional profile.

≠ the data of profile 2 were smoothed by the Method of Summary Values B; the remainder were smoothed using version A.

鲚 E denotes data from the structure groups of Table 4.4 which have been corrected for ellipticity.

‡ denotes a situation in which $\chi^2$ was just outside the range $u \pm \sqrt{u}$, but the solution was allowed to stand. $u$ is the number of degrees of freedom.

//= The one event beyond 93.7° was not included when fitting a curve through the data.
**Table 4.6 List of Summary Points of Final dT/d Δ Curve**

<table>
<thead>
<tr>
<th>Range (deg)</th>
<th>First Summary Point</th>
<th>Second Summary Point</th>
<th>Third Summary Point</th>
<th>Total $^2$</th>
<th>No. of Degrees of Freedom</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>$dT/d\Delta$</td>
<td>$\Delta$</td>
<td>$dT/d\Delta$</td>
<td>$\Delta$</td>
<td></td>
</tr>
<tr>
<td></td>
<td>(sec/deg)</td>
<td>(deg)</td>
<td>(sec/deg)</td>
<td>(deg)</td>
<td></td>
</tr>
<tr>
<td>27.9-30.3</td>
<td>9.141 $^+0.026$</td>
<td>27.947</td>
<td>9.036 $^+0.043$</td>
<td>28.957</td>
<td>8.762 $^+0.038$</td>
</tr>
<tr>
<td>30.9-34.8</td>
<td>8.771 $^+0.014$</td>
<td>31.545</td>
<td>8.715 $^+0.010$</td>
<td>32.922</td>
<td>8.634 $^+0.017$</td>
</tr>
<tr>
<td>34.9-42.3</td>
<td>8.642 $^+0.014$</td>
<td>35.527</td>
<td>8.483 $^+0.016$</td>
<td>38.325</td>
<td>8.244 $^+0.016$</td>
</tr>
<tr>
<td>43.4-48.9</td>
<td>8.022 $^+0.019$</td>
<td>43.883</td>
<td>7.940 $^+0.015$</td>
<td>45.971</td>
<td>7.630 $^+0.017$</td>
</tr>
<tr>
<td>49.0-67.0*</td>
<td>7.488 $^+0.013$</td>
<td>50.860</td>
<td>7.125 $^+0.014$</td>
<td>56.192</td>
<td>6.422 $^+0.015$</td>
</tr>
<tr>
<td>69.8-78.2</td>
<td>6.192 $^+0.014$</td>
<td>70.546</td>
<td>5.936 $^+0.015$</td>
<td>73.499</td>
<td>5.649 $^+0.021$</td>
</tr>
<tr>
<td>79.4-88.2</td>
<td>5.267 $^+0.015$</td>
<td>80.279</td>
<td>5.089 $^+0.020$</td>
<td>83.585</td>
<td>4.944 $^+0.029$</td>
</tr>
<tr>
<td>Total</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>49.0-56.1</td>
<td>7.501 $^+0.014$</td>
<td>50.670</td>
<td>7.234 $^+0.015$</td>
<td>54.769</td>
<td>-</td>
</tr>
<tr>
<td>56.6-67.0</td>
<td>6.948 $^+0.024$</td>
<td>58.108</td>
<td>6.430 $^+0.015$</td>
<td>65.813</td>
<td>-</td>
</tr>
<tr>
<td>Total</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

* denotes distance range over which three summary points were replaced by four points obtained using version A.

The coordinates of these four summary points are listed in the second portion of the table.
<table>
<thead>
<tr>
<th>Range, (deg)</th>
<th>First Summary Point</th>
<th>Second Summary Point</th>
<th>Total</th>
<th>No. of Degrees of Freedom</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>dT/dΔ (sec/deg)</td>
<td>Δ (deg)</td>
<td>dT/dΔ (sec/deg)</td>
<td>Δ (deg)</td>
</tr>
<tr>
<td>27.9 - 31.6</td>
<td>9.067 ± 0.052</td>
<td>28.594</td>
<td>8.774 ± 0.026</td>
<td>31.328</td>
</tr>
<tr>
<td>31.8 - 37.7</td>
<td>8.723 ± 0.030</td>
<td>32.567</td>
<td>8.583 ± 0.025</td>
<td>36.648</td>
</tr>
<tr>
<td>39.1 - 46.3</td>
<td>8.263 ± 0.025</td>
<td>40.448</td>
<td>7.973 ± 0.019</td>
<td>45.567</td>
</tr>
<tr>
<td>46.8 - 52.6</td>
<td>7.732 ± 0.018</td>
<td>47.882</td>
<td>7.387 ± 0.022</td>
<td>51.673</td>
</tr>
<tr>
<td>52.9 - 60.5</td>
<td>7.305 ± 0.031</td>
<td>53.132</td>
<td>6.607 ± 0.046</td>
<td>60.330</td>
</tr>
<tr>
<td>64.5 - 71.3</td>
<td>6.410 ± 0.035</td>
<td>65.423</td>
<td>6.172 ± 0.022</td>
<td>71.086</td>
</tr>
<tr>
<td>71.4 - 80.0</td>
<td>6.123 ± 0.020</td>
<td>71.645</td>
<td>5.612 ± 0.029</td>
<td>77.361</td>
</tr>
<tr>
<td>80.3 - 82.8</td>
<td>5.265 ± 0.020</td>
<td>80.400</td>
<td>5.176 ± 0.032</td>
<td>82.609</td>
</tr>
<tr>
<td>83.9 - 87.8</td>
<td>5.102 ± 0.028</td>
<td>84.736</td>
<td>4.983 ± 0.036</td>
<td>87.021</td>
</tr>
<tr>
<td>Total</td>
<td></td>
<td></td>
<td>170.83*</td>
<td></td>
</tr>
</tbody>
</table>

* The value of $\chi^2$ is just outside the range $153 \pm 17.5$, but the solution was allowed to stand.
As in the preliminary method it was necessary to reject a small proportion of the dT/dΔ data. However, it must be remembered that this rejected information has been used in deriving the structures of Table 4.4. Ideally these structures should be worked out again with the bad data removed. Since the method of correcting the data was being tried for the first time, no new set of structures was derived. Consequently some systematic errors are present in the structures; these small errors will be transmitted to the corrected dT/dΔ values, but will be removed to some extent when the portions of the dT/dΔ curve are constrained to the 1968 Seismological Tables for P Phases.

Finally a smooth curve was fitted through the corrected dT/dΔ values using the Method of Summary Values B. In fitting the curve only 515 points were used, and a χ² test gave 522.8 on 494 degrees of freedom, which is well inside the range 494.0 ± 31.4. Also, owing to their large separation, the three summary points covering the distance interval 49.0° to 67.0° were rejected, and replaced by four points derived by splitting the distance interval into two sections and smoothing by version A of the Method of Summary Values. These four points were used in preparing the final smooth dT/dΔ curve used for calculating a velocity distribution. The refined version of the dT/dΔ curve is plotted in Figure 8.2, and the summary points are listed in Table 4.6. The smooth curve was fitted only to a distance of 88.2°; beyond 88.2° dT/dΔ decreases sharply and subsequently remains almost constant (see Chapter 8). Two regional dT/dΔ curves have also been prepared. The first has been obtained simply by interpolating between the corrected summary points of profiles 2, 5 and 6 of Table 4.5 with one small modification described in Chapter 8. The events of profiles 2, 5 and 6 define the region called for convenience the west Pacific margin. The second has been prepared by smoothing all the corrected data from events to the west of the first regional profile; the summary points are listed in Table 4.7.

At this stage it is interesting to compare the parameters of the preliminary structures of Table 3.3 with those of Table 4.4 determined by summing dip vectors. For the vector summation method the dip angles are
Figure 4.1  Section through the earth to illustrate the ideas of seismic ray theory and the process of 'stripping the earth'. 
expected to be larger, since the interface has been placed at the base of the crust instead of within the crust. Comparing Tables 3.3 and 4.4, the structures worked out from Aleutian Islands events and from Mariana Islands earthquakes are similar and fairly useful in correcting $dT/d\Delta$ measurements; however, between distances of about $48^\circ$ and $80^\circ$ and over the azimuth range $8^\circ$ to $16^\circ$, the structure affecting the P arrivals seems quite different. The structure derived from the recording of the Novaya Zemlya explosion cannot be used for events from similar azimuths but at shorter distances. The preliminary structures of Table 3.3 are therefore of limited value owing to the complex series of changes in structure between azimuths of $330^\circ$ and $30^\circ$.

4.5 THE CONSTRUCTION OF A VELOCITY MODEL.

Mathematical Theory. The development of high speed digital computers in recent years has led to more effective use of ray theory in studying the structure of the earth's interior. In developing the theory concerning the transmission of body waves within the earth, it is assumed that the earth is spherical and symmetrical about its centre in all its properties; in addition diffraction effects are ignored. Consider a ray travelling from a source E at the earth's surface and emerging at a recording station S as shown in Figure 4.1. Owing to the assumed symmetry of the earth, a ray path lies in a plane defined by E, S and the earth's centre O. At any point Q along a ray path Bullen (1963, p. 109) has shown that

$$p = r \sin i / v \quad (4-2)$$

where $r$ is the radial distance OQ, $i$ in the acute angle between OQ and the direction of the ray at Q, and $v$ is the ray velocity at Q. $p$ is a constant along a particular ray path. Then, if $T$ is the total travel time along the ray path, from Bullen (1963, p. 110),

$$p = dT/d\Delta \quad (4-3)$$
T-Δ Relations. Let the parameter \( \eta = r/v \). Bullen (1963, pp. 111-112) has shown that the travel times and distances can be computed from a velocity distribution by evaluation of integrals of the form

\[
\frac{1}{2} \Delta_{12} = p \int_{r_2}^{r_1} r^{-1} (\eta^2 - p^2)^{-\frac{1}{2}} \, dr \\
\frac{1}{2} T_{12} = \int_{r_2}^{r_1} \eta^2 r^{-1} (\eta^2 - p^2)^{-\frac{1}{2}} \, dr
\] (4-4)

where \( \frac{1}{2} \Delta_{12} \) and \( \frac{1}{2} T_{12} \) represent the distance and travel time respectively for a ray of parameter \( p \) terminating at levels \( r_1 \) and \( r_2 \) where \( r_1 > r_2 \). The factor of one half on the left hand side of these equations accounts for the fact that \( \Delta_{12} \) and \( T_{12} \) refer to both the upgoing and downgoing portions of the ray path within the layer. Thus, if \( \Delta_{12} \) and \( T_{12} \) are required for a segment such as EF in Figure 4.1, this factor is removed.

A simple and effective way of evaluating expressions (4-4) and (4-5) for a specified \( p \) and a tabulated velocity model has been described by Bullen (1961). The brief account of this method outlined below is based on Bullen's work and follows closely the treatment given by Engdahl, Taggart, Lobdell, Arnold and Clawson (1968). Suppose the velocity between each point of the table is represented by a power law of the form \( v = a r^b \) with \( a \) and \( b \) constants. Thus \( v(r) \) is continuous, but \( dv/dr \) is in general discontinuous at each tabulated point. So if \( v = a r^{b_{12}} \) in the layer between radii \( r_1 \) and \( r_2 \), substitution in (4-4) and (4-5) yields (see Engdahl et al., 1968)

\[
\frac{1}{2} \Delta_{12} = (1 - b_{12})^{-1} \left[ \cos^{-1} \left( \frac{p}{\eta_1} \right) - \cos^{-1} \left( \frac{p}{\eta_2} \right) \right] \\
\frac{1}{2} T_{12} = (1 - b_{12})^{-1} \left[ \left( \eta_1^2 - p^2 \right)^{\frac{1}{2}} - \left( \eta_2^2 - p^2 \right)^{\frac{1}{2}} \right]
\] (4-6)

where \( \eta_1 \) and \( \eta_2 \) are the values of \( \eta \) at \( r_1 \) and \( r_2 \) respectively. If a ray reaches its deepest point between radii \( r_1 \) and \( r_2 \)
\[ \Delta_{12} = 2 \left( 1 - b_{12} \right)^{-1} \cos^{-1} \left( \frac{p}{\eta_1} \right) \]  
(4.8)

\[ T_{12} = 2 \left( 1 - b_{12} \right)^{-1} \left( \eta_1^2 - p^2 \right)^{\frac{1}{2}} \]  
(4.9)

The radius of penetration of this ray is given by

\[ r_p = (a_{12} \eta_1) \]  
(4.10)

where

\[ c_{12} = (1 - b_{12})^{-1} \]  
(4.11)

Equations (4.6) to (4.9) enable the travel time and distance of any ray or ray segment to be determined by simple summation, each layer being represented by new values of \( a, b \) and \( \eta \). One advantage of this method of evaluating travel times and distances is that the parameter \( b \) is the same as the parameter \( \zeta \) defined by Bullen (1963, p. 112).

Thus

\[ \zeta = b = \frac{r}{v} \frac{d v}{d r} \]  
(4.12)

Using a tabulated velocity distribution \( b_{12} \) and \( a_{12} \) are determined from the relations

\[ b_{12} = \ln \left( \frac{v_1}{v_2} \right) / \ln \left( \frac{r_1}{r_2} \right) \]  
(4.13)

\[ a_{12} = v_1 r_1^{b_{12}} \]  
(4.14)

This simple theory has four important applications to the work described in this thesis. First it can be used to derive T-\( \Delta \) and p-\( \Delta \) curves for a specified velocity distribution, and therefore can be used to fit a velocity model to a p-\( \Delta \) or T-\( \Delta \) curve by trial and error. It can also be used for ray tracing to investigate the complexity in travel-time curves caused by regions of high velocity gradients or low velocity layers. These two processes can be performed using computer programme ZCW101C6. As a check on the accuracy obtainable by this power law interpolation method, the P wave velocity distribution listed in the 1968 Seismological Tables for P Phases was used to recalculate the travel times. The original times were reconstructed
to within 0.02 seconds at 30°. One important point in connection with this programme needs to be mentioned. If, in the vicinity of a discontinuity in the velocity model, the velocity gradient \( dv/dr \) is very large and negative in a specific layer, it often happens that \( r^b < \exp(-174.673) \). In this case the computer cannot cope with the data, so that the region of high velocity gradients has to be modified slightly or a first order discontinuity has to be introduced. In practice this is not a serious disadvantage. The theory can also be used to correct the epicentral distance \( \angle FOS \) in Figure 4.1, of an earthquake at a focal depth \( h \), to an adjusted surface to surface distance \( \angle EOS \). The only information required is the focal depth and an approximate value of \( dT/d\Delta \). Finally the technique can be used to "strip" the earth to any depth in the manner described in the next section on the inversion of body wave data. Both of these processes can be implemented using computer programme ZCW101C2.

The Herglotz-Wiechert Method. It was remarked in Chapter 1 that if a \( dT/d\Delta \) curve is prepared a velocity distribution for the earth's mantle can be calculated. Provided that \( \eta \) is a monotonic decreasing function of depth below the surface, the integral equation (4.4) for \( \eta \) can be solved to yield the integral (see Bullen, 1963, p. 119 or Stacey, 1969, p. 271).

\[
\ln \left( \frac{r_0}{r_1} \right) = \frac{1}{\pi} \int_0^\Delta \cosh^{-1} \left( \frac{p}{p_1} \right) d \Delta 
\]

(4.15)

\[
v_1 = \frac{r_1}{p_1} \quad (4.16)
\]

where \( r_0 \) is the mean radius of the earth,

\( r_1 \) is the radius where the velocity of P or S waves is \( v_1 \),

\( p_1 = \eta_1 \), and is the parameter of a ray that penetrates to depth \( r_0 - r_1 \),

and reaches the surface at a distance \( \Delta_1 \) from the source.

For ease of computation (4.15) can conveniently be rewritten as

\[
r_1 = r_0 \exp \left[ -\frac{1}{\pi} \int_0^\Delta \ln \left( \frac{p}{p_1} + \sqrt{\left( \frac{p}{p_1} \right)^2 - 1} \right) d \Delta \right] \quad (4.17)
\]
Therefore, to evaluate $v_1$, $dT/d\Delta$ is required as a function of $\Delta$ from 0 to $\Delta_1$. Expression (4.17) enables $r_1$ to be calculated and hence $v_1$ using (4.16). In practice the data extend only from $\Delta_1$ down to a distance $\Delta_2$, so that a velocity model based on other work must be used to fill in the gap. This assumed velocity model is used to find where the ray path of parameter $p(\Delta_2)$ reaches its maximum depth $h$, and then the contributions to the distance $\Delta$ of the region down to radius $r_0 - h$ for values of $p$ between $p(\Delta_2)$ and $p(\Delta_1)$ are calculated. This enables another $dT/d\Delta$ curve characteristic of an earth of radius $r_0 - h$ to be derived - a process known as 'stripping the earth'. This technique for deriving velocity distributions for P or S waves is known as the Herglotz-Wiechert method. In general the velocities of P or S waves increase with increasing depth throughout the earth, though there are notable exceptions.

If $d\eta/dr > 0$, i.e. $\xi > 1$, so that the velocity decreases with increasing depth at a sufficiently high rate, the Herglotz-Wiechert method formally breaks down. In this case the trouble arises because no ray paths reach their maximum depth within the low velocity layer itself. The method can, however, be adapted if the velocity decrease is discontinuous as at the mantle-core boundary. The problem of low velocity layers will be discussed in greater detail in Chapter 6.

A smooth $dT/d\Delta$ curve can be inverted to obtain a velocity distribution using computer programme ZCW101C8. The details of the inversion technique are as follows. First the $dT/d\Delta$ data are smoothed by the Method of Summary Values, using either programme ZCW101C1 or ZCW101C4. It is difficult to find a velocity model that will fit smoothly on to the $dT/d\Delta$ curves for the lower mantle. This problem is overcome by choosing a velocity model with an accompanying $dT/d\Delta$ curve that will almost fit on to the smooth $dT/d\Delta$ curve to be inverted. Then two values of the $dT/d\Delta$ curve derived from the 'stripping' model that correspond to rays that have penetrated almost to the base of this model are chosen separated by a distance interval similar to the average spacing of the summary points; the second value is chosen with an abscissa very close to that of the first summary point. The next step is to
Table 4.8 Data to Illustrate the Smoothing of the Calculated Velocity Distribution on to the Velocity Model Assumed for the Upper Layers.

<table>
<thead>
<tr>
<th>Summary Points</th>
<th>dT/dΔ Curve Calculated from Stripping Model</th>
<th>Points Used to Interpolate dT/dΔ Curve</th>
</tr>
</thead>
<tbody>
<tr>
<td>Distance (deg)</td>
<td>dT/dΔ (sec/deg)</td>
<td>Distance (deg)</td>
</tr>
<tr>
<td></td>
<td>etc.</td>
<td>31.960</td>
</tr>
</tbody>
</table>
interpolate the curve to 0.2° intervals by the method of divided differences using the two points derived as explained above, together with the calculated summary points except the first. This procedure is illustrated by the data of Table 4.8. In this way it is ensured that the calculated velocity distribution fits smoothly on to the 'stripping' model. The tabulated velocities used to 'strip' the earth are read in, and the interpolated dT/dΔ curve is then 'stripped', using the power law technique, to obtain another dT/dΔ curve characteristic of an earth of radius \( r_0 - h \), where \( h \) is the depth at which the assumed model ends. The new dT/dΔ curve is then interpolated to equal distance intervals of 0.2°. Next the Herglotz-Wiechert integral is evaluated by Simpson's rule, thus giving a velocity and depth corresponding to each 0.4° interval of the 'stripped' dT/dΔ curve.

Although it might appear simpler and more efficient to 'strip' the summary points and interpolate afterwards, it was found that there are numerical difficulties owing to the very wide distance intervals between the 'stripped' summary points at small distances. The velocity distribution is finally interpolated to intervals of 5 km. To check the numerical values of the velocity model travel times and dT/dΔ values as a function of Δ are calculated by the power law technique. The programme was checked using the dT/dΔ data given in the 1968 Seismological Tables for P Phases stripped to a depth of 40 km. The original travel times were reconstructed from the derived velocity model to well within 0.1 seconds at a distance of about 80°.

4.6 DISCUSSION.

Two methods of deriving an average dT/dΔ curve for the distance range 27.9° to 98.8° have been described. In both of these techniques the ultimate aim is to fit the smoothest possible curve through the data and simultaneously to ensure that the smooth curve satisfies the \( \chi^2 \) test. This involved the somewhat arbitrary procedure of rejecting data that appeared to have considerable bias. It is not certain that this bias or systematic error for a particular event is due to irregularities in structure close to the array
site or in the source region. It is observed that there are specific portions of the \(dT/d\Delta\) curve within which the scatter of the data is great. This scatter might be due to velocity anomalies at the deepest point of the ray path, and therefore fitting a single smooth curve to the data so that \(d^2T/d\Delta^2\) is continuous everywhere may tend to smear out some important structural features. Consequently in Chapters 6 and 7 the possibility of triplications or other types of complexity in the \(dT/d\Delta\) curve, and of regional differences in structure, are examined. In Chapter 8 the velocity models derived using an assumed upper mantle velocity distribution to 'strip' the earth to a depth of 700 km are presented and compared with other published work. It was not possible to derive a reliable average upper mantle model from the \(dT/d\Delta\) data obtained over the distance range 13.1° to 26.8°. For this reason the construction of upper mantle models has not been considered in this chapter, but two regional models are presented in Chapter 5.

It is relevant here to comment briefly on the relative merits of the two versions of the Method of Summary Values. Version B has been found to be an improvement on version A, but it appears to break down if the data are unevenly distributed or are not capable of precisely defining a curvature. A combination of both methods in a single computer programme is now required so that a range of data within a series of ranges can be smoothed by either technique at will.
CHAPTER 5

UPPER MANTLE STRUCTURE

5.1 PREVIOUS WORK.

The upper mantle was defined in section 1.4, and is considered to extend from the base of the crust to a depth of about 700 km. The first study of upper mantle structure using an array was published by Niazi and Anderson (1965) who investigated the upper mantle structure below western North America using P arrivals from 70 events at distances between 10° and 30°, recorded at the original 10 km Tonto Forest array. There was appreciable scatter in their data, but they found that although dT/dΔ decreased considerably with increasing distance, two relatively abrupt changes were observed at distances close to 17° and 24°. These were interpreted as two second order discontinuities at depths of about 320 km and 640 km. They did not detect any azimuthal dependence in their dT/dΔ values, and therefore made no attempt to correct for local structure. However, the results of a seismic refraction survey (Warren, Roller and Jackson, 1965) indicated an irregularly dipping M-discontinuity in the vicinity of the array. Johnson (1967) obtained a velocity structure for P waves in the uppermost 750 km of the mantle, also below the western part of North America, derived from dT/dΔ measurements using the extended TFSO. Data from 52 earthquakes and explosions in the distance range 0° to 30° were used, and corrections for the effects of the crust below the array were applied using crustal models derived from the seismic refraction results of Warren et al. and gravity data. Two regions with high velocity gradients were found near depths of 400 km and 650 km, consisting of 9 to 10 per cent increases in velocity spread out over 50 to 100 km. Kanamori (1967) made measurements of apparent velocities of P waves for about 50 earthquakes recorded at the Wakayama Microearthquake Observatories in Japan over the distance range 5° to 55°. A velocity model was constructed which had a rapid increase in velocity at 375 km and a slight increase in velocity gradient between
640 and 770 km followed by an almost constant velocity down to 950 km.

Recent P wave velocity models derived from travel times from chemical and nuclear explosions, published by Green and Hales (1968), Lewis and Meyer (1968), Iyer, Pakiser, Stuart and Warren (1969) and Archambeau, Flinn and Lambert (1969), also show these two regions of high velocity gradients. The model of Lewis and Meyer extends only to a depth of 500 km and has a sharp discontinuity at 450 km. Green and Hales and Archambeau et al. also found evidence for regional differences in the low velocity layer below the United States and the western continental United States respectively. Evidence for two regions of rapidly changing velocities for P waves within the upper mantle had been found previously in several travel-time studies for different parts of the Earth, and have been compared and summarised in Table 4 of Johnson (1967). In particular Doyle and Webb (1963) studied travel times of P waves to Australian stations from Pacific nuclear explosions. They found bends in the travel-time curve at 19° and near 25° or 26°. These regions of high velocity gradients at depths close to 400 km and 650 km have been found in the S wave investigations of Ibrahim and Nuttli (1967) and Kovach and Robinson (1969), as well as in the S wave model derived by Anderson and Toksöz (1963) from observations of Love wave dispersion. The apparent discrepancies between surface wave and body wave models were examined and explained by Julian and Anderson (1968). Thus recent seismological work has established the existence of two major discontinuities in the upper mantle. Further, these discontinuities can be satisfactorily explained by phase transformations in the mineral phases likely to occur in the mantle, (see for example Ringwood, 1969a, b and Ringwood and Major, 1969). Regional differences in upper mantle structure below the North American continent have now been established, but comparatively little work has been accomplished in other parts of the world.

Direct measurements of travel-time gradients for events at distances less than 30° have only been made for the mantle beneath western North America and the Japan region. It is therefore important that P wave travel-time gradient
Figure 5.1 Epicentres used in investigating upper mantle structure.
studies at short distances should be attempted for the Australasian region, particularly the structurally complex region to the north of Australia. WRA is ideally situated for a study of this kind, and the objective in this chapter is to present some preliminary results and to see whether regional differences in upper mantle structure or any unusual features of the mantle can be detected. Unfortunately, however, the effect of the local structure beneath the array imposes severe limitations on the interpretation of the dT/dΔ data.

5.2 dT/dΔ AND AZIMUTH DATA.

A total of 110 measurements of dT/dΔ and azimuth of both first and second arrivals from 45 events over the distance range 13.1° to 26.8°, from two separate regions, has been obtained. These dT/dΔ and azimuth values have been derived in exactly the same manner as the data for lower mantle structure, and are listed together with the epicentre data in Tables 1 to 3 of Appendix 2. All event numbers quoted in this chapter and in the remainder of the thesis refer to those of Appendix 2. The epicentres shown in Figure 5.1 can conveniently be split into two regions: first those to the north west of the array and secondly those to the north east. An examination of the dT/dΔ and azimuth values shows that the P arrivals appear to be affected by local structures rather different from those affecting more distant events within the same azimuth range. Further, the work on the local structure described in Chapter 3 has shown that it would be futile to assume that any of the structures of Table 3.4 or Table 4.4 can be applied to the upper mantle data. To illustrate this point the azimuth anomalies for both first and second arrivals have been plotted in Figures 5.2 and 5.3 for comparison with Figures 3.5 to 3.7. In Figure 5.3 a different symbol has been used for points belonging to each separate branch of the travel-time curve; the cause and identification of these branches are discussed later. Since widespread regional differences in upper mantle structure probably occur, it is not justifiable at present to derive correcting structures from the dT/dΔ data itself, or to constrain the data to a travel-time curve, as was done for the lower mantle. Thus, although
Figure 5.2 Azimuth anomalies for Banda Sea earthquakes. (Standard errors vary between 0.6° and 1.2°.)
Figure 5.3 AZIMUTH ANOMALIES for first and second arrivals of events at distances between 18.0° and 26.8°. (Standard errors vary between 0.3° and 1.5°.)
Figure 5.4 \( \frac{dT}{d\Delta} \) measurements for first and second arrivals for events of regions A and B.
systematic errors in \( \frac{dT}{d\Delta} \) undoubtedly exist, there is no way of telling whether they are produced by structure at the array site, changes in structure along the transmission path, or irregularities in structure in the source region. In addition, for the structurally complex region to the north of Australia, epicentre mislocations and errors in focal depths may be important. It seems, therefore, that the only plausible way of correcting the \( \frac{dT}{d\Delta} \) data is by use of the travel-time residuals of events recorded at WRA.

(a) Data from Region A: Distance Range 13.1° to 26.8°; Azimuth Range 310.8° to 355.8°. (See Figure 5.1).

The 27 earthquakes are from the Banda and Flores Seas, the Molucca Sea, Halmahera, Celebes, the Talaud and Philippine Islands region and the West New Guinea region. For a study of this kind shallow events should really be used, owing to the difficulty of correcting epicentral distances for focal depth for deep focus earthquakes. One limitation imposed was that for close events from the Banda Sea area there were very few shallow earthquakes that gave clear \( P \) arrivals at the array; consequently some reasonably deep earthquakes had to be used. The events can be divided into three groups according to focal depth \( h \): (a) 9 with \( h \leq 60 \text{ km} \), (b) 9 with \( 60 \text{ km} < h \leq 130 \text{ km} \) and (c) 9 with \( h > 130 \text{ km} \). It is of course virtually impossible to obtain reliable focal depth corrections for those events at depths greater than about 60 km; all epicentral distances have nevertheless been corrected approximately to correspond to a surface to surface distance, using the upper mantle model used to define the 1968 Seismological Tables for \( P \) Phases at distances of less than 20°. The errors in these corrections could be larger than 2° for some Banda Sea events, but owing to other uncertainties any further refinements are not justifiable at present. No ellipticity corrections have been included.

The \( \frac{dT}{d\Delta} \) measurements are displayed graphically in Figure 5.4. In spite of the scatter of the data, it is clear that there are three distinct branches defined by the first and second arrivals; these are denoted \( ab, cd \) and \( ef \) in Figures 5.3 - 5.5. It is possible to ascribe each datum to one of these branches. Such an interpretation may be correct, but it seems unsatisfactory owing to the
Figure 5.5 Diagram to illustrate the possibility of additional branches of the travel-time curve between branches ab and cd and between cd and ef.
scatter of the data. Moreover this scatter can be explained if the presence of additional branches of the dT/dA curve between ab and cd and between cd and ef is postulated; these extra branches will be called pq and xy respectively as indicated in Figure 5.5. It is also postulated that branches pq and xy may be seen for some events near 19° and 24° but not for others, and that they can be the first arrivals for a distance range of a few tenths of a degree. The physical significance of this will be discussed later. The P arrival of event 224 and the second dT/dA measurement of event 225 were ascribed to branch pq. The dT/dA measurements for events 235, 236, 240 and two of those for event 234 were assumed to belong to branch xy. The dT/dA values believed to belong to either of these branches were ignored in the subsequent analysis.

A straight line was fitted by least squares through each of the segments ab, cd and ef; each value of dT/dA was given a weight proportional to the reciprocal of the square of its standard error. No data from earthquakes at depths greater than 225 km were used in fitting the straight lines, and dT/dA measurements from Banda Sea events 215, 216 and 221 were not used to define branch ab. A χ² test on each of the straight line segments gave the results listed in Table 5.1 (a). Owing to the scatter of the dT/dA data due to azimuthal variations in local structure, the total χ² for branch ab is excessively large; note that the straight line fit to branch cd is excellent. The J-B residuals for each branch are plotted in Figure 5.6, and a straight line has been fitted through the three branches ab, cd and ef in order to define approximately the total travel time at any point. Each residual has been ascribed to one of these three branches, for simplicity ignoring branches pq and xy.

The next step is to fit a velocity model to the data. Very little is known about the P wave velocities at the base of the crust and the top of the upper mantle in the Australasian region. P wave velocities for the upper mantle below the New Guinea-Solomon Islands region have been published by Brooks (1962), but little else has been accomplished. It is evident, however,
Table 5.1. Analysis of dT/\Delta Data.

(a) Straight Line Segments : Parameters and \( \chi^2 \) Tests.

<table>
<thead>
<tr>
<th>Region</th>
<th>Branch</th>
<th>Distance Range, deg</th>
<th>Equation</th>
<th>Total ( \chi^2 )</th>
<th>No. of Degrees of Freedom</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>ab</td>
<td>15.8 - 19.3</td>
<td>( p = -0.1009 \Delta + 14.435 )</td>
<td>26.5</td>
<td>7</td>
</tr>
<tr>
<td></td>
<td>cd</td>
<td>19.3 - 25.1</td>
<td>( p = -0.0717 \Delta + 12.382 )</td>
<td>32.0</td>
<td>27</td>
</tr>
<tr>
<td></td>
<td>ef</td>
<td>24.0 - 26.8</td>
<td>( p = -0.0549 \Delta + 10.998 )</td>
<td>5.5</td>
<td>7</td>
</tr>
<tr>
<td>B</td>
<td>ab</td>
<td>18.0 - 19.2</td>
<td>( p = -0.6510 \Delta + 24.937 )</td>
<td>7.4</td>
<td>6</td>
</tr>
<tr>
<td></td>
<td>cd</td>
<td>20.2 - 24.4</td>
<td>( p = -0.1664 \Delta + 14.387 )</td>
<td>43.9</td>
<td>14</td>
</tr>
<tr>
<td></td>
<td>ef</td>
<td>24.1 - 25.3</td>
<td>( p = -0.2156 \Delta + 14.846 )</td>
<td>1.1</td>
<td>4</td>
</tr>
</tbody>
</table>

(b) Corrections to the Three Branches of the dT/\Delta Curve Used in Deriving a Velocity Model for Region A.

<table>
<thead>
<tr>
<th>Branch</th>
<th>Distance Range, deg</th>
<th>Correction, sec/deg</th>
</tr>
</thead>
<tbody>
<tr>
<td>ab</td>
<td>15.8 - 19.3</td>
<td>-0.600</td>
</tr>
<tr>
<td>cd</td>
<td>19.3 - 25.1</td>
<td>-0.346</td>
</tr>
<tr>
<td>ef</td>
<td>24.0 - 26.8</td>
<td>-0.308</td>
</tr>
</tbody>
</table>

(c) \( \chi^2 \) Tests on Data Ascribed to Branches pq and xy.

<table>
<thead>
<tr>
<th>Region</th>
<th>Branch</th>
<th>dT/\Delta (Weighted Mean), sec/deg</th>
<th>Total ( \chi^2 )</th>
<th>No. of Degrees of Freedom</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>pq</td>
<td>11.390 ± 0.049</td>
<td>0.6</td>
<td>3</td>
</tr>
<tr>
<td></td>
<td>xy</td>
<td>10.160 ± 0.032</td>
<td>4.3</td>
<td>10</td>
</tr>
<tr>
<td>B</td>
<td>pq</td>
<td>12.044 ± 0.085</td>
<td>5.1</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>xy</td>
<td>9.835 ± 0.047</td>
<td>3.9</td>
<td>3</td>
</tr>
</tbody>
</table>
Figure 5.6 J-B residuals for three branches of the travel-time curve:
(a) region A, (b) region B.
from the J-B residuals from the Banda Sea events that provided the epicentre locations are not subject to large systematic errors, the upper mantle velocities below the northern part of Australia are fairly fast and characteristic of a stable region. It is clear from the raw \( \frac{dT}{d\Delta} \) data that a derived velocity model will not yield the correct travel times, so the travel-time residuals have been used to correct each branch of the \( \frac{dT}{d\Delta} \) curve. The straight lines fitted through the J-B residuals were used to define the travel times at the end points of the branches concerned, and then the base line of each straight line segment of the \( \frac{dT}{d\Delta} \) curve was adjusted to make the area beneath it equal to the travel-time difference between the end points, in exactly the same way as the lower mantle data were corrected. This procedure could not be applied to branch ef owing to the limited amount of observational data available. Consequently a correction was chosen to give the same fractional change in \( \frac{dT}{d\Delta} \) as for branch cd:

i.e. correction for branch ef

\[
= \text{correction for branch cd} \times \frac{\frac{dT}{d\Delta} \text{ at mid-point of ef}}{\frac{dT}{d\Delta} \text{ at mid-point of cd}}
\]

These corrections are listed in Table 5.1 (b).

The data belonging to branch ab were of too poor quality to enable a reliable velocity model to be derived. To strip the earth to a depth of 410 km, a model of the crust and upper mantle was chosen by trial and error to fit branch ab and the travel-time residuals approximately; this process was accomplished with the aid of programme ZCW101C6. The stripping model was made a little less arbitrary by stripping the earth to 87 km, using branch ab to derive a velocity distribution from 87 to 210 km by the Herglotz-Wiechert technique, and then incorporating these velocities in the stripping model itself. The data are also too sparse to enable anything to be deduced about the low velocity layer. No low velocity layer has been included in the stripping model at present, as the work of Green and Hales (1968) and Barr (1967) indicate that
Table 5.2. Velocity Model and Corresponding Travel-Time and dT/dA Curves for Region A.

(a) Velocity Model.

<table>
<thead>
<tr>
<th>(V_p) (\text{km/sec})</th>
<th>Depth (km)</th>
<th>(V_p) (\text{km/sec})</th>
<th>Depth (km)</th>
<th>(V_p) (\text{km/sec})</th>
<th>Depth (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>6.2000</td>
<td>0.0</td>
<td>9.6909</td>
<td>470.0</td>
<td>10.0321</td>
<td>600.0</td>
</tr>
<tr>
<td>6.3000</td>
<td>33.0</td>
<td>9.7078</td>
<td>475.0</td>
<td>10.0548</td>
<td>610.0</td>
</tr>
<tr>
<td>8.1365</td>
<td>33.0</td>
<td>9.7242</td>
<td>480.0</td>
<td>10.0776</td>
<td>620.0</td>
</tr>
<tr>
<td>8.1533</td>
<td>90.0</td>
<td>9.7402</td>
<td>485.0</td>
<td>10.1004</td>
<td>630.0</td>
</tr>
<tr>
<td>8.1630</td>
<td>120.0</td>
<td>9.7558</td>
<td>490.0</td>
<td>10.4194</td>
<td>630.0</td>
</tr>
<tr>
<td>8.1892</td>
<td>150.0</td>
<td>9.7709</td>
<td>495.0</td>
<td>10.4788</td>
<td>635.0</td>
</tr>
<tr>
<td>8.2263</td>
<td>180.0</td>
<td>9.7857</td>
<td>500.0</td>
<td>10.5129</td>
<td>640.0</td>
</tr>
<tr>
<td>8.2690</td>
<td>210.0</td>
<td>9.8002</td>
<td>505.0</td>
<td>10.5408</td>
<td>645.0</td>
</tr>
<tr>
<td>8.3417</td>
<td>310.0</td>
<td>9.8143</td>
<td>510.0</td>
<td>10.5653</td>
<td>650.0</td>
</tr>
<tr>
<td>8.5500</td>
<td>410.0</td>
<td>9.8282</td>
<td>515.0</td>
<td>10.5875</td>
<td>655.0</td>
</tr>
<tr>
<td>9.3589</td>
<td>410.0</td>
<td>9.8418</td>
<td>520.0</td>
<td>10.6081</td>
<td>660.0</td>
</tr>
<tr>
<td>9.4262</td>
<td>415.0</td>
<td>9.8551</td>
<td>525.0</td>
<td>10.6274</td>
<td>665.0</td>
</tr>
<tr>
<td>9.4652</td>
<td>420.0</td>
<td>9.8682</td>
<td>530.0</td>
<td>10.6457</td>
<td>670.0</td>
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Branch ef  

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*denotes distance values at end points of branches.


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Table 5.2 (Contd)

(c) Measured Travel Times.
Figure 5.7  P wave velocity models for the upper mantle: region A, region B and model CIT 208 of Johnson (1969).
such a layer is vestigial or absent beneath stable continental shields. Then a velocity distribution was calculated for branch cd using the Herglotz-Wiechert technique. Owing to the discontinuity in dT/dΔ between branches cd and ef, the complete velocity model was used to strip the earth again so that the Herglotz-Wiechert integral could be evaluated for branch ef. The parameters of the velocity model and the associated travel times and travel-time gradients are listed in Table 5.2, and the model itself is shown in Figure 5.7. In Figure 5.8 the actual travel times are compared with the times calculated from the model.

It is worth commenting that little meaning can be attached to the dT/dΔ measurements for branch ab since few of the events are shallow. The travel-time residuals between 13.1° and 20.2° suggest a reasonably high subcrustal velocity and a slow rate of increase of P velocity with depth. Because of this, the stripping model is such that surface to surface ray paths of 16.0° and 18.0° penetrate only to depths of 103 km and 144 km respectively. Therefore events at depths greater than 100 km and at epicentral distances less than 20° should have corrected distances much larger than the values obtained from the 1968 Seismological Tables for P Phases, if the stripping model is approximately correct. A remark on the technique of stripping the earth is also relevant at this stage. First, since no retrograde branches of the travel-time curve have yet been identified, the upper mantle transition layers have been assumed to be first order velocity discontinuities. The ab branch of the dT/dΔ curve of Figure 5.4 was extrapolated in order to give a maximum depth of penetration of 410 km and a P wave velocity of 8.55 km/sec above the discontinuity. This depth and the corresponding velocity were chosen by trial and error. Velocities of 8.55, 8.6 and 8.65 km/sec and depths between 350 km and 420 km were tried, and the values ultimately chosen gave the best fit to the travel times and the correct cross-over distance between branches ab and cd. Branch cd was extrapolated similarly to give depths of penetration between 600 km and 650 km when deriving the velocities below the lower discontinuity.
Table 5.3. Velocity Model and Corresponding Travel-Time and dT/dΔ Curves for Region B.

(a) Velocity Model.

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*denotes distance values at end points of branches.
Table 5.3 (Contd)

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<td>ab</td>
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<td>18.0</td>
</tr>
<tr>
<td>ef</td>
<td>330.1</td>
<td>25.3</td>
</tr>
</tbody>
</table>
Figure 5.8 A comparison of the travel times calculated from the velocity models derived for regions A and B with the actual travel times.
(b) Data from Region B: Distance Range $18.9^\circ$ to $25.3^\circ$; Azimuth Range $17.5^\circ$ to $67.2^\circ$.

The 18 earthquakes are from the north and north west coasts of New Guinea, New Ireland, New Britain and the Solomon Islands region, as shown in Figure 5.1. For this region it is easier to find good shallow events, but there are no data at distances less than $18.0^\circ$. In this case there are (a) 10 events with $h \leq 60$ km, (b) 7 with $60$ km $< h \leq 130$ km, and (c) 1 with $h > 130$ km. The method of analysing the data is the same as for region A, except that the $dT/d\Delta$ measurements were not used to define any parts of branch ab. Three distinct branches of the $dT/d\Delta$ curve can be identified, and once again a straight line has been fitted through each segment (see Figure 5.4). The relevant data are given in Table 5.1. It is also possible to reduce the scatter in the data by ascribing some of the points to branches pq and xy. The second arrival of event 242 and the first arrival of 249 have been assumed to belong to branch pq. The first arrival of event 257 and the second arrivals of 255, 258 and 260 have been ascribed to branch xy. The $dT/d\Delta$ data were not subject to as much systematic error as the values for region A, and no attempt was made to correct them. A velocity distribution was chosen somewhat arbitrarily by fixing the subcrustal velocity at $8.08$ km/sec, and adjusting the underlying velocities to give a total travel time close to 252.1 seconds at $18.0^\circ$. The derived velocity model shown in Figure 5.7 was used to generate the expected travel-time and $dT/d\Delta$ curves, and the results are displayed in Figure 5.8 and Table 5.3. In Figure 5.8 the total travel times calculated from the derived model for region B become progressively too long beyond $20^\circ$, thus showing that the $dT/d\Delta$ values need small corrections. Also the magnitude of the velocity anomaly near 600 km is very small. This effect is probably not real; it is most likely to be due to the poor control over the rate of decrease of the cd branch of the $dT/d\Delta$ curve caused by the scatter and paucity of the basic $dT/d\Delta$ data.
5.3 DISCUSSION.

The salient features of this short investigation of the upper mantle will now be considered. For both regions A and B, high velocity gradients or sharp velocity changes for P waves at depths of about 400 km and 620 km have been identified for the region of the mantle below the northern and north eastern edges of the Australian continent, but it has not been possible to deduce reliable P wave velocities owing to the complexity of the crustal structure at the array site. The velocity model derived by Johnson (1967, 1969) has been included in Figure 5.7 for comparison with the two upper mantle models derived in this study. The model for region B is obviously unsatisfactory.

It is important to consider whether the dT/dA data reveal differences in structure between regions A and B. The azimuthal differences in dT/dA for a specific distance range are most probably due to differences in local structure, and do not merit any further consideration. For region A, ef (or xy) is the first arrival beyond 24.0°, but for region B, cd appears to be the first arrival up to 24.4°. Thus the branch ef (or xy) may become the first arrival at a shorter distance for region A than for region B, though systematic errors in epicentre and focal depth determinations could easily account for this discrepancy. It is not possible with the limited data available to say much about the cross-over between branches ab and cd, except that ab is probably the first arrival beyond 19° for both regions. Other studies have generally suggested a slightly smaller distance (see Johnson, 1967). In Figure 11 of Johnson (1967) the cross-over is given at 17°. It is of some interest that branch ab has been positively identified by its dT/dA value as the first arrival for event 223 at a distance of 19.3°, since the second arrival 2.0 seconds later definitely belongs to branch cd. However, the cross-over distance between cd and ef given by Johnson does not differ significantly from the value found in this study. Therefore, unless Johnson has misinterpreted his data, there appear to be significant differences in mantle structure between
Table 5.4. dT/dΔ and Azimuth Data for All Events at Distances Close to 24° Illustrating the Possibility of the Presence of an Additional Branch in the Travel-Time Curve.

<table>
<thead>
<tr>
<th>Event No.</th>
<th>Azimuth, deg</th>
<th>Corrected Distance, deg</th>
<th>Measured Azimuth, deg</th>
<th>Measured dT/dΔ, sec/deg</th>
<th>Method</th>
<th>Branch</th>
<th>Separation of Phases, sec</th>
</tr>
</thead>
<tbody>
<tr>
<td>232</td>
<td>340.4</td>
<td>23.8</td>
<td>341.5 ± 0.7</td>
<td>10.622 ± 0.092</td>
<td>Peak A</td>
<td>cd</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>342.8 ± 0.6</td>
<td>10.752 ± 0.086</td>
<td>Peak B</td>
<td>cd</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>342.1 ± 0.6</td>
<td>10.640 ± 0.085</td>
<td>Trace</td>
<td>cd</td>
<td></td>
</tr>
<tr>
<td>233</td>
<td>326.4</td>
<td>24.0</td>
<td>333.3 ± 1.0</td>
<td>9.717 ± 0.108</td>
<td>Peak A</td>
<td>ef</td>
<td>0.6</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>327.7 ± 0.8</td>
<td>10.496 ± 0.100</td>
<td>Peak B*</td>
<td>cd</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>328.4 ± 1.0</td>
<td>10.487 ± 0.118</td>
<td>Second Zero*</td>
<td>cd</td>
<td></td>
</tr>
<tr>
<td>234</td>
<td>339.1</td>
<td>24.2</td>
<td>340.1 ± 0.9</td>
<td>10.461 ± 0.113</td>
<td>Peak B</td>
<td>cd</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>344.0 ± 0.7</td>
<td>10.227 ± 0.087</td>
<td>Peak C</td>
<td>xy</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>343.6 ± 0.8</td>
<td>10.183 ± 0.105</td>
<td>Trace</td>
<td>xy</td>
<td></td>
</tr>
<tr>
<td>235</td>
<td>329.3</td>
<td>24.3</td>
<td>330.6 ± 1.3</td>
<td>10.115 ± 0.150</td>
<td>Peak A</td>
<td>xy</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>330.0 ± 1.3</td>
<td>10.181 ± 0.153</td>
<td>Peak C</td>
<td>xy</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>330.0 ± 1.3</td>
<td>10.215 ± 0.154</td>
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<td></td>
</tr>
<tr>
<td>236</td>
<td>310.8</td>
<td>24.6</td>
<td>313.6 ± 0.6</td>
<td>10.197 ± 0.075</td>
<td>Peak A</td>
<td>xy</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>313.2 ± 0.8</td>
<td>10.170 ± 0.096</td>
<td>First Zero</td>
<td>xy</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>313.4 ± 0.7</td>
<td>10.200 ± 0.087</td>
<td>Trace</td>
<td>xy</td>
<td></td>
</tr>
<tr>
<td>237</td>
<td>327.5</td>
<td>24.7</td>
<td>333.5 ± 0.6</td>
<td>9.606 ± 0.070</td>
<td>Peak A</td>
<td>cd</td>
<td>0.6</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>326.5 ± 0.8</td>
<td>10.639 ± 0.103</td>
<td>Peak B*</td>
<td>ef</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>327.2 ± 1.0</td>
<td>10.511 ± 0.132</td>
<td>Second Zero*</td>
<td>ef</td>
<td></td>
</tr>
<tr>
<td>238</td>
<td>342.1</td>
<td>24.9</td>
<td>344.1 ± 1.0</td>
<td>9.426 ± 0.118</td>
<td>Peak A</td>
<td>ef</td>
<td>1.0</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>343.1 ± 0.7</td>
<td>10.772 ± 0.098</td>
<td>Peak B*</td>
<td>cd</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>344.0 ± 0.7</td>
<td>10.577 ± 0.099</td>
<td>Trace*</td>
<td>cd</td>
<td></td>
</tr>
<tr>
<td>239</td>
<td>326.5</td>
<td>25.1</td>
<td>330.8 ± 0.8</td>
<td>9.661 ± 0.091</td>
<td>First Zero</td>
<td>ef</td>
<td>1.3</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>331.7 ± 0.8</td>
<td>9.674 ± 0.096</td>
<td>Peak B</td>
<td>ef</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>330.8 ± 0.8</td>
<td>9.701 ± 0.091</td>
<td>Trace</td>
<td>ef</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>325.7 ± 1.0</td>
<td>10.733 ± 0.135</td>
<td>Peak A*</td>
<td>cd</td>
<td></td>
</tr>
</tbody>
</table>
Table 5.4 (Contd)

<table>
<thead>
<tr>
<th>Event No.</th>
<th>Azimuth, deg</th>
<th>Corrected Distance, deg</th>
<th>Measured Azimuth, deg</th>
<th>Measured dT/dΔ, sec/deg</th>
<th>Method</th>
<th>Branch</th>
<th>Separation of Phases, sec</th>
</tr>
</thead>
<tbody>
<tr>
<td>240</td>
<td>325.3</td>
<td>26.1°</td>
<td>324.1° ± 0.9</td>
<td>10.067° ± 0.110</td>
<td>Peak A</td>
<td>xy</td>
<td>1.0</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>325.2° ± 1.0</td>
<td>10.037° ± 0.123</td>
<td>First Zero</td>
<td>xy</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>325.1° ± 0.9</td>
<td>10.034° ± 0.114</td>
<td>Trace</td>
<td>xy</td>
<td></td>
</tr>
<tr>
<td>255</td>
<td>50.5</td>
<td>23.9</td>
<td>52.9° ± 0.4</td>
<td>10.471° ± 0.099</td>
<td>Peak A</td>
<td>cd</td>
<td>1.0</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>50.5° ± 0.4</td>
<td>9.996° ± 0.096</td>
<td>Peak B*</td>
<td>xy</td>
<td></td>
</tr>
<tr>
<td>256</td>
<td>40.0</td>
<td>24.1</td>
<td>40.6° ± 0.5</td>
<td>10.788° ± 0.138</td>
<td>Peak A</td>
<td>cd</td>
<td>1.5</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>40.7° ± 0.6</td>
<td>10.755° ± 0.145</td>
<td>Trace</td>
<td>cd</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>38.5° ± 0.4</td>
<td>9.609° ± 0.097</td>
<td>First Zero*</td>
<td>ef</td>
<td></td>
</tr>
<tr>
<td>257</td>
<td>55.9</td>
<td>24.3</td>
<td>55.4° ± 0.5</td>
<td>9.802° ± 0.120</td>
<td>Peak A</td>
<td>xy</td>
<td>1.6</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>59.2° ± 0.7</td>
<td>9.551° ± 0.175</td>
<td>Peak A*</td>
<td>ef</td>
<td></td>
</tr>
<tr>
<td>258</td>
<td>51.8</td>
<td>24.4</td>
<td>51.9° ± 0.4</td>
<td>10.203° ± 0.098</td>
<td>Peak B</td>
<td>cd</td>
<td>1.0</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>51.8° ± 0.4</td>
<td>9.755° ± 0.091</td>
<td>Peak A*</td>
<td>xy</td>
<td></td>
</tr>
<tr>
<td>259</td>
<td>58.3</td>
<td>24.5</td>
<td>58.2° ± 0.4</td>
<td>9.555° ± 0.107</td>
<td>Peak A</td>
<td>ef</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>58.4° ± 0.3</td>
<td>9.631° ± 0.083</td>
<td>First Zero</td>
<td>ef</td>
<td></td>
</tr>
<tr>
<td>260</td>
<td>67.2</td>
<td>25.3</td>
<td>67.0° ± 0.3</td>
<td>9.398° ± 0.063</td>
<td>Peak A</td>
<td>ef</td>
<td>1.3</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>67.3° ± 0.4</td>
<td>9.361° ± 0.085</td>
<td>First Zero</td>
<td>ef</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>68.0° ± 0.3</td>
<td>9.798° ± 0.083</td>
<td>Peak A*</td>
<td>xy</td>
<td></td>
</tr>
</tbody>
</table>

* denotes second phase.

# denotes a deep focus event for which the corrected distance may be seriously in error.

Events 232 to 240 and 255 to 260 belong to regions A and B respectively.
Figure 5.9 Array record for the Northern Celebes earthquake of 8.10.67 showing the arrival of two distinct phases at almost the same time. (Event 233). The output of R2 is inverted.
Figure 5.10 Array record for the Northern Celebes earthquake of 5.6.66 showing the arrival of two distinct phases at almost the same time. (Event 237).
Figure 5.11 Array record for the Northern Celebes earthquake of 16.7.66 showing the arrival of two distinct phases at almost the same time. (Event 239).
Figure 5.12 Map showing the region of deep focus Banda Sea earthquakes and the ray paths of the anomalous Northern Celebes events.
northern Australia and western North America.

There is an important feature of the results which may indicate lateral variations in structure below 600 km. Events 233, 237 and 239 all show two distinct arrivals separated by less than 1.5 seconds. In each case the branch ef is the first arrival, but with a significant azimuth anomaly of 4° - 7°; branch cd arrives from approximately the right azimuth. The results are listed in Table 5.4, and array records of events 233, 237 and 239 are shown in Figures 5.9 - 5.11; all three events are at almost the same epicentral distance, but at different focal depths. Since events 233 and 237 are almost exactly at the cross-over point it is evident that the focal depth determinations are unreliable. It is hard to see how such a rotation of one phase can be produced by the local structure beneath the array, but this cannot be dismissed entirely at present. There is, however, another important consideration which makes it reasonable to suggest that this rotation effect is produced near the deepest portion of the ray path for the first phase. At an azimuth close to 325° from WRA and at distances of about 14° to 18°, is the region of deep focus Banda Sea earthquakes (see Figure 5.12). The rays from the three Northern Celebes earthquakes will pass through the base of the deep focus earthquake zone. Thus it is suggested that the large anomalies in azimuth for these three events are produced when the waves pass through or pass close to this zone in which the phase transition region may be dipping or may vary laterally. There is, however, an apparent inconsistency: Event 240 from the Banda Sea, at a depth of 656 km and an azimuth of 325.3°, does not show this effect, though an error in focal depth of 100 km would mean that the rays would not penetrate deep enough to reach the phase transformation region. Evidence for such an error is provided by the dT/dΔ values for the first arrival, which are characteristic of the branch xy discussed below. Clearly more work is required on events within this narrow azimuth range to ascertain whether the explanation of the large azimuth shift for branch ef is reasonable. It does seem that there are serious errors in the U.S.C.G.S. focal depth determinations for events in the Banda Sea, Celebes and surrounding regions.
Table 5.5. List of Values of the Ratio \((dT/d\Delta)_{1} / (dT/d\Delta)_{2}\)
for Both Regions A and B.

<table>
<thead>
<tr>
<th>Event Number</th>
<th>((dT/d\Delta)<em>{1} / (dT/d\Delta)</em>{2})</th>
<th>Corrected Distance, deg</th>
<th>Azimuth, deg</th>
</tr>
</thead>
<tbody>
<tr>
<td>233</td>
<td>1.08</td>
<td>24.0</td>
<td>326.4</td>
</tr>
<tr>
<td>237</td>
<td>1.10</td>
<td>24.7</td>
<td>327.5</td>
</tr>
<tr>
<td>238</td>
<td>1.18</td>
<td>24.9</td>
<td>342.1</td>
</tr>
<tr>
<td>239</td>
<td>1.11</td>
<td>25.1</td>
<td>326.5</td>
</tr>
<tr>
<td>255</td>
<td>1.05</td>
<td>23.9</td>
<td>50.5</td>
</tr>
<tr>
<td>256</td>
<td>1.12</td>
<td>24.1</td>
<td>40.0</td>
</tr>
<tr>
<td>257</td>
<td>1.03</td>
<td>24.3</td>
<td>55.9</td>
</tr>
<tr>
<td>258</td>
<td>1.05</td>
<td>24.4</td>
<td>51.8</td>
</tr>
<tr>
<td>260</td>
<td>1.05</td>
<td>25.3</td>
<td>67.2</td>
</tr>
</tbody>
</table>
It is now pertinent to examine the problem of whether the hypothetical branches of the travel-time curve, pq and xy, are real, and to provide an explanation. Suppose it is assumed that the relative changes in $dT/d\Delta$ for branches cd and ef are small over a distance range of about 1° on either side of the cross-over. For regions A and B the ratio $(dT/d\Delta)_1/(dT/d\Delta)_2$ has been calculated for each event for which two distinct phases were discernible, and the results are shown in Table 5.5. The suffices 1 and 2 refer to the phases with higher and lower $dT/d\Delta$ values respectively. The ratio is between 1.08 and 1.18 for the events of region A and event 256 at an azimuth of 40° in region B; the average change in $dT/d\Delta$ at the cross-over point is 10%. However, for the remaining four events of region B, at azimuths greater than 50°, the ratio is between 1.03 and 1.05, giving an average change in $dT/d\Delta$ at the cross-over of about 5%. Thus the $dT/d\Delta$ contrast between the two branches is considerably less for events at azimuths between about 50° and 70° from WRA. This could suggest regional variations in the zone of high velocity gradients near 620 km, but another interpretation seems more probable.

The fractional change in $dT/d\Delta$ in moving from branch cd to ef appears to be twice as great for region A. This, taken in conjunction with the $dT/d\Delta$ values for events 255, 257, 258 and 260, suggests that for region B there is an intermediate branch xy which is the first arrival over a very small distance interval of a few tenths of a degree, but is a clear second arrival at other distances. This is illustrated in Figure 5.5, and this effect would be produced by two distinct, but fairly closely spaced, discontinuities. It is also significant that for region A the data ascribed to branch xy are first arrivals for three events. An average value of $dT/d\Delta$ for branch xy is listed in Table 5.1 for both regions. It is therefore suggested that at a depth of about 620 km, for both regions A and B, there are two closely spaced discontinuities which give rise to an additional branch xy in the travel-time curve; why the intermediate branch should be seen for some events and not for others is not yet clear. In addition the large azimuth shift of the first
Figure 5.13  Suggested model of the upper mantle showing discontinuities at 320, 400, 600 and 680 km.
Figure 5.14 Travel-time and dT/dΔ curves for the velocity model of Figure 5.13.
arrival of the Northern Celebes events of Table 5.4 may be related to this phenomenon. The evidence for two distinct transition layers near 400 km is more tenuous.

A velocity model has been chosen with first order discontinuities at 320, 400, 600 and 680 km to illustrate the way in which the extra branches pq and xy are produced. The velocity model and the corresponding travel-time and p-Δ curves are shown in Figures 5.13 and 5.14. Branches pq and xy are never the first arrivals, but a few minor adjustments of the parameters of the model could make both pq and xy the first arrivals over a small distance interval. An alternative explanation of the appearance of the branch pq might be that the first phase, say, for event 249 has been missed, and that the second arrival is really the third arrival belonging to the retrograde branch bc; this is improbable, and definitely cannot be the explanation of the observations of branch xy.

The two major discontinuities in the upper mantle at depths of about 400 km and 620 km have been satisfactorily explained as phase changes (Anderson, 1967a, Ringwood, 1969a, b, Ringwood and Major, 1969). But at present the dT/dΔ observations do not enable one to say whether the discontinuities are of first or second order. Petrological evidence suggests that the discontinuity near 400 km may be associated with two important phase transformations: first the pyroxene-garnet transformation and secondly the olivine-spinel transformation (Ringwood, 1969a, b). The second of these has recently been shown to be capable of producing a first order seismic discontinuity at a depth of about 400 km when the olivine and spinel react to form a beta phase (Ringwood and Major, 1969). It is also likely that the discontinuity at a depth of about 620 km is of first order; Engdahl and Flinn (1969) have observed precursors to waves of the PKPPKP type in the distance range 55° to 75°, and attribute them to reflections from the lower side of this discontinuity. The ratio of dT/dΔ values for adjacent branches of the dT/dΔ curve should ultimately show whether there are in fact two regions of sharply changing velocity gradients either near depths of 400 km or near 600 km. Two discontinuities
Figure 5.15 The effect of focal depth on the separation of two closely spaced phases produced by a region of high velocity gradients. (a) surface focus, (b) depth h.
near 400 km are to be expected, and it is likely that a complex series of phase changes occurs around the deeper discontinuity (Ringwood, 1969a, b). The problem of the sharpness of the velocity changes can be examined by trying to detect reflections from them using the correlation technique described in Chapter 2, and this is discussed further in Chapter 9.

It is relevant to remark that it is very difficult to obtain any useful $dT/d\Delta$ data at WRA for the region of the mantle shallower than 300 km, owing to the absence of natural earthquakes at distances less than $17^\circ$ that give clear P arrivals. The only region closer to WRA than $17^\circ$ for which earthquakes are often reported on the U.S.C.G.S. preliminary determination of epicentre cards is the Banda Sea and adjacent area. Most events from this region are either deep or give very complex P wave arrivals for which it is not easy to obtain reliable $dT/d\Delta$ and azimuth values. Explosion studies are therefore required to elucidate the structure of the uppermost regions of the upper mantle beneath the northern part of Australia. Very recently earthquakes have occurred in central Australia in a region where seismic activity was completely unknown, and the P arrivals at WRA will provide $dT/d\Delta$ measurements at distances of about $7^\circ - 10^\circ$.

5.4 THE EFFECT OF FOCAL DEPTH ON THE RELATIVE DISPLACEMENT OF TWO CLOSELY SPACED BRANCHES OF THE TRAVEL-TIME CURVE.

There is an important effect produced at a recording station by a deep focus earthquake, when two distinct P phases associated with a region of high velocity gradients are observed. Referring to Figure 5.15, the T-$\Delta$ and p-$\Delta$ curves will look slightly different because of the difference $GH$ in corrected distances for the two phases and the difference in travel times along portions FA and F'A in (a) and (b). It is therefore useful to calculate the magnitude of this effect. As an example consider a surface focus earthquake recorded at a distance of $19.75^\circ$, and let the crust and upper mantle for the ray path correspond to the velocity model of Table 5.2. In Table 5.6 the parameters of the first and second arrivals are 10.62 and 13.04 sec/deg
Table 5.6. dT/dA and Travel-Time Values for Two Phases Recorded at the Same Distance.

(a) Surface Focus.

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<th>Δ, sec</th>
<th>Separation of Phases, sec</th>
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(b) h = 120 km.

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respectively, and the phases are separated by 1.01 seconds. However, if a first phase of parameter 10.62 sec/deg is observed from an event at a depth of 120 km, the second arrival has a parameter of 12.89 sec/deg and arrives 2.31 seconds later. The focal depth alone causes a relative displacement of the two phases, relative to a fixed $dT/d\Delta$ for the first arrival, of 1.3 seconds. This also means that the cross-over point with respect to the surface distance for branch cd is reduced by 0.5° from 19.3° to 18.8°. If a similar calculation is performed for an event at a depth of 300 km, the relative shift of the two phases associated with the 630 km discontinuity, at a distance of 24° - 25°, is negligible (about 0.1 seconds). This will not be true for an event occurring well below the 410 km discontinuity.

This discussion gives rise to two ideas which have not yet been fully investigated. It is suggested that this focal depth effect which results in a relative shift of the different branches of the $dT/d\Delta$ curve could in principle be used to resolve the fine details of upper mantle structure, provided that accurate focal depth and epicentre determinations can be obtained. This phenomenon will be particularly significant for events occurring just above one of the upper mantle discontinuities. In addition, for an event at a depth just below the 410 km discontinuity and at such a distance from a recording station that the ray path reaching the surface leaves the source horizontally or almost horizontally, only the branch cd will be seen. If there are two discontinuities at a depth close to 400 km, it might be possible to see branch pq as a first arrival by choosing an event at the right focal depth and epicentral distance from WRA. Banda Sea earthquakes could well provide the required information. These ideas are unfortunately difficult to apply in practice, since reasonably accurate upper mantle velocities are required before accurate earthquake locations and focal depth determinations can be obtained.
Figure 6.1  Distribution of epicentres of earthquakes at corrected distances between 27.9° and 42.0° from WRA.
CHAPTER 6
THE LOWER MANTLE

PART 1: THE REGION BETWEEN 700 KM AND 1000 KM (28° < Δ < 42°).

6.1 INTRODUCTION.

Perhaps the most interesting and at present the most controversial part of the mantle is the region between depths of 700 km and 1000 km. Earthquakes have not been recorded below about 700 km, and evidence from nuclear explosion studies suggests that the mantle has become comparatively homogeneous at depths of around 800 km. Moreover, it is certain that there is an anomalous region close to 800 km, and explanations of its nature given in recent array investigations are in conflict. Johnson (1969) finds rapid increases in P velocity near 830 km and 1000 km, and Corbishley (1970) an increase between 850 and 900 km. Chinnery and Toksöz (1967) and Greenfield and Sheppard (1969), however, favour low velocity gradients or possibly a decrease in P velocities with increasing depth near 800 km. Evidence that the second interpretation is more probable was advanced by the writer (Wright, 1968). Conventional travel-time investigations and amplitude studies have also suggested the presence of anomalous regions between 700 and 1000 km. Bugayevski (1964) has presented evidence of a discontinuity in the travel-time curve at 36° - 37° suggesting a low velocity zone at a depth of about 900 km. Evidence for increased curvature in the travel-time curve near 40° indicating rapid changes in P wave velocity at a depth close to 1000 km has been put forward in the nuclear explosion studies of Carder (1964) and Archambeau et al. (1969). Gutenberg (1959, p. 95) found a minimum in the P amplitude curve close to 35°. Since, over a limited distance range, the amplitude is proportional to $\sqrt{|d^2 T/d\Delta^2|}$, a flattening of the $dT/d\Delta$ curve should coincide with a minimum in the amplitude curve, unless the first arrival corresponds to two or more closely spaced branches of the $dT/d\Delta$ curve. Carpenter, Marshall and Douglas
(1967), however, found a sharp peak in the amplitude curve between $33^\circ$ and $36^\circ$ followed by a slight minimum between $36^\circ$ and $39^\circ$.

In view of the outstanding interest of the region of the $dT/d\Delta$ curve between distances of $30^\circ$ and $40^\circ$ and the small amount of data over this distance interval used in other array studies, well over one third of the lower mantle $dT/d\Delta$ measurements are within the distance range $28^\circ$ to $42^\circ$. Exhaustive $dT/d\Delta$ measurements from earthquakes of the Caroline and Mariana Islands and surrounding regions have been compared with similar data from the Solomon, Santa Cruz, New Hebrides and Loyalty Islands regions, and also with less copious data from other areas. For each region or group of $dT/d\Delta$ data in this chapter and in Chapter 7, the corrected $dT/d\Delta$ profiles obtained by both the preliminary and the refined methods of sections 4.3 and 4.4 have been discussed. In this way any inconsistencies and ambiguities in interpretation can be demonstrated, since slightly different assumptions have been made in preparing the two versions of the corrected profiles. There are difficulties in the interpretation of the data owing to the occurrence of an apparent increase in $dT/d\Delta$ with increasing distance beyond $32^\circ$ for one azimuth range. Further, $\chi^2$ tests on the smooth curves fitted through the $dT/d\Delta$ values show that for this distance range, particularly close to $40^\circ$, the scatter of the data is much greater than one would expect if the observational points were normally distributed about the curve. It is not known whether this effect is caused largely by the structure near the array site, but some of the scatter can be attributed to complexity in structure at the deepest point of the ray path. Evidence in favour of this interpretation is provided by $dT/d\Delta$ values at larger distances, which generally show a little less scatter; further, the scatter appears to be concentrated into specific distance ranges. Thus, in this chapter and in Chapter 7, large scatter has been assumed to be diagnostic of complexity in the behaviour of the $dT/d\Delta$ curve due to real anomalies at great depths, unless supplementary evidence suggests an alternative explanation.
Figure 6.2  Corrected $dT/d\Delta$ data for first arrivals from earthquakes of the Mariana and Caroline Islands and surrounding regions. Triangles denote first arrivals not used in fitting the smooth curves; open circles and triangles denote data from events at depths greater than 130 km.
Using simple ray theory, several velocity models that might explain the observed distribution of $dT/d\Delta$ values for specific distance ranges have been presented; these models contain low velocity layers at depths close to 800 km, but the results can be applied to similar anomalies at greater depths. Finally, evidence for regional differences in structure and possible petrological and tectonic implications have been discussed.

6.2 DATA FROM REGION C: EARTHQUAKES OF THE MARIANA AND CAROLINE ISLANDS AND SURROUNDING REGIONS.

AZIMUTH RANGE 9.9° TO 25.6°. (See Figure 6.1 and Appendix 2).

The two $dT/d\Delta$ profiles shown in Figure 6.2 are from (a) groups 2A and 2B of Table 4.1, and (b) profile 2R of Table 4.5. There are two important differences in the methods of correcting the data. In the preliminary work, owing to the increase in the measured values of $dT/d\Delta$ between about 31.8° and 34.1°, the data were corrected in two separate distance ranges: 31.8° to 37.5° and 39.2° to 46.4°. In the final analysis the data were corrected to first order using an average structure for the distance range 32.6° to 47.9°. Also events 39, 41 and 42 have been placed in profile 4 of Table 4.5, since the opposite sense of the azimuth anomalies for these three events shows that they have been affected by a rather different local structure (see also section 3.2 and Table 3.1). However, the most important difference between the two profiles of Figure 6.2 is in the distance interval 37° to 39°. All the original $dT/d\Delta$ measurements for region C are systematically too large, and the correction for group 2A of the preliminary $dT/d\Delta$ curve is 0.1 sec/deg greater than for group 2B. Therefore the final $dT/d\Delta$ profile is steeper than the preliminary profile between about 36° and 40°. Otherwise there is little difference between the curves prepared by the two methods. It should be remarked that the $dT/d\Delta$ data of Figure 6.2 (b) have not been corrected in the same manner as the data of Figure 1 of Wright (1968). Nevertheless, the differences in the corrections are quite small. The values of $dT/d\Delta$ show considerable scatter, especially beyond 40°, which is reduced when the deeper focus events are
Figure 6.3: J-B residuals for events used in Wright (1968).

- RESIDUAL = ARRIVAL TIME AT ARRAY - J-B TRAVEL TIME - ORIGIN TIME
- DISTANCE, deg.
- RESIDUAL, sec.

○ DENOTES EVENTS OCCURRING AT DEPTHS GREATER THAN 130 KM
removed. The general similarity of the curves of Figure 6.2 indicates that the scatter does not appear to be significantly azimuth dependent. The remarkable feature of the results is that $\frac{dT}{d\Delta}$ increases with increasing distance from $32.8^\circ$, reaching a maximum at about $34.1^\circ$; then $\frac{dT}{d\Delta}$ starts to decrease again beyond $35^\circ$. The effect of crustal corrections on the maximum in the $\frac{dT}{d\Delta}$ curve is negligible. Thus there is an anomalous region of travel-time gradients between $32.8^\circ$ and $35.0^\circ$. The important consideration is that the $\frac{dT}{d\Delta}$ increases with increasing distance the Herglotz-Wiechert method of deriving velocity-depth distributions breaks down. There is no positive indication that the increase is due to structure near the array site.

**Provisional Interpretation.** The first attempt to explain the data of Figure 6.2 was published by Wright (1968), and the basic idea is as follows. If the $\frac{dT}{d\Delta}$ curve flattens over a limited distance range, it is clear that the P wave velocity must either increase very slowly or decrease slightly with increasing depth. Further, if the velocity increases with depth continuously and sufficiently rapidly, $r/v$ increases with depth (where $r$ is the distance from the centre of the earth and $v$ the velocity), and the Herglotz-Wiechert method formally breaks down. It thus appears that the increase in $\frac{dT}{d\Delta}$ with increasing distance can be interpreted in terms of some kind of low velocity layer close to 800 km. Dowling and Nuttli (1964) have indicated that a significant low velocity layer may manifest itself either as a shadow zone or as an overlap of two distinct branches of the T-\(\Delta\) curve. The deviations of the travel times from the J-B times are shown in Figure 6.3. There is certainly no definite evidence of a break in the travel-time curve, although three events close to $33^\circ$ have late arrivals with respect to J-B, whereas all others at distances less than $40^\circ$ tend to arrive early. Each event between $32^\circ$ and $36^\circ$ was examined for second arrivals. Three events showed what might be a second arrival about two seconds after the P onset.

In order to explain the increase in $\frac{dT}{d\Delta}$, together with the absence of any significant change in J-B residual or any second arrivals, it is first assumed that the rate of increase of velocity with increasing depth below the low velocity
Figure 6.4 Sketch of T-A curves for a wedge-shaped low velocity layer near 800 km that becomes thicker as the distance from the recording station increases.
layer is sufficiently great to cause two overlapping branches of the T-Δ curve.
Following Dowling and Nuttli, the travel-time curves for low velocity layers
of differing thicknesses would appear as in Figure 6.4, where lines A and D
correspond to the thinnest and thickest layers respectively. Now suppose
that the low velocity layer near 800 km becomes thicker towards the north.
Then the travel times between 32° and 36° might appear as shown by points
1 to 6 in Figure 6.4. From Figure 6.2(b) the increase in dT/dΔ between 32.6°
and 34.1° is about 0.26 sec/deg spread over 1.5°; for simplicity this increase
is taken as the maximum possible value, 0.3 sec/deg. Assuming this occurs
over the distance interval represented by points 1 to 4 in Figure 6.4, the
total offset of the travel-time curve between point 4 and the continuous curve
XYZ is less than 0.45 sec. Between points 4 and 6, dT/dΔ starts to decrease
again in the normal fashion. Note that the slope at any point on line D is less
than that of the tangent at Y. Let the J-B residual at Y be t. Between 32.6°
and 34.3° the J-B dT/dΔ curve decreases by only 0.08 sec/deg (see Figure 1
of Wright, 1968). Then at point 4 in Figure 3, the J-B residual will be less
than t + 0.6 sec. Thus the increase in dT/dΔ found by the array measurements
is far too small to be detected by measuring travel times. It appears that the
low velocity layer responsible for these results is either very thin or involves
a fairly small velocity decrease; further, the increase in dT/dΔ with increasing
distance is adequately explained if the layer becomes thicker towards the
north. The ray paths for different forms of low velocity layer can be complex,
and before considering alternative explanations of the increase in dT/dΔ with
increasing distance, it is important to examine the T-Δ and dT/dΔ curves
produced by low velocity layers at depths of about 800 km.

6.3 T-Δ AND P-Δ CURVES FOR LOW VELOCITY LAYERS.

Some time after the publication of my evidence for a slight low
velocity zone near 800 km, two dT/dΔ curves showing a flat region beyond 30°
were published by Greenfield and Sheppard (1969). Moreover, one of these
curves increases slightly with increasing distance, thus giving strong support
for the idea that the apparent increase in \( \frac{dT}{dA} \) with increasing distance was caused by structure at the deepest point of the ray path. In addition, no increase or decrease in \( \frac{dT}{dA} \) was found between distances of 30.9° and about 33° for P arrivals from earthquakes to the east of WRA. Therefore it was important to look for an explanation of an increase in \( \frac{dT}{dA} \) that does not invoke the rather artificial postulate of a layer that is of variable thickness. Consequently, a detailed investigation of possible ray paths for different forms of low velocity layer near 800 km was undertaken using computer programme ZCW101C6 described in Chapter 4. Before the widespread use of high speed digital computers, the only type of low velocity layer discussed in seismological literature was one that produced a shadow zone. Dowling and Nuttli (1964), however, showed that a low velocity layer need not produce a shadow zone, but could produce two (or three) overlapping branches of the T and \( \frac{dT}{dA} \) curves, provided that the rate of increase of velocity below the low velocity layer was sufficiently large. My own examination of the problem of low velocity layers has not only demonstrated that there are important forms of the T-Δ and p-Δ curve that had not been considered previously, but has also enabled an explanation of an apparent increase in \( \frac{dT}{dA} \) to be provided.

The procedure was as follows. The velocity distribution given at 5.0 km intervals in the 1968 Seismological Tables for P Phases was perturbed by inserting low velocity layers of different kinds, in each case starting at a depth of 800 km. This at first required the introduction of first or second order discontinuities at two or three depths. These trial models were used to give T-Δ and p-Δ curves, and if these curves were of interest, the unwanted discontinuities were removed by smoothing the perturbed portion of the velocity model on to the unperturbed portion by fitting a Lagrange interpolation polynomial through selected points on either side of the contact.
Table 6.1. Parameters of Earth Models Incorporating a Low Velocity Layer at a Depth of about 800 km. (Velocities above 780 km are those of the 1968 Seismological Tables for P Phases).

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Table 6.1 (Contd)

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Figure 6.5 Case 1. Sharp decrease in velocity followed by a steady increase.
Figure 6.6 Case 2. Sharp decrease in velocity followed by a rapid increase without first or second order discontinuities.
Case 1. (See Figure 6.5 and Table 6.1).

**Sharp Decrease in Velocity Followed by a Steady Increase.**

A first order discontinuity involving a decrease in P wave velocity of 0.036 km/sec has been introduced at a depth of 790 km. The velocities beyond 855 km are those of the 1968 Seismological Tables for P Phases. The T-Δ and p-Δ curves correspond to the case usually discussed in the standard texts (see for example Bullen, 1963, pp. 117-119, Kaula, 1968, pp. 83-84). There is a shadow zone between 31.6° and 33.3°. There are very low amplitudes along the retrograde branch cd, and a caustic or focus occurs at about 33.3°. Therefore with this model it is unlikely that the retrograde branch will have any effect on the observed values of dT/dΔ, even though the separation of branches cd and de is less than 0.3 seconds below 35°.

Case 2. (See Figure 6.6 and Table 6.1).

**Sharp Decrease in Velocity Followed by a Rapid Increase without First or Second Order Discontinuities.**

At 800 km there is a sharp decrease in P wave velocity of 0.0766 km/sec, and the subsequent rate of increase is greater than in Case 1; \[
\frac{r}{v} \frac{dv}{dr}
\]
reaches - 4.8 between 835 and 837.5 km. This has the effect of pulling the retrograde branch cd back towards ab, so that there are three overlapping branches of the T-Δ and p-Δ curves. Consider the p-Δ diagram. Branch ab ends at 32.3° at which the corresponding ray of parameter 8.779 sec/deg penetrates to a depth of exactly 800 km. Then as p decreases further there is a retrograde branch cd extending back to 32.0°. The most important features are the larger amplitudes (higher \( d^2T/dΔ^2 \)) of the retrograde branch cd in comparison with de between 32.0° and about 34.5°, and the small separation of branches cd and de (less than 0.3 seconds between 31.5° and 34.7°). Thus, if such a low velocity layer existed in the earth, the retrograde branch would contribute appreciably to the measured value of dT/dΔ for the first arrival.
Figure 6.7 Case 3. Sharp decrease in velocity followed by a constant velocity which ends with an abrupt increase.
Figure 6.8 Case 4. Second order discontinuity at 800 km. Rate of decrease of velocity above the critical rate followed by fairly rapid positive velocity gradients.
Figure 6.9 Case 5. As for case 4 except that the rate of decrease of velocity is below the critical rate.
In addition, the offset between branches ab and cd or de is about 0.6 seconds, and therefore the two arrivals would be difficult to separate.

Case 3. (See Figure 6.7 and Table 6.1).

**Sharp Decrease in Velocity Followed by a Constant Velocity which Ends with an Abrupt Increase.**

This model, which has a sharp velocity decrease of 0.0766 km/sec at 800 km and a sharp increase of 0.1726 km/sec at 850 km, is similar to those discussed by Dowling and Nuttli (1964). It yields the two overlapping branches of the T-Δ and p-Δ curve between distances of 30.9° and 32.3° that were invoked to explain the apparent increase in dT/dΔ near 33°. The offset between branches ab and de is about 0.9 sec at 32.2°. It is worth noting that a retrograde branch exists, but only between distances of 45.4° and 44.3°.

Case 4. (See Figure 6.8 and Table 6.1).

**Second Order Discontinuity at 800 km. Rate of Decrease of Velocity above the Critical Rate Followed by Fairly Rapid Positive Velocity Gradients.**

The T-Δ and p-Δ curves are basically no different from those of Case 2. With the parameters chosen, the caustic occurs at 32.9°. Thus there is a slight shadow zone extending over a distance of 0.6°. If such a shadow zone really exists near 33° it would be difficult to identify. As in Case 2 the amplitudes along the retrograde branch predominate between the caustic and about 34.5°.

Case 5. (See Figure 6.9 and Table 6.1).

As for Case 4 except that the Rate of Decrease of Velocity is below the Critical Rate.

Although there is never a break in the T-Δ or p-Δ curves, the amplitudes on the forward branch bc are so small that no detectable waves beyond 32.3° are likely to be produced, except very close to the point c at 43.4°. The retrograde branch ends at 32.5°. Other features of the curves are similar
to Cases 1 and 4. The T-Δ curve shows a triplication, and is therefore fundamentally no different from the case of an increase in velocity gradients without a low velocity layer.

There are two reasons for presenting these velocity models at this stage. First, these models incorporating low velocity layers produce two or three branches of the T-Δ and p-Δ curves that are separated by less than one second over the distance range of interest; thus if such low velocity layers really exist, the individual branches could not be resolved by visual examination of normal seismograms. Further, it has been shown that the retrograde branches produced can either mask or be indistinguishable from the forward branches that are theoretically the first arrivals. Secondly, the results show that the T-Δ and p-Δ curves for possible velocity anomalies in the earth are more closely related to one another than the standard literature on seismic ray theory would lead one to believe. The T-Δ and p-Δ curves of Figures 6.5 to 6.9 can all be regarded as variants of a single standard velocity anomaly - the triplication produced by rapid increases of velocity gradients. The low velocity layers involving sharp velocity decreases or rates of decrease of velocity above the critical rate yield T-Δ and p-Δ curves whose forms can be obtained merely by removing portions of the standard p-Δ and T-Δ curves for a significant increase in velocity gradients. This phenomenon is illustrated by the sequence of p-Δ and T-Δ curves of Figure 6.10. It is considered important to look at the problem of velocity anomalies in this way for the following reasons; first the gradual transition from Case 1 of Figure 6.5 to the standard triplication of Figure 6.10 then becomes apparent, and secondly, there is evidence that the type of velocity anomaly occurring in the lower mantle is intermediate between these two extremes. Thus the dT/dΔ data discussed in this chapter and in Chapter 7 need to be considered in terms of pairs of anomalies: low velocity gradients followed by high velocity gradients or abrupt velocity increases.
- - - - denotes low amplitudes
A sequence of $T-\Delta$ and $p-\Delta$ curves to illustrate the gradual change in form of these curves as the parameters of the velocity anomalies responsible are varied. $v$ denotes velocity and $h$ depth. The corresponding points on successive $T-\Delta$ and $p-\Delta$ curves are denoted by the letters a, b, c and d. The six cases considered are as follows:

1. High velocity gradients.
2. Sharp increase in velocity.
3. Low velocity layer. Rate of decrease below critical rate followed by rapid increase.
4. Low velocity layer. As for (c) but decrease sharp or above the critical rate.
5. Low velocity layer. Sharp decrease in velocity or rate of decrease above the critical rate followed by a sharp increase.
6. Low velocity layer. Sharp decrease in velocity or rate of decrease above the critical rate followed by a gradual increase.
Table 6.2. Parameters of a Velocity Model that Will Partly Explain the Increase in $dT/d\Delta$ Shown in Figure 6.2.

<table>
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All velocities above 790 km and below 905 km correspond to those of model ANUW1 given in Table 8.1.
Figure 6.11 T-Δ and p-Δ curves for a velocity model that will partly explain the dT/dΔ data of Figure 6.2.
Figure 6.12 Diagram to show how an apparent increase in $dT/d\Delta$ can be produced near 32.8°. Dotted line shows how a smooth curve through observed $dT/d\Delta$ values might appear.
6.4 DATA FROM REGION C: A POSSIBLE VELOCITY MODEL

A velocity model that will explain an observed increase in \( \frac{dT}{d\Delta} \) has been constructed in the following way. A low velocity zone starting with a first order discontinuity at 797 km has been inserted into the preliminary velocity model ANUW1 (see Chapter 8) in the same manner as the low velocity layer models discussed in section 6.3. The parameters of this model are listed in Table 6.2, and the T-\( \Delta \) and p-\( \Delta \) curves are sketched in Figure 6.11. There is a shadow zone between 32.1° and 32.45°. Also the arrival times of branches cd and de are less than 0.4 seconds apart below 35°. Since there is a caustic at 32.45°, and \( \frac{d^2T}{d\Delta^2} \) is large along the retrograde branch cd up to about 34.5°, there will be a tendency for the \( \frac{dT}{d\Delta} \) values to be characteristic of branch cd. Branch de has comparatively low amplitudes from 33.3° up to about 34.5°, and consequently will not contribute much to the measured \( \frac{dT}{d\Delta} \) values. Beyond 34.5° the measured \( \frac{dT}{d\Delta} \) values will start to decrease rapidly as branch de starts to take over from the retrograde branch. This situation is depicted graphically in Figure 6.12. The dotted line shows how a smooth curve through the measured \( \frac{dT}{d\Delta} \) values might appear. Errors in epicentre determinations might tend to mask the shadow zone. There is a steady decrease in \( \frac{dT}{d\Delta} \) which reaches a minimum at about 32.5°, increases again to a maximum at about 34.5°, and then decreases steadily. This is similar to what is observed, though the model will only give an increase in \( \frac{dT}{d\Delta} \) of about 0.13 sec/deg. The travel-time curve of Figure 6.11 shows that a significant offset in the travel times of 2.5 seconds should occur between about 32° and 32.5°. Since there are travel-time observations for only one event for the Mariana Islands region below 32.6°, such an offset might exist for the Mariana Islands - Tennant Creek path. The velocity model of Table 6.2 does at least provide an explanation of how an apparent increase in \( \frac{dT}{d\Delta} \) can be produced, though a low velocity zone that would give an increase of 0.3 sec/deg would also produce a very large anomaly in the travel times at about 32°. Therefore the \( \frac{dT}{d\Delta} \) data near 33° from
Figure 6.13 Corrected dT/dΔ data for first arrivals from earthquakes of the Solomon, Santa Cruz, New Hebrides and Loyalty Islands Region. Circles and squares denote data from structure groups 1 and 2 respectively of Table 4.4 (or groups 1A and 1B of Table 4.1). Triangles denote data not used in fitting the smooth curve. Open circles, squares and triangles refer to events at depths greater than 130 km.
Mariana Islands events cannot be satisfactorily explained by any simple velocity model.

The smooth curve of Figure 6.2(b) shows a flat region between 37.2° and 38.2° followed by another plateau region between 40.6° and 42.0°. It is not clear at present whether these anomalies are real. A number of \( \frac{dT}{d\Delta} \) values were rejected in fitting the smooth curve (see Table 4.5) because of the very large scatter of the data. This scatter is possibly caused by complex structure between 850 km and 1000 km. These complexities may cause a number of triplications or duplications of the \( \frac{dT}{d\Delta} \) curve which are so close together that the measured \( \frac{dT}{d\Delta} \) values might be characteristic of any one branch or a mixture of branches, depending on their relative amplitudes. The situation will be further confused by errors in epicentre determinations and in focal depths. It is therefore not possible at this stage to identify any distinct anomalies, but the data do suggest small irregularities in structure between 850 km and 1000 km. Two events, numbers 70 and 71 at distances of 40.6° and 41.2°, show two distinct arrivals separated by about 2 seconds; in each case the measured \( \frac{dT}{d\Delta} \) value of the later phase is about 5% lower than for the first phase. This effect is probably associated with a sharp velocity increase well below 1000 km, and will be discussed in Chapter 7.

6.5 DATA FROM REGION D: EARTHQUAKES OF THE SOLOMON, SANTA CRUZ, NEW HEBRIDES AND LOYALTY ISLANDS REGION.

AZIMUTH RANGE 73.0° to 100.8°. (See Figure 6.13 and Appendix 2).

The \( \frac{dT}{d\Delta} \) data fall into two distinct groups: those events at azimuths between 73.0° and about 94°, and those between 94° and 100.8°. A casual glance through the uncorrected data suggests a sharp offset in the \( \frac{dT}{d\Delta} \) curve near 34.0°. The events at azimuths south of 94°, however, are generally at larger distances. Moreover, in Chapter 3 it was pointed out that curious wave forms were observed for events with azimuths of arrival between about 88° and 100° - an effect attributed to local structure (see Table 3.2). It is of interest that the distorted arrivals are observed at azimuths as low as
88.5°, but there is no sharp change in \(dT/d\Delta\) until about 94.0°. Whatever the true cause of these effects, it is very unlikely that structure near the deepest point of the ray path is responsible. Event 23 at an azimuth of 94.4° and a distance of 32.7° supports this view; the \(dT/d\Delta\) value for the second arrival is close to the values obtained for events at similar azimuths and larger distances. The first arrival is small and precedes the large second arrival by about 1.5 seconds, and is most probably due to a small shock preceding the main earthquake. It could be argued that the second phase is a different branch of the travel-time curve, but this seems unlikely as a similar effect is not observed for other events in the same distance range.

In correcting the \(dT/d\Delta\) data two different approaches have been used, and these give slightly different results.

When preparing the preliminary \(dT/d\Delta\) curve the \(dT/d\Delta\) data were split into two groups, 1A and 1B of Table 4.1. In group 1A, covering the distance range 27.9° to 34.9°, there is considerable scatter of the data, suggesting complexity in the \(dT/d\Delta\) curve. The smooth curve fitted through the data is shown in Figure 6.13(a). There is a fairly rapid decrease in \(dT/d\Delta\) from 28° to 30°. Beyond 30.6° \(dT/d\Delta\) changes very little until 34.0°, when it starts to decrease steadily again. The important feature is the flat region of the \(dT/d\Delta\) curve over at least 3°, indicating very low velocity gradients at depths of about 780 to 800 km. The \(dT/d\Delta\) values of group 1B indicate only a very slight decrease in \(dT/d\Delta\) between 34.1° and 35.7°, and a straight line fits the data well. The data of both groups 1A and 1B, when plotted on the same graph, reveal a portion of the \(dT/d\Delta\) curve between 31.0° and 35.7° over which \(d^2T/d\Delta^2\) is small and negative.

In correcting the \(dT/d\Delta\) data by the refined method of section 4.4, the \(dT/d\Delta\) values were once again divided into two groups, and corrections to \(dT/d\Delta\) were made using the dip vector technique described in Chapter 4. Referring to Table 4.4, for structure group 2, Fisher's 'K' is only 2.83 so that the average structure is not particularly useful. Further, the same data
Figure 6.14 Corrected $dT/d\Delta$ data for earthquakes of the Philippines, Sumatra and other regions. Triangles denote data not used in fitting the smooth curves. Open circles and triangles refer to events at depths greater than 130 km.
which cover a distance range of only $1.6^\circ$ gave a correction to $dT/d\Delta$ of $0.152 - 0.090$ when corrected by the preliminary method. Bearing in mind that slight irregularities or regional differences in structure within the lower mantle are likely to cause perturbations of a standard travel-time curve of the same order of magnitude as the random errors in the tables themselves, the lack of coherence of the dip vectors and the value of the correction quoted above indicate that there is no good reason for applying any corrections at all. Therefore the data of group 1 of Table 4.5 were corrected to first order by the vector technique, but the data of group 2 of the same table were not corrected at all. Finally profile 1R of Table 4.5 was adjusted using a smooth curve fitted through the combined data of these two groups. The shape of the smooth curve differs slightly from the preliminary curve. It is flat between $31.7^\circ$ and $32.7^\circ$, and then becomes relatively steep between $32.8^\circ$ and $33.8^\circ$; there is then another plateau region beyond $34.0^\circ$. The steep region between $32.8^\circ$ and $33.8^\circ$ probably represents a slight discontinuity between the two groups of data due to inadequate compensation for local structure. Thus the preliminary profile of Figure 6.13(a) seems to give a better indication of the form of the $dT/d\Delta$ curve between $27.9^\circ$ and $35.7^\circ$, as the steep region of Figure 6.13(b) does not appear in other profiles covering the same distance interval.

The travel-time residuals of Figure 6.3 show no evidence of a break. The seismograms were in fact examined for distinct second arrivals; some events showed such arrivals within two seconds of the P onsets, but no coherent pattern could be detected.

6.6 DATA FROM OTHER REGIONS: EARTHQUAKES OF THE PHILIPPINES, SUMATRA AND OTHER REGIONS. (See Figures 6.14 and 15 and Appendix 2).

In preparing the preliminary $dT/d\Delta$ curve the remaining data between $27.9^\circ$ and $42.0^\circ$ were split into three groups: 3A, 8A and 9 of Table 4.1. The smooth curve through the data between azimuths of $333.0^\circ$ and $344.5^\circ$ shows
Figure 6.15 Corrected $dT/d\Delta$ data for earthquakes of the Philippines, Sumatra and other regions. Triangles denote data not used in fitting the smooth curves. Open circles and triangles refer to events at depths greater than 130 km.
a plateau region between $32^\circ$ and $38^\circ$, though the amount of data available is very limited (see Figure 6.14(a)). The $dT/d\Delta$ curve for the Sumatra region decreases almost uniformly between $27.9^\circ$ and $40.6^\circ$ (Figure 6.14(b)). The data of group 9 are too sparse to warrant any further comment.

Before the final $dT/d\Delta$ curve was prepared it was noticed that the data between azimuths of $333.0^\circ$ and $344.5^\circ$ appeared to be affected by two different structures; a sharp break in structure appears to occur at an azimuth close to $338^\circ$. Consequently the raw $dT/d\Delta$ values were corrected for structure in two groups; these are 6 and 7 of Table 4.4. These groups were then combined to give profile 4 of Table 4.5. Events 210 and 211, from azimuths of $147.7^\circ$ and $168.2^\circ$ respectively, were included in group 7, as they seem to be affected by a similar structure. Similarly, for group 6, events 39, 41, and 42, at azimuths between $9.9^\circ$ and $11.8^\circ$, were used in deriving the correcting structure. In fitting the smooth curve for profile 4, event 210 was rejected because it appeared to be incorrectly adjusted for local structure. The smooth curve through this data (Figure 6.15(a)) shows a flat region between $31^\circ$ and $36^\circ$ in excellent agreement with the preliminary curve of Figure 6.13(a). $dT/d\Delta$ subsequently decreases fairly rapidly beyond $36^\circ$. This curve is clearly an improvement on that of Figure 6.14(a). As $dT/d\Delta$ remains constant over a distance range of about $3^\circ$, the data imply a slight low velocity layer. The westerly profile 9 of Table 4.5 and Figure 6.15(b) gives an almost uniform decrease in $dT/d\Delta$ very similar to that of Figure 6.14(b) prepared from similar raw data. Owing to the paucity of data, the final corrected $dT/d\Delta$ values for both profiles 4 and 9R, excluding events 39, 41, 42, and 210, have been combined to yield an average profile representative of the mantle to the northwest of Australia. The results are plotted in Figure 6.15(c), and the summary points are listed in Table 4.7; there is a flat region of the smooth curve between $31.5^\circ$ and $35.0^\circ$. 
Figure 6.16  Residuals of all events at distances between 27.9° and 45° relative to the 1968 Seismological Tables for P Phases, with the effects of local structure removed as indicated in section 3.4.
6.7 **TRAVEL-TIME RESIDUALS.**

The travel-time residuals for three azimuth ranges have been plotted in Figure 6.16. These residuals have been obtained by taking the residuals relative to the 1968 Seismological Tables for P Phases and correcting for local structure using equation A4.13. Thus the quantity

\[ A_i + D_i \left( \Delta_j - 27.9 \right) + B_i \sin \left( \Phi_j + E_i \right) \]

has been subtracted from each residual using the values of \( A_i, B_i, D_i \) and \( E_i \) given in the second line of Table A4.10; \( i \) and \( j \) have the same meaning as in equation 3.12. The data inevitably show great scatter, but it is evident that any offset in the travel-time curve between about 30° and 40° is less than one second, though it has been pointed out earlier that a larger offset could exist for events from the Mariana Islands region at distances less than 32.6°.

6.8 **DISCUSSION.**

\( dT/d\Delta \) measurements from earthquakes of three separate regions have shown that the \( dT/d\Delta \) curve is in each case very flat between distances of about 31.5° and 36.0°. It is therefore virtually certain that there is a world wide region of low velocity gradients, possibly including a low velocity layer, at a depth of about 800 km. Although great difficulties have been encountered in correcting the data for local structure, the shapes of the \( dT/d\Delta \) curves for events firstly to the east of WRA and secondly in the north west quadrant relative to WRA, are in good agreement. Further, the \( dT/d\Delta \) data from the Mariana Islands and surrounding regions suggest a prominent low velocity zone at a depth of approximately 815 km.

A comparison of my own results with those of other recent array investigations is appropriate at this juncture. In their study of P wave velocities in the mantle below 700 km, **Chinnery and Toksöz** (1967) found evidence for an anomalous region of the mantle at a depth of about 800 km, where the velocity changes slowly with depth. They remarked that the preliminary evidence available suggests that the velocity decreases with increasing
depth. Their $dT/dA$ curve flattens close to $32^\circ$, but they had no good data between about $32^\circ$ and $37^\circ$. Chinnery (1969), in his study of P wave velocities in the lower mantle using the LASA in Montana, found that the $dT/dA$ values between $25^\circ$ and $30^\circ$ decreased very little, but between $37^\circ$ and $40^\circ$ they decreased much more rapidly. The flat region of his $dT/dA$ curve implies a region of the mantle where the velocities change very slowly with depth. Chinnery was not able to determine the exact form of the anomaly owing to the shortage of data between $30^\circ$ and $37^\circ$, but his evidence, as far as it goes, is in excellent agreement with mine. Greenfield and Sheppard (1969) obtained $dT/dA$ curves for both NW and SE azimuths over the distance range $30^\circ$ to $100^\circ$, again using the LASA in Montana; both curves show a flat region beyond $30^\circ$. Johnson (1969), in his study of P velocities in the lower mantle using the extended TFSO, found steep regions of the $dT/dA$ curve near $34.5^\circ$ and $40.5^\circ$. The first of these was poorly defined. The offset at $40^\circ$ is not well supported by my data or by any other array study, though it is evident in my work that between $37^\circ$ and $42^\circ$ the $dT/dA$ curve is comparatively steep. This point will be discussed further in Chapter 8. It is of some interest that Johnson's anomaly at $40.5^\circ$ occurs exactly where there is a regional break in his $dT/dA$ data from the southeast quadrant to the northwest quadrant. It is thus suggested that the anomaly, though it may well exist, has been exaggerated by inadequate allowance for local structure for the two azimuth ranges. The slight anomaly near $34.5^\circ$ detected by Johnson may be due to the comparatively rapid increase in velocity below the low velocity gradients inferred in this chapter. Consequently it seems that due to the relatively small amount of data near $35^\circ$, Johnson has not been able to identify the region of low velocity gradients. Further, if a low velocity zone of the kind suggested in section 6.4 really exists, it is unlikely that it could be identified easily by an array $300$ km in aperture. Referring to Figure 6.12, with a large aperture array, $dT/dA$ would effectively be averaged over nearly $3^\circ$, so that between $31^\circ$ and $35^\circ$ only a general flattening of the $dT/dA$ curve would be observed; the details, such as the increase in $dT/dA$ along the retrograde branch, would be lost. It must, however, be stressed that an increase in $dT/dA$, whatever its cause, is present in one of
Figure 6.17  Distribution of epicentres of events used by Wright (1968).

Having identified a region at a depth of about 800 km where it is likely that there is a low velocity zone, it is important to find a plausible explanation. Moreover, the dT/dA data from earthquakes of the Mariana Islands region do suggest that such a low velocity layer may be more prominent in some parts of the earth. This consideration makes it reasonable to ask whether there may be some feature or characteristic peculiar to the Mariana Islands - Tennant Creek path. The possibility that the different form of the dT/dA curve for the Mariana Islands events is caused by irregularities in the crust or upper mantle must not be overlooked. There is, however, some petrological evidence outlined below which might eventually support the idea that regional differences in structure at depths of 800 km really exist.

There are two well known causes of low velocity layers that need to be considered. First, low velocity zones can be produced by high temperature gradients. Increases in temperature and pressure produce decreases and increases in seismic velocities respectively. Thus if the temperature effect predominates a low velocity layer might occur. Further, if partial melting occurs a decrease in P and S wave velocities is extremely likely. Neither of these two explanations or a combination of them is plausible at a depth of 800 km (see, for example, Stacey, 1969, pp. 258-261). Another possible explanation would be a rapid increase in the iron to magnesium ratio, which is again very unlikely owing to the required density increase. So a less obvious explanation must be sought.

The P waves from the Mariana Islands earthquakes at distances of 32.6° to 33.0° reach their maximum depth immediately below the northern edge of the Central Highlands of the island of New Guinea (Figure 6.17). Ringwood and Green (1966), on the basis of experimental work on the gabbro-eclogite transition, concluded that eclogite is thermodynamically stable under the P, T conditions in the normal continental crust, provided the partial pressure of water vapour is low. Thus large crustal intrusions and extrusions of gabbro and basalt may ultimately transform on cooling into bodies of eclogite. Such
bodies, if large enough, would subside into the mantle because of their high density; this density is also substantially greater than the mean density of the ultramafic upper mantle. So a mechanism has been proposed by which large blocks of eclogite might be introduced into the mantle, and would start to sink slowly into the deeper regions.

Ringwood (1967, 1969a, b) has suggested that a pyroxene-garnet transition occurs in the mantle between 350 and 400 km. As a consequence of this transition an eclogite sinker would continue to fall through the mantle as garnetite to a depth of at least 700-800 km. According to Ringwood, whether these sinkers subside any further depends on the sequence and properties of other phase transitions that are not yet well established. It is likely that somewhere in this region of the mantle garnet-ilmenite or garnet-perovskite transformations take place (Ringwood and Major, 1967, Ringwood, 1969a, b), though Ringwood (1969a, b) places both transformations above 700 km. The depths of these transformations, however, are not known precisely. At about 620 km, the velocity of P waves increases rapidly (see, for example, Johnson, 1967, Green and Hales, 1968, or Chapter 5 of this thesis). At this depth, the beta or 'spinel-like' form of magnesian olivine may transform to a material having approximately the properties of the component oxides (Anderson, 1967a), or more probably to a strontium plumbate or closely related structure, (Ringwood, 1969a, b). Alternatively, the beta phase may disproportionate to an ilmenite-type phase and magnesium oxide (Ringwood, 1969a). Provided the garnet-ilmenite transformation occurs below 800 km, the garnetite blocks might be expected to accumulate at a depth somewhere between 620 km and 800 km or so. Anderson (1967a) has indicated that a garnet content of 10% below 620 km in the mantle would reduce the P wave velocity by 0.2 - 0.6 km/sec. It appears that a low velocity layer at a depth of about 800 km could result from an increase in garnet content with increasing depth. Thus it is suggested that this low velocity layer below New Guinea results from the accumulation of eclogite sinkers originating in the eugeosynclinal zone to the north.
Since the publication of this explanation of a low velocity layer at about 800 km (Wright, 1968), Press (1969) has suggested that eclogite is widespread at a depth of about 100 km in the suboceanic mantle. He has also proposed a mechanism in which eclogite fractionation from the underlying partially molten asthenosphere plays a key role in the creation and the spreading of the rigid, lithospheric plate. If this is correct, it seems more likely that eclogite sinkers are real. If the garnet-ilmenite or garnet-perovskite transformation is subsequently found to occur well above 800 km, it is not obvious at present whether the explanation of the low velocity layer is still valid.

A region of low velocity gradients at a depth of about 800 km is most likely to be explained in terms of chemical inhomogeneity; phase changes generally produce sharp velocity increases. Anderson and Julian (1969) have attempted to reconcile the discrepancies between the upper mantle P wave velocity model of Johnson (1967) and the S wave velocities of Ibrahim and Nuttli (1967). By inserting a low velocity layer immediately above each of the two major discontinuities in the upper mantle, even more pronounced than those of Ibrahim and Nuttli's original model, they were able to make the depths of these discontinuities agree more closely with those of Johnson's P velocity model. They also point out that Barsch and Anderson, in a personal communication to the authors, have recently suggested that a negative $d\mu/dP$ implies an unstable crystal lattice, where $\mu$ is the rigidity. Therefore, a decrease in shear wave velocity with increasing depth should occur above each phase transformation in the mantle. This phenomenon might be expected to have some effect on P wave velocities; it is perhaps unlikely that it will produce a low velocity zone, but a general reduction in velocity gradients above each phase change in the mantle seems a distinct possibility. This point will be raised again when velocity anomalies at greater depths are discussed.
6.8 **CONCLUSIONS.**

Low velocity gradients for P waves at a depth of about 800 km in the mantle are widespread. In addition there is probably a slight decrease in P wave velocity with increasing depth just below 800 km. This decrease can be explained in terms of the eclogite sinker model proposed by Ringwood (1967); this idea is an extension of the modification of the ocean floor spreading hypothesis of Hess (1962), using the basalt-eclogite transformation as a tectonic engine, put forward by Ringwood and Green (1966). The low velocity layer is possibly more prominent beneath a tectonically active region, though the dT/dΔ measurements from northwest azimuths also sample the mantle beneath an active region (see Figure 6.1). There is no evidence for a single velocity anomaly at about 1000 km depth of the kind suggested by Johnson (1969) or Archambeau et al. (1969). However, there is some evidence for more rapidly changing velocities close to 1000 km which will be discussed in Chapter 8. The velocity gradients apparently change slowly, or alternatively there is a series of small closely-spaced discontinuities between 850 km and 1000 km. Also a caustic or focus may have been located at a distance of 32.6°, using dT/dΔ data from earthquakes of the Mariana Islands region. An apparent increase in dT/dΔ can be explained by considering T-Δ and p-Δ curves for small low velocity layers.
Figure 7.1 Epicentres of events at corrected distances greater than 420 from WRA.
CHAPTER 7
THE LOWER MANTLE

PART 2: 1000 KM TO 2890 KM (42° < Δ ≤ 99°).

7.1 INTRODUCTION.

Until recently there has been a tendency either to believe or assume that the region of the mantle below 1000 km to about 200 km above the mantle-core boundary (i.e. Bullen's region D') is both approximately chemically homogeneous and devoid of phase changes. Bullen (1963, pp. 229-239), for example, has used the assumption of chemical homogeneity in the D' region to derive both his earth models A and B. Before the development of seismic arrays it was not possible to test this view owing to the lack of resolution obtainable by the conventional travel-time method, though a number of investigators in the past have suggested the existence of irregularities in the structure of the lower mantle. Gutenberg (1958), using amplitude data, has suggested a region between 1400 and 1500 km where the velocity increases relatively slowly with depth. Bugayevski (1964) has put forward evidence of first order discontinuities in the travel-time curve near 50° - 54° and 70° - 72° which he has suggested are due to low velocity zones at depths close to 1200 and 1800 km. Recent array studies (Chinnery and Toksöz, 1967, Chinnery, 1969, Johnson, 1969, Corbishley, 1970) have also suggested the presence of velocity anomalies below 1000 km in the mantle, but there are unresolved discrepancies when the results are compared.

Therefore the objective in this chapter is to examine the evidence for inhomogeneities or discontinuities in the mantle below 1000 km without comparing the results with other investigations; the approach is to attempt to identify such anomalies by discussing the data on a regional basis. As in Chapter 6, each group of data is presented together with a smooth curve fitted by the Method of Summary Values. It is stressed that the presentation is
Figure 7.2 $dT/d\Delta$ data from region E: azimuth range $342.9^\circ$ to $357.8^\circ$; distance range $41.7^\circ$ to $53.3^\circ$. Triangles denote points rejected in fitting a smooth curve. Open circles and triangles refer to events at depths greater than 130 km.
designed primarily to indicate how the problem of lower mantle inhomogeneities should be approached in the future. Although the results suggest anomalies and possibly even regional differences in structure, they are not sufficiently detailed to make a completely rigorous discussion possible. Owing to the complexity of the observed data and the consequent difficulty of exact quantitative analysis, some of the interpretation must be qualitative. This introduces an undesirable, but unavoidable, arbitrariness into the interpretation of the data.

7.2 DATA FROM REGION E: EARTHQUAKES OF THE EAST CHINA SEA, TAIWAN, THE RYUKYU ISLANDS AND SURROUNDING REGIONS.

AZIMUTH RANGE 342.9° TO 357.8°. DISTANCE RANGE 41.7° to 53.3°.
(See Figures 7.1, 7.2 and Appendix 2).

In the preliminary dT/dΔ curve of Figure 7.2(a) there is a reasonably steep region between 41.7° and 46.7°, followed by a comparatively flat region between 47.5° and 53.3°, though the amount of data used is limited. It is worth remarking that these observations (group 3A of Table 4.1) lie in the same azimuth range as the data of group 3C of the same table, which cover the distance range 30.1° to 39.1°. The corrections for the two groups are $-0.672 \pm 0.035$ and $-0.643 \pm 0.045$ sec/deg respectively, so that correcting the data in two separate distance intervals has not introduced a discontinuity into the overall dT/dΔ curve.

Referring to the refined dT/dΔ curve of Figure 7.2(b) (structure group 3R of Table 4.5), five events (Nos. 263-267) have been added which alter the shape of the curve in the following way. Between 44.9° and 46.4°, the curve is very flat; it then falls rapidly between 46.7° and 48.7° and flattens again between 49.0° and 52.0°. In addition the dT/dΔ curve may be discontinuous, the break occurring between 47.2° and 48.5° where there are unfortunately no dT/dΔ measurements. In preparing the smooth curve, two of the three dT/dΔ measurements from event 84 at a distance of 46.0° were much too low, and were consequently rejected. Further, this event has a fairly clear P onset, and the dT/dΔ values are therefore reliable. There are two possible explanations of this phenomenon. First it could be purely an azimuthal
Figure 7.3 $dT/d\Delta$ data from regions C and F: azimuth range $4.8^\circ$ to $16.1^\circ$; distance range $41.2^\circ$ to $63.9^\circ$. Triangles denote points rejected in fitting a smooth curve. Open circles and triangles refer to events at depths greater than 130 km.
effect, as the event occurs at an azimuth of 352.0°, and the other events at
distances between 44.9° and 47.2° are at azimuths between 344.3° and
348.1°. On the other hand the low values of dT/dA could be caused by confusion
of two closely spaced arrivals; the second arrival would have a comparable
amplitude and a lower dT/dA value due to the duplication or triplication of the
dT/dA curve caused by a sudden increase in velocity or velocity gradient.

Data from events of other regions discussed in sections 7.3 and 7.5 provide
some support for the second interpretation. Unfortunately the apparent break
in the dT/dA curve between 47.2° and 48.5° coincides with a shift in azimuth
of 8 – 10°. Thus, like the anomalous dT/dA values for event 84, the abrupt
fall in dT/dA could be attributed to changes in local structure with azimuth.

However, there is independent evidence from other regions, which will be
described later, that the break is real.

7.3 DATA FROM REGIONS C AND F. EARTHQUAKES OF THE MARIANA
AND VOLCANO ISLANDS REGIONS (C); THE BONIN ISLANDS, THE
JAPAN REGION, THE KURILE ISLANDS, KAMCHATKA AND SURROUND­
ING REGIONS (F).

AZIMUTH RANGE 4.8° TO 16.8°. DISTANCE RANGE 41.8° TO 80.1°.
(See Figures 7.1, 7.3 and 7.4, and Appendix 2).

Great care needs to be taken in interpreting the data in the
azimuth range 5° to 20°, owing to the sudden change in the azimuth anomalies
of P arrivals close to 10° (see, for example, Table 3.1 and section 3.2). On
examining the raw dT/dA data one outstanding feature is apparent. There
appears to be a sharp break in the dT/dA curve near 46° or 47°, in quali­
tative agreement with the results of section 7.2. There is, however, a perplex­
ing feature; the fall in dT/dA between 44° and 50° is about 0.9 sec/deg. The
travel-time residuals (see section 7.7 ) indicate that a real break of this
magnitude is highly improbable; if it were real a significant anomaly in the
travel times should be detectable. It therefore seems probable that the observed
effect is due partly to a rapid change in structure either in the crust or the upper
mantle in the vicinity of the array. In correcting the dT/dA data, the magnitude
of the possible dT/dA anomaly must be reduced to some extent.
To prepare the preliminary $dT/d\Delta$ curve, the data were split into three groups. The first group (Figure 7.3) is 2B of Table 4.1, for which the data corresponding to distances of less than $42^\circ$ have already been interpreted in Chapter 6. The group terminates at event 80 at a distance of $46.4^\circ$. The scatter of the data is very great, but the smooth curve does become very steep between $45.0^\circ$ and $46.4^\circ$. The second group (4A of Table 4.1) covers the distance range $47.9^\circ$ to $59.5^\circ$. The smooth curve through this $dT/d\Delta$ data decreases rapidly between $47.9^\circ$ and $49.0^\circ$, and then gradually flattens so that there is a comparatively flat region between about $50.5^\circ$ and $57.0^\circ$. Beyond $57^\circ$ the curve becomes slightly steeper, though the paucity of data beyond $56^\circ$ indicates that this is probably insignificant. In the distance range $50^\circ$ to $59^\circ$, the $dT/d\Delta$ curve has very little detail, and it is of some interest that the scatter of the data is considerably less than would be expected from the error estimates in $dT/d\Delta$. $\chi^2$ is only 36.19 on 51 degrees of freedom, but it is unlikely that the data are undersmoothed. The reason for this is quite clear. In calculating the standard deviation on each onset time the residuals on the least-squares fit to $dT/d\Delta$ and azimuth for all working seismometers have been used (see section 2.2(i)). However, the seismometers of the red line tend to show larger residuals, and further, the wave fronts of the P arrivals for earthquakes of the Japan region are almost parallel to the red line. Therefore these seismometers contribute little towards the $dT/d\Delta$ values, and their residuals will cause the errors in the $dT/d\Delta$ measurements to be slightly over-estimated. Whether events 80, 106 and 107, at distances of $46.4^\circ$, $47.9^\circ$ and $47.9^\circ$ respectively, are placed in group 2B or 4A is somewhat arbitrary. Events 106 and 107 were included in group 4A when determining corrections, but in preparing the preliminary world average $dT/d\Delta$ curve they were corrected by $-0.829 \text{ sec/deg}$, the correction to group 2B. This procedure, although unsatisfactory, was adopted to avoid smoothing out the anomaly, and is justifiable since it is not clear how to separate the true anomaly in $dT/d\Delta$ from the apparent local structural effect. The uncertainty in the corrections to be applied for these events is indicated in Figure 6.3(a) by the arrow, question mark and
Figure 7.4 $dT/d\Delta$ data from regions F and G: azimuth range 6.5° to 25.6°; distance range 63.9° to 81.2°. Triangles denote points rejected in fitting a smooth curve. Open circles, squares and triangles refer to events at depths greater than 130 km.
two sets of points. The data of group 5A of Table 4.1 cover the distance range 65.4° to 78.2° (Figure 7.4). The dT/dΔ curve shows little fine structure, and no distinct anomalies are visible. In contrast to the large difference of 0.5 sec/deg between the corrections for groups 2B and 4A, the corrections to groups 4A and 5A do not differ significantly.

In preparing the final revised dT/dΔ curve some distinct improvements were possible (Figures 6.3(b) and 6.4(b)). First, the gap in the data between 59.5° and 65.4° was removed by the addition of seven events at distances between 57.6° and 67.0°, and secondly, two extra events at corrected distances of 44.2° and 44.5° were added to the data of region C. There are some slight differences in the way the data have been split into groups. First, events 106 and 107 were included in profile 2R of Table 4.5. Also peak B of event 80 was corrected using structure 8 of Table 4.4 in an attempt to account for both the large decrease in dT/dΔ between the two Volcano Islands events, 80 and 262, and the systematic changes in wave form as event 80 crosses the array. In profile 5R the data of structure groups 8E and 9E of Table 4.4 have been combined. Since there is evidence for a sharp change in local structure at azimuths close to 20°, event 144 at a distance of 80.1° was included in this profile because it occurred at an azimuth of 16.8°. The final refined dT/dΔ curve for the distance range 41.8° to 80.1° shows a few significant features that were not clearly revealed in the preliminary curves. The steep region between 42.5° and 46.5° is followed by a very flat region around 47.5°. Beyond 51° the dT/dΔ curve gradually becomes steeper up to 56° and then progressively becomes flatter again. An important result is that the refined dT/dΔ profile 5R does not show any evidence for the anomaly near 60° reported by Johnson (1969) and Corbishley (1970). There is also a rapid decrease between the dT/dΔ values of regions F and G which shows up clearly at about 80° (Figure 7.4). Evidence presented in sections 7.4 and 7.6 indicate that this anomaly is probably real.

Some important clues on the nature of the inhomogeneities in the lower mantle between 1000 and 1200 km are provided by careful consideration
of the $dT/d\Delta$ measurements from individual events. First, for events 70, 71, 78 and 79 $dT/d\Delta$ measurements have been made for distinct second arrivals between 1.0 and 3.2 seconds after the P onset. In each instance the $dT/d\Delta$ values for the second phase are lower, suggesting penetration below a region of high velocity gradients. This is additional, but by no means decisive, support for the existence of a real anomaly in the $dT/d\Delta$ curve close to $47^\circ$. However, events 78 and 79, although they both correspond to a larger corrected distance, give wider separations of the two phases than events 70 or 71. Further, the differences in $dT/d\Delta$ and azimuth for the two phases for each event are due largely to differences in arrival times along the red line of the array. Thus their origin is uncertain. Events 79, 261 and 262 all give anomalously high $dT/d\Delta$ values for the first arrival and are all at similar azimuths, distances and focal depths. This phenomenon could possibly be caused by irregularities in structure in the source region, though the azimuths of all these events are close to $10^\circ$, so that the results could equally well be related to those of Table 3.1. Event 80 is at only a slightly larger distance and almost the same azimuth as events 79, 261 and 262, but has $dT/d\Delta$ values between about 0.4 and 1.0 sec/deg lower, in closer agreement with the $dT/d\Delta$ value for the second phase of event 79. Thus, for all of the data of profile 2R beyond $40^\circ$, there is a large scatter in the $dT/d\Delta$ values particularly for deep focus events. It is possible that in addition to local structural effects this scatter is caused partly by the differences in relative amplitudes of the multiple arrivals in the vicinity of a cusp, and partly by irregularities in the source region.

7.4 DATA FROM REGION G: EVENTS OF THE ALEUTIAN ISLANDS AND ALASKA REGIONS.

AZIMUTH RANGE 21.4° TO 33.4°. DISTANCE RANGE 79.4° TO 98.8°.

(See Figures 7.1, 7.5 and Appendix 2).

The preliminary $dT/d\Delta$ curve of Figure 7.5(a) (see group 6A of Table 4.1) decreases steadily beyond $80^\circ$ and gradually flattens, becoming very flat beyond $90^\circ$. No distinct anomaly can be detected.
Figure 7.5 $dT/d\Delta$, sec/deg

Figure 7.5 $dT/d\Delta$ data from region G: azimuth range 21.4° to 33.4°; distance range 79.4° to 98.8°. Triangles denote points rejected in fitting a smooth curve. Open squares and triangles refer to events at depths greater than 130 km.
In calculating structure 10 of Table 4.4, the observational points used differed from those of the preliminary curve in two respects. First, event 144 at an azimuth of 16.8° was placed in region F owing to the apparent sharp change in local structure between azimuths of about 17° and 20°. Until the preparation of the refined dT/dA data, it was not clear whether the apparent change in structure was real, as there is a break in the data with respect to both azimuth and distance. However, the dT/dA measurements for event 279, at a corrected distance of 74.2° and an azimuth of 28.5°, show conclusively that the change in structure is real. Secondly, five events at distances between 79.7° and 88.2° have been used to supplement the data. The most obvious feature is the apparent increase in dT/dA in the distance range 84° to 88°. In order to obtain a physically meaningful dT/dA curve, some of the points have been rejected so that the smooth curve decreases monotonically. Initially there is a fairly steep region of the curve between 79.4° and 81.5° which flattens rapidly revealing a plateau region between 83.0° and about 88°. The curve then becomes steep between 90.0° and 93.7°. Owing to the sparseness of the data beyond 88°, the smooth curve is perhaps misleading. dT/dA might possibly fall sharply close to 88°. The apparent increase in dT/dA could be due to bias introduced near the source or close to the array; alternatively it could be caused by confusion of two or three branches of the dT/dA curve produced by an anomaly of the kind illustrated in Figures 6.9 or 6.10(3). The latter interpretation seems more probable, since the data from other azimuth ranges to be discussed in section 7.6 yield similar results.

7.5 DATA FROM REGION H: EARTHQUAKES OF THE FIJI, TONGA AND KERMADEC ISLANDS REGION.

AZIMUTH RANGE 83.6° TO 119.5°. DISTANCE RANGE 40.9° TO 50.1°.
(See Figures 7.1, 7.6 and Appendix 2).

The difficulties in interpreting dT/dA data between azimuths of about 90° and 100° have been described in Chapter 3. Consequently no dT/dA values from this azimuth range have been used in preparing the preliminary dT/dA curve. In Figure 7.6 (a) corrected dT/dA values from the
Figure 7.6  $dT/d\Delta$ data from region H: azimuth range $83.6^\circ$ to $119.5^\circ$; distance range $40.9^\circ$ to $49.4^\circ$. Triangles denote points rejected in fitting a smooth curve. Open circles, squares and triangles refer to events at depths greater than 130 km.
troublesome azimuth range have been included for illustrative purposes.

The refined dT/dΔ profile of Figure 7.2(b) was prepared by taking the dT/dΔ values of event 38 and those of group 11 of Table 4.4, corrected to first order for structure. Also, dT/dΔ values from events 161 - 163, 165 and 167, within the azimuth range associated with structural complexities, were corrected by applying the structure of group 11 of Table 4.4 and used in deriving the smooth curve. This curve shows a comparatively flat region between 43.5° and 46.5° followed by a steep region beyond 47.0°. Although the amount of data is limited, the results are in qualitative agreement with those of Figure 7.2. It is worth remarking that events 155 and 156 give anomalously low values of dT/dΔ; for event 156 the lower dT/dΔ values are associated with the third peak of the onset, and it may therefore be associated with a distinct second arrival. Note also that a dT/dΔ displacement of a similar amount was observed for event 84 at an azimuth and distance of 352.0° and 46.0° respectively.

7.6 DATA FROM OTHER REGIONS: EVENTS FROM THE ASIAN CONTINENT AND SURROUNDING REGIONS.

AZIMUTH RANGES 297.5° TO 327.3°; 338.5° TO 358.5°; 28.5°.
DISTANCE RANGE 44.6° TO 90.9°. (See Figures 7.1, 7.7 and Appendix 2).

The corrected dT/dΔ values used in preparing the preliminary dT/dΔ curve are displayed in Figure 7.7(a). The data were corrected in five separate groups, and because of the lack of detail present in each separate group, no further discussion is warranted, except to remark that there is some evidence for a steep region of the dT/dΔ curve near 80°. With the addition of more dT/dΔ measurements beyond 80°, it became possible to obtain more reliable corrections for structure for the data used in the refined dT/dΔ curve. To construct the dT/dΔ profile between 64.5° and 80.4° (profile 7 of Table 4.5), data corrected to first order from several different azimuth ranges were combined. One dT/dΔ value from each of events 202 to 204 was used in obtaining the smooth curve. The remaining dT/dΔ measurements
Figure 7.7  $dT/d\Delta$ data from other regions: azimuth ranges $297.7^\circ$ to $327.3^\circ$; $338.5^\circ$ to $358.5^\circ$; $28.5^\circ$; distance range $44.6^\circ$ to $90.9^\circ$. Triangles denote points rejected in fitting a smooth curve. Open circles, squares and triangles refer to events at depths greater than $130$ km.
for these events were placed in profile 11. The smooth curve for profile 7 shows a flat region near 65°; it then becomes progressively steeper up to 80.4°.

The most interesting and effective application of the refined method of correcting dT/dΔ described in section 4.2 is exemplified by the data of profile 11 of Table 4.5. Although the dT/dΔ measurements are drawn from an azimuth range of nearly 45°, it was possible to obtain satisfactory corrections using only one structure. In contrast, for the construction of the preliminary dT/dΔ curve, the data over this distance and azimuth range were corrected in three separate groups. dT/dΔ data from 8 events in the distance range 80.4° to 90.8° were added before the preparation of the final dT/dΔ profile. The scatter of the raw data, especially between 80.0° and 90.9°, is very great indeed. However, all the data in the distance range 71.2° to 90.9° and within the azimuth range 283.4° to 327.3° have been assumed to be affected by the same structure, and this has been designated structure group 14 of Table 4.4. The vector sum of the individual structures associated with all dT/dΔ and azimuth values gives a value of 89.5 for Fisher's 'K'. Therefore the 57 structure vectors used in working out this average form by far the most coherent set of structure vectors. This average structure introduces very strongly azimuth dependent corrections, and has very effectively reduced the scatter.

The scatter of the corrected dT/dΔ data between 80.0° and 81.6° has made it necessary to reject several points in order to obtain a statistically acceptable smooth curve. The dT/dΔ and azimuth values for events 201, 285 and 286, at azimuths of 325.0°, 324.7° and 324.6° and at distances of 80.0°, 80.4° and 81.6° respectively, show systematically high dT/dΔ values when corrected for structure. Of more importance are the systematically larger values of dT/dΔ for event 286 in comparison with 285. This, together with the scatter of the remaining data, suggests that an anomaly in the dT/dΔ curve of the kind illustrated in Figure 6.9 may occur close to 80°, perhaps with point b near 81.6°. Consequently the smooth dT/dΔ curve of Figure 7.7(b)
Figure 7.8 Residuals of all events at distances between 40.2° and 98.8° relative to the 1968 Seismological Tables for P Phases with the effects of local structure removed as indicated in section 3.4.
may be a little misleading in this distance range. It does show a steep region between 81.2° and 82.8°, but the flat region below 81° is unrealistic. There is a plateau region between 83.0° and 85.3°, and finally the curve becomes much steeper between 86.0° and 90.9°. These features are in reasonable agreement with the general shape of the curve for the Aleutian Islands region events. There are few observations beyond 85°, but an examination of the dT/dΔ values for events 207-209 and 291-292 suggests a sudden or rapid decrease in dT/dΔ between 87.8° and 88.9°.

7.7 TRAVEL-TIME RESIDUALS.

The travel-time residuals displayed in Figure 7.8 show that there are no anomalies in the lower mantle that are sufficiently large to produce an anomaly in the travel-time curve greater than a few lengths of a second, even when the data are examined on a regional basis. These residuals were calculated relative to the 1968 Seismological Tables for P Phases in exactly the same manner as those of Figure 6.16. It is therefore reasonable to assume that any irregularities in lower mantle structure will produce anomalies in a modern travel-time curve not significantly greater than the random errors in the travel times themselves. This result provides ample justification for the use of the empirical and refined techniques for correcting dT/dΔ values that were described in sections 4.3 and 4.4.

7.8 DISCUSSION.

It has been possible to detect anomalies in the dT/dΔ curve for P waves beyond 40° for different distance ranges and for data from events occurring in entirely different regions. Hence, without considering anybody else's work, it can be stated confidently that irregularities in the structure of the lower mantle have been identified. At this stage only a brief summary and discussion of the results presented in this chapter are necessary; a detailed comparison with other work is given at the end of Chapter 8.
The dT/dΔ values for regions E and H both indicate a very flat portion of the dT/dΔ curve between 44° and 46 - 47°, followed by a rapid decrease near 48°. The rapid decrease is also supported by the data of region F, but the preceding flat region does not show up. Thus the following picture emerges. At about 47° there is definitely a sharp decrease in dT/dΔ for first arrivals. For the regions to the northwest and to the east of WRA, the nature of the velocity anomaly is very similar. However, for events to the north of WRA, at distances between 40.6° and 44.4°, distinct second arrivals between 1.0 and 3.2 seconds later than P and having appreciably lower dT/dΔ values have been observed. Therefore the anomaly below the west Pacific margin to the north of WRA might be more pronounced than elsewhere. This is perhaps surprising, but it was also suggested that the possible low velocity layer near 800 km for this azimuth range was more pronounced than in other regions. Alternatively, this enhancement of dT/dΔ anomalies may really be due to structure fairly close to the array; nevertheless, it is tempting to speculate that regional variations in lower mantle discontinuities may be related to tectonic processes more obviously manifested near the surface of the earth.

Beyond 49.0° events of both regions E and F, slightly west of and east of north relative to WRA respectively, indicate a plateau region of the dT/dΔ curve extending over several degrees. From 53° to 83° the only region providing a detailed dT/dΔ profile is the west Pacific margin. There is no offset near 60° as Johnson (1969) and Corbishley (1970) have suggested, though the smooth curve is significantly steeper around 56°. Because of the limited amount of data the situation near 70°, where both Johnson and Corbishley find an anomaly, is unclear.

There is most probably a triplication of the kind suggested by Chinnery (1969) near 80°. The dT/dΔ data from the Aleutian Islands and from the Asian Continent both indicate a comparatively rapid decrease in dT/dΔ between 78° and 81°, and the large scatter in the data over this distance interval suggests multiplicity in the dT/dΔ curve. Both sets of data also
indicate a relatively flat region of the curve between 82.0° and about 88.0°, which is probably followed by a sharp fall in \( \frac{dT}{d\Delta} \) beyond 88°. Thus it is tentatively concluded that there is another anomaly in the \( \frac{dT}{d\Delta} \) curve between about 84° and 89° of the kind illustrated in Figure 6.9.

The interesting feature suggested by these results is that relatively sudden increases in P velocity in the mantle are preceded by regions of anomalously low velocity gradients. Therefore a plausible explanation must be sought. At the end of the last chapter it was mentioned that Barsch and Anderson have suggested that a negative \( \frac{du}{dP} \) implies an unstable crystal lattice. If this is correct then each sharp decrease in \( \frac{dT}{d\Delta} \) is associated with a phase transformation in the lower mantle, and the slowly changing \( \frac{dT}{d\Delta} \) values represent the slowly changing velocities in the region above the phase change, where the crystal lattice is becoming unstable. Where the data for a particular distance range cover more than one region of the earth, there is no conclusive evidence for any regional differences in structure below 1000 km, though there is some evidence that such differences may persist to about 1200 km. This is not meant to imply that such differences do not exist. It is just that arrays are not capable at present of detecting such subtle variations in structure, owing to the difficulty of separating locally induced effects. The sharpness of each rapid velocity change is not known from the seismic data; relatively abrupt increases in velocity probably give rise to triplications in the travel-time curve as suggested by Johnson (1969) and Chinnery (1969) in which the individual branches are too close together to be observed on seismograms. The interpretation of \( \frac{dT}{d\Delta} \) data is much more difficult if such multiplicities really exist, since a single valued smooth curve fitted through the data will not provide a realistic indication of the behaviour of the \( \frac{dT}{d\Delta} \) curve.
CHAPTER 8

P WAVE VELOCITIES IN THE MANTLE BELOW 700 KM.

8.1 PRELIMINARY REMARKS.

The methods by which a $dT/d\Delta$ curve is corrected for structure, smoothed and inverted were described in Chapter 4, and in Chapters 6 and 7 detailed evidence for the existence of regions of anomalous velocity change were presented by examining the $dT/d\Delta$ data region by region. Since each identifiable anomaly in the data was present for more than one region of the earth, it was concluded that structure at the deepest point of the ray path was the most likely explanation. Therefore to complete this research project, it is necessary to derive some world average velocity models for the lower mantle. Because the $dT/d\Delta$ profiles have been constrained to the 1968 Seismological Tables for P Phases, it is expected that the actual P velocities will be similar to those associated with the tables, modified at shallower depths according to the velocity distribution used to 'strip' the earth to the top of the $dT/d\Delta$ curve.

The preliminary $dT/d\Delta$ curve, together with the original interpretation presented by Wright (1970), is discussed first. Then the final 'smooth' model ANUW2, derived using additional data and the refined method of correcting $dT/d\Delta$ measurements, is compared with the preliminary version ANUW1 to illustrate how the careful selection of extra data and the use of a more elaborate scheme for its analysis have answered some of the problems raised when interpreting the preliminary data. In fitting a smooth curve through the $dT/d\Delta$ data there is a tendency to smear out possible sharp changes in the $dT/d\Delta$ curve. Consequently, a modified version of ANUW2, denoted ANUW3, is presented, in which the anomalies have been made sharp. Then two regional $dT/d\Delta$ curves for the west Pacific margin and for the area in the northwest quadrant relative to WRA are shown. In Chapter 2 the events used
Table 8.1. P Wave Velocities for Models ANUW1, ANUW2 and ANUW3.

(a) ANUW1 and ANUW2.

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Figure 8.1 dT/dA curve for preliminary lower mantle model ANUW1.
in this study were classified into a number of groups according to focal depth and the quality of the recorded information (see Table 2.4 and section 2.5). Since it is useful to examine whether the scatter of the dT/dA data is related to the quality of this information or to focal depth, some statistical significance tests on the dT/dA values are included. Finally, the results are compared with other investigations of travel times, travel-time gradients and amplitudes; an attempt is also made to explain the discrepancies between the results of this and other array studies.

8.2 AN AVERAGE P WAVE VELOCITY MODEL.

(a) Preliminary Model ANUW1. (See Figure 8.1 and Table 8.1).

A P wave velocity distribution from a depth of 700 km to 2790 km was calculated from the preliminary dT/dA curve of Figure 8.1 by the Herglotz-Wiechert method, using computer programme ZCW101C8. A modified version of Early Rise Model 2 of Green and Hales (1968) was used to 'strip' the earth to a depth of 700 km, and the parameters of this model are listed in Table 8.2.

The most prominent feature of Figure 8.1 is the flat portion of the dT/dA curve between 32° and 37° corresponding to anomalously low velocity gradients, - d v/ d r, between 800 and 850 km. The velocity model shows a slight decrease in the P wave velocity with increasing depth at 820 km. As indicated in section 4.3, the dT/dA data have been over-smoothed in this distance range to avoid an apparent increase in dT/dA which would cause the Herglotz-Wiechert method to formally break down. Such an increase is produced largely by data from Mariana Islands earthquakes, and has already been explained in terms of a low velocity layer. Figure 8.1 also shows the anomalous region beyond 46° corresponding to a small increase in - d v/ d r at a depth of about 1150 km, followed by a region of low velocity gradients between 1260 and 1300 km. The smooth curve suggests a slight increase in - d v/ d r near 1600 km, which agrees with the results of Johnson (1969) and Corbishley (1970). However, subsequent work has suggested that the effect is produced largely by
Table 8.2. Modified Version of Early Rise Model 2 of Green and Hales used to 'Strip' the Earth to a Depth of 700 km.

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Figure 8.2 $dT/d\Delta$ curve prepared by the refined method of section 4.4. Dotted line denotes average $dT/d\Delta$ value for events at distances between $88.9^\circ$ and $98.8^\circ$. 
inadequate compensation for structure for two events at distances of 58.5° and 60.5° belonging to group 8B of Table 4.1. The existence of a rapid increase in velocity near 1600 km has not been substantiated by the additional data used in preparing the refined dT/dΔ curve. There also appears to be a comparatively flat region of the dT/dΔ curve between 63° and 66° corresponding to low velocity gradients between depths of 1700 and 1750 km. Finally, there is a fairly steep portion of the curve between 78° and 83.5° yielding comparatively high velocity gradients between about 2200 and 2500 km. Since the preparation of the preliminary velocity model, additional dT/dΔ data over the distance ranges 44.2° to 47.2°, 57.6° to 67.0° and 79.7° to 80.8° have been compiled in order to map out the anomalous regions in more detail. To illustrate the depths of the anomalies, the parameter \( \left( \frac{r}{v} \frac{dv}{dr} \right) \) has been calculated from the velocity distribution interpolated at 5 km intervals, and displayed graphically in Figure 8.3.

(b) **Refined Model ANUW2.** (See Figures 8.2, 8.5 and Table 8.1).

The smooth curve of Figure 8.2 cannot be used to derive a velocity model owing to the slight increase in dT/dΔ between 30.7° and 32.0° and between 34.8° and 36.0°. These two increases can, incidentally, be explained by the random errors in the data points. The two plateau regions in the dT/dΔ curve are introduced when data from several different azimuth ranges are combined. The differences in the form of the dT/dΔ curve for regions C and D of Chapter 6 (i.e. to the north and to the east of WRA respectively), are largely responsible for this, and there seems no justification for assuming that the two flat regions are real. Regional differences in structure or inadequate compensation for local structure are the most likely explanations. Consequently, the dT/dΔ curve was adjusted in the following way to enable a velocity model to be derived. The ordinates of the third and sixth summary points of Table 4.6 were increased by 0.040 and 0.014 sec/deg respectively, and the ordinate of the fourth summary point was decreased by 0.006 sec/deg. These adjustments are arbitrary, but since the model ANUW2 derived from
Figure 8.3 - $\frac{r}{v} \frac{dv}{dr}$ for the 5 km layers of models ANUW1 and ANUW2 and the velocity model derived from the 1968 Seismological Tables for P Phases.
the adjusted curve is not unique, this does not matter. The smooth \( \frac{dT}{d\Delta} \) curve was fitted only to a distance of 88.2° owing to the sudden decrease in \( \frac{dT}{d\Delta} \) beyond 88.2°, the general paucity of data beyond 88°, the discrepancy in the \( \frac{dT}{d\Delta} \) values between events from the west (Gulf of Aden and Iran) and from the northeast (Aleutian Islands and Alaska), and the effect of the core on the P arrivals as the depth of penetration approaches the mantle-core boundary (see Johnson, 1969). In addition, it was not possible to obtain a smooth \( \frac{dT}{d\Delta} \) curve beyond 90° in which there was not an increase in \( \frac{dT}{d\Delta} \) with increasing distance.

The velocity distribution was derived in exactly the same manner as the preliminary model, and the parameter \( \zeta \) is plotted in Figure 8.3. The derived model shows a region of high velocity gradients below 700 km, followed by low velocity gradients near 790 km. There is another plateau region between 825 and 860 km. \( \zeta \) then reaches a slight maximum at 885 km, decreases slightly, and increases to another maximum at 1030 km. Thus the velocity anomaly near 1000 km suggested by Johnson (1969) and Archambeau et al. (1969) may be present, but only as a comparatively small irregularity. \( \zeta \) subsequently decreases to a minimum at about 1090 km, and increases steadily to a maximum at 1210 km; these features correspond to the flat region of the \( \frac{dT}{d\Delta} \) curve between 43° and 46°, followed by the sharp offset or rapid decrease beyond 47°. There is also a region of low velocity gradients between 1260 and 1330 km corresponding to the flat region of the \( \frac{dT}{d\Delta} \) curve between about 51° and 54°. The parameter \( \zeta \) reaches a maximum again at about 1500 km corresponding to a distance of 58°. Beyond 1500 km, \( \zeta \) decreases steadily so that there is an extended region of low velocity gradients down to about 1850 km; it then increases slowly, reaching a maximum at 2000 km, at which the equivalent distance is 72.8°. This may indeed suggest a slight anomaly in the \( \frac{dT}{d\Delta} \) curve at 70.5° as reported by Johnson (1969) and Corbishley (1970). Since the data for any one region give a poor coverage between 67° and 72°, more \( \frac{dT}{d\Delta} \) measurements over this distance range are required to resolve the problem.
Figure 8.4 A possible velocity model for the D'' region.
Table 8.3. Parameters of a Possible Velocity Model for the D'' Region of the Mantle and Associated Travel Times.

(a) Velocities.

<table>
<thead>
<tr>
<th>Depth, km</th>
<th>Velocity, km/sec</th>
</tr>
</thead>
<tbody>
<tr>
<td>2595</td>
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</tr>
<tr>
<td>2735</td>
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</tr>
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<tr>
<td>2800</td>
<td>13.8800</td>
</tr>
<tr>
<td>2894</td>
<td>13.7000</td>
</tr>
</tbody>
</table>

Velocities above 2595 km correspond to those of model ANUW3 (see Table 8.1).

(b) Travel Times.

<table>
<thead>
<tr>
<th>p, sec/deg</th>
<th>Δ, deg</th>
<th>T, sec</th>
<th>Depth of Penetration, km</th>
</tr>
</thead>
<tbody>
<tr>
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<td>763.65</td>
<td>2530.1</td>
</tr>
<tr>
<td>4.90</td>
<td>88.13</td>
<td>772.25</td>
<td>2593.3</td>
</tr>
<tr>
<td>4.80</td>
<td>90.64</td>
<td>784.47</td>
<td>2658.8</td>
</tr>
<tr>
<td>4.69</td>
<td>92.76</td>
<td>794.53</td>
<td>2731.1</td>
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<tr>
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<td>80.32</td>
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<td>2740.1</td>
</tr>
<tr>
<td>4.50</td>
<td>86.71</td>
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<td>2794.2</td>
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<td>88.77</td>
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<td>2800.6</td>
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<td>96.68</td>
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<td>4.47</td>
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<td>828.24</td>
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</tr>
</tbody>
</table>
Below 2000 km, $\xi$ decreases to a minimum at about 2100 km, and increases again to a maximum at 2360 km for which the corresponding distance of emergence of a ray is $80.0^\circ$. The velocity gradients are very low between 2460 and about 2600 km.

No attempt was made to derive a velocity distribution below 2600 km. Beyond $88.2^\circ$, the corrected $dT/d\Delta$ values were all significantly lower, but the data were not sufficiently copious to enable a reliable smooth curve to be obtained. A straight line fitted through the ten $dT/d\Delta$ measurements at distances greater than $88.2^\circ$ shows a positive gradient, which is obviously unacceptable. So, as an alternative procedure, a weighted mean of the $dT/d\Delta$ values was calculated and found to be $4.468 \pm 0.028$ sec/deg; further, $x^2$ was 10.62 on 9 degrees of freedom, which indicates that there is no statistical evidence for any increase or decrease in $dT/d\Delta$ between $88.9^\circ$ and $98.8^\circ$. If it is assumed that the average value of $dT/d\Delta$ indicates $r/v$ immediately below a first order discontinuity at a depth of 2700 km, the calculated velocity is 14.34 km/sec, which is too high. However, if the value of $dT/d\Delta$ is assumed to be representative of the value of $r/v$ at the mantle-core boundary, taking the depth of the core to be 2894 km (Taggart and Engdahl, 1968), the velocity is 13.60 km/sec, which is reasonable. It is likely that the $dT/d\Delta$ values have not been adequately corrected for local structure. But for events to the west of the array, on the limited data available, the sharp decrease in $dT/d\Delta$ appears to be real. In addition, recent P wave velocities provided in the 1968 Seismological Tables for P Phases, and by Hales et al. (1968), differ significantly below 2700 km, so that the velocities in the lowest 200 km of the mantle are not known accurately. A possible interpretation of the $dT/d\Delta$ data might be that there is a sharp increase in P wave velocity at about 2735 km, followed by a decrease to about 13.6 to 13.7 km/sec at the mantle-core boundary. Such a model that gives acceptable $dT/d\Delta$ values is illustrated in Figure 8.4, and the relevant parameters are listed in Table 8.3. The low $dT/d\Delta$ values mean that the travel times become too fast beyond $90^\circ$. At $86.5^\circ$ and $96.5^\circ$, the travel
Figure 8.5  Velocity models of Jeffreys, Johnson (CIT208) and models ANUW1, ANUW2 and ANUW3.
Figure 8.6: dT/dΔ curve for model ANUW3.

dT/dΔ, sec/deg
Table 8.4.  \( p-\Delta \) and \( T-\Delta \) Curves for Model ANUW3.

<table>
<thead>
<tr>
<th>( p ), sec/deg</th>
<th>( \Delta ), deg</th>
<th>( T ), sec</th>
<th>Depth of Penetration, km</th>
<th>( p ), sec/deg</th>
<th>( \Delta ), deg</th>
<th>( T ), sec</th>
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<td>1570.9</td>
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</table>
times are 764.2 and 809.3 seconds respectively, whereas the corresponding
times from the 1968 Seismological Tables for P Phases are 764.0 and 810.7
seconds. Further, a ray emerging at a distance of 96.7° penetrates to a depth
of only 2816 km, so that the model gives only a crude representation of the
P velocities at the base of the mantle.

Regions of anomalous velocity gradients are certainly present in
velocity model ANUW2, but since a smooth continuous curve with a continuous
first derivative has been fitted through the dT/dΔ data, the regions of rapidly
changing velocities are spread over an appreciable depth interval. Moreover,
it seems more likely that the rapid velocity changes, if due to phase transform-
ations, should be concentrated into narrower depth intervals (A.E. Ringwood,
personal communication, 1969). Therefore an alternative model ANUW3,
which will eventually be refined to satisfy the dT/dΔ data equally well, has
been constructed.

(c) Refined Model with Adjustment of Anomalous Regions. ANUW3.
(See Figures 8.5 and 8.6 and Tables 8.1 and 8.4).

Model ANUW3 was prepared by adjusting the parameters of models
ANUW1 and ANUW2 in the light of the evidence for regions of anomalous velocity
gradients discussed in Chapters 6 and 7. First, the velocities of model ANUW1
to a depth of 815 km have been used, and then a low velocity zone, similar to
case 2 of Figure 6.6 and Table 6.1, was added below 815 km. Thus a sharp
discontinuity has been introduced at 815 km, and the velocities subsequently
increase rapidly between 835 and 855 km, and fit smoothly on to model ANUW3
at 880 km. Below 900 km, all of the regions of anomalously high velocity
gradients have been replaced by small first order discontinuities. These
discontinuities have been introduced at the depths at which the velocity gradients
in model ANUW2 are maximum (viz. 1032.5 km, 1205 km, 1505 km, 2000 km
and 2355 km). A velocity model similar in form to that of Figure 8.4 has been
used to represent the region below 2600 km. These first order discontinuities
were inserted by increasing and decreasing the velocities by approximately
Figure 8.7  Regional $dT/d\Delta$ curves.
equal amounts at the depths concerned. The velocity model and the associated
dT/dΔ curve are shown in Figures 8.5 and 8.6, and the travel times and
dT/dΔ values are listed in Table 8.4. In Figure 8.5 the three models ANUW1,
ANUW2 and ANUW3 are shown for comparison.

It is important to stress the fact that the velocity changes for the
discontinuities at 1205, 2355 and 2735 km have probably been made too large.
Each produces two phases with significantly different travel times more than
two degrees away from the cross-over points at about 46.5°, 78.8° and 88.8°.
The existence of such phases whose onsets are separated by more than one
second cannot be dismissed entirely at present. The parameters of model
ANUW3 will be adjusted in the future to yield better travel times for the two
branches of the dT/dΔ curve associated with each of the discontinuities at
1205 km and 2355 km. The model itself has been introduced primarily to
present an alternative, and perhaps physically more realistic, interpretation
of the dT/dΔ data, and to emphasise the non-uniqueness of the solution obtained
by the Herglotz-Wiechert method.

8.3 REGIONAL dT/dΔ PROFILES.

Eventually it will be possible to derive lower mantle models on a
regional basis, but at present the dT/dΔ data are not sufficiently detailed to
make this worthwhile. However, two regional dT/dΔ curves are presented
in Figure 8.7, and the profile names refer to the geographic regions relative
to Tennant Creek. Since the dT/dΔ data were examined by region in Chapters
6 and 7, no discussion of the detailed shapes of the smooth curves is necessary,
though it is important to realise that the differences in these shapes are due
largely to the uneven distribution of the dT/dΔ data, particularly for the north­
west regional profile.

(a) West Pacific Margin. (See Table 8.5).

The dT/dΔ curve was derived by interpolating between the summary
points of the dT/dΔ profiles 2R, 5R and 6R of Table 4.5, adjusted to give the
correct travel times. The last two summary points of profile 5R and the first
Table 8.5. Summary Points Used for Interpolating the $dT/d\Delta$ Curve for the West Pacific Margin.

<table>
<thead>
<tr>
<th>Distance Range,</th>
<th>First Summary Point</th>
<th>Second Summary Point</th>
<th>Third Summary Point</th>
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<td>$dT/d\Delta$, sec/deg</td>
<td>$\Delta$, deg</td>
<td>$dT/d\Delta$, sec/deg</td>
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<td>35.965</td>
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<td>41.917</td>
<td>8.004</td>
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two of 6R were replaced by three points obtained by smoothing the corrected values of \( \frac{dT}{d\Delta} \) from both profiles between 77.6° and 83.7°.

(b) Northwest Regional Profile.

The smooth curve was fitted through all the corrected \( \frac{dT}{d\Delta} \) values for events in the northwesterly quadrant, and the summary points are listed in Table 4.7. Unfortunately the data are very sparse between 53° and 64°.

8.4 CLASSIFICATION OF EVENTS AND STATISTICAL SIGNIFICANCE TESTS ON THE DATA.

In Chapter 2 an event classification scheme was devised according to the quality of the P onset and the number of seismometers working. The question arises as to how the scatter of the data points is related to the quality of the recorded information. Does the 'external' consistency of a set of onset times provide an adequate means of weighting the data? If not, is there a tendency for the errors in the poorer quality data to be underestimated, due to fortuitous coherence of the relative onset times, perhaps caused by coherent noise or mismatching of different portions of the wave form? Last, and probably most important, is the scatter of the data greater for deep focus earthquakes?

No attempt has been made to present a really rigorous statistical investigation of these problems, but a simple calculation will enable any obvious effects to be detected. All of the data of Figure 8.2 up to a distance of 88.2° were used to define a \( \frac{dT}{d\Delta} \) curve fitted in exactly the same way as the curve shown in Figure 8.2, except that no data were rejected. Thus the distance ranges used were those of Table 4.6. The total \( \chi^2 \) of 1386.3 on 574 degrees of freedom is much larger than one would expect if the data were normally distributed. The individual \( \chi^2 \) values were grouped according to the eight main divisions of Table 2.4, and the ratio total \( \chi^2 \)/number of data points was calculated. The results are displayed in Table 8.6. It appears that the poorer quality data of groups C, D, \( \gamma \) and \( \delta \) may have been given too much weight, since the errors have probably been slightly underestimated. However, many events
Table 8.6. $\chi^2$ Values for All dT/d\(\Delta\) Measurements Grouped According to the Classification Scheme of Table 2.4.

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<td>1.71</td>
</tr>
<tr>
<td>All shallow events</td>
<td>873.83</td>
<td>442</td>
<td>1.98</td>
</tr>
<tr>
<td>(\alpha)</td>
<td>23.91</td>
<td>6</td>
<td>3.98</td>
</tr>
<tr>
<td>(\beta)</td>
<td>93.10</td>
<td>45</td>
<td>2.07</td>
</tr>
<tr>
<td>(\gamma)</td>
<td>228.43</td>
<td>71</td>
<td>3.11</td>
</tr>
<tr>
<td>(\delta)</td>
<td>33.70</td>
<td>29</td>
<td>1.16</td>
</tr>
<tr>
<td>All events at depths &gt; 130 km</td>
<td>379.14</td>
<td>151</td>
<td>2.51</td>
</tr>
<tr>
<td>One rejected event *</td>
<td>133.37</td>
<td>2</td>
<td></td>
</tr>
<tr>
<td>All events</td>
<td>1386.35</td>
<td>595</td>
<td></td>
</tr>
</tbody>
</table>

* Event 261 was used in deriving the smooth curve, but the dT/d\(\Delta\) values were subject to a very large bias. Consequently it was not included in the calculations shown in the table.
obviously subject to considerable bias were of high quality. Also, the scatter of the \( \frac{dT}{d\Delta} \) data from events at depths greater than 130 km seems to be greater, though it is emphasised that because deep focus earthquakes are concentrated in specific regions, the observed effect is not necessarily related to focal depth. It is therefore concluded that the weighting scheme used in preparing the \( \frac{dT}{d\Delta} \) curves is not entirely satisfactory, but apart from focal depth, no single factor likely to cause bias for a particular event has been detected. This is perhaps to be expected if multiple arrivals in the first second or so of an event, due to structure deep in the mantle, really exist.

8.5 **COMPARISON WITH OTHER INVESTIGATIONS.**

(a) **Travel Times and Amplitudes.**

In Chapters 6 and 7, when discussing the identification of specific anomalies in the lower mantle, brief comparisons were made with the results of other studies, particularly array investigations. Now that certain anomalies have been identified or suggested, and their depths have been calculated, it is necessary to provide a more detailed comparison to enable a plan for future studies of the deep interior of the earth to be evolved.

Initially the results are compared with recent P wave travel-time investigations. The detailed travel-time studies of Herrin et al. (1968) and Cleary and Hales (1966) have not revealed any distinct anomalies in the lower mantle. Thus it is likely that any irregularities in the structure of the lower mantle are either too small to be detected by the conventional travel-time method, or can only be detected by an extensive study of one or more events, preferably nuclear explosions, in a particular region of the earth. Of some interest in this respect is the shell model of Carder (1964) derived from travel times of central Pacific nuclear explosions. He found apparent breaks in the P travel-time curve near \( 39^0, 52^0, 69^0, 79.5^0, \) and \( 89.5^0 \), which he interpreted in terms of sharp velocity increases at depths of 1066, 1371, 1916, 2371 and 2771 km. These results are interesting as the first and third discontinuities
correspond to the smaller less clear velocity anomalies of model ANUW3 (Figure 8.5). The fourth and fifth discontinuities agree well with the dT/dΔ anomalies at 80° and beyond 88°. The second break in Carder's travel-time curve, close to 52°, could easily correspond to the sharp decrease in my dT/dΔ data near 47°, in which case his second discontinuity is about 150 km too deep. Bugayevski (1964) found evidence for first order discontinuities in the travel-time curve near distances of 35° - 38°, 50° - 54° and 70° - 72°, and suggested that they were due to low velocity zones; the first two of these do agree quite well with the regions of low velocity gradients around 830 km and 1300 km inferred from WRA dT/dΔ measurements. Regions of anomalous velocity gradients in the lower mantle have been suggested in many other travel-time studies, but no coherent pattern has emerged. Such studies have been reviewed by Anderson (1967b), Chinnery (1969) and Johnson (1969).

Since, over a limited distance range, the amplitude of P waves is approximately proportional to \( \sqrt{|d^2 T/d\Delta^2|} \), a flattening or a rapid decrease of the dT/dΔ curve should correspond to a minimum or a maximum in the amplitude curve respectively (see for example Carpenter, 1966). The amplitudes of short-period P waves from nuclear explosions between 30° and 102° have been studied by Carpenter et al. (1967). Their preliminary curve shows a sharp peak between 33° and 36°, a slight minimum followed by a marked increase at about 75°, and a sharp minimum between 93° and 96°. It is tempting to speculate that the peak beyond 30° is associated either with the rapid increase in velocity below the region of low velocity gradients near 850 km, or perhaps with a region of overlap of two or three branches of the travel-time curve, or a caustic, caused by a low velocity zone; further, the increase in amplitude beyond 75° might be associated with the rapid decrease in dT/dΔ near 80°. However, the prominent features found by Carpenter et al. were not present in the study by Cleary (1967) of the amplitudes of short-period P waves recorded by LRSM stations. The relation between amplitudes and dT/dΔ will be complicated if the multiplicities in the dT/dΔ curve, separated by less than a second, that were suggested earlier are real. Thus high amplitudes do
not necessarily imply high velocity gradients. Therefore a more explicit comparison of my results with amplitude studies is not justified at present.

(b) Array $dT/d\Delta$ Measurements.

In view of the difficulties involved in interpreting array data, the only way of establishing or denying the presence of the less clear anomalous regions in the lower mantle is to compare the results of several investigations. The aim in this final section is to show that the major discrepancies in the results of recent array studies of the lower mantle are largely due to differences in interpretation resulting from the scatter and irregular distribution of the basic $dT/d\Delta$ data, and not necessarily to subtle irregularities in structure at the array site, or to regional differences in structure at depths below 900 km.

The first array study of the lower mantle was published by Chinnery and Toksöz (1967). They used 167 events in the distance range $27^\circ$ to $90^\circ$ recorded at the LASA in Montana. Their $dT/d\Delta$ curve shows three regions between $32^\circ$ and $40^\circ$, $51^\circ$ and $55^\circ$, and between $72^\circ$ and $78^\circ$ where there are relatively few points. In each case the curve appears to flatten before the gap. In their paper they discussed only the flat portions of their curve, but note also that there is a comparatively steep region near $78^\circ$; this seems to be exaggerated as a result of having little data between $72.5^\circ$ and $77.5^\circ$. Four additional papers concerned with array data and lower mantle structure were published during 1969 and early 1970 when my own work was nearing completion.

Greenfield and Sheppard (1969) were concerned largely with Moho depth variations under the LASA and their effect on $dT/d\Delta$ measurements. They obtained two $dT/d\Delta$ curves: one for NW azimuths and one for SE azimuths. They remarked that they found small absolute values of $d^2T/d\Delta^2$ between $30^\circ$ and $34^\circ$, and their $dT/d\Delta$ curve for NW azimuths shows an increase in $dT/d\Delta$ in this distance interval. Both of their curves apparently show flat regions near $46^\circ$, though this is not obvious from their curve. Thus up to $46^\circ$ their results are in excellent agreement with mine. Beyond $60^\circ$, their NW curve is in reasonable agreement with both of my world average curves in the sense that
it does show some flattening before $70^\circ$, but their SE curve is rather puzzling. It is also worth pointing out that the numerical values of $dT/d\Delta$ in Greenfield and Sheppard's Figure 9 and Table 1 are seriously in error. $dT/d\Delta$ should be 9.0 sec/deg near $30^\circ$ and not 8.0 sec/deg. The authors claim that their $dT/d\Delta$ curves have been corrected for local structure. If this were true their $dT/d\Delta$ curves should appear as small perturbations of the J-B curve, but this is clearly not the case. Chinnery (1969) followed up the earlier work of Chinnery and Toksöz with another paper in which the author presented $dT/d\Delta$ measurements on P arrivals at the LASA from about 400 events at a northwesterly azimuth from the array. He found regions of anomalous velocity change at depths near 700, 1150 and 2000 km, and perhaps an additional anomaly at about 2500 km. There is a fairly flat region of his $dT/d\Delta$ curve near $30^\circ$, but unfortunately there are few observations between $31^\circ$ and $37^\circ$ so that the exact form of the anomaly is unclear. His data show an anomalous region between about $46^\circ$ and $49^\circ$, and he mentions the possibility of a triplication of the $dT/d\Delta$ values, presumably implying a fairly abrupt increase in velocity or velocity gradients. The data also suggest a flattening of the curve near $50^\circ$.

Chinnery also finds an extended region of slowly changing $dT/d\Delta$ between $65^\circ$ and $75^\circ$, and he indicates a possible triplication of the $dT/d\Delta$ values at about $77^\circ$. The possible velocity anomaly at a depth of about 2500 km may correspond to the flat region beyond $83^\circ$ followed by the sharp decrease beyond $88^\circ$, found in my own work. It is important to remark that Chinnery does not indicate clearly the evidence for the existence of triplications, and his velocity-depth profiles only give a rough idea of how they are produced.

My own work on $dT/d\Delta$ curves for regions of anomalous velocity changes, summarised in the sequence of diagrams of Figure 6.10, gives a clear physical picture of how such multiplicities can be produced. The results of Chinnery's $dT/d\Delta$ study correspond more closely with mine than those of any other array investigation. The only real discrepancy is in his interpretation of the data between $70^\circ$ and $80^\circ$. Chinnery emphasises the anomalously low velocity gradients before $75^\circ$, but his data do not preclude a very rapid decrease in
dT/dA between $75^\circ$ and $80^\circ$. In these triplications suggested by Chinnery, Johnson and myself, the individual branches are likely to be too close together to be observed separately on seismograms.

Johnson (1969) used the extended array at the TFSO to measure dT/dA for direct P waves from 212 earthquakes over the distance range 30$^\circ$ to 100$^\circ$. His curve fit shows anomalous regions near epicentral distances of 34.5$^\circ$, 40.5$^\circ$, 49.5$^\circ$, 59.5$^\circ$, 70.5$^\circ$ and 81.5$^\circ$ which may correspond to increased velocity gradients near depths of 830, 1000, 1230, 1540, 1910 and 2370 km. He remarks that the offsets near 34.5$^\circ$ and 70.5$^\circ$ are small. Also, his scheme used to fit the CIT 208 and CIT 210 curves to the dT/dA data tended to emphasise the anomalously steep portions of the curve. The scheme for fitting the CIT 206 curve contains no previous assumptions about the shape of the curve, and although it contains regions that are both steeper and flatter than average, the steeper regions predominate. The anomalous regions of his dT/dA data at 49.5$^\circ$ and 81.5$^\circ$ are in reasonable agreement with Chinnery’s results and mine, and the slight anomaly at 34.5$^\circ$ may result from the rapid increase in velocities below the region of low velocity gradients near 830 km. However, the anomalies at 40.5$^\circ$ and 59.5$^\circ$ must be regarded with some suspicion; in both cases the irregularity is introduced when the data points change from the northeast to the southwest quadrant. These two anomalies could therefore result from inadequate allowance for structure in the vicinity of the array.

The only published array dT/dA study for the lower mantle, apart from mine (Wright, 1968, 1970), using a single medium aperture array, was undertaken by Gopalakrishnan (1969). He made dT/dA measurements over the distance range 25$^\circ$ to 95$^\circ$ using P arrivals from 88 earthquakes recorded at the Gauribidanur Array in India. He used too little data to enable a really detailed velocity model to be deduced, but his observations do suggest low velocity gradients near 750 km and relatively high velocity gradients near 1200 km. Corblishley (1970) derived a dT/dA curve for the lower mantle using all four
Table 8.7. Anomalous Features of the Lower Mantle Inferred from Array dT/dΔ Investigations.

<table>
<thead>
<tr>
<th>Approximate Depth Range, km</th>
<th>Corresponding Distance Range, deg</th>
<th>Velocity Gradients</th>
</tr>
</thead>
<tbody>
<tr>
<td>800 - 850</td>
<td>32 - 37</td>
<td>Low and probably negative over part of the range. Regional differences possibly important.</td>
</tr>
<tr>
<td>1070 - 1110</td>
<td>44 - 46</td>
<td>Low</td>
</tr>
<tr>
<td>1160 - 1220</td>
<td>47 - 48.5</td>
<td>High</td>
</tr>
<tr>
<td>1260 - 1330</td>
<td>50 - 53</td>
<td>Low</td>
</tr>
<tr>
<td>1750 - 1850</td>
<td>65.5 - 69.5</td>
<td>Low</td>
</tr>
<tr>
<td>2180 - 2370</td>
<td>77.5 - 80.5</td>
<td>High</td>
</tr>
<tr>
<td>2460 - 2600</td>
<td>83 - 88</td>
<td>Low</td>
</tr>
<tr>
<td>2700 - 2750</td>
<td>Beyond 88°</td>
<td>High?</td>
</tr>
</tbody>
</table>

High velocity gradients or sharp increases of velocity may also occur at depths of approximately 1000, 1500 and 1900 km.
Figure 8.8  $dT/\Delta$ curves derived by Chinnery and Toksöz (1967), Johnson (1969), Corbishley (1970) and Herrin et al. (1968) compared with the $dT/\Delta$ curve of Figure 8.2.
U.K.A.E.A. arrays. His method of analysis of the relative arrival times at
the arrays involves attempting to estimate dT/dΔ and the crustal effects
beneath each array simultaneously. He found regions of rapidly changing
dT/dΔ at 35° - 36°, 48° - 49°, 68° - 70° and 84° - 85°.

When a comparison is made of all the array studies discussed
above, a reasonably coherent picture emerges. The anomalous features of
the lower mantle, either established or suggested by my own work supplement­
ed by the results of other studies, are summarised in Table 8.7. The irregul­
arities suggested by Johnson at depths of 1000, 1540 and 1910 km are not
confirmed either by my data or by data from the LASA in Montana, though the
dT/dΔ data presented in this thesis do yield slight maxima in the velocity
gradients close to these three depths. The dT/dΔ curves from Chinnery and
Toksöz (1967), Corbishley (1970) and the 1968 Seismological Tables for P
Phases, and my own refined dT/dΔ curve are plotted in Figure 8.8, together
with the dT/dΔ curve derived from model CIT 208 of Johnson (1969). Because
of the differences in the mathematical techniques used in smoothing the dT/dΔ
data, a visual comparison of these curves is a little misleading. It is also a
pity that Chinnery (1969) did not supply any parameters of his dT/dΔ curves,
particularly as his results agree fairly closely with mine. The alarming
difference in the features of Corbishley's curve when compared with other
studies leads one to suspect that there may be a fallacy in his method of analysis.
My own view is that the mathematical model he uses to give site corrections
(equation 4 of Corbishley, 1970) is completely inappropriate; it is clear from
the results of Chapter 3 that such a model does not correct adequately for
local structure at WRA, and this seems to me to be the cause of the trouble.
As far as other array studies are concerned, the differences in array dT/dΔ
curves and their corresponding velocity models can be adequately explained
by the scatter and irregular distribution of the basic dT/dΔ data. Regional
differences in structure below 900 km do not have to be invoked, though this
is not meant to imply that they do not exist. There is indeed some evidence
in my data that regional differences may persist to a depth of 1200 km. Although
certain velocity anomalies have been identified, the sharpness of velocity increases is not known; relatively abrupt but fairly small increases in velocity may give rise to duplications or triplications in the travel-time curve in which the proximity of the individual branches will cause just a large scatter in the $dT/dA$ data, thus tending to smear out or even mask the anomaly itself.

In Figure 8.5 the velocity model of Jeffreys, derived from the J-B travel times for P, and Johnson's model CIT 208 have been plotted for comparison with models ANUW1, ANUW2 and ANUW3. From depths of 700 km to 2300 km the shapes of my velocity distributions ANUW1 and ANUW2 are similar to Johnson's, but rather different from that of Jeffreys. Model ANUW1 generally shows higher velocities than ANUW2. The refined model ANUW2 is closer to Johnson's model than ANUW1, and also yields travel times in better agreement with those of Herrin et al. (1968); it is therefore a substantial improvement on ANUW1, though the random errors in the $dT/dA$ curves can explain the differences between the models.
CHAPTER 9

FURTHER APPLICATIONS OF THE ARRAY

9.1 INTRODUCTION.

Although arrays were designed specifically to improve the signal to noise ratio of distant seismic events, their signal enhancing ability has not yet been extensively adapted to studies of the structure of the earth's interior. Applications to crustal studies have been described by Agger and Carpenter (1964) and by Birtill and Whiteway (1965). Key (1967) investigated the signal-generated noise at the Eskdalemuir array. Whitham and Weichert (1968) used velocity filtering of Yellowknife records of the Early Rise chemical explosions to investigate upper mantle structure beneath the Canadian Shield. Hannon and Kovach (1966) used velocity filtering of TFSO records combined with adjacent LRSM stations forming a large array of 400 km in aperture to study core phases, and obtained results that appeared to support the core model of Adams and Randall (1964). Towards the end of 1967 it was evident that a large quantity of valuable information stored in array tape records was virtually unexploited. Consequently a short research project designed to exploit some of this information and to advance some ideas for future array investigations was undertaken in collaboration with Dr. K.J. Muirhead. We attempted to show, using the WRA record of the Novaya Zemlya nuclear explosion of October 27, 1966, that both visual matching and array phasing techniques can be used to great advantage in studying the usual P phases following the first arrival and in investigating coherent signals that do not correspond to any of the conventional phases, particularly precursors to PP (see Wright and Muirhead, 1969).
9.2 THE NOVAYA ZEMLYA EXPLOSION.

The Novaya Zemlya nuclear explosion of October 27, 1966, at a distance of 106.0° (event 96 of Appendix 2), was well recorded by the array, and a number of later phases as well as P were visible. The high quality of the record of P, together with accurate knowledge of the expected apparent velocity of diffracted P, makes this explosion ideal for determining a crustal structure beneath the array. A discussion of possible crustal structures derived from the measured apparent velocity and azimuth of the P onset has been given in Chapter 3. These structures, although of limited value in correcting measurements of $dT/d\Delta$ for events in the distance range $30^0 - 100^0$, do at least provide first order corrections to apparent velocities for the later P phases in the explosion record. The WRA tape recording of the explosion was bandpass-filtered between 0.4 and 2.0 cps and transferred on to paper at a speed of about 8 mm/sec; this revealed some exceptionally clear later P phases, particularly PKiKP and PKKp, and numerous other signals that appeared to be coherent across the array. All coherent signals following P were picked out visually from the paper record, and the correlation method described by Birtill and Whiteway (1965) was used to give an estimate of the apparent velocity and azimuth of each signal. The computer technique used for this has already been described briefly in section 2.2 (ii).

Jeffreys (1962a, pp. 104-105) has pointed out the difficulty of explaining the continuous irregular oscillations that follow P and S. Several hypotheses have been put forward to explain these oscillations. For example, it has been suggested that the general irregular movement is impressed in the immediate neighbourhood of the observing station because of scattering. Key (1967) has investigated the signal-generated noise at the Eskdalemuir array, and has shown a close correlation between low velocity apparent noise sources and topographic features; these signals are Rayleigh waves. The exceptional quality of the record of the Novaya Zemlya explosion has enabled some of the oscillations following P that are coherent across the array to be explained.
Table 9.1. Correlation Measurements.

This table gives a selection of correlation measurements of apparent velocity and azimuth for some of the later P phases and coherent signals. A 3-second integration time was used for correlation measurements except for entries otherwise indicated.

<table>
<thead>
<tr>
<th>Name of Phase</th>
<th>Time for Start of Correlation, (GMT)</th>
<th>Expected Apparent Velocity, km/sec</th>
<th>Expected Apparent Velocity Assuming Two Dipping Interfaces, km/sec</th>
<th>Measured Apparent Velocity, km/sec</th>
<th>Measured Azimuth, deg</th>
</tr>
</thead>
<tbody>
<tr>
<td>P</td>
<td>6h 12m 12.0s*</td>
<td>24.55(^a)</td>
<td>22.10</td>
<td>21.0</td>
<td>344.2</td>
</tr>
<tr>
<td></td>
<td>6h 13m 18.0s(^x)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>6h 13m 22.0s(^x)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>6h 13m 33.5s</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>6h 15m 00.0s</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>6h 15m 24.8s</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>6h 15m 27.4s</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>6h 15m 32.5s</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>6h 15m 38.5s</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>6h 15m 50.0s</td>
<td></td>
<td></td>
<td></td>
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</tr>
<tr>
<td></td>
<td>6h 15m 58.0s</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>PKiKP</td>
<td>6h 16m 24.2s</td>
<td>59.0(^b)</td>
<td>46.8</td>
<td>45.9</td>
<td>334.3</td>
</tr>
<tr>
<td>PP</td>
<td>6h 16m 34.0s</td>
<td>15.07(^c)</td>
<td>14.0</td>
<td>14.0(^d)</td>
<td>342.0</td>
</tr>
<tr>
<td></td>
<td>6h 16m 54.0s</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>6h 17m 19.5s</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>6h 17m 46.0s</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>6h 17m 53.0s</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>PcPPcP</td>
<td>6h 18m 54.0s</td>
<td>29.6(^d)</td>
<td>26.1</td>
<td>31.3</td>
<td>343.2</td>
</tr>
<tr>
<td></td>
<td>6h 18m 59.5s</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>6h 19m 07.3s</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>6h 24m 02.5s</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>6h 25m 17.0s(^x)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>6h 25m 30.0s</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>6h 25m 37.5s</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>6h 26m 04.5s</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>PKKP A</td>
<td>6h 27m 47.1s</td>
<td>39.0(^d)</td>
<td>47.0</td>
<td>46.8</td>
<td>162.4</td>
</tr>
<tr>
<td>PKKP B</td>
<td>6h 27m 50.5s</td>
<td></td>
<td></td>
<td></td>
<td>165.1</td>
</tr>
<tr>
<td>PKKP C</td>
<td>6h 28m 01.8s(^x)</td>
<td>62.0(^d)</td>
<td>84.3</td>
<td>76.0</td>
<td>157.0</td>
</tr>
<tr>
<td>PKKP D</td>
<td>6h 28m 03.1s(^x)</td>
<td>25.0(^d)</td>
<td>28.2</td>
<td>29.6</td>
<td>167.1</td>
</tr>
<tr>
<td></td>
<td>6h 31m 52.5s</td>
<td></td>
<td></td>
<td></td>
<td>145.9</td>
</tr>
<tr>
<td>PcPPKP</td>
<td>6h 32m 00.0s(^x)</td>
<td>25.4(^d)</td>
<td>28.8</td>
<td>23.0</td>
<td>187.0</td>
</tr>
<tr>
<td>PKPPKP</td>
<td>6h 36m 09.5s(^x)</td>
<td>58.0(^d)</td>
<td>78.0</td>
<td>82.0</td>
<td>201.3</td>
</tr>
</tbody>
</table>
Table 9.1 (Contd)

* 1-second integration time used.
\( \neq \) 2-second integration time used.
\( \neq \) 1.2-second integration time used.
\( \phi \) For PP the correlation peak occurred at a velocity of 39.6 km/sec and an azimuth of 304°, and the values quoted in the table correspond to the maximum of a slightly smaller peak. These results are due to interference from the waves associated with PKiKP or PKIKP.

\( a \) Sacks (1967).
\( b \) Bolt and O'Neill (1965).
\( c \) Cleary and Hales (1966).
\( d \) Jeffreys and Bullen (1958).
Table 9.2. Analysis of Phases from the Novaya Zemlya Explosion

<table>
<thead>
<tr>
<th>Name of Phase</th>
<th>Travel Time</th>
<th>Least-Squares Measurements</th>
<th>Matching Technique Used</th>
<th>Mean Square Residual, sec^{-1}</th>
<th>Correlation Method</th>
<th>Expected Apparent Velocity, km/sec</th>
<th>Expected Apparent Azimuth, deg</th>
<th>Azimuth Allowing for Structure, deg</th>
</tr>
</thead>
<tbody>
<tr>
<td>P</td>
<td>14m13.5s</td>
<td>22.14\pm0.20 344.4\pm0.7 First peak 0.135</td>
<td>First peak 0.494 21.0 344.2 22.10 344.5</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>P</td>
<td>17m26.8s</td>
<td>19.64\pm0.27 347.7\pm1.0 First peak 0.399 18.8 350.0</td>
<td>Second peak 0.355</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>P</td>
<td>17m29.4s</td>
<td>21.02\pm0.35 344.5\pm1.0 First peak 0.467 18.9 349.6</td>
<td>Second peak 0.750</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>PKiKP</td>
<td>18m26.2s</td>
<td>45.25\pm1.73 322.4\pm3.2 First peak 0.771 45.9 334.3 46.7 346.2</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>PKiKP</td>
<td>18m28.1s</td>
<td>56.62\pm2.42 349.1\pm3.0 First peak 0.412</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>PP</td>
<td>18m37.2s</td>
<td>13.73\pm0.39 342.7\pm2.5 First peak 2.435 14.0 342.0 14.0 343.9</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>PKKP A29m49.1s</td>
<td></td>
<td>42.16\pm1.27 161.8\pm2.3 First peak 0.490 46.8 162.4 47.0 159.5</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>PKKP B29m52.5s</td>
<td></td>
<td>38.83\pm1.08 170.1\pm1.9 First peak 0.435 47.4 165.1</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>PKKP C30m03.8s</td>
<td></td>
<td>46.54\pm2.52 178.0\pm3.5 First peak 0.922 76.0 157.0 84.3 156.7</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>PKKP D30m05.1s</td>
<td></td>
<td>26.92\pm0.90 175.8\pm2.1 First peak 0.875 29.6 167.1 28.8 160.8</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

257.
in terms of reflections at a considerable distance from the array, as well as providing some indication of the reliability of measurements of $dT/d\Delta$ for identifiable phases using the correlation method. It must be stressed, however, that large unexplained signals are common at a distance of about 106°, so that the seismogram is probably not typical of other distance ranges; this point will be discussed later.

The results of this study of coherent signals following $P$ are given in Tables 9.1 and 9.2. In Table 9.1 only the clearest unidentified signals have been listed. The expected apparent velocities and azimuths allowing for structure, listed in Table 9.2, have been calculated using structure 2(a) of Table 3.3. For the large later phases it was possible to supplement the correlation method with hand measurements of the onset times and a least-squares azimuth and apparent velocity determination. The errors in the total travel times to the origin of the array are believed to be about 0.2 seconds for the phases of Table 9.2; for the other coherent signals the errors are larger. To give some indication of the quality of the least-squares fit, the mean square time residual is also displayed in Table 9.2. For these later phases seismometers were rejected only if the magnitude of the residuals exceeded 0.05 seconds. The least-squares fit for $PP$ was so poor, however, that it was not possible to adopt this rejection procedure. The reason for this is provided by the apparent velocity measurement of $PP$ using the correlation technique; in this case the correlation peak occurred at a velocity of 39.6 km/sec and an azimuth of 304°, and the apparent velocity and azimuth quoted in Table 9.1 correspond to the maximum of a slightly smaller peak. Thus, even 10 seconds after the onset of PKiKP there is still a significant amount of energy associated with waves transmitted through the core.

9.3 IDENTIFIED PHASES.

The travel times of all identified phases without ellipticity corrections are listed in Table 9.3. The ellipticity corrections calculated as described by Bullen (1937, 1938a, b) have been included in the table for all
Table 9.3. Travel Times of P Phases for the Novaya Zemlya Nuclear Explosion.

<table>
<thead>
<tr>
<th>Phase</th>
<th>Expected Travel Time</th>
<th>Ellipticity Correction</th>
<th>Actual Travel Time</th>
</tr>
</thead>
<tbody>
<tr>
<td>P</td>
<td>14m 14.9s</td>
<td>-0.6s</td>
<td>14m 13.5s</td>
</tr>
<tr>
<td>PKiKP</td>
<td>18m 25.8s*</td>
<td>-0.8s</td>
<td>18m 26.2s</td>
</tr>
<tr>
<td>PP</td>
<td>18m 41s</td>
<td>-0.2s</td>
<td>18m 37.2s</td>
</tr>
<tr>
<td>PcPPcP</td>
<td>20m 58.8s</td>
<td>-0.2s</td>
<td>20m 56s</td>
</tr>
<tr>
<td>PKKP A</td>
<td>29m 47.3s</td>
<td></td>
<td>29m 49.1s</td>
</tr>
<tr>
<td>PKKP B</td>
<td></td>
<td></td>
<td>29m 52.5s</td>
</tr>
<tr>
<td>PKKP C</td>
<td>29m 57.2s</td>
<td></td>
<td>30m 03.8s</td>
</tr>
<tr>
<td>PKKP D</td>
<td>30m 06.2s</td>
<td></td>
<td>30m 05.1s</td>
</tr>
<tr>
<td>PcPPKP</td>
<td>34m 04.2s</td>
<td></td>
<td>34m 02s</td>
</tr>
<tr>
<td>PKPPKP</td>
<td>38m 12.2s</td>
<td>-0.3s</td>
<td>38m 11.5s</td>
</tr>
</tbody>
</table>

*Bolt and O'Neill (1965).*
Figure 9.1 Some processed records of P. Array phased to an azimuth of 344.2° and to an apparent velocity of 21.0 km/sec. A 2-second square window of integration was used for trace 5 and for all correlator outputs in Figures 9.2 - 9.5.
Figure 9.2: PKiKP and PP. Array phased to an azimuth of 334.3° and a velocity of 45.9 km/sec for traces 1 and 2 and to 342.0° and 14.0 km/sec for traces 3 and 4.
phases except PKKP and PcPPKP. Owing to the uncertainty in the expected travel times and also in the ellipticity corrections themselves it was not considered worthwhile calculating ellipticity corrections for these phases. All phases arrived early relative to the expected travel times, except PKiKP and PKKP. It seems likely that the core model used by Bolt and O'Neill (1965) to calculate the travel times for PKiKP is slightly in error in comparison with the J-B times for other phases, giving PKiKP about 2 seconds earlier than observed. For P, PKiKP, PP, the PKKP phases and two large unidentified phases, d A/dT and azimuth were measured by the least-squares method as well as by correlation, thus enabling a direct comparison between the two methods. In the case of these larger phases, except PP, the values of velocity and azimuth determined by the least-squares method are undoubtedly more reliable than the values obtained by the correlation technique. The values of velocity and azimuth obtained by the two methods are not in good agreement in all instances. Figures 9.1-9.5 show summed and correlator outputs for P, two large unidentified phases, PKiKP, PP, PcPPcP and PKKP. These processed records have been produced on a digital computer using the correlation technique described by Muirhead (1968a).

Individual Phases.

PKiKP. The amplitude is nearly as large as P and is much clearer than previous observations of this phase at similar distances, such as those of Ergin (1967). Other reports of PKiKP have been given by Melik-Gajkazan (1955) and Caloi (1961). Melik-Gajkazan (1955) gives an empirical travel-time curve for PKiKP with forty-nine observations between 22^0 and 140^0, but does not reproduce any records. Bolt and O'Neill (1965) indicate that this phase is most likely to be observed for the distance range 105^0 < Δ < 110^0, and they consider Caloi's identification of PKiKP at 15^0 as unlikely. These authors have indicated that positive identification of this phase would enable an independent estimate of the rigidity of the inner core to be made. The measurements of d A /dT and azimuth suggest that PKiKP may be two separate
Figure 9.3  PePPeP.
Array phased to an azimuth of 343.2° and a velocity of 31.3 km/sec.
Traces 1 and 2 are phased to 167.1° and a velocity of 29.6 km/sec for traces 3 and 4.

Figure 9.4 PKKP. Array phased to an azimuth of 162.4° and a velocity of 46.8 km/sec for traces 1 and 2, and to 167.1° and 29.6 km/sec for traces 3 and 4.
phases. It is perplexing that measurements of the arrival times of the first two peaks of this phase give values of $dA/dT$ close to the expected value but azimuths quite different from the true value. Moreover, a later peak gives an azimuth close to the expected value but a rather high velocity. These results may be due to interference with a scattered signal.

PP. The values of velocity and azimuth obtained by both the correlation and the least-squares methods are in good agreement, but interference from other signals, most probably waves from the earth's core, resulted in a very poor least-squares fit. An interesting fact is that two unidentified phases show up more clearly than PP, and these are discussed later.

PcPPcP. Previous identifications of this phase are either very rare or absent. The evidence for its identification is very convincing; the azimuth and apparent velocity are in reasonable agreement with the expected values, and the onset time differs from the calculated arrival time (J-B) by less than 3 seconds. It is of some interest that this phase has a travel time that differs from the expected time for PPP by only 1 second. However, the measured apparent velocity rules out the possibility that it is PPP.

PKKP. According to the J-B tables, there are three possible paths for PKKP, each phase travelling in the reverse direction. The onset of the first PKKP is remarkably sharp. Both correlation and least-squares measurements of velocity and azimuth agree fairly well, but the results suggest four phases. The first and fourth of these have values of $dA/dT$ that agree well with the values for the first and third phases given by Jeffreys and Bullen; the arrival of the fourth phase is not clearly defined. One interpretation of these results is that the middle branch of PKKP given by Jeffreys and Bullen is really two separate branches. Evidence for this is provided by the J-B travel time which is almost halfway between the travel times for two of the observed phases. Further work is required to ascertain whether this interpretation is compatible with recent core models. Engdahl (1968) provides surface focus travel times
Figure 9.5  Two large unidentified phases. Phase A: P reflected at M-discontinuity. Phase B: P reflected at surface. See Table 9.4.
of PKKP calculated from the core model T2 of Bolt (1964), in which there are four branches at 106°, showing reasonably good agreement with the results presented in Table 9.2. The measured azimuths differ from the expected azimuths, suggesting that the structures derived from P are not valid for signals arriving from the opposite azimuth.

**PcPPKP and PKPPKP.** These phases are both quite small, but the correlation method gives apparent velocities in reasonable agreement with those expected; although these phases cross the array in the reverse direction the measured azimuths differ considerably from the expected values.

An unsuccessful search was made for the phases PKJKP and PKIIKP.

**9.4 UNIDENTIFIED SIGNALS.**

Between P and PP there are ten unidentified coherent signals all with measured apparent velocities between 17.4 and 22.9 km/sec and with azimuths within 45° of the azimuth of the explosion. These signals are spread over a time interval of 3 minutes 47 seconds, and it is unlikely that they are due to scattering or multiple reflections in the vicinity of the array. They appear to have originated, presumably by reflection, at a considerable distance from the array, though reverberation effects at the explosion site may be partly responsible. After PP there is a small unidentified phase, possibly a core phase, travelling in the reverse direction at 47.4 km/sec, and with a total travel time of 19 minutes 21.5 seconds; no adequate explanation of this phase has yet been found. Between PP and PKKP there are several signals arriving in the reverse direction with velocities of less than 30 km/sec. It is suggested that these signals are associated with reflection and scattering of core phases reaching the surface or a near-surface discontinuity between 106° and 180° from Novaya Zemlya.

There are two large phases, illustrated in Figure 9.5 and marked A and B, with travel times of 17 minutes 26.8 seconds and 17 minutes 29.4 seconds; the onset of the second of these is very sharp. Hand measurements
Figure 9.6  Diagram illustrating the reflection of the phase arriving at 6 h 15 m 24.8 s.
Table 9.4. Two Large Unidentified Phases Arriving between P and PP.

<table>
<thead>
<tr>
<th></th>
<th>P Reflected at 33 km</th>
<th>P Reflected at Surface</th>
</tr>
</thead>
<tbody>
<tr>
<td>Arrival time of phase (GMT)</td>
<td>6h 15m 24.8s</td>
<td>6h 15m 27.4s</td>
</tr>
<tr>
<td>Travel time</td>
<td>17m 26.8s</td>
<td>17m 29.4s</td>
</tr>
<tr>
<td>Corrected apparent velocity</td>
<td>21.65 km/sec</td>
<td>22.80 km/sec</td>
</tr>
<tr>
<td>Corrected azimuth</td>
<td>347.1°</td>
<td>343.8°</td>
</tr>
<tr>
<td>Point of reflection</td>
<td>60.7°N, 107.6°E</td>
<td>62.1°N, 98.1°E</td>
</tr>
<tr>
<td>$\Delta_1$, distance from WRA to point of reflection</td>
<td>83.2°</td>
<td>86.7°</td>
</tr>
<tr>
<td>$t_1$, travel time from point of reflection to WRA</td>
<td>12m 24s</td>
<td>12m 47s</td>
</tr>
<tr>
<td>$\Delta_2$, distance from Novaya Zemlya to point of reflection</td>
<td>23.2°</td>
<td>19.3°</td>
</tr>
<tr>
<td>$t_2$, travel time from Novaya Zemlya to point of reflection</td>
<td>5m 05s</td>
<td>4m 29s</td>
</tr>
<tr>
<td>Calculated travel time</td>
<td>17m 29s</td>
<td>17m 16s</td>
</tr>
</tbody>
</table>
of the onset times give azimuths fairly close to the true value of $342.9^\circ$. First the original interpretation and calculations given in Wright and Muirhead (1969) are considered, and then these results are discussed in the light of subsequent work on the structure beneath the array. With the assumption that these phases result from a simple reflection at a dipping interface, the points of reflection have been deduced from the azimuths and apparent velocities corrected for structure. The distance was obtained by comparing the measured velocity, corrected using structure 2(a) of Table 3.3, with the apparent velocities derived from the Cleary-Hales travel-time curve. The distances of these points of reflection from the epicentre were then calculated, and the P travel times for the two paths were added. These calculations on the two asymmetrical ray paths were performed using computer programme ZCW101 B2. For the first of these phases the travel time for P reflected at a dipping M-discontinuity agrees with the actual value to within 3 seconds, but the second larger phase does not fit so well, the travel time being 13 seconds longer than for a surface reflection. By taking the first phase, the angle of dip required to give such a reflection has been found to be about $10^\circ$ (see Figure 9.6 and Table 9.4). Further, the dip would have to be toward the mountain ranges of central Asia; the points of reflection are situated just to the north of the Tannu Ola and Yablanovy ranges.

One important question arises in connection with this interpretation; how reliable is the structure used to correct the azimuths and apparent velocities? In Chapter 4 it was shown that the structures derived from the Novaya Zemlya explosion are not really applicable even for events at $75^\circ$ from WRA and at the same azimuth as Novaya Zemlya. It therefore appears that structure 5 of Table 4.4 might be more appropriate. However, this structure gives virtually the same corrected azimuths as those listed in Table 9.4, but results in larger corrected apparent velocities. This makes the value of $\Delta_1$ slightly larger so that the calculated total travel times become smaller. Owing to the uncertainty in the apparent velocities due to other causes, this modification does not affect
the validity of the explanation of the two phases A and B. The problem of which is the better structure remains unresolved.

Points of reflection for all other phases, assuming they are some form of reflected P, have been determined in the manner outlined above, but attempts to fit travel times and ray paths have not been entirely successful. When the paper discussing these unidentified signals (Wright and Muirhead, 1969) was completed, the occurrence of a major seismic discontinuity in the mantle at a depth close to 400 km had already been established (see section 5.1), but whether the existence of first order seismic discontinuities in the mantle was possible was not then known. Consequently it was remarked that many of the unidentified phases could be explained if reflections at the lower sides of discontinuities in the upper mantle were postulated. While our paper was in press the problem of precursors to PP in the $105^\circ < \Delta < 110^\circ$ time window was discussed by Bolt, O'Neill and Qamar (1968). It has been believed for some time that an olivine-spinel phase transformation takes place at a depth of about 400 km in the mantle. Recently Ringwood and Major (1969) have found experimentally that this phase transformation is more complicated than was previously expected; they have indicated that olivine transforms via spinel into a beta or spinel-like phase, and that at the spinel-beta phase reaction point a discontinuous change in density and mineralogy occurs. These effects are likely to produce a first order seismic discontinuity near 400 km in the mantle. This evidence makes the possibility of mantle reflections, called PdP waves by Bolt et al. (1968), more plausible. Recently a controversy has arisen over the true nature of the waves observed by Wright and Muirhead, and by Bolt et al., and this is discussed in the next section.

9.5 DISCUSSION.

This study of later phases for a single event has demonstrated that hand measurements of relative onset times can give good results for apparent velocities and azimuths of later P phases as well as initial P. The correlation method described by Birtill and Whiteway (1965) has been shown
to be unsuitable for precise determinations of $d\Delta/dT$, but is very useful in picking out small phases when the signal to noise ratio is small; the values of $d\Delta/dT$ obtained by this method probably have an error of 15%, even when the signal to noise ratio is large. A new correlation technique that appears to give more reliable measurements of $d\Delta/dT$ and azimuth has been described by Muirhead (1968a), and is now being thoroughly investigated.

The importance of the measurements of $dT/d\Delta$ and azimuth for the precursors to PP listed in Table 9.2 was stressed by Bolt and Qamar (1969) in a letter to the Editor of the *Journal of Geophysical Research* commenting on the paper by Wright and Muirhead (1969). It is therefore pertinent at this stage to outline briefly what Bolt et al. (1968) have published on these unexplained signals, and then discuss their comments on our data. Bolt et al. have discussed the history of observations of unexplained arrivals between P and PP in the distance range $105^0 - 110^0$. This phenomenon had been described earlier by Gutenberg (1960) and also by Hai (1963) who attributed some of his observations to reflections of the PP type from the lower side of discontinuities in the upper mantle; others were explained in terms of upward reflection of the down-going P wave at a depth of about 920 km followed by downward reflection from the earth's surface. Bolt et al. also considered the possibility that some of these waves may have originated from the outer boundary of the main transition zone between the outer and inner cores, but concluded that waves of the PdP type provided a more probable explanation.

Bolt and Qamar (1969) were interested in the seven phases preceding PKiKP that are listed in Table 9.2 and in their Table 1. On the PdP hypothesis of Bolt et al. (1968) the phase A of Figure 9.5 (the fifth precursor of Table 9.2) would be interpreted as a wave reflected at a point almost symmetrically situated between Novaya Zemlya and WRA; according to Bolt and Qamar the reflecting layer would be horizontal and at a depth of about 325 km (353 km according to the J-B tables and 329 km according to the 1968 Seismological Tables for P Phases). Similarly Bolt and Qamar indicate
that the fourth precursor of Table 9.2 would be reflected at a depth of about 450 km, and that if this explanation is correct the measured slowness should be a little less than the value for PP (\( \simeq 7.1 \) sec/deg). The measured slowness is 6.1 sec/deg. They also remarked that the explanation in terms of PdP still deserves consideration owing to the uncertainties in the estimates of \( dT/d\Delta \), and because corrections for crustal structure need to be applied. This is true, but it must be pointed out that the structures of Table 3.3 or structure 5 of Table 4.4 will always decrease the measured value of \( dT/d\Delta \). Further, the values of \( dT/d\Delta \) for the phases A and B of Figure 9.5 were measured by the least-squares method and are therefore of high precision. The effect of the structure beneath the array is sufficiently well known to make it unlikely that the value of \( dT/d\Delta \) for either phase is greater than 6.0 sec/deg. Therefore phases A and B cannot be satisfactorily explained on the PdP hypothesis in its present form, as the expected value of \( dT/d\Delta \) is 7.2 sec/deg; this argument obviously does not apply to the other precursors.

It is believed that a large dip exists in the boundaries of a low velocity layer in the upper mantle below central California (Nuttli and Bolt, 1969). It seems plausible, therefore, that the M-discontinuity and layers within the upper mantle, including the possible first order seismic discontinuity near 400 km, may have dips of several degrees or more locally, thus producing asymmetric reflections of the PdP type. The correct explanation of these precursors thus seems most likely to be a combination of the original PdP and the dipping interface hypotheses, as indeed Bolt and Qamar have remarked. A detailed study of these waves using one of the U.K.A.E.A. type arrays would be invaluable for elucidating any fine structure or lateral variations in structure in the upper mantle. Bolt and Qamar have also reported that attempts to correlate precursors to PP in the 105° - 110° source window across the LASA in Montana have not been successful, even though the signals were clearly visible. It thus appears that only a medium aperture device will yield satisfactory results in this kind of study. Finally it is worth
commenting that it is significant that unidentified signals still arrive after PP, but in the reverse direction. These waves may be associated with reflection and scattering of core phases reaching the surface between $106^0$ and $180^0$ from Novaya Zemlya, and it is possible that their mode of production is related to that of the precursors. When examining these signals and the precursors to PP, it was noticed that if the ray paths were projected back to their apparent sources, these sources showed a tendency to cluster around major tectonic features of the earth. The data were neither sufficiently accurate nor sufficiently plentiful to enable any real significance to be attached to this trend at this stage. Nevertheless this feature should be looked for in future investigations.
CHAPTER 10

CONCLUSIONS

10.1 SUMMARY OF RESULTS.

The major objective in this thesis has been to examine the evidence for the existence of inhomogeneities in the lower mantle. In addition some useful results for the upper mantle have been obtained, and the signal enhancing ability of a medium aperture array has been shown to be particularly useful in studying both upper mantle and core structure. Throughout this work great difficulties have been encountered because of the complexity of the local structure beneath the array. No complete explanation of these irregularities has been possible owing to the inadequacy of the data available, but a technique for eliminating the effects of local structure from $dT/d\Delta$ measurements, as far as possible, has been described and applied with some degree of success. Bearing in mind the original aims of this work, it can be said with confidence that these aims have been largely fulfilled, in spite of the unsatisfactory nature of the array site. Moreover, the problem of correcting $dT/d\Delta$ values for local structure has at least led to a better appreciation of the effects of this structure on $dT/d\Delta$ measurements. The principal results and conclusions will subsequently be examined, together with some pertinent remarks on the approach to the problem when this is substantially different from that of other workers, or when there is some unusual feature. These comments can conveniently be divided into three sections: those concerned with (a) techniques and interpretation in terms of local structure, (b) interpretation in terms of mantle structure, and (c) further applications of the array.

(a) Techniques and Local Structure.

Chapter 2. An examination of the basic theory of the measurement of $dT/d\Delta$ and azimuth has led to the adoption of a least-squares technique for preparing a set of $dT/d\Delta$ values. Random and possible sources of systematic
errors have been discussed in detail, and the criteria for selecting earthquake and explosion data have been enunciated.

**Chapter 3.** An investigation of the effect of local structure at WRA on the measurement of $dT/d\Delta$ has shown that no simple crustal structure can explain the complex pattern of $dT/d\Delta$ and azimuth anomalies, though the $dT/d\Delta$ data indicate a predominating southwesterly dipping structure. The existence of specific azimuth ranges for which the wave forms of the P arrivals were unusual or varied from one seismometer to another, and for which the arrival times at the seismometers of the red line change abruptly, suggest that diffraction effects produced near the base of the crust are an additional source of $dT/d\Delta$ and azimuth anomalies. Thus, although it has been possible to provide a plausible explanation of the $dT/d\Delta$ and azimuth anomalies, reliable corrections to $dT/d\Delta$ could not be obtained by any simple mathematical approach. Finally, a technique called the method of projected planes has been devised to enable the regions of the crust responsible for possible diffracted arrivals to be mapped; the principle of this technique could be adapted to help interpret the results of a seismic reflection survey in the vicinity of an array.

**Chapter 4.** Two methods of constructing either a world average or a regional $dT/d\Delta$ curve for the lower mantle have been described. The first technique involves constraining the $dT/d\Delta$ data for a specific azimuth range to a modern travel-time curve, so that each $dT/d\Delta$ value in the range concerned is corrected by a constant amount. The second technique has been devised as a direct result of the work on the structure surface described in Chapter 3. The $dT/d\Delta$ data are grouped according to the dip angle and dip direction of the apparent structure responsible for the anomalies; this structure is determined from the azimuth anomaly and the ratio of the measured value of $dT/d\Delta$ to that obtained from a modern travel-time curve. Then an average structure for each group is calculated by vector summation, and finally the $dT/d\Delta$ values, corrected to first order using the structures, are adjusted
again using the first method. Thus the second refined method seeks to reduce systematic errors by using a series of plane dipping interfaces to represent the local structure; the fine details of the $dT/d\Delta$ curve are regarded as perturbations of a $dT/d\Delta$ curve derived by travel-time studies. Both techniques yield similar results, but the refined method is an improvement.

To smooth the $dT/d\Delta$ profiles, the two versions of the Method of Summary Values invented by Jeffreys have been used. The second of these involves locating three summary points in an interval in which there is a measurable curvature. I believe this second technique is superior to the simpler method of locating two summary points in each interval, since it fails if the data are not capable of estimating a curvature, and is therefore less arbitrary. It is also important to stress that a smooth curve, fitted either by the Method of Summary Values or by the usual polynomial regression technique, is not likely to provide an adequate representation of $dT/d\Delta$, owing to the possibility that every velocity anomaly in the lower mantle produces a duplication or triplication in the $dT/d\Delta$ curve.

(b) Interpretation in Terms of Mantle Structure.

Chapter 5. The Upper Mantle. $dT/d\Delta$ measurements for two separate provinces were made for the distance range 13.1° to 26.8°. The two groups of two distinct branches of the $dT/d\Delta$ curve associated with regions of rapid velocity change near 400 and 620 km were identified, and the cross-over points between the branches were found to be close to 20° and 24° for both provinces. Two preliminary velocity models for the upper mantle were derived. In addition there was some indication in the data of either regional differences in the detailed structure of the regions of high velocity gradients, or alternatively, that each of the transition layers is really two separate discontinuities. The effect of focal depth on the relative displacement of two overlapping branches of the travel-time curve was considered. Finally, a method of examining the evidence for the presence of two discontinuities near 400 km and near 620 km was suggested.
Chapters 6 - 8. The Lower Mantle. dT/d\(\Delta\) measurements for the distance range 28° to 99° were presented on a regional basis in Chapters 6 and 7, in an attempt to establish or deny the presence of velocity anomalies in the lower mantle. Seismic ray theory was used to study the forms of dT/d\(\Delta\) curve that could be produced by possible velocity anomalies. Evidence for the existence of low velocity gradients, or possibly a low velocity layer, at a depth of about 820 to 860 km, was found for three different regions of the earth. There was also some evidence that this anomalous region is more pronounced beneath the northern edge of the island of New Guinea, and an explanation of a low velocity layer at such depths was advanced in terms of the eclogite sinker model discussed by Ringwood (1967). The dT/d\(\Delta\) data also suggested complexity of structure down to depths of at least 1000 km. Sharp increases in velocity or velocity gradient were identified at depths of about 1200 km and 2350 km, and another possible rapid increase was suggested at 2735 km. In addition, very slowly changing velocity gradients were found at depths of approximately 1070 to 1110 km, 1260 to 1330 km, 1750 to 1850 km and 2460 to 2600 km. Regional differences in structure may persist to 1200 km. The sharp increases in velocity are almost certainly due to phase transformations. The velocity model ANUW2, presented in Chapter 8, also shows slight maxima in \(-\frac{r}{v} \frac{dv}{dr}\) at depths of 1030, 1505 and 1900 km, but no distinct anomaly is visible in the dT/d\(\Delta\) data. There was little evidence for the anomaly in dT/d\(\Delta\) near 60° suggested by Johnson (1969) and Corbishley (1970), though the data close to 60° are sparse. A comparison was made of this study with other array, travel-time and amplitude investigations, and a summary of the main features of the structure of the lower mantle that have now been identified or tentatively suggested was given in Table 8.7. The presence of inhomogeneities in the lower mantle has therefore been established, and a reasonably coherent picture of lower mantle structure emerges when my own work is compared with the evidence from other recent array investigations.

I believe that the most significant advantage of my approach to the problem of lower mantle structure is in analysing the dT/d\(\Delta\) data in groups
corresponding to narrow azimuth ranges, thus largely eliminating the problem of azimuthal bias. I would also like to emphasise the point that any velocity anomaly in the lower mantle is more likely to produce a triplication or duplication of \( \frac{dT}{d\Delta} \) than a simple steepening of the \( \frac{dT}{d\Delta} \) curve. Another important consideration is the size of an array. The advantage of using a medium aperture array is that \( \frac{dT}{d\Delta} \) is being averaged over a distance interval of only \( 0.2^\circ \) to \( 0.3^\circ \). However, with a large aperture array, \( \frac{dT}{d\Delta} \) is in effect being averaged over \( 2^\circ \) to \( 3^\circ \), even when the variation of \( \frac{dT}{d\Delta} \) with distance is allowed for. Therefore the \( \frac{dT}{d\Delta} \) data presented in this thesis have shown that a medium aperture device is not only as useful as, but is in many ways preferable to, a large aperture array in studying the structure of the earth's deep interior, since small anomalies will not tend to be smoothed out across an array with an aperture of only about 25 km.

(c) **Further Applications of the Array.**

Chapter 9. The ability of an array to enhance a signal at the expense of noise was used to study the identified phases and other coherent signals following \( P \) for a large nuclear explosion in Novaya Zemlya at a distance of \( 106^\circ \). This work enabled \( \frac{dT}{d\Delta} \) measurements for precursors to PP in the \( 105^\circ < \Delta < 110^\circ \) source window to be obtained, and an explanation of their origin was advanced in terms of asymmetric reflections at the surface of the earth, the base of the crust, or at deeper discontinuities. The alternative explanation of these waves is the PdP hypothesis of Bolt et al. (1968), and at present a combination of both explanations seems to be correct. The phased array was also used to identify and study core phases; in particular, the phase PcPPcP and four branches of the PKKP travel-time curve were identified.
10.2 RELATIONSHIP TO CURRENT RESEARCH IN OTHER BRANCHES OF GEOPHYSICS AND GEOCHEMISTRY.

Current research into phase transformations and the chemical composition of the mantle has enabled the two major discontinuities within the upper mantle to be explained in terms of phase transformations in an assumed pyrolite upper mantle, in which the low pressure silicate phases transform to denser structures involving sixfold coordination of the silicon. The phase transformations likely to occur near 400 km have been established by direct experiment, but those likely to occur near 620 km have been inferred indirectly, largely by studying phase transformations in germanate analogues. At present the static experimental techniques enable pressures of 200 kb, corresponding to a depth of about 600 km in the mantle, to be developed (see Ringwood, 1969a, b, Ringwood and Major, 1969). Consequently for higher pressures only shock wave experiments can be used, and these do not provide any information on the structures of high pressure phases. Their main application is in identifying phase changes in silicates, and in providing information on the equations of state of the high pressure phases (Ahrens, Anderson and Ringwood, 1969). Thus at present, experimental work indicates that phase transformations probably occur in the lower mantle, but their exact nature is not yet known.

The identification of P wave velocity anomalies in the lower mantle will no doubt help to resolve the controversy as to whether there is any need to postulate an increase in the Fe/(Fe + Mg) ratio throughout the mantle. Anderson (1967a, 1968) and Press (1968), for example, have suggested that a twofold increase in this ratio is required across the two transition zones in the upper mantle. Ringwood (1969b), however, has pointed out that a change in iron content with depth is not essential to achieve consistency between phase changes and accompanying velocity changes in a pyrolite upper mantle.

The extent to which the assumption of spherical symmetry in the earth is true is still not known. The systematic errors in dT/d\(\Delta\) and azimuth observed for WRA, and the errors in dT/d\(\Delta\) observed at the LASA in Montana, could be partly due to lateral variations in upper mantle structure. The
problems of seismic ray theory need to be re-examined in relation to plate
tectonics and geoid undulations.

10.3 SUGGESTIONS FOR FUTURE WORK IN ARRAY SEISMOLOGY.

An elaborate reflection survey in the Tennant Creek area, together
with an extensive gravity survey, is really required to elucidate the crustal
complexities of the WRA site. Unfortunately, a small scale refraction experi­
ment is unlikely to yield much valuable information, and only an expensive
reflection programme, supplemented by refraction data, would enable a clearer
understanding of the local crustal structure to be achieved.

A detailed study of the upper mantle in the Australasian region
should be undertaken with an emphasis on detecting regional differences in
structure and trying to correlate any such differences with tectonic processes,
particularly for the region to the north of the Australian continent. Explosions
are required to investigate the structure of the uppermost 200 km of the mantle.
It is also important to find out whether the splitting of each of the two transition
layers at depths of about 400 and 620 km into two separate discontinuities,
tentatively suggested in Chapter 5, is real. For the lower mantle more dT/dΔ
data for the distance ranges 37° to 42° and 67° to 73° are required, both for
the west Pacific margin and for other regions, to ascertain whether there
really are dT/dΔ anomalies at about 40° and 70°. A systematic search for
second arrivals associated with the anomalous regions near 1200 km and 2350
km would also be worthwhile, and would enable model ANUW3 to be refined.
The properties of the bottom 300 km of the mantle are not yet fully understood,
so that dT/dΔ data from events over the distance range 83° to 100° are required,
especially for the region beyond 88°. Such an investigation would also involve
an examination of dT/dΔ for PcP waves. Unfortunately, the number of large
events recorded at WRA at distances between 88° and 100° is small, though the
other U.K.A.E.A. arrays record numerous events over the same distance
range.
Using large events at distances between 100° and 110°, it is important to follow up the work of Chapter 9 with more travel-time and dT/dΔ measurements of precursors to PP to obtain a more complete explanation of their origin. Some work of this nature has already been started by Dr. R.M. Clowes using a WRA record of a large earthquake in the South Sandwich Islands. Many of these precursors are present, but their arrival times do not show much resemblance to those of the Novaya Zemlya explosion discussed in Chapter 9 (R.M. Clowes, personal communication, 1970). This seems to be evidence against the PdP hypothesis of Bolt et al. (1968), at least in its original form. Events in the same distance range should also enable the distance at which PKiKP (or PKIKP) waves are no longer observable to be determined using the correlation technique. A medium aperture array can also be used to look for PKJKP, PKIIKP, PKKKP and other unusual phases. The problem of dT/dΔ measurements for the various branches of PKP has so far received little attention. The structure of the core should now be examined using an array, to resolve the discrepancies between the core models of Bolt (1964), Adams and Randall (1964) and Ergin (1967), and to check with dT/dΔ measurements whether the various PKP branches identified by Bolt (1968) are correct. It might also be possible to examine core structure using PKKP phases; this approach could be especially useful for an array such as YKA, for which there is no major seismic zone at distances between 130° and 150°.
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EVIDENCE FOR A LOW VELOCITY LAYER FOR P WAVES
AT A DEPTH CLOSE TO 800 KM

C. WRIGHT

Department of Geophysics and Geochemistry,
The Australian National University, Canberra, A.C.T., Australia

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A study of the travel time gradients, $dT/d\Delta$, for P waves recorded at the Warramunga Seismic Array has been made over the distance range 32.8° to 43.6° using thirty earthquakes occurring in the Mariana Islands and surrounding regions. It was found that $dT/d\Delta$ increases with increasing distance between 32.8° and about 35°. This is interpreted in terms of a low velocity zone, at a depth of approximately 800 km, which becomes thicker towards the north. Ringwood and Green have proposed that large blocks of eclogite can be introduced into the mantle in tectonic regions. The accumulation of these eclogite sinkers below areas of tectonic activity may be responsible for this low velocity layer.

1. INTRODUCTION

In their study of P wave velocities in the mantle below 700 km, Chinnery and Toksöz [1] measured the slope of the travel time curve, $dT/d\Delta$, over the distance range 27°—90° using the LASA array in Montana. Their velocity-depth model shows an anomalous region at a depth of about 800 km where the velocity changes slowly with depth. They remarked that the preliminary evidence available suggests that the velocity actually decreases with increasing depth. Their $dT/d\Delta$ curve flattens close to 32°, but they had no good data between about 32° and 37°. Bugayevski [2] has presented evidence of a discontinuity in the travel time curve at 36°—37°. Gutenberg [3] found a minimum in the $P$ amplitude curve close to 35°. Since, over a limited distance range, the amplitude is proportional to $\sqrt{|dT/d\Delta^2|}$, a flattening of the $dT/d\Delta$ curve should coincide with a minimum in the amplitude curve. Carpenter et al. [4], however, found a sharp peak in the amplitude curve between 33° and 36° followed by a slight minimum between 36° and 39°.

In view of the interest of the region of the $dT/d\Delta$ curve between 30° and 40°, a study of $dT/d\Delta$ has been made for $P$ arrivals from thirty earthquakes occurring in the Mariana Islands and surrounding regions, recorded on the Warramunga Seismic Array (WRA), near Tennant Creek in the Northern Territory of Australia. These events correspond to a distance range of 32.8° to 43.6° and an azimuth range of 11.9° to 25.6°.

2. PROCEDURE

The values of apparent velocity, $d\Delta/dT$, and azimuth were calculated using a least squares fit to relative onset times measured in the manner described by Wright and Muirhead [5]. For a study of this kind the effect of the local structure beneath the array on the $P$ arrivals makes it essential to use a narrow range of azimuths. However, structures have previously been determined for correcting $dT/d\Delta$ for the azimuth ranges 345° ± 10° and 25° ± 5°. The problems encountered in deriving accurate values of $dT/d\Delta$ are extensively discussed by Niazi [6], Cleary et al. [7] and Wright and Muirhead [5].

The crustal model derived by Cleary et al. [6], to give corrections to $dT/d\Delta$ and azimuth for Aleutian Islands events at distances close to 80° and at azimuths of 25° ± 5°, gave corrections to $dT/d\Delta$ that were too small and azimuth corrections that were slightly too...
large when Mariana Islands earthquakes were used. For this reason a crustal model involving a single dipping interface within the crust, to give approximate corrections to $dT/dA$, was derived in the following way: seven clear records of shallow events were selected, corresponding to distances of between $36.0^\circ$ and $43.6^\circ$; the effects on apparent velocities and azimuths of a number of different structures with velocity contrasts of 0.7 were investigated. The best structure was taken to be that which gave corrected azimuths closest to the true azimuth, and apparent velocities closest to the values obtained by Cleary and Hales [8] for this distance range. This structure differs from the model of Cleary et al. by $20^\circ$ in dip direction and $0.5^\circ$ in dip angle, and is given below.

Thus, in deriving this crustal model, it has been assumed that the form of the $dT/dA$ curve beyond $36^\circ$ was in good agreement with that of Cleary and Hales.

For an earthquake occurring at depth $h$ and epicentral distance $A$, the value of $dT/dA$ measured at an array will correspond to an adjusted distance $A_0$, where $A_0$ is measured from the point where the extension of the ray path beyond the focus reaches the surface. For array measurements of $dT/dA$ it is preferable to adjust all distance measurements to correspond to $A_0$, and this procedure has been adopted for all earthquakes used in this paper. [The values of $A_0$ were calculated using an assumed crustal and upper mantle structure for shallow events, or using the extended distance tables of Hodgson and Storey [9] for earthquakes occurring at depths greater than 60 km.]

### 3. RESULTS

The measured values of $dT/dA$ were systematically high, and the azimuth anomalies were positive, in

<table>
<thead>
<tr>
<th>Dip direction (deg)</th>
<th>$P$ velocity in lower medium (km/sec)</th>
<th>$P$ velocity in upper medium (km/sec)</th>
<th>Dip angle (deg)</th>
</tr>
</thead>
<tbody>
<tr>
<td>215.0</td>
<td>6.0</td>
<td>4.2</td>
<td>7.0</td>
</tr>
</tbody>
</table>

Fig. 1. Corrected values of $dT/dA$ for thirty earthquakes in the Mariana Islands and surrounding regions.
EVIDENCE FOR A LOW VELOCITY LAYER

qualitative agreement with the results for Aleutian Islands earthquakes. The corrected values of $dT/d\Delta$ for thirty events are displayed graphically in fig. 1: three or four independent measurements of $dT/d\Delta$ have been made for each event. The values of $dT/d\Delta$ show considerable scatter, especially beyond 40°, which is reduced when the deeper focus events are removed. This scatter does not appear to be azimuth-dependent. The remarkable feature of the results is that $dT/d\Delta$ increases with increasing distance from 32.8° reaching a maximum at about 34°; then $dT/d\Delta$ starts to decrease again beyond 35°. The effect of crustal corrections on the maximum in the $dT/d\Delta$ curve is negligible. Thus there is an anomalous region of travel time gradients between 32.8° and 35.0°. The important consideration is that if $dT/d\Delta$ increases with increasing distance, the Herglotz-Wiechert method of deriving velocity-depth distributions breaks down.

4. INTERPRETATION

If the $dT/d\Delta$ curve flattens over a limited distance range it is clear that the P wave velocity must either increase very slowly or decrease slightly with increasing depth. Further, if the velocity decreases with depth continuously and sufficiently rapidly, $r/v$ increases with depth (where $r$ is the distance from the centre of the earth and $v$ is the velocity), and the Herglotz-Wiechert method formally breaks down. It would thus appear that the increase in $dT/d\Delta$ with increasing distance must be interpreted in terms of some kind of low velocity layer close to 800 km. Dowling and Nuttli [10] have indicated that a significant low velocity layer may manifest itself either as a shadow zone or as an overlap of two distinct branches of the $T-\Delta$ curve. The deviations of the travel times from the J-B times [11] are shown graphically in fig. 2. There is certainly no definite evidence of a break in the travel time curve, although three events close to 33° have late arrivals with respect to J-B, whereas all others at distances less than 40° tend to arrive early. Each event between 32° and 36° was examined for second arrivals. Three events showed what might be a second arrival about two seconds after P.

In order to explain the increase in $dT/d\Delta$, together with the absence of any significant change in J-B resid-
subside into the mantle because of their high density; this density is also substantially greater than the mean density of the ultramafic upper mantle. So a mechanism has been proposed by which large blocks of eclogite might be introduced into the mantle, and would start to sink slowly into the deeper regions.

Ringwood [13] has suggested that a pyroxene-garnet transition occurs in the mantle between 350 and 400 km. As a consequence of this transition an eclogite sinker would continue to fall through the mantle as garnetite to a depth of at least 700-800 km. According to Ringwood, whether these sinkers subside any further depends on the sequence and properties of other phase transitions that are not yet well established. It is possible that somewhere in this region of the mantle a garnet-ilmenite or garnet-perovskite transformation takes place (Ringwood and Major [14]). At about 620 km the velocity of P waves increases rapidly (Johnson [15], and Green and Hales [16]); at this depth the spinel form of magnesian olivine may transform to a material having approximately the properties of the component oxides (Anderson [17]), or to ilmenite, perovskite and other dense structures (Ringwood [13]). Provided the garnet-ilmenite transformation occurs well below 800 km, the garnetite blocks might be expected to accumulate at a depth somewhere between 620 km and 800 km or so. Anderson [17] has indicated that a garnet content of 10% below 620 km in the mantle would reduce the P wave velocity by 0.2–0.6 km/sec. It appears that a low velocity layer at a depth of about 800 km could result from an increase in garnet content with increasing depth. Thus it is suggested that this low velocity layer below New Guinea results from the accumulation of eclogite sinkers originating in the eugeosynclinal zone to the north.
6. CONCLUSIONS

It has been inferred that an increase in $dT/d\Delta$ between 32.8° and 34.3° is due to a low velocity zone at a depth of about 800 km, which lies immediately below the northern part of the Central Highlands of western New Guinea; this low velocity zone possibly becomes thicker towards the north. The measurements of $dT/d\Delta$ have also demonstrated how relative onset times of high precision, measured across a small array, can be used to investigate the fine structure of the mantle, provided that adequate allowances can be
made for the crustal and upper mantle structure beneath the array. Moreover, the results give some support to the hypothesis of Ringwood and Green that eclogite sinkers may be introduced into the mantle during an orogenic cycle. It is now important to establish whether this low velocity zone is a world wide phenomenon or merely a local effect beneath certain regions of tectonic activity.

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The Effects of Local Structure upon Measurements of the Travel Time Gradient at the Warramunga Seismic Array

J. R. Cleary, C. Wright and K. J. Muirhead

(Received 1968 February 26)

Summary

Measurements of $dT/d\Delta$ across the Warramunga seismic array are perturbed by variations in the structure beneath the array. On the assumption that the structure consists of uniformly dipping interfaces, an estimate of the effect was obtained by combining data from opposite azimuths. The results were not entirely satisfactory, however, because of the over-simplicity of the model. It seems likely that accurate array determinations of $dT/d\Delta$ can only be obtained by 'calibrating' the array, using events at distances for which $dT/d\Delta$ is already well known.

Introduction

The Warramunga seismic array was first put into operation in 1965 October, in time to record the nuclear explosion Longshot. When the array records of this event were analysed, they provided a measurement of the $P$ travel time gradient $dT/d\Delta$ across the array, which was about 11 per cent lower than that predicted by the Jeffreys-Bullen travel time curve. Studies by Cleary & Hales (1966) and by Carder et al. (1966) indicated that the J–B value for $dT/d\Delta$ was in error by less than 1 per cent at this distance. It seemed probable, therefore, that the discrepancy was due to crustal structure beneath the array, and it was decided to obtain a first approximation to this structure by means of the data from Longshot and some selected earthquakes.

Although the results described here are only of a preliminary nature, it seemed worthwhile to present details of the method of approach, and of the difficulties encountered. The use of arrays in $dT/d\Delta$ studies is increasing, and the considerable effect that local structure may have on the measurements is perhaps not yet fully appreciated.

The Warramunga array (WRA)

WRA is a U.K.A.E.A. array near Tennant Creek, Northern Territory, Australia, operated by the Australian National University. It consists of twenty seismometers arranged in two arms which are approximately 22 km in length and roughly at right angles to each other (Fig. 1). Signals from each seismometer are telemetered to a central recording station and recorded simultaneously on magnetic tape. The telemetry system is similar to that of the Yellowknife array, which has been described in detail by Keen et al. (1965).

The array is situated on granite outcrops of the lower Proterozoic Warramunga Geosyncline, and is about 500 km from the nearest part of the Australian coast.
Fig. 1. The Warramunga seismic array.

Fig. 2. Noise spectra at WRA, uncorrected for seismometer response. 100 units on the amplitude scale corresponds to 10 m$\mu$/s.
The effects of local structure

The noise level is correspondingly low, although the degree of microseismic activity is still noticeably affected by low pressure atmospheric systems near the coast. Each seismometer is operated at a displacement magnification of about 250 000. Typical noise spectra observed on 'quiet' and 'noisy' days are shown in Fig. 2.

A computer system for automatic digital processing of the magnetic tape records from WRA has been put into operation at the Australian National University, and has been described in some detail by Muirhead & Newstead (1968).

Previous work

Niazi & Anderson (1965) calculated values of $dT/d\Delta$ across the Tonto Forest array (TFSO), using data from earthquakes at epicentral distances between $12^\circ$ and $30^\circ$. TFSO consists of two lines of seismometers in the form of a cross, each line being approximately 10 km long. Their values of $dT/d\Delta$ were subject to a scatter of about 1 s/deg, most of which may be attributed to an uncertainty of up to 0.1 s in the measurements of arrival times upon which their calculations were based.

Later Niazi (1966) considered the effect of a dipping $M$-discontinuity beneath TFSO. He constructed tables of the deviations in $dT/d\Delta$ and azimuth which would be produced by a dipping interface, and compared them with deviations actually observed in these parameters from events at distances between $40^\circ$ and $85^\circ$. In this way he calculated a structure beneath the array which was in reasonable agreement with that found by a refraction survey of the area.

Chinnery & Toksoz (1967) studied $dT/d\Delta$ across the LASA array from earthquakes northwest of the array in the distance range $27^\circ$ to $90^\circ$. The breadth of LASA is about 200 km, so that although arrival times were measured only to the nearest 0.1 s, as they were at TFSO, the precision of the $dT/d\Delta$ determinations was an order of magnitude greater. To compensate for systematic errors due to structure beneath the array, Chinnery & Toksoz 'calibrated' their $dT/d\Delta$ curve by comparing a derived travel time curve with data from the Longshot nuclear explosion. This device was not entirely successful, because the Longshot data were themselves biased by azimuthal variation in the source term (Cleary 1967). The earthquake data are sparse in some distance ranges from LASA, indicating the need for additional information from arrays in other locations.

Measurement techniques at WRA

It has been seen that the large dimensions of the LASA array give it a considerable advantage over smaller arrays in the determination of $dT/d\Delta$. The aperture of the WRA array is about one-tenth that of LASA; therefore in order to determine $dT/d\Delta$ with similar precision it was necessary to find a means of measuring times to an accuracy of about 0.01 s. It is virtually impossible to measure $P$ onset times to this accuracy. On the other hand, if the $P$-waves recorded at each pit are matched with each other the measurement of relative $P$ times to 0.01 s becomes quite feasible. Accurate measurement of the relative arrival times of surface waves by matching waveforms has been a common technique in the calculation of the phase velocities of surface waves across tripartite arrays since the method was first described by Evernden (1958). Matching $P$-waves having periods of about 1 s to an accuracy of 0.01 s is no more difficult than matching surface waves with periods of 20 s and greater to an accuracy of 0.2 s, provided a sufficiently expanded time scale is used.

The relative merits of various matching techniques are still being studied. The method used for the present study was a manual technique, which has been illustrated by Longshot records in Fig. 3. The procedure is as follows:

1. Magnetic tape records of the event from each pit are transcribed on to paper, at a speed of about 40 mm/s.
2. In order to compensate for variations in amplitude between records, the $P$-waves from one record are used to construct, on transparent paper, a family of curves having a range of amplitudes.

3. Using this paper as an overlay, the curves are matched with the $P$-waves recorded at each pit. The part of the record found to be most suitable for accurate matching is the section between the first peak and the first trough of the $P$ train. Measurements made in this way were repeatable to better than 0.01 s.

The desirability of using a technique involving direct computer processing of the magnetic tape data is obvious. It has been found that the correlation method of Birtill & Whiteway (1965) gives results inferior in precision to the manual method described above. However, more promising results have been obtained recently using a new technique developed by Muirhead (1968).

**Results**

Five events were chosen for this preliminary study. They include the Longshot explosion and three earthquakes in the Aleutian Islands, at distances between 79.4° and 84.7° from WRA, and an earthquake south of Africa at a distance of 84.0°. The azimuth of the last event from the array was approximately opposite to those of the Aleutian Islands.

The relative times of arrival at the individual array seismometers were measured in the manner described in the preceding section, and were compared with J-B
times on the assumption that the U.S.C.G.S. locations were correct. The residuals (on an arbitrary baseline) are shown in Fig. 4. Certain features are worth noting:

1. The trend of the residuals is in the same direction for opposite azimuths, indicating that the deviations are caused by a structure dipping to the southwest, rather than by errors in the J-B times.

2. The trend is not linear, although it is roughly symmetrical along both lines with respect to the crossover point of the array. This suggests that the structure consists of a series of monoclinal folds striking northwest. However, the possibility that these departures from linearity have been produced by small variations in the instrumental constants cannot be discounted at this stage.

3. The residuals from the South of Africa event are similar to those from the Aleutian Islands events except at pits R1, B1 and B2 near the crossover point, where they correspond to earlier arrivals. It would be desirable to confirm these discrepant observations by means of readings from other events to the south of the array. It seems likely, however, that the dipping interface rises abruptly in the vicinity of the crossover point, so that waves travelling to pits R1, B1 and B2 from the south pass through a structure that is not continuous with that traversed by waves approaching from the north. For this reason, readings from these three pits were excluded from further analysis of the South of Africa event.

Underwood et al. (1968) have described a seismic experiment which involved firing an explosive charge in a mine close to the array. From an analysis of arrival times from this shot, Underwood (1968) has inferred the presence of a structure immediately beneath the array, which he estimates as a first approximation to dip 5° in a direction 200°E of N. It is likely that this structure is mainly responsible for the observed deviations in the residuals.

Niazi (1966) has provided equations for calculating the effect of a dipping interface on the $dT/d\Delta$ and azimuth of an incoming wave. With the aid of tables constructed from these equations, data from earthquakes at almost equal distances and at opposite azimuths from the array can be used to separate the effects of structure
### Table 1

*dT*/dΔ and azimuth measurements at WRA

<table>
<thead>
<tr>
<th>Event</th>
<th>Date</th>
<th>Latitude (deg)</th>
<th>Longitude (deg)</th>
<th>Distance to WRA (deg)</th>
<th>Azimuth from WRA</th>
<th>J-B (s/deg)</th>
<th>Measured (s/deg)</th>
<th>Adjusted* (s/deg)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Near Is.</td>
<td>1966 14 July</td>
<td>53.1 N</td>
<td>171.1 E</td>
<td>79.4</td>
<td>21.5</td>
<td>5.39</td>
<td>6.07</td>
<td>5.29</td>
</tr>
<tr>
<td>Rat Is.</td>
<td>1966 1 Aug.</td>
<td>51.5 N</td>
<td>177.6 E</td>
<td>80.6</td>
<td>25.7</td>
<td>5.29</td>
<td>5.91</td>
<td>5.12</td>
</tr>
<tr>
<td>Longshot</td>
<td>1965 29 Oct.</td>
<td>51.4 N</td>
<td>179.2 E</td>
<td>81.2</td>
<td>26.5</td>
<td>5.24</td>
<td>5.88</td>
<td>5.09</td>
</tr>
<tr>
<td>S. of Africa</td>
<td>1966 14 July</td>
<td>52.9 S</td>
<td>27.5 E</td>
<td>84.0</td>
<td>215.7</td>
<td>5.05</td>
<td>4.07</td>
<td>4.94</td>
</tr>
<tr>
<td>Andreanof Is.</td>
<td>1966 19 July</td>
<td>51.7 N</td>
<td>173.3 W</td>
<td>84.7</td>
<td>29.7</td>
<td>5.00</td>
<td>5.79</td>
<td>4.98</td>
</tr>
</tbody>
</table>

* Using corrections derived from single-layer structure described on page 27 of text.
The effects of local structure

beneath the array from errors in the J–B tables, in much the same way that 'split spreads' are used in refraction seismology.

The relative arrival times at each pit were used to calculate a value of $dT/d\Delta$ and an azimuth of arrival at WRA for each event. These are compared in Table 1 with values calculated from U.S.C.G.S. locations and J–B tables. Assuming a standard deviation of 0.01 s on each arrival time, standard errors of the $dT/d\Delta$ and azimuth computations were calculated for each event, in the manner described by Kelly (1964). In each case, standard errors of about 0.4° in azimuth and about 0.1 s/deg in $dT/d\Delta$ were obtained.

Using Niazi's equations, corrections to $dT/d\Delta$ and azimuth were calculated for a number of structures having various dip angles and dip corrections. The structure giving the closest approximations to the U.S.C.G.S. azimuths, together with reasonable correspondence between values of $dT/d\Delta$ from the events in both azimuths, was as follows:

- Dip angle = 6.5°
- Dip direction = 235°
- Velocity ratio = 0.7

The 'corrected' $dT/d\Delta$ and azimuth values, based on this structure, are also given in Table 1. The errors in azimuth, after this compensation for structure, are between 0.5° and 1.1°. The values of $dT/d\Delta$ are displaced by about 0.1 s/deg from the J–B values (Fig. 5). As mentioned earlier, the values found for $dT/d\Delta$ in this distance range by Cleary & Hales (1966) and by Carder et al. (1966) are in very good agreement with J–B, so this discrepancy probably represents a systematic error in the analysis due to some inadequacy in the model.

The dip angle and dip direction of the structure are in broad agreement with the results of Underwood, mentioned earlier. However, the structure gives a rather poor fit to Underwood's data. It is therefore likely that the model involving a single dipping interface is too simple. An attempt to obtain a more realistic solution was made by extending the model to include a dipping Mohorovicic discontinuity in addition to Underwood's structure. The following two-layer structure gave the best fit to the data:

### Interface I (Underwood 1967)
- Dip angle: 5.3°
- Dip direction: 205.5°
- Velocity ratio: 0.9

### Interface II
- Dip angle: 7.8°
- Dip direction: 246.0°
- Velocity ratio: 0.7

![Fig. 5. $dT/d\Delta$ at WRA from five events between 79° and 85°, after compensation for structure (see text). The Jeffreys–Bullen $dT/d\Delta$ curve is shown for comparison.](image)
This structure and the single-layer model give corrected values of $dT/d\Delta$ that fit the data equally well. Because of the low velocity contrast in Underwood’s model, most of the effect has been transferred to the second layer. It seems probable that this solution is less realistic than the first one. However, it would be pointless to revise the model further without a more detailed study of the structure immediately beneath the array.

The next stage of the investigation was to establish whether the calculated structure gave valid results for events from other azimuths. A preliminary study has indicated that for some azimuths the corrections obtained from the structure are of the right order of magnitude. However, a disturbing result was obtained by Wright & Muirhead (1968) from their examination of the WRA recordings of diffracted $P$ from a large nuclear explosion at Novaya Zemlya. This event differed in azimuth from the Aleutian Islands events by about 40°. Using a reliable value of $dT/d\Delta$ for diffracted $P$ obtained by Sacks (1967), the authors calculated a structure which differed from the one given here by about 4° in dip angle and 80° in dip direction. It follows from this result that the assumption of uniformly dipping interfaces is not valid for all azimuths, and hence that the structural effect cannot be properly deduced from observations in opposite azimuths. This would account for the discrepancy between the ‘corrected’ and expected values of $dT/d\Delta$ noted above.

Finally, then, a single-layer structure was calculated from the Aleutian Islands events on the assumption that the J-B $dT/d\Delta$ curve is correct between 80° and 85°. For all four events, the corrections derived from the model gave azimuths within 0-5° of the true azimuths, and values of $dT/d\Delta$ within 0-07 s/deg of J-B. A structure was also calculated from the South of Africa event. The two structures are as follows:

<table>
<thead>
<tr>
<th></th>
<th>Aleutian Islands</th>
<th>S. of Africa</th>
</tr>
</thead>
<tbody>
<tr>
<td>Dip angle</td>
<td>6-0°</td>
<td>7-0°</td>
</tr>
<tr>
<td>Dip direction</td>
<td>243-0°</td>
<td>230-0°</td>
</tr>
<tr>
<td>Velocity ratio</td>
<td>0-7</td>
<td>0-7</td>
</tr>
</tbody>
</table>

It should be emphasized that these are simply formal solutions, which give at best only an approximation to the true structure. Nevertheless they should be useful in providing corrections to data from the same azimuths but at different distances.

Conclusions

The attempt to account for the discrepancies in measurements of $dT/d\Delta$ and azimuths at WRA in terms of uniformly dipping structures has met with only partial success. The ‘reversal’ technique provides only an approximate correction for structure, and is therefore not suitable for obtaining accurate determinations of $dT/d\Delta$. A more satisfactory approach is to ‘calibrate’ the array for a particular azimuth using observations at distances for which $dT/d\Delta$ is accurately known, and then to apply this calibration to measurements of arrivals from the same azimuth but different distances. The variation from the simple structural model is so pronounced that its cause is probably not close to the surface, and it is possible that it resides in the upper mantle. If this is so, then local seismic surveys will not provide sufficient information for complete corrections to be made. It follows that results from this and other arrays are subject to an uncertainty which, it seems, can only be resolved by reference to travel-time data obtained in the traditional manner.

Department of Geophysics and Geochemistry,
and Department of Engineering Physics,
Australian National University,
Canberra,
Australia.

1968 February.
The effects of local structure

References


Longitudinal Waves from the Novaya Zemlya Nuclear Explosion of October 27, 1966, Recorded at the Warramunga Seismic Array

C. WRIGHT

Department of Geophysics and Geochemistry
Australian National University, Canberra, Australia

K. J. MUIRHEAD

Department of Engineering Physics, Australian National University
Canberra, Australia

The apparent velocity and azimuth of $P$ from the Novaya Zemlya nuclear explosion of October 27, 1966, have been measured for the Warramunga array and have been used to work out possible crustal structures for correcting measured apparent velocities and azimuths in the azimuth range 335° to 355°. Comparison of the measured apparent velocities and azimuths for seven earthquakes at different distances, but at similar azimuths to Novaya Zemlya, has shown that the computed structures give satisfactory corrections. Measurements of apparent velocity and azimuth have been used as an aid in identifying the phases $PKiKP$, $PP$, $PcPPcP$, $PKKP$, $PcPPKP$, and $PKPPKP$ in the explosion record; similar measurements have been made for coherent signals not corresponding to any of the conventional phases.

INTRODUCTION

The Warramunga Seismic Array (WRA), near Tennant Creek in the Northern Territory of Australia, consists of twenty short-period vertical component seismometers arranged in two lines of ten, 22.5 km long and approximately at right angles to each other. All data are recorded on magnetic tape and can readily be transcribed onto paper. The apparent surface velocities, $dA/dT$, and azimuths of $P$ waves from teleseisms recorded at WRA show systematic differences from the true azimuth and the Jeffreys-Bullen (J-B) apparent velocities; these discrepancies are believed to be due to a dipping Mohorovicic discontinuity and other dipping interfaces within the crust.

An important application of an array of short-period vertical component seismometers is the determination of a velocity-depth distribution for $P$ waves throughout the whole of the earth's mantle. The measurements of $dA/dT$ need, however, to be corrected for the effects of structure beneath the array. The apparent velocities and azimuths have been calculated by means of a least-squares program designed by E. W. Carpenter and written by B. S. Gopalan. This least-squares method has been summarized by Otsuka [1966a, b]. A subroutine has been added to the least-squares program that calculates the standard errors in velocity and azimuth by using the formulas given by Kelly [1964]. An alternative, but less precise, means of velocity and azimuth determination is a correlation method, which has been described by Birtill and Whiteway [1965]. Niazi [1966] considered the effect of a dipping $M$ discontinuity on measurements of the parameter $dT/dA$ and azimuth. By using the theory outlined in Niazi's paper, a computer program has been written that will work out the effect on apparent velocity $dA/dT$ and azimuth of any combination of dipping interfaces for a seismic event whose azimuth and distance are accurately known. The time residuals for each seismometer of the array are also calculated.

The Novaya Zemlya nuclear explosion of October 27, 1966, at a distance of 106.0°, was well recorded by the array, and a number of later phases as well as $P$ were visible. The high quality of the record of $P$, together with accurate knowledge of the expected apparent velocity of diffracted $P$, makes this explosion ideal for determining a crustal structure beneath the array. Discussion of possible crustal structures...
and corrections to apparent velocities and azimuths constitute the first part of this paper. The second part is concerned first with the identification of later phases by supplementing travel times with apparent velocity and azimuth measurements and second with an investigation of unidentified coherent signals following $P$. This investigation of later phases was conducted to demonstrate the advantages of an array in studying the travel time curves of these later $P$ phases.

**Crustal Structure Beneath Array**

**Procedure.** In working out a crustal structure consisting of one or more dipping interfaces using a single teleseism or several teleseisms at similar azimuths, accurate knowledge of both the azimuths and the expected apparent velocities of the $P$ waves crossing the array are required. For the majority of earthquakes the azimuth is known accurately, but the apparent velocities, determined by differentiating the J-B travel-time curve for $P$, is not, owing to the presence of errors in the tables. Working out a crustal structure will provide corrections to $dT/dA$. Hence, the procedure adopted is to select a realistic crustal model and to adjust the dip angle and dip direction, keeping the $P$-wave velocities constant, until the apparent velocity and azimuth calculated by using the structure agree with the measured values. Obviously, an infinite number of models would fit the data, but any differences in the corrections from one model to another would only become significant at azimuths widely separated from the azimuths of the data used.

**Previous work.** Jeffreys [1962b] realized from a study of Pacific nuclear explosions that the Jeffreys and Bullen [1948] travel times for $P$ required small modifications. A recent modification of the $P$ travel-time curve prepared by Cleary and Hales [1966] gives values of $d\Delta/dT$ that differ significantly in some distance ranges from the values obtained from the J-B tables. New travel times for $P$ have also been proposed by Carder et al. [1966] and Herrin et al. [1968]. The results of Carder et al. show deviations from the J-B values similar to the deviations shown by Cleary and Hales.

The effects of structure can be separated from errors in the J-B tables if data from earthquakes occurring at similar distances, but at opposite azimuths, are used [see Niazi, 1966]. The distribution of regions of high seismicity with respect to WRA is such, however, that approximately equidistant events at opposite azimuths are rare occurrences. One such occurrence has been used to determine two possible structures. The first is a simple structure involving a single dipping interface between sediments and basement beneath the array. The second structure involves two dipping interfaces; a dipping M discontinuity is placed below a superficial structure derived by Underwood [1967] from the results of the seismic experiment WRAMP. These structures were determined by using three earthquakes from the Aleutian Islands and the nuclear explosion Longshot, corresponding to a distance range of $79.4^\circ$-$84.7^\circ$ and an azimuth range of $20^\circ$-$30^\circ$, together with one earthquake south of Africa at a distance of $84.0^\circ$ and azimuth of $215.7^\circ$. The structures are listed in Table 1 for comparison with the results obtained in this study of the Novaya Zemlya nuclear explosion.

**Measurement techniques.** The techniques used in measuring relative onset times on WRA records are described, together with an account of the work on the structure, by Cleary et al. [1968]. As well as the method of matching waveforms described in the paper cited above, the first prominent peak and the first zero (or crossover point) of each seismometer output were also matched. Thus three independent sets of relative onset times for $P$ were obtained, both for the explosion and for the other events used in this paper.

Measurements of $d\Delta/dT$ and azimuth for a wide range of azimuths and distances have revealed that the crustal structure can change quite sharply over a small azimuth range. In particular, it has been discovered that the structure affecting $P$ waves arriving between $330^\circ$ and $10^\circ$ in azimuth is quite different from that between $20^\circ$ and $30^\circ$. Moreover, for some azimuths the structure appears to be different for $P$ waves arriving from different distances. In view of these considerations, the Novaya Zemlya nuclear explosion has been used to determine two possible types of structure, involving a single dipping interface and two dipping interfaces respectively, which should
WRIGHT AND MUIRHEAD

TABLE 1. Structures Determined for the Warramunga Array

<table>
<thead>
<tr>
<th>Structure</th>
<th>Dip Direction, deg</th>
<th>P-Wave Velocity</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>In Lower</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Medium, km/sec</td>
</tr>
<tr>
<td>1 (a)*</td>
<td>One dipping interface using Sacks velocity</td>
<td>179.0</td>
</tr>
<tr>
<td>1 (b)*</td>
<td>One dipping interface using J-B velocity</td>
<td>175.0</td>
</tr>
<tr>
<td>2 (a)*</td>
<td>Two dipping interfaces using Sacks velocity</td>
<td>205.5</td>
</tr>
<tr>
<td>2 (b)*</td>
<td>Two dipping interfaces using J-B velocity</td>
<td>205.5</td>
</tr>
<tr>
<td>3†</td>
<td>One dipping interface</td>
<td>235.0</td>
</tr>
<tr>
<td>4†</td>
<td>Two dipping interfaces</td>
<td>205.5</td>
</tr>
</tbody>
</table>

* Structures determined from the Novaya Zemlya nuclear explosion.
† Structures determined from three Aleutian Islands earthquakes, the nuclear explosion Longshot, and one south of Africa earthquake.

give reasonably good seismometer corrections in the azimuth range from 335° to 355°.

WRA structure from the Novaya Zemlya explosion. The initial P onset of the explosion was sharp, and a large quantity of energy was present in the first 3 seconds of the record. Figure 1 shows the traces of P on fifteen seismometers, and Figure 2 shows some processed records including both summed lines and correlator outputs produced on a digital computer in the manner described by Birtill and Whitehead [1965]. It is not clear whether P is diffracted or direct. Recent work on P in the shadow zone has been published by Sacks [1966, 1967]. Ergin [1967] has suggested that a reduction in the velocity of P is required at the base of the mantle, on evidence from PKP travel times near 180° and beyond; this reduction would cause direct P to propagate up to 130° and beyond.

To determine a reliable structure it is essential that the apparent velocity of diffracted or direct P at 106.0° is known to within 0.2 km/sec. The J-B P curve straightens at about 90° and has very little curvature from there on; dA/dT for A > 100° is 25.3 km/sec. Sacks [1967] has, however, given preliminary travel times of diffracted P to 167° and has given dA/dT as 24.55 ± 0.08 km/sec. This value has been used as a reference in this study. From their travel time curve for P, Cleary and Hales [1966] give an apparent velocity of 24.72 km/sec at 95°, and Carder et al. [1966] give a value of 24.44 km/sec for the distance range 94° to 100°. From the new Seismological Tables for P Phases [1968] the apparent velocity is 24.38 km/sec for P beyond 97.5°. Structures have been determined by assuming that the apparent velocity given by Sacks is correct; for comparison similar structures are given by assuming the J-B value is correct. The results are given in Table 1 and have been obtained by averaging the three independent least-squares determinations for P from the nuclear explosion.

Corrections to dA/dT and azimuth. The extent to which the computed structures are applicable to other events has been investigated by using seven earthquakes occurring at similar azimuths to the Novaya Zemlya explosion, but at different distances. Information about these events, given in Table 2, has been taken from the United Kingdom Atomic Energy Authority
from the U. S. Coast and Geodetic Survey information cards. The expected apparent velocity and azimuth using two of the structures of Table 1 have been determined for each earthquake. For the four closest events the J-B values of $d\Delta/dT$ have been used, and for the remaining three, both the Cleary-Hales and the J-B velocities have been used. Least-squares values of velocity and azimuth have also been determined; the standard errors in both these quantities have been estimated from a standard deviation on each onset time calculated from the set of seismometer residuals obtained for each event. For each set of onset times the values of $d\Delta/dT$ and azimuth have been computed from all seismometers that were working.

Fig. 1. Traces of $P$ on fifteen seismometers. Red, R, and blue, B, lines run approximately E–W and N–S, respectively. See Cleary et al. [1968].
Fig. 2. Some processed records of $P$. Array phased to an azimuth of 344.2° and to an apparent velocity of 21.0 km/sec. A 2-second square window of integration was used for trace 5 and for all correlator outputs in Figures 3-6.

<table>
<thead>
<tr>
<th>Date</th>
<th>Origin Time (GMT)</th>
<th>Distance from WRA, deg</th>
<th>Corrected Distance, deg</th>
<th>Azimuth, deg</th>
<th>Depth, km</th>
<th>Region</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sept. 15, 1967</td>
<td>19h 15m 53.8s</td>
<td>22.6</td>
<td>23.8</td>
<td>340.8</td>
<td>119</td>
<td>Halmahera</td>
</tr>
<tr>
<td>July 30, 1966</td>
<td>17h 39m 18.8s</td>
<td>29.9</td>
<td>30.1</td>
<td>344.5</td>
<td>36</td>
<td>Mindanao, Phillipines</td>
</tr>
<tr>
<td>July 27, 1966</td>
<td>5h 10m 36.1s</td>
<td>24.0</td>
<td>24.2</td>
<td>339.1</td>
<td>33</td>
<td>Molucca Passage</td>
</tr>
<tr>
<td>Aug. 4, 1966</td>
<td>20h 23m 02.5s</td>
<td>23.0</td>
<td>24.0</td>
<td>340.4</td>
<td>95</td>
<td>Molucca Passage</td>
</tr>
<tr>
<td>Oct. 27, 1966</td>
<td>5h 57m 58.0s</td>
<td>106.0</td>
<td>106.0</td>
<td>342.9</td>
<td>0</td>
<td>Novaya Zemlya (nuclear explosion)</td>
</tr>
<tr>
<td>Jan. 2, 1967</td>
<td>14h 47m 11.6s</td>
<td>41.5</td>
<td>41.7</td>
<td>342.9</td>
<td>33</td>
<td>Philippine Islands region</td>
</tr>
<tr>
<td>July 6, 1966</td>
<td>20h 21m 43.5s</td>
<td>45.9</td>
<td>46.0</td>
<td>352.0</td>
<td>23</td>
<td>Ryukyu Islands</td>
</tr>
<tr>
<td>July 29, 1966</td>
<td>6h 25m 35.2s</td>
<td>48.9</td>
<td>49.0</td>
<td>354.2</td>
<td>21</td>
<td>Ryukyu Islands</td>
</tr>
</tbody>
</table>
well. Then, if necessary, a new computation was performed with suspect seismometers removed to give an improved least-squares fit. A seismometer was rejected if the magnitude of its residual exceeded 0.04 second. Table 3 compares the expected and the measured values of apparent velocity and azimuth.

For the four teleseisms the azimuth anomalies tend to be lower than the values derived from the computed structures, but the discrepancy is only really significant for the Ryukyu Islands event of 29.7.66; this is probably due to a rather different structure affecting P waves arriving at azimuths close to 354°. For the three closer events the azimuth anomalies are larger than expected, and for two of them the spread in azimuth for the different methods of measurement is significantly greater.

The epicenters of the two Molucca Passage and the Halmahera earthquakes are very close to each other and are at distances close to 24°. For an earthquake occurring at depth \( h \) and epicentral distance \( \Delta \), the apparent velocity measured at an array will correspond to a corrected distance \( \Delta' \), where \( \Delta' \) is measured from the point at which the extension of the ray path beyond the focus reaches the surface. For each of the events close to 24°, \( \Delta' \) was determined from the epicentral distance \( \Delta \) and focal depth \( h \), by using an assumed crustal and upper mantle structure for the shallow event and making use of the tables of extended distances prepared by Hodgson and Storey [1953] for the two deeper events. The corrected distances for both Molucca Passage earthquakes differ by only 0.1°; yet, the apparent velocities differ by 0.4 km/sec, which is at least 4 times the estimated standard errors. If the epicenter locations are correct, these results suggest an irregularity in the slope of the travel-time curve close to 24°. This would be in conformity with the presence of a discontinuity close to 24°, as found by Niazi and Anderson [1965] and by Green and Hales [1968].

**Study of Later Phases**

**Previous Work**

The WRA record of the Novaya Zemlya explosion provided some exceptionally clear later P phases, particularly PKiKP and PKKP. Coherent signals following P were picked out visually from an array record bandpass-filtered between 0.4 and 2.0 cps, and the correlation method described by Birtill and Whiteway [1965] was used to give an estimate of apparent velocity for each signal. Jeffreys [1962a] has pointed out the difficulty of explaining the continuous irregular oscillations that follow P and S. Several hypotheses have been put forward to explain these oscillations. For example, it has been suggested that the general irregular movement is impressed in the immediate neighborhood of the observing station because of scattering. Key [1967] has investigated the signal generated noise at the United Kingdom Atomic Energy Authority seismometer array station at Eskdalemuir, Scotland, and has shown a close correlation between low-velocity apparent noise sources and topographic features; these signals are Rayleigh waves. The exceptional quality of the record of the Novaya Zemlya explosion has enabled some of the oscillations following P that are coherent across the array to be explained in terms of reflections at a considerable distance from the array, as well as providing some indication of the reliability of measurements of \( d\Delta/dT \) for identifiable phases using the correlation method. The results of this study of coherent signals following P are given in Tables 4 and 5.

For the large later phases it was possible to supplement the correlation method with hand measurements of the onset times and a least-squares azimuth and velocity determination. The errors in the total travel times to the crossover point are believed to be about 0.2 second for the phases of Table 5; for the other coherent signals the errors are larger. To give some indication of the quality of the least-squares fit, the mean square time residual is also displayed in Table 5. For these later phases seismometers were rejected only if the magnitude of the residuals exceeded 0.05 second. The least-squares fit for PP was so poor, however, that it was not possible to adopt this rejection procedure.

**Identified Phases (See Table 6)**

All phases arrived early relative to the expected travel times, except PKiKP and PKKP. It seems likely that the core model used by Bolt and O'Neill [1965] to calculate the travel times for PKiKP is slightly in error by com-
<table>
<thead>
<tr>
<th>Event</th>
<th>True Azimuth, deg.</th>
<th>Apparent Velocity from (a) J-B, (b) Cleary-Hales, (c) Sacks, km/sec</th>
<th>Expected Apparent Velocity Corrected for Structure, km/sec</th>
<th>Expected Azimuth Anomaly Corrected for Structure, deg</th>
<th>Measured Apparent Velocity, km/sec</th>
<th>Measured Azimuth, deg</th>
<th>Azimuth Anomaly, deg</th>
<th>Matching Technique Used</th>
</tr>
</thead>
<tbody>
<tr>
<td>Halmahera, Sept. 9, 1967</td>
<td>340.8</td>
<td>(a) 11.5</td>
<td>10.90</td>
<td>10.76</td>
<td>1.0</td>
<td>0.9</td>
<td>10.39 ± 0.08</td>
<td>342.3 ± 0.6</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>10.37 ± 0.07</td>
<td>342.8 ± 0.5</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>10.38 ± 0.07</td>
<td>342.7 ± 0.5</td>
</tr>
<tr>
<td>Novaya Zemlya Explosion, Oct. 27, 1966</td>
<td>342.9</td>
<td>(a) 25.3</td>
<td>22.13</td>
<td>22.13</td>
<td>1.6</td>
<td></td>
<td>22.14 ± 0.19</td>
<td>344.4 ± 0.6</td>
</tr>
<tr>
<td></td>
<td></td>
<td>(c) 24.55</td>
<td>22.11</td>
<td>22.13</td>
<td></td>
<td></td>
<td>22.21 ± 0.36</td>
<td>344.2 ± 1.2</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>21.96 ± 0.27</td>
<td>344.9 ± 0.9</td>
</tr>
<tr>
<td>Mindanao, Philippines, July 30, 1966</td>
<td>344.5</td>
<td>(a) 12.5</td>
<td>11.79</td>
<td>11.63</td>
<td>0.8</td>
<td>0.8</td>
<td>11.59 ± 0.09</td>
<td>345.1 ± 0.5</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>11.59 ± 0.08</td>
<td>344.5 ± 0.5</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>11.59 ± 0.09</td>
<td>344.4 ± 0.6</td>
</tr>
<tr>
<td>Molucca Passage, July 27, 1966</td>
<td>339.1</td>
<td>(a) 11.5</td>
<td>10.91</td>
<td>10.76</td>
<td>1.1</td>
<td>1.1</td>
<td>10.67 ± 0.09</td>
<td>340.6 ± 0.7</td>
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<td></td>
<td>10.74 ± 0.08</td>
<td>344.4 ± 0.5</td>
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<td></td>
<td>10.77 ± 0.09</td>
<td>343.8 ± 0.6</td>
</tr>
<tr>
<td>Molucca Passage, Aug. 4, 1966</td>
<td>340.4</td>
<td>(a) 11.5</td>
<td>10.90</td>
<td>10.76</td>
<td>1.0</td>
<td>1.0</td>
<td>10.35 ± 0.08</td>
<td>341.8 ± 0.5</td>
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<td></td>
<td>10.27 ± 0.06</td>
<td>343.3 ± 0.4</td>
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<td></td>
<td>10.38 ± 0.08</td>
<td>342.4 ± 0.5</td>
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<tr>
<td>Philippine Islands region, Jan. 2, 1967</td>
<td>342.9</td>
<td>(a) 13.5</td>
<td>12.70</td>
<td>12.50</td>
<td>1.0</td>
<td>0.9</td>
<td>12.38 ± 0.10</td>
<td>343.7 ± 0.6</td>
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<tr>
<td></td>
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<td>(b) 13.74</td>
<td>12.91</td>
<td>12.71</td>
<td>1.0</td>
<td>0.9</td>
<td>12.37 ± 0.09</td>
<td>343.7 ± 0.5</td>
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<td></td>
<td>12.35 ± 0.10</td>
<td>343.2 ± 0.6</td>
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<tr>
<td>Ryukyu Islands, July 6, 1967</td>
<td>352.0</td>
<td>(a) 14.1</td>
<td>13.20</td>
<td>13.00</td>
<td>0.5</td>
<td>0.3</td>
<td>13.36 ± 0.11</td>
<td>351.9 ± 0.5</td>
</tr>
<tr>
<td></td>
<td></td>
<td>(b) 14.29</td>
<td>13.37</td>
<td>13.17</td>
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<td>0.1</td>
<td>13.41 ± 0.11</td>
<td>351.5 ± 0.6</td>
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<td>13.12 ± 0.11</td>
<td>351.8 ± 0.5</td>
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<td>Ryukyu Islands, July 29, 1966</td>
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<td>13.47</td>
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<td>0.1</td>
<td>13.47 ± 0.12</td>
<td>353.3 ± 0.6</td>
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<td>(b) 14.54</td>
<td>13.59</td>
<td>13.38</td>
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<td>0.1</td>
<td>13.71 ± 0.13</td>
<td>353.0 ± 0.7</td>
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<td></td>
<td>13.62 ± 0.10</td>
<td>352.4 ± 0.5</td>
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</table>

* See Table 1.
TABLE 4. Correlation Measurements

This table gives a selection of correlation measurements of apparent velocity and azimuth for some of the later \( P \) phases and coherent signals. A 3-second integration time was used for correlation measurements except for entries otherwise indicated.

<table>
<thead>
<tr>
<th>Name of Phase</th>
<th>Time for Start of Correlation, (GMT)</th>
<th>Expected Apparent Velocity, km/sec</th>
<th>Expected Apparent Velocity Assuming Two Dipping Interfaces, km/sec</th>
<th>Measured Apparent Velocity, km/sec</th>
<th>Measured Azimuth, deg.</th>
</tr>
</thead>
<tbody>
<tr>
<td>( P )</td>
<td>6h 12m 12.0s*</td>
<td>24.55*</td>
<td>22.10</td>
<td>21.0</td>
<td>344.2</td>
</tr>
<tr>
<td></td>
<td>6h 13m 18.0s†</td>
<td></td>
<td></td>
<td>17.4</td>
<td>345.1</td>
</tr>
<tr>
<td></td>
<td>6h 13m 22.0s†</td>
<td></td>
<td></td>
<td>16.4</td>
<td>325.7</td>
</tr>
<tr>
<td></td>
<td>6h 13m 33.5s</td>
<td></td>
<td></td>
<td>18.8</td>
<td>306.4</td>
</tr>
<tr>
<td></td>
<td>6h 15m 00.0s</td>
<td></td>
<td></td>
<td>18.2</td>
<td>356.3</td>
</tr>
<tr>
<td></td>
<td>6h 15m 24.8s</td>
<td></td>
<td></td>
<td>18.8</td>
<td>350.0</td>
</tr>
<tr>
<td></td>
<td>6h 15m 27.4s</td>
<td></td>
<td></td>
<td>18.9</td>
<td>349.6</td>
</tr>
<tr>
<td></td>
<td>6h 15m 32.5s</td>
<td></td>
<td></td>
<td>17.6</td>
<td>342.8</td>
</tr>
<tr>
<td></td>
<td>6h 15m 39.5s</td>
<td></td>
<td></td>
<td>22.9</td>
<td>343.0</td>
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<tr>
<td></td>
<td>6h 15m 50.0s</td>
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<td></td>
<td>6h 15m 58.0s</td>
<td></td>
<td></td>
<td>17.2</td>
<td>351.4</td>
</tr>
<tr>
<td>( PKiKP )</td>
<td>6h 16m 24.2s</td>
<td>59.0‡</td>
<td>46.8</td>
<td>45.9</td>
<td>334.3</td>
</tr>
<tr>
<td>( PP )</td>
<td>6h 16m 34.0s</td>
<td>15.07e</td>
<td>14.0</td>
<td>14.0§</td>
<td>342.0</td>
</tr>
<tr>
<td></td>
<td>6h 17m 19.5s</td>
<td></td>
<td></td>
<td>16.0</td>
<td>352.3</td>
</tr>
<tr>
<td></td>
<td>6h 17m 46.0s</td>
<td></td>
<td></td>
<td>14.6</td>
<td>326.0</td>
</tr>
<tr>
<td>( P_{CP}P )</td>
<td>6h 18m 54.0s</td>
<td>29.6d</td>
<td>26.1</td>
<td>31.3</td>
<td>343.2</td>
</tr>
<tr>
<td></td>
<td>6h 18m 59.5s</td>
<td></td>
<td></td>
<td>24.6</td>
<td>167.8</td>
</tr>
<tr>
<td></td>
<td>6h 19m 07.3s</td>
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<td>20.4</td>
<td>330.7</td>
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<tr>
<td></td>
<td>6h 24m 02.5s</td>
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<td>18.0</td>
<td>144.5</td>
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<td></td>
<td>6h 25m 17.0s†</td>
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<td>17.0</td>
<td>169.1</td>
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<td></td>
<td>6h 25m 30.0s</td>
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<td>15.2</td>
<td>180.8</td>
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<td></td>
<td>6h 25m 37.5s</td>
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<td></td>
<td>17.4</td>
<td>179.5</td>
</tr>
<tr>
<td></td>
<td>6h 26m 04.5s</td>
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<td></td>
<td>18.2</td>
<td>187.8</td>
</tr>
<tr>
<td>( PKKP )</td>
<td>6h 27m 47.1s</td>
<td>39.0d</td>
<td>47.0</td>
<td>46.8</td>
<td>162.4</td>
</tr>
<tr>
<td>( PKKP ) B</td>
<td>6h 27m 50.5s</td>
<td></td>
<td></td>
<td>47.4</td>
<td>165.1</td>
</tr>
<tr>
<td>( PKKP ) C</td>
<td>6h 28m 01.8s‡</td>
<td>62.0d</td>
<td>84.3</td>
<td>76.0</td>
<td>157.0</td>
</tr>
<tr>
<td>( PKKP ) D</td>
<td>6h 28m 03.1s‡</td>
<td>25.0d</td>
<td>28.2</td>
<td>29.6</td>
<td>167.1</td>
</tr>
<tr>
<td>( P_{CPPK} )</td>
<td>6h 32m 00.0s‡</td>
<td>25.4d</td>
<td>28.5</td>
<td>37.3</td>
<td>145.9</td>
</tr>
<tr>
<td>( PKKPPK )</td>
<td>6h 36m 09.5s‡</td>
<td>58.0d</td>
<td>78.0</td>
<td>82.0</td>
<td>201.3</td>
</tr>
</tbody>
</table>

* 1-second integration time used.
† 2-second integration time used.
‡ 1.2-second integration time used.
§ For \( PP \) the correlation peak occurred at a velocity of 39.6 km/sec and an azimuth of 304°, and the values quoted in the table correspond to the maximum of a slightly smaller peak. These results are due to interference from the waves associated with \( PKiKP \) or \( PKKP \).

Sacks [1967].
Bolt and O'Neill [1965].
Cleary and Hales [1966].
Jeffreys and Bullen [1948].

Comparison with the J-B times for other phases, giving \( PKiKP \) about 2 seconds earlier than observed. For \( P, PKiKP, PP \), the \( PKKP \) phases and two large unidentified phases, \( d\Delta/dT \) and azimuth were measured by the least-squares method as well as by correlation, thus enabling a direct comparison between the two methods. In the case of these larger phases, except \( PP \), the values of velocity and azimuth determined by the least-squares method are undoubtedly more reliable than the values obtained by the correlation method. The values of velocity and
<table>
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<tr>
<th>Name of Phase</th>
<th>Travel Time</th>
<th>Apparent Velocity, km/sec</th>
<th>Measured Azimuth, deg</th>
<th>Matching Technique Used</th>
<th>Mean Square Residual, sec^2 × 10^-3</th>
<th>Correlation Method</th>
<th>Expected Apparent Velocity, km/sec</th>
<th>Azimuth, deg</th>
<th>Expected Azimuth, deg</th>
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<tr>
<td>P</td>
<td>14m 13.5s</td>
<td>22.14 ± 0.19</td>
<td>344.4 ± 0.6</td>
<td>First zero</td>
<td>0.135</td>
<td>21.0</td>
<td>344.2</td>
<td>22.10</td>
<td>344.5</td>
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<tr>
<td>Unidentified</td>
<td>17m 26.8s</td>
<td>19.64 ± 0.27</td>
<td>347.7 ± 1.0</td>
<td>First peak</td>
<td>0.399</td>
<td>18.8</td>
<td>350.0</td>
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<tr>
<td>Unidentified</td>
<td>17m 29.4s</td>
<td>21.02 ± 0.35</td>
<td>344.5 ± 1.0</td>
<td>First peak</td>
<td>0.467</td>
<td>18.9</td>
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<td>PKiKP</td>
<td>18m 26.2s</td>
<td>45.25 ± 1.73</td>
<td>322.4 ± 3.2</td>
<td>First peak</td>
<td>0.771</td>
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<td>334.3</td>
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<td>346.2</td>
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<td>56.62 ± 2.42</td>
<td>349.1 ± 3.0</td>
<td>First peak</td>
<td>0.412</td>
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<td>342.7 ± 2.5</td>
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<td>342.0</td>
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<td>343.9</td>
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<td>PKKP A</td>
<td>29m 49.1s</td>
<td>42.16 ± 1.27</td>
<td>161.8 ± 2.3</td>
<td>First zero</td>
<td>0.490</td>
<td>46.8</td>
<td>162.4</td>
<td>47.0</td>
<td>159.5</td>
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<td>PKKP B</td>
<td>29m 52.5s</td>
<td>43.29 ± 1.17</td>
<td>166.3 ± 2.0</td>
<td>First peak</td>
<td>0.357</td>
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<tr>
<td>PKKP C</td>
<td>30m 03.8s</td>
<td>38.83 ± 1.08</td>
<td>170.1 ± 1.9</td>
<td>First peak</td>
<td>0.435</td>
<td>47.4</td>
<td>165.1</td>
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<tr>
<td>PKKP D</td>
<td>30m 05.1s</td>
<td>39.57 ± 1.22</td>
<td>172.2 ± 2.4</td>
<td>First peak</td>
<td>0.472</td>
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<td>P</td>
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<td>344.4 ± 0.6</td>
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<td>21.0</td>
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<td>0.399</td>
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<td>344.5 ± 1.0</td>
<td>First peak</td>
<td>0.467</td>
<td>18.9</td>
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<td>First peak</td>
<td>0.771</td>
<td>45.9</td>
<td>334.3</td>
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<td>346.2</td>
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<td>PKiKP?</td>
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<td>56.62 ± 2.42</td>
<td>349.1 ± 3.0</td>
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<td>PP</td>
<td>18m 37.2s</td>
<td>13.73 ± 0.39</td>
<td>342.7 ± 2.5</td>
<td>First peak</td>
<td>2.435</td>
<td>14.0</td>
<td>342.0</td>
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<td>0.490</td>
<td>46.8</td>
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<td>PKKP D</td>
<td>30m 05.1s</td>
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<td>172.2 ± 2.4</td>
<td>First peak</td>
<td>0.472</td>
<td></td>
<td></td>
<td></td>
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</tbody>
</table>
Individual Phases

**PKiKP.** The amplitude is nearly as large as P and is much clearer than previous observations of this phase at similar distances, such as those of Ergin [1967]. Other reports of PKiKP have been given by Melik-Gajkazan [1955] and Caloi [1961]. Melik-Gajkazan [1955] gives an empirical travel-time curve for PKiKP with forty-nine observations between 22° and 140° but does not reproduce any records. Bolt and O'Neill [1965] indicate that this phase is most likely to be observed for the distance range 105° < A < 110°, and they consider Caloi's identification of PKiKP at 15° as unlikely. These authors have indicated that positive identification of this phase would provide an independent estimate of the rigidity of the inner core. The measurements of \(d\Delta/dT\) and azimuth suggest that PKiKP may in fact be two separate phases. It is perplexing that measurements of the arrival times of the first two peaks of this phase give values of \(d\Delta/dT\) close to the...
expected value but azimuths quite different from the true value. Moreover, a later peak gives an azimuth close to the expected value but a rather high velocity. These results may be due to interference with a scattered signal. If two phases are present, one interpretation is that these waves correspond to the reflected phase PKiKP, closely followed by the phase PKIKP resulting from refraction through the inner core.

**PP.** The values of velocity and azimuth obtained by both the correlation and the least-squares methods are in good agreement, but interference from other signals resulted in a very poor least-squares fit. An interesting fact is that two unidentified phases show up more clearly than PP, and these are discussed later.

**PcP**. Previous identifications of this phase are either very rare or absent. The evidence for its identification is very convincing; the azimuth and apparent velocity are in reasonable agreement with the expected values, and the travel time is within 3 seconds of that expected value. It is of some interest that this phase has a travel time that differs from the expected time for PPP by only 1 second. However, the apparent velocity rules out the possibility that it is PPP.

**PKKP.** According to the J-B tables, there are three possible paths for PKKP, each phase traveling in the reverse direction. The onset of the first PKKP is remarkably sharp. Both correlation and least-squares measurements of velocity and azimuth agree fairly well, but the results suggest four phases. The first and fourth of these have values of $\Delta A/\Delta T$ that agree well with the values for the first and third phases given by Jeffreys and Bullen; the arrival of the fourth phase is not clearly defined. One interpretation of these results is that the mid-

Fig. 4. *PcP**. Array phased to an azimuth of 343.2° and a velocity of 31.3 km/sec.
An unsuccessful search was made for the phases PKJKP and PKIIKP.

**Unidentified Signals**

Between P and PP there are ten unidentified coherent signals all with measured apparent velocities between 17.4 and 22.9 km/sec and with azimuths within 45° of the azimuth of the explosion. These signals are spread over a time interval of 3 minutes 47 seconds, and it is unlikely that they are due to scattering or multiple reflections in the vicinity of the array. They appear to have originated, presumably by reflection, at a considerable distance from the array, though reverberation effects at the explosion site may be partly responsible. After PP there is a small unidentified phase, possibly a core phase, traveling in the reverse direction at 47.4 km/sec, and with a total travel time of 19 minutes 21.5 seconds; no adequate explanation of this phase has been found. Between PP and PKKP there are several signals arriving in the reverse direction with velocities of

---

![Figure 5](image-url)  
*PKKP. Array phased to an azimuth of 162.4° and a velocity of 46.8 km/sec for traces 1 and 2 and to 167.1° and 29.6 km/sec for traces 3 and 4.*
Fig. 6. Two large unidentified phases. Phase A, P reflected at M discontinuity. Phase B, P reflected at surface. See Table 7.

less than 30 km/sec. It is suggested that these signals are associated with reflection and scattering of core phases reaching the surface somewhere between 106° and 180° from Novaya Zemlya.

There are two remarkable phases with travel times of 17 minutes 26.8 seconds and 17 minutes 29.4 seconds; the onset of the second of these is very sharp. Hand measurements of the onset times give azimuths fairly close to the true value of 342.9°. With the assumption that these phases result from a simple reflection at a dipping interface, the points of reflection have been deduced from the azimuths and apparent velocities corrected for structure. The distance was obtained by comparing the measured velocity, corrected for structure, with the apparent velocities derived from the Cleary-Hales travel-time curve. The distances of these points of reflection from the epicenter were then calculated, and the P travel times for the two paths were added. For the first of these phases the travel time for P reflected at a dipping M discontinuity agrees with the actual value to within 3 seconds, but the second larger phase does not fit so well, the travel time being 13 seconds longer than for a surface reflection. By taking the first phase, the angle of dip required to give such a reflection has been found to be about 10° (see Figures 6 and 7 and Table 7). Further, the dip would have to be toward the mountain ranges of central Asia; the points of reflection are situated just to the north of the Tannu Ola and Yablanovy ranges. Points of reflection for all other phases, assuming they are some form of reflected P, have been determined, but attempts to fit travel times and ray paths have not been entirely successful.
LONGITUDINAL WAVES FROM NOVAYA ZEMLYA EXPLOSION

Fig. 7. Diagram illustrating the reflection of the phase arriving at 6h 15m 24.8s.

Many of the unidentified phases can be explained if reflections at discontinuities in the upper mantle are postulated; this idea is now being investigated.

DISCUSSION

This study has shown how a large nuclear explosion can be used to determine crustal models suitable for correcting apparent velocities and azimuths over a limited azimuth range. These crustal models have been tested by using the measured apparent velocities and azimuths for seven earthquakes. For none of these seven events is the apparent velocity sufficiently well known to enable one to say definitely whether the assumed apparent velocity for diffracted $P$ of 24.55 km/sec is correct. However, there is no evidence to suggest that the velocity used is seriously in error, and it is unlikely to be out by more than 0.2 km/sec. The least-squares method has been shown to be ideal for obtaining precise measurements of azimuth and apparent velocity; with crustal models derived in the manner described, the method should be capable of giving accurate estimates of $dT/d\Delta$. Moreover, the results have shown that careful work can reveal minor irregularities in the slope of the travel-time curve. There are differences in the azimuths and velocities obtained by using different matching techniques for the same event, which are due to systematic changes of wave form from one seismometer to another; these differences for some earthquakes are greater than the differences due to any measurement errors.

It has been demonstrated that hand measurements of relative onset times can give good results for apparent velocities and azimuths of later $P$ phases as well as initial $P$. The correlation method described by Birtill and Whiteway [1965] has been shown to be unsuitable for precise determinations of apparent velocities but is very useful in picking out small phases where the signal to noise ratio is small; the values of $d\Delta/dT$ obtained by this method probably have an error of 15%, even when the signal to noise ratio is large. A new correlation technique that appears to give more reliable measurements of $d\Delta/dT$ and azimuth has been described by Muirhead [1968] and is now being thoroughly investigated.

TABLE 7. Two Large Unidentified Phases Arriving between $P$ and $PP$

<table>
<thead>
<tr>
<th></th>
<th>$P$ Reflected at 33 km</th>
<th>$P$ Reflected at Surface</th>
</tr>
</thead>
<tbody>
<tr>
<td>Arrival time of phase (GMT)</td>
<td>6h 15m 24.8s</td>
<td>6h 15m 27.4s</td>
</tr>
<tr>
<td>Travel time</td>
<td>17m 26.8s</td>
<td>17m 29.4s</td>
</tr>
<tr>
<td>Corrected apparent velocity</td>
<td>21.65 km/sec</td>
<td>22.80 km/sec</td>
</tr>
<tr>
<td>Corrected azimuth</td>
<td>347.1°</td>
<td>343.8°</td>
</tr>
<tr>
<td>Point of reflection</td>
<td>60.7°N, 107.6°E</td>
<td>62.1°N, 98.1°E</td>
</tr>
<tr>
<td>$\Delta_1$, distance from WRA to point of reflection</td>
<td>83.2°</td>
<td>86.7°</td>
</tr>
<tr>
<td>$t_1$, travel time from point of reflection to WRA</td>
<td>12m 24s</td>
<td>12m 47s</td>
</tr>
<tr>
<td>$\Delta_2$, distance from Novaya Zemlya to point of reflection</td>
<td>23.2°</td>
<td>19.3°</td>
</tr>
<tr>
<td>$t_2$, travel time from Novaya Zemlya to point of reflection</td>
<td>5m 05s</td>
<td>4m 29s</td>
</tr>
<tr>
<td>Calculated travel time</td>
<td>17m 29s</td>
<td>17m 16s</td>
</tr>
</tbody>
</table>
Measurements of azimuth and apparent velocity for most later phases yield azimuths different from the expected values, and for the larger phases there is reasonable agreement between correlation and least-squares measurements. Correlation techniques have shown that several coherent signals arriving between PP and PP are not associated with the structure in the vicinity of the array but may be the result of reflections between the array and the source. The signals that arrive after PP in the reverse direction may be associated with reflection and scattering of core phases reaching the surface between 106° and 180° from Novaya Zemlya.

Acknowledgments. We are indebted to Dr. J. R. Cleary and Dr. R. Underwood for critically reading the manuscript and offering suggestions for its improvement. We would also like to thank Dr. E. W. Carpenter and Mr. B. S. Gopalakrishnan for supplying the onset time analysis program.

References


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P-WAVE TRAVEL TIME GRADIENT
MEASUREMENTS AND LOWER MANTLE STRUCTURE

C. WRIGHT
Department of Geophysics and Geochemistry,
Australian National University, Canberra

Received 25 January 1970

535 values of the P-wave travel time gradient have been derived from 192 events recorded at the Warramunga Seismic Array over the distance range 28°–99°. These dT/dA measurements have indicated the presence of regions of anomalous velocity gradient in the lower mantle. The results are compared with those of other recent array investigations.

The travel time gradient, dT/dA, for P wave arrivals from earthquakes and explosions, measured as a function of distance, enables a velocity distribution for the Earth’s mantle to be derived by the Herglotz-Wiechert method [1]. The installation of arrays of short period vertical component seismometers in various parts of the world during the last decade has allowed dT/dA to be evaluated directly instead of smoothing and differentiating a travel time curve. Thus in principle arrays permit the fine details of mantle structure to be better elucidated. Consequently several dT/dA studies have been published recently [2–5], and these strongly suggest the existence of anomalous regions below 700 km in the mantle. The interpretation of array data, however, is beset with ambiguities and uncertainties. The fine details present in dT/dA curves may not be due to velocity anomalies at great depths; irregularities in both crustal and upper mantle structure in the vicinity of an array and possibly even lateral variations in mantle structure near an earthquake focus can introduce complexity into a dT/dA curve. These difficulties can only be resolved by comparing the results of several investigations. The purpose of this communication is to present the results of another dT/dA study for the lower mantle, and to show that the discrepancies in the previous investigations are largely due to differences in interpretation and not necessarily to subtle irregularities in local structure or differences in structure at great depth.

The Warramunga Seismic Array, WRA, near Tennant Creek in the Northern Territory of Australia, consists of twenty short period vertical component seismometers arranged in two lines of ten, each 22.5 km long and approximately at right angles. The travel time gradients and azimuths of P-wave onsets from 192 events recorded at the array over the distance range 27.9° to 98.8° have been measured using the techniques described by Cleary et al. [6] and by Wright and Muirhead [7]. Two or three measurements of these quantities were made for each event by matching different portions of the P-wave train, so that a total of 535 values of both dT/dA and azimuth has been obtained. These values often differ significantly from the true azimuth and the dT/dA values predicted by the J-B tables [8] or the 1968 Seismological tables for P Phases [9]. Further, errors in the measurements of onset times, in epicentre determinations or in the travel time tables themselves can only account for anomalies an order of magnitude smaller than those observed. A crustal refraction experiment performed by Underwood [10] and attempts to fit a single or a series of dipping interfaces below the array in the manner described by Niazi [11], Cleary et al. [6], Wright [12] and Wright and Muirhead [7] have
yielded results of limited value. Hence the anomalies in $d^2T/d\Delta$ and azimuth cannot be explained by any relatively simple crustal model, and may be partly due to upper mantle structure beneath the array. The only previous lower mantle $d^2T/d\Delta$ study using an array similar in size to WRA was published by Gopalakrishnan [13], and his results give little indication of whether irregularities in local structure are particularly serious.

At WRA the azimuth and $d^2T/d\Delta$ anomalies are for most azimuths slowly changing functions of azimuth and of the angle of incidence of the P waves at the array site. Recent travel time studies [14—16] have shown marked deviations from the J-B times without enabling distinct anomalous regions of the lower mantle to be identified. Thus it is reasonable to suppose that any irregularities in lower mantle velocities will cause perturbations of a modern travel time curve no larger than the random errors in the travel times themselves. Therefore by constraining the $d^2T/d\Delta$ data to an acceptable travel time curve, the effect on the P wave arrivals of the local structure can be largely eliminated.

The $d^2T/d\Delta$ data have been divided into seventeen groups each corresponding to a narrow range of azimuths and a distance range of about 10°. Each value of $d^2T/d\Delta$ was treated as an independent observation and was weighted according to the reciprocal of the square of its standard error. A smooth curve was fitted through each group by the Method of Summary Values [17—19], and a $\chi^2$ test was used to check the smoothing procedure. The area under each smooth curve was evaluated and subtracted from the travel time differences between the end points of the range of integration derived from the 1968 Seismological tables for P Phases [9]. This enabled a correction to be applied to every value of $d^2T/d\Delta$ in a particular group. The method outlined above merely involves changing the baseline of each set of $d^2T/d\Delta$ measurements, and is similar to using a single plate dipping interface to correct each group of data. A smooth curve was then fitted through the corrected values of $d^2T/d\Delta$ by the Method of Summary Values, and is shown in fig. 1. Only 492 of the corrected points were used in deriving the smooth curve, and a $\chi^2$ test gave 487.5 on 470 degrees of freedom. A P wave velocity distribution from a depth of 700 km to 2790 km was calculated from the $d^2T/d\Delta$ curve using the Herglotz-Wiechert method. A modified version of Early Rise Model 2 of Green and Hales [20] was used to 'strip' the earth to a depth of 700 km. Although there are probably systematic errors in the 1968 Seismological Tables for P phases [21,22] they were used in preference to the surface focus travel times of Hales et al. [23] because standard errors in the travel times were supplied [19]; these errors are essential if a statistical significance test is to be applied in the smoothing of the corrected data.

The most interesting feature of fig. 1 is the flat portion of the $d^2T/d\Delta$ curve between 32° and 37° corresponding to anomalously low velocity gradients, $da/dz$, between 800 and 850 km. The velocity model shows a slight decrease in P wave velocity with increasing depth at 820 km. The $d^2T/d\Delta$ data have been over-smoothed in this distance range to avoid an apparent increase in $d^2T/d\Delta$ which would cause the Herglotz-Wiechert method to formally break down. Such an increase is best explained in terms of a low velocity layer [12]. Fig. 1 also shows an anomalous region between 46° and 52° corresponding to a small increase in $da/dz$ at a depth of about 1150 km followed by a region of low velocity gradients between 1260 and 1300 km. There may also be a slight increase in $da/dz$ near 1600 km, and there seems to be a comparatively flat region of the $d^2T/d\Delta$ curve between 63° and 66° corresponding to low velocity gradients between depths of 1700 km and 1750 km.
Unfortunately there are no data covering the region between 1600 km and 1730 km so that the evidence for an anomalous region is somewhat limited. Finally there is a fairly steep portion of the curve between 78° and 83.5° yielding comparatively high velocity gradients between about 2200 and 2500 km. It is important to point out that the anomalous regions of the $dT/d\Delta$ curve are apparent in the uncorrected data, and with the possible exception of the anomaly near 80° have not been affected when combining the corrected $dT/d\Delta$ values. Although the uncorrected data do suggest an increase in the slope of the curve near 80°, the effect may be exaggerated slightly when the results from different azimuth ranges are combined.

It is now pertinent to compare these results with those of other workers to see if a coherent picture emerges. Chinnery and Toksöz [2] discussed only the flat portions of their $dT/d\Delta$ curve. The plateau region at 52° is well established by their data, but those near 35° and 70° are spread out over a considerable distance range. It is also worth pointing out that Chinnery and Toksöz's $dT/d\Delta$ curve has relatively steep regions near 60° and 78°. Johnson [4] stated that the anomalous regions of this $dT/d\Delta$ data near 34.5° and 70.5° were poorly defined; his CIT206 curve reveals both regions that are relatively flat and relatively steep, but the steep regions at distances close to 40.5°, 49.5° and 81.5° appear to be more outstanding. The two $dT/d\Delta$ curves of Greenfield and Sheppard [3] show flat regions between 30° and 34°, and one of them has a flat region near 70°. The $dT/d\Delta$ data of Chinnery [5] show an anomalous region near 30°, but unfortunately there are few observations between 31° and 37°. His data do indicate a relatively flat part of the curve near 50° which may be preceded by a comparatively steep region. The anomalous features of lower mantle structure suggested by the data of fig. 1 taken in conjunction with other array investigations are summarised in table 1. It is may view that the differences in P wave velocity models are adequately explained by the scatter and irregular distribution of the basic $dT/d\Delta$ data. Regional differences in structure below 900 km do not have to be invoked, though this is not meant to imply that they do not exist. However, the sharpness of the velocity changes is not known; relatively abrupt increases in velocity may give rise to triplications in the travel time curve as suggested by Johnson [4] and Chinnery [5] in which the individual branches are too close together to be observed on seismograms.

A more detailed discussion of the method of analysis and the results presented here, supplemented with upper mantle data, will be published later.

I am indebted to Mr. H.A. Doyle for critically reading the manuscript.

<table>
<thead>
<tr>
<th>Approximate depth range (km)</th>
<th>Corresponding distance range (deg)</th>
<th>$d\alpha/dz$</th>
</tr>
</thead>
<tbody>
<tr>
<td>800—850</td>
<td>32—37</td>
<td>Low and probably negative over part of the range. Regional differences possibly important.</td>
</tr>
<tr>
<td>1150—1260</td>
<td>46—49.5</td>
<td>High</td>
</tr>
<tr>
<td>1260—1300</td>
<td>49.5—52</td>
<td>Low</td>
</tr>
<tr>
<td>1540—1660</td>
<td>59—62</td>
<td>High</td>
</tr>
<tr>
<td>2150—2450</td>
<td>77—83</td>
<td>Relatively high</td>
</tr>
</tbody>
</table>

Other anomalous regions probably exist but have only been suggested by one or two authors.

* The lower mantle has been taken as the region from a depth of 700 km to the mantle-core boundary.

References


COMMENTS ON A PAPER BY E. P. ARNOLD, "SMOOTHING TRAVEL TIME TABLES"

BY CEDRIC WRIGHT AND DEREK J. CORBISHLEY

Arnold (1968) discusses various techniques for smoothing travel-time tables, and concludes that the Method of Summary Values (Jeffreys, 1961) is the most useful. He then applies this method to the raw travel times of Herrin et al. (1968) that form the basis of the 1968 Seismological Tables for P Phases.

Unfortunately there are some numerical errors in Arnold's computations which were discovered independently by the two authors. We feel it is worthwhile to draw attention to these errors because many seismologists may wish to use the Method of Summary Values and check their calculations using Arnold's data.

1. The distance value of the second summary point in Arnold's Table 3 is incorrectly printed as 20.486. The true value is 28.486.

2. The numerical values of the ordinates of the summary points in Arnold's Table 3 are all too large by approximately 0.3 seconds and could not have been used in obtaining the smoothed travel times of Arnold's Table 2. In other words, if Arnold's summary values are interpolated, a new travel time curve emerges with values approximately 0.3 seconds greater than those quoted in Table 2.

3. Values of \((O - C)\) in Table 2 differ significantly from values obtained if the smoothed travel times (Table 2) are subtracted from the unsmoothed values (Table 1) at the distances 27°, 31°, 62° and 101°. At these distances the differences should read 0.06, 0.04, 0.04 and -0.10 seconds respectively, in place of -0.01, 0.10, 0.13 and -0.15 seconds if the data of Table 1 are correct. The overall effect is to reduce the total value of \(\chi^2\) by 4.9, resulting in a total \(\chi^2\) of 45.5 on 63 degrees of freedom (as opposed to \(\chi^2 = 50.4\) quoted in Arnold's paper).

4. The large differences between the smoothed and raw travel times at 20.25° and 21.0° indicate that the smoothed curve was not obtained from the tabulated data at these distances—note the high value of \(\chi^2\) at 20.25° which accounts for nearly 20 per cent of the total \(\chi^2\). Our own calculations of a smoothed curve from the tabulated data differ from Arnold's between 20.25° and 24°. With the corrections listed in section 3, \(\chi^2\) is now only 36.4 on 63 degrees of freedom which is well below the acceptable lower limit of 51.8.

5. Arnold interpolated the summary points using third divided differences. However, one of us (C.W.) found that fourth divided differences do make a small but significant contribution to the results below 37°. Contributions from fifth and sixth divided differences are virtually negligible. With these further corrections, \(\chi^2\) becomes 37.5.

The data in Arnold's tables thus indicate that the travel times have not been smoothed sufficiently, even allowing for possible over-estimation of the standard errors.

References
