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SEISMIC AND GRAVITY INVESTIGATIONS
OF THE CENTRAL VOLCANIC REGION,
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Abstract

Gravity and seismic refraction studies were undertaken in order to investigate the geological structure of the Central Volcanic Region. A detailed analysis of density determinations from bore-hole rock samples, three seismic refraction surveys and a spectral analysis of the magnetic anomaly field are described.

Interpretation of the observed gravity anomaly field for the Central Volcanic Region is initially undertaken by analytically separating the observed anomaly field into its regional and residual components; the almost entirely negative residual anomaly field is then interpreted in terms of varying thicknesses of near-surface, low-density volcanic rocks. At Mangakino and just west of Taupo, however, it is found that the calculated gravity anomaly effect of the seismically determined thickness of low-velocity, and hence low-density, volcanic rocks is less negative than the observed residuals; at both locations "secondary residuals" of about $-200 \mu\text{N/kg}$ remain unexplained. Models are presented that account for these secondary residuals as being due to discrete volumes of low-density molten rhyolite emplaced within the seismic basement.

The second method of gravity interpretation used in this study involves modelling all components of the observed gravity anomaly field. This necessitated giving consideration to both the gravity effect of the subducted Pacific plate and to seismic data bearing upon the variation of crustal thickness and mantle density throughout the central North Island. A gravity model for the central North Island is developed for which the important features are:

(iii)

i) The crust of the Central Volcanic Region is deduced to be only about half the normal continental thickness, and underlying the crust is an "anomalous", low-density upper mantle. This finding from the gravity model is supported by the results of a previous study of upper mantle seismic velocities and from the interpretation of a long-range seismic refraction survey carried out within the Region. These seismic data indicate the depth to, and the velocity of the upper mantle beneath the Region to be 15 km and 7.4 km/s respectively.

ii) The positive gravity anomalies that predominate over the western and northwestern North Island can largely be explained by gravity edge-effects associated with variations in the crustal thickness and mantle density within the back-arc areas of the North Island.

The gravity model is interpreted as lending support for a previously made proposal that the Region is the site of asymmetric back-arc spreading, and that the crustal rocks now being created are transitional in character between typical oceanic and typical continental.

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CHAPTER 1 INTRODUCTION

The North Island of New Zealand lies in one of the zones of great geophysical and geological activity which have variously been described as orogenic zones, mobile belts, island arcs or active continental margins. In terms of plate tectonics New Zealand straddles the Indian-Pacific boundary in the manner depicted by fig. 1.1. The shift-pole position and rotation direction for this plate boundary is deduced by Chase (1978) to show dextral-convergent motion in the New Zealand region. In the vicinity of the North Island the Pacific plate is subducting beneath the Indian plate forming a northwestward dipping Benioff zone, and subduction terminates at a latitude of about 41°S (see fig. 1.2).

A convergent boundary where subduction occurs has been termed an asymmetric active region (Evison, 1968) because geophysical features, including the level of the solid surface, display an asymmetry that contrasts with the symmetry of oceanic rises. The side of the subduction zone where the solid surface is high-standing is characterised by regions of volcanism and high heat-flow, and often by extensional tectonics; regions where extension occurs have variously been described as marginal basins, inter-arc basins (Karig, 1971) or back-arc basins (Uyeda, 1977). The term back-arc basin will be used here. Back-arc basins that form at continental margins have been called ensialic back-arc (or inter-arc) basins (Scholz *et al.*, 1971; Hovarth *et al.*, 1975; Tarney & Windley, 1981). Ensialic back-arc basins show a wide range of forms and features from simple extensional faulting, to rupturing of the continental lithosphere and separation of the volcanic arc from the stable continental region by a basin beneath which the crust is oceanic in type.

Fig. 1.1 A mercator projection of the New Zealand region showing the velocity of the Pacific plate, in mm per year, relative to the Indian plate as inferred by Chase (1978). Bathymetry after Lawrence (1967) and the Central Volcanic Region is shown by the stippled pattern.

Fig. 1.2 Map of the North Island showing the isobaths (dotted lines) of deep seismic activity (after Adams & Ware, 1977) and some of the tectonic units of the North Island. Note that the Taupo Volcanic Zone (TVZ) forms the eastern half of the wedge-shaped Central Volcanic Zone (CVR). Low-potash andesite volcanoes that have been active in the last 10,000 years are shown by stars. The single open star represents the high-potash andesite volcano of Mt Egmont.

Seismic evidence presented by Thompson & Evison (1962) and Garrick (1968) point to the overall crustal structure of the North Island as being continental in character. However, analogies of the geological features that are developed within oceanic island arc systems can also be found in the North Island. Fig. 1.3 shows a generalised model of the principal structural elements typically found within island arc systems of the western Pacific (after Karig, 1974) and how various features of this model can be related to geological structures of the North Island.

Karig (1970) placed what he termed the Taupo Volcanic Zone of the North Island into its plate-tectonic context when he noted that the Zone is an apparent continuation of the Havre trough - the back-arc basin of the Tonga-Kermadec trench-arc system. The area that Karig calls the Taupo Volcanic Zone closely resembles the area referred to in this thesis as the Central Volcanic Region; the distinction in terminology is an important one and will be explained shortly. On the basis of the Central Volcanic Region and other examples, Karig & Jansky (1972) have postulated a general progression, with increasing amounts of extension, from volcano-tectonic rift zones, which they also describe as the earliest phase of extension in a trench-arc system, to back-arc basins with oceanic crust. In this progression they class the Central Volcanic Region as a volcano-tectonic rift zone.

The location of what will be referred to in this thesis as the Taupo Volcanic Zone is shown in fig. 1.3. Past geophysical and geological studies of the volcanic areas of the North Island have tended to concentrate on the Taupo Volcanic Zone, for it is here that the economically important and visually impressive geothermal activity is found. Large volumes of rhyolitic rocks have been erupted from the

Fig. 1.3 Diagram showing some of the associations made by Cole & Lewis (1981) between various geological structures of the North Island and the generalised structural model of a western Pacific island arc system as given by Karig (1974). Upright triangles represent low-potash andesite volcanoes that have been active within the last 10,000 years. Note the apparent continuation of the Havre trough with the Central Volcanic Region as remarked upon by Karig (1970).

Zone in the last million years (Healy, 1962; Cole, 1979) and the average heat flow in the Zone is calculated by Studt & Thompson (1969) to be 0.83 W/m^2 ; a value some 14 times higher than the continental average of 0.06 W/m^2 . Recent analyses of repeated geodetic surveys show high strain rates (Walcott, 1978) and what has been interpreted by Sissons (1980) to be back-arc spreading. The Taupo Volcanic Zone forms part of a much larger area of Quaternary volcanism in the North Island - the Central Volcanic Region. Thompson (1964) first used the term Central Volcanic Region when describing an expanse of some $26\,000 \text{ km}^2$ that lies between the southward converging Mesozoic greywacke ranges of the North Island and is covered by ignimbrite sheets and other rocks of a rhyolitic or andesitic composition. The present usage of the term Central Volcanic Region, shown in fig. 1.3, is slightly different from that used by Thompson, and reflects the structural concepts that will be developed from the gravity and seismic studies of this thesis.

Although there is both geophysical and geological (Lensen, 1977) evidence to suggest that the Taupo Volcanic Zone is an area of tension, how much extension or spreading has occurred is difficult to establish. Typical oceanic spreading centres are characterised by the following features: ocean ridge topography; high heat flow; linear, symmetrically arranged magnetic anomalies; and transform faults. On land, however, the identification of spreading centres is rendered difficult by a thick sedimentary or volcanic cover and magnetic anomalies which are diffuse and of low amplitude (Elders *et al.*, 1972; Lawver & Hawkins, 1978). Hence the patterns of linear magnetic anomalies that characterise the ocean floor are not present and other lines of evidence must be used to identify spreading centres. Among these might be heat flow and gravity anomalies.

Large dipolar free-air and isostatic gravity anomalies are developed across subduction zones. These anomalies have been ascribed by Hatherton (1969a), Grow (1973) and others to mass anomalies directly associated with subducted plates, and theoretical models (Minear & Toksöz, 1970) indeed predict that a subducted plate will be colder and therefore more dense than the surrounding asthenosphere at depths greater than about 80 km. On the other hand, Watts & Talwani (1974, 1975) have proposed that the positive component of the gravity anomaly pair observed across subduction zones may, in regions of extension, be due not so much to the excess mass of the subducted plate but to gravity anomaly effects associated with extension. The North Island of New Zealand presents a favourable location to examine this question for the following reason. Because the southern limit of the Central Volcanic Region occurs within the middle of the North Island yet the subducted Pacific plate is inferred, from seismic evidence, to be present for a further 350 km to the southwest (fig. 1.2), it should be possible to separate from the gravity anomaly map of the North Island those effects arising from the subducted plate and those associated with extension in the back-arc regions of the central North Island.

This thesis describes seismic refraction and gravity interpretation work carried out in the Central Volcanic Region. The principal aims of the study are:

- i) To determine from seismic refraction surveys and analyses of bore-hole densities the thickness, seismic velocity and density structure of the superficial volcanic rocks at various localities within the Central Volcanic Region.

ii) To present a gravity interpretation model for the Central Volcanic Region which accounts for all components of the gravity anomaly field, and is consistent with current geological concepts.

iii) To identify the western boundary of the Central Volcanic Region and relate its position to the tectonic evolution of the North Island since late Tertiary times.

CHAPTER 2 PREVIOUS GEOPHYSICAL STUDIES OF THE NORTH ISLAND
WITH SPECIAL REFERENCE TO THE CENTRAL VOLCANIC REGION

In this chapter a summary is presented of previous geophysical observations and interpretations relevant to the work undertaken in this thesis. The first section concentrates on regional aspects of North Island geophysics; the second deals more specifically with geophysical aspects of the Central Volcanic Region.

2.1 North Island Geophysics

2.1.1 Seismic studies of the North Island

Seismological topics relevant to this study are the pattern of earthquake hypocentres and seismic velocity and attenuation properties of the upper mantle. The results are summarised as follows:

1) Seismicity pattern: Crustal seismicity is manifest in a 200-300 km wide band passing through a large part of the North Island and the northern end of the South Island; this is the Main Seismic Region described by Eiby (1971). Earthquakes deeper than 100 km are represented by the isobaths shown in fig. 1.2.

2) Seismic properties of the upper mantle: From a study of the transmission of high-frequency seismic waves beneath the North Island Mooney (1970) found two distinct regions - a high-frequency transmitting region and an attenuating region - as shown in fig. 2.1. The interpretation offered by Mooney is that there is a 50 km thick tabular attenuating region centred at a depth of 100 km beneath the western North Island. He adds, however, that the data could alternatively be reconciled with a dipping attenuating zone parallel to and above the Benioff Zone.

Fig. 2.1 Regions of the upper mantle beneath the North Island with attenuating and transmitting properties (after Mooney, 1970). Also shown is the region found by Haines (1979) where anomalously low velocities of P_n and S_n are found.

Haines (1979) reports upper mantle compressional and shear wave velocities (Pn and Sn velocities respectively) calculated from interval velocities of earthquake waves travelling beneath various stations of the New Zealand Seismograph Network. Anomalously low velocities of Pn (7.4 km/s) and Sn (3.95 km/s) were found to correspond to a region which includes most of the Central Volcanic Region (fig. 2.1). Moreover, Haines noted that Sn phases from earthquakes located beneath the East Coast did not appear on records at stations to the west and northwest of the Central Volcanic Region. Haines related these anomalous seismic properties of the upper mantle to possible variations in petrology or temperature.

2.1.2 Regional gravity anomalies

The Bouguer and isostatic gravity anomalies of the central North Island are shown in figs. 2.2 and 2.3 respectively. The isostatic anomalies are calculated on the basis of an Airy-Heiskanen earth model with a crustal thickness of 30 km and a density of compensation equal to -0.6 Mg/m^3 (Reilly *et al.*, 1977). Dominating both maps is the intense band of negative anomalies, named the Rangitikei-Waiapu anomaly by Robertson & Reilly (1958), that coincide, in part, with thick deposits of low density Tertiary-Quaternary sediments. Positive Bouguer and isostatic gravity anomalies of 500 and 650 $\mu\text{N/kg}$, respectively, occur on the western side of the North Island. These positive anomalies will be referred to as the North Island back-arc gravity high.

While the Bouguer and isostatic gravity anomalies show many similarities the isostatic anomaly map brings the principal gravity anomalies of the North Island into sharper focus. Moreover, several

Fig. 2.2 Bouguer gravity anomalies of the central North Island as would be observed at 1 km above sea level (after Reilly *et al.* 1977).

Fig. 2.3 Airy-Heiskanen ($T=30$ km, density of compensation $= -0.6 \text{ Mg/m}^3$) isostatic gravity of the central North Island as would be observed at 1 km above sea level (after Reilly *et al.* 1977).

geological and geophysical boundaries of the North Island are highlighted by the isostatic anomalies. For example, boundaries of the Central Volcanic Region receive emphasis from the isostatic anomalies and the path of the zero isostatic anomaly contour shows a close correspondence with the attenuating-transmitting boundary shown in fig. 2.1.

It was noted by Reilly (1978) that there is a lack of correlation of individual isostatic anomalies with station height for New Zealand as a whole: he interprets this to mean that isostatic compensation of New Zealand topography is achieved in a manner not differing too greatly from that assumed by the Airy hypothesis. An alternative interpretation of this lack of correlation is that the gravity effect of the compensation mechanism for New Zealand topography is not markedly different from the gravity effect of the compensating masses predicted by the Airy model of local isostatic compensation. This distinction is important for, as Jeffreys (1976, p.234) points out, the gravity effect of a local compensation mechanism at one depth can always be reproduced by the gravity effect of a more regional form of compensation at a shallower depth.

In any event, whichever form the true compensation mechanism is for New Zealand topography, the ubiquitous gravity gradient associated with a transition from oceanic to continental crust has largely been eliminated by the isostatic correction and isostatic anomalies can play a useful role for gravity interpretations made within New Zealand (e.g. Hatherton, 1966).

In examining the relationship between isostatic gravity anomalies and seismicity Hatherton (1969a) pointed to the close association that exists at a number of active margins between an asymmetric pair of positive and negative anomalies, on the one hand, and the Benioff zone, on the other,

Fig. 2.4. Two possible models proposed by Hatherton (1970) to explain the isostatic gravity anomalies (along A-A' of fig. 2.3) of the central North Island in terms of the subducted Pacific plate.

despite wide variations in the relative position and form of the associated oceanic trench. Two models were proposed by Hatherton (1970) (fig. 2.4) that satisfactorily explained a SE - NW isostatic gravity anomaly profile along A-A' in fig. 2.3. In model A the transition of negative to positive density contrast occurs at around 80 km - the depth at which andesitic volcanism is thought to be first generated (Dickinson & Hatherton, 1967). Model B, on the other hand, incorporates crustal downwarping and therefore goes some way towards accounting for the sedimentary basins that are more or less coincident with the Rangitikei-Waiapu negative gravity anomaly.

While Hatherton's studies highlight with impressive simplicity the association of gravity and seismicity at asymmetric active regions, a closer examination of the North Island association reveals a difficulty. While subduction of the Pacific plate is occurring as far south as latitude 41°S , the positive component of the asymmetric gravity pair, which according to the models of fig. 2.4 has its source within the subducted plate, can only be identified northwest of Mt Ruapehu (fig 2.3). Unexplained by Hatherton's models is an intense north-south gravity gradient in the region between Mt Egmont and Mt Ruapehu. This is a major feature of the gravity anomaly map and renders any two-dimensional treatment of the whole North Island gravity anomaly map inappropriate.

2.1.3 Crustal Structure

Knowledge of crustal structure for the North Island is incomplete. From the results of an unreversed crustal refraction profile near Wellington, Eiby (1957) concluded that the crust had a thickness of about 18 km, but has since (Eiby, 1968) accepted as definitive a reinterpretation of the

original data by Garrick (1968) giving a total crustal thickness of 36 km.

From a study of dispersive earthquake waves, Thompson & Evison (1962) propose a typical continental thickness of crust between 30-40 km over much of New Zealand. A similar result was found by Reilly (1962) when he applied to New Zealand correlations, made for other continental areas, between Bouguer gravity anomalies and seismically determined crustal thickness. While these methods provide an average crustal thickness for New Zealand there may be substantial variations from this average value. This is particularly so for the North Island where in some places the gravity anomalies and the velocity of Pn show large variations within distances of 200 km or so.

During the 1960's intensive long-range seismic refraction work, particularly in North America, showed that the velocity of Pn in continental areas may be correlated to crustal thickness. For example, Woollard (1970) from about 100 observations made in North America, found a systematic relationship between Pn velocity and depth to the mantle (fig. 2.5). Of direct relevance to this study of the Central Volcanic Region are the seismic interpretations of the tectonically active Basin & Range Province of the Western United States. Here the Pn velocities are low (7.6-7.8 km/s) and the crust is found to be about half the thickness that would be predicted by Airy isostasy for the average 2 km elevation of the Province. This has been generally interpreted (e.g. Pakiser, 1963; Thompson & Burke, 1974) to imply that the low velocity of Pn is indicative of a low-density mantle which provides isostatic support for the anomalously thin crust. If the relationship of fig. 2.5 is applied to New Zealand the Pn velocity of 7.4 km/s for the Central Volcanic Region would imply a crustal thickness of about 16 km.

DEPTH OF MANTLE AS FUNCTION OF MANTLE VELOCITY

Fig. 2.5 The relationship found by Woollard (1970) between depth to the mantle and the velocity of Pn. Based on over 100 observations from North America.

Fig. 2.6 Hatherton's (1969b) map showing the distribution of andesites within the North Island. The NW-SE trending arcs are predominantly Tertiary in age whereas The Active Volcanic Front is Recent.

The idea of a distinct "crust" and "mantle" may be inappropriate in a region where rifting and spreading is occurring. Regions of the world where Pn velocities of 7.1 to 7.5 km/s occur are often associated with magmatic activity, high heat flow and extension. Three such examples are the Mid-Atlantic Ridge (Talwani *et al.*, 1965), the East African Rifts (Fairhead, 1976) and Iceland (Bott, 1965). The term "crust-mantle mix" has been applied by Cook (1962) to regions where these low Pn velocities are observed, implying that some intermingling of crust and mantle material is taking place. However, Talwani *et al.*, in their integrated geophysical study of the Mid-Atlantic Ridge, considered the term crust-mantle mix to be unsuitable as their gravity models required the 7.3 km/s layer to be, in some places, placed above and below normal mantle material with a velocity of 8.1 km/s. They regarded the 7.3 km/s material as normal mantle which had been subjected to a phase change and accordingly term it "anomalous mantle". Whatever the cause of these low Pn velocities it is possible that where they occur the boundary between crust and mantle will be gradational rather than sharp and discrete. For instance, Haines (1979) noted that many crustal earthquakes that are located beneath the central North Island failed to generate Pn or Sn phases suggesting, therefore, that there may not be clear boundary between the crust and mantle.

2.1.4 North Island Volcanism

Andesitic volcanism commonly associated with asymmetric active regions is well developed in the North Island. The distribution of calc-alkaline andesites with respect to the major tectonic features of the North Island is shown in fig. 2.6. Dickinson and Hatherton (1967)

have found that the percentage of potash in andesites, corrected for silica content, correlates with the depth of the Benioff zone under the volcano. Fig. 1.2 depicts the presently active volcanoes of low-potash affinity straddling the 100 km isobath of seismic activity, yet low-potash volcanoes of an older age are shown in fig. 2.6 to exist 80 to 400 km northwest of the volcanic front. Whereas the presently active volcanoes fall on a N.E. line the older low-potash volcanoes of the Coromandel and Northland Regions trend N.W.. Such a disparity in age, position and trend between andesites of similar chemistries prompted Hatherton (1969b) to suggest that the source zone of the low-potash andesites extended under Northland and Waikato in Tertiary times and has rotated in a clockwise manner since (fig. 2.6).

Subsequent availability of K-Ar ages of andesites led Calhaem (1973) to propose a quantitative scheme for the migration of volcanic activity across the North Island with time. The proposed scheme is shown in fig. 2.7. A $2^\circ/\text{My}$ clockwise rotation of the andesite axis through the North Island is postulated during the period 16-4 My B.P. and a 6° per My rotation rate thereafter. The 4 My axis coincides with a marked change in geology. West of the 4 My boundary surface geology consists predominantly of Mesozoic sediments interspersed with Tertiary-Quaternary volcanics; the volcanics are mostly andesites and basalts. In contrast, surface geology between the 4 and 0 My axes consists almost entirely of products derived from Quaternary rhyolite volcanism with a sole outcrop of Mesozoic greywacke exposed near Otamarakau. Accordingly, Calhaem proposed that the 4 My boundary represents a change in style of crustal extension. Volcanic intrusion into an existing lithosphere is postulated prior to 4 My and extension by active spreading and the

PROPOSED MIGRATION OF PLATE BOUNDARY

Fig. 2.7 Andesite ages (in My) and the proposed migration of the plate boundary as suggested by Calhaem (1973). The solid lines represent the proposed positions of the andesite axes (in My BP) and triangles represent subsequent andesite age determinations presented by Midha (1979).

creation of a new lithosphere from 4 My to the present day. Under this proposal the Otamarakau greywacke outcrop is viewed not to be part of a continuous greywacke basement beneath the volcanic region, but the surface expression of an isolated block, possibly allochthonous.

In terms of plate tectonics Calhaem considers the eastern boundary of the Taupo Volcanic Zone to represent an accreting margin of the Indian plate, and the region between this boundary and the Hikurangi trench to be part of a separate minor plate. Since Calhaem's proposal both Ballance (1976) and Eiby (1977) have advanced support for the existence of such a minor plate with Ballance terming it the Hawkes Bay Microplate. However, the southern and northern boundaries of the minor plate cannot unequivocally be identified and it is here considered to be a plate in the making, and accordingly will be referred to as the Hawkes Bay proto-plate.

More recent andesite age determinations supplied by Adams *et al.* (1974), Adams (1975) and Ballance (1976) are presented by Midha (1979) (see fig. 2.6). These data show that andesite ages from the Coromandel area vary from about 3 to 16 My whereas the data available when Calhaem made his proposal showed the oldest andesite on the Coromandel Peninsula to be only 7 My. Adams *et al.* (1974) suggest that the 16 My B.P. determinations, which correspond to andesites that are the least hydrothermally altered, represent the eruption age while younger ages represent subsequent hydrothermal events. Thus prior to 4 My the andesite migration history may have been somewhat different to what Calhaem initially envisaged. Nevertheless, the data of Adams *et al.* (1974) do not invalidate that part of Calhaem's proposal that is of interest to this study; namely that there has been an

apparent, rapid migration and rotation of the andesite axis in the last 4-5 My from a position adjacent to the Coromandel Peninsula to its present position today.

Although Calhaem's proposal was the first quantitative one for the development of the Central Volcanic Region posed within the context of plate tectonics, it has not met widespread acceptance from New Zealand geologists and geophysicists. For example, Ballance (1976) contends that while the volcanoes of the Taupo Volcanic Zone represent a continuation of the Tonga-Kermadec volcanic ridge, there is no continuation into the North Island of an extensional back-arc basin.

Subsequent to Ballance's work northwest-southeast extensional strain rates of $2 \times 10^{-7} \text{ y}^{-1}$ within the Central Volcanic Region have been calculated by Walcott (1978), who notes that this rate is consistent with Calhaem's spreading proposal. A recent analysis of repeated geodetic surveys in the Taupo Volcanic Zone by Sissons (1980) show secular widening of the Zone at the Bay of Plenty coast at a rate of $7 \pm 4 \text{ mm y}^{-1}$; Sissons interprets this as back-arc spreading.

Calhaem contends - "that the Benioff zone, trench and minor plate (Hawkes Bay proto-plate) have migrated together, relative to the Indian plate in a manner indicated by the andesite ages." Under this proposal the spreading rate at the Bay of Plenty coast should be $20\text{-}30 \text{ mm y}^{-1}$; a figure considerably greater than the geodetically determined rate of 7 mm y^{-1} . The difference may just reflect the fact that the geodetic rate was determined by data that only spans the last 70 years or, alternatively, the difference may be real, suggesting that the andesite migration rate is greater than that of the spreading.

2.2 Geological and geophysical background to the Central Volcanic Region

2.2.1 Boundaries of the Central Volcanic Region and the Taupo Volcanic Zone

The terms Central Volcanic Region and Taupo Volcanic Zone were introduced in Chapter 1, although no firm definition of these terms were given. It was pointed out however, that the Taupo Volcanic Zone occupies the eastern section of the Central Volcanic Region and they therefore both share a common eastern boundary. This boundary is represented by The Active Volcanic Front (Adams & Hatherton, 1973) which is a line joining the active andesitic volcanoes of Mt Ruapehu and White Island. All other andesitic or dacitic volcanoes of the Taupo Volcanic Zone which have been active within the last 10,000 years fall within 10 km of The Active Volcanic Front.

Fig. 2.8 shows a generalised geological map that places emphasis on the features that distinguish the western boundaries of the Central Volcanic Region and the Taupo Volcanic Zone. The area shown in fig. 2.8 as the Taupo Volcanic Zone is that defined by Healy (1964) and specifically excludes the active thermal areas around Rotorua. Gregg (1964) criticised Healy's definition, arguing that the Rotorua geothermal areas be included within the boundaries of the Zone. However, later workers (Cole, 1979) have tended to more or less follow Healy's definition and it will therefore be used here.

The Central Volcanic Region described by Thompson (1964) is a large area of the central North Island that represents "...the dominantly volcanic area lying between the two main diverging greywacke ranges...." The present usage of the term Central Volcanic Region closely follows

Fig. 2.8 A generalised geological map of the Central Volcanic Region (after New Zealand Geological Survey, 1972).

that given by Thompson except for the western boundary which is here taken as the line dividing Quaternary rhyolites to the east from Mesozoic greywackes or Tertiary andesites and rhyolites to the west. As well as having a firm geological basis this choice of the western boundary closely corresponds to the 4 My boundary of Calhaem (see fig. 2.7).

2.2.2 Rhyolite volcanism within the Taupo Volcanic Zone

The western boundary of the Taupo Volcanic Zone takes in the three rhyolite volcanic centres of Okataina, Maroa and Taupo and the andesitic volcanic centre of Tongariro (Cole, 1979; see fig. 2.9). The Rotorua volcanic centre is a circular basin occupied by Lake Rotorua, surrounded by rhyolite domes and lies on the western margin of the Taupo Volcanic Zone. Within the Taupo Volcanic Zone and the Rotorua Volcanic centre lie all the major geothermal fields of the central North Island.

There is a clear spatial separation of rhyolites from andesites within the Taupo Volcanic Zone. Rhyolites, in addition to small volumes of high-aluminium basalts, are restricted to the rhyolite volcanic centres whereas 95% of the andesites found at the surface are contained within the Tongariro Volcanic Centre to the south of Lake Taupo - Cole (1979) estimates the volume relationship for volcanic rocks within the Taupo Volcanic Zone to be: rhyolite = 12 000 km³, andesite = 260 km³, dacite = 5 km³ and high aluminium basalt = 2 km³.

West of the Taupo Volcanic Zone, but still within the Central Volcanic Region are the Mamaku and Kaharoa ignimbrite plateaux (Healy *et al.*, 1964). To the east of The Active Volcanic Front is the elevated yet remarkably flat Kaingaroa plateau. Drill-holes and seismic work (Hochstein & Hunt, 1970) show a thin veneer of ignimbrite overlying

Fig. 2.9 Map of the volcanic centres associated with the Taupo Volcanic Zone (modified from Cole, 1979). Dots represent recently active andesitic-dacitic volcanoes and squares represent recent basaltic deposits. Heavily shaded volcanic centres are rhyolitic; the light shaded Tongariro centre is andesitic. The fine line enclosing the Taupo Volcanic Zone represents the spatial extent of erupted ignimbrite sheets.

greywacke here, and structurally the plateau has been interpreted by Healy (1962) as a slice of the eastern greywacke ranges that has subsided towards the Taupo Volcanic Zone.

The earliest ignimbrites associated with North Island volcanism were erupted about 5-6 My ago from the Coromandel Peninsula (Ballance, 1976; Cole, 1979) and activity has progressively moved south since. The first ignimbrite eruptions from the Taupo Volcanic Zone are thought to have taken place about 1.1 My ago from the Maroa Volcanic Centre (Cole, 1979) and at one time there may have been an ignimbrite surface right across the Zone. This is suggested by the occurrence of ignimbrite in bore-holes at Wairakei at depths between 600 and 1200 m (Grindley, 1965). Surface exposures of ignimbrite are now, however, found mainly outside of the Taupo Volcanic Zone.

2.2.3 Residual gravity interpretations

Interpretations of four residual gravity profiles across the eastern margin of the Taupo Volcanic Zone were presented by Modriniak & Studt (1959).

Implicit in these interpretations are the following assumptions:

- i) A regional gravity anomaly field could be obtained by visually extrapolating greywacke-based Bouguer gravity anomaly values across the Central Volcanic Region. The resulting residuals are then considered to solely represent the gravity effect of varying thicknesses of low-density volcanics overlying basement.
- ii) That greywacke exposed on the flanks also constitutes the basement rocks of the Region.
- iii) That a mean density contrast of -0.5 Mg/m^3 exists between the proposed greywacke basement and the overlying volcanic cover.

Fig. 2.10 The residual gravity anomaly map and interpretations of Modriniak & Studt (1959). Map is in units of mgal and a density contrast of -0.5 Mg/m^3 was used for the interpretation models.

The residual gravity anomaly field and profile interpretations of Modriniak & Studt (1959) are shown in fig. 2.10. Although their interpretational models show the thickness of the low-density volcanic cover when considered in contrast to a basement with the density of greywacke, the models are not evidence that a greywacke basement beneath the Region actually exists. Greywacke has been found in drill-holes at Broadlands and Kawerau at distances of up to 5 km from the eastern boundary of the Taupo Volcanic Zone and at a variety of depths. Whereas Hochstein & Hunt (1970) interpreted these occurrences as being due to a continuous but severely block-faulted basement, Evison *et al.* (1976) consider them to be blocks which have become detached from the Kaingaroa Plateau by extensive igneous intrusions.

At Wairakei, some 15 km within the Region, an extensive drilling programme failed to locate greywacke or greywacke derived sediments. Thus it is of importance to note that the deepest hole, which was put down subsequent to the work of Modriniak & Studt, terminated in andesite at a depth of 2.2 km (Smith, 1970) yet the interpretational model along profile D (fig. 2.10) predicts greywacke at a depth of 1.6 km beneath Wairakei.

Calhaem's plate tectonic model of the Central Volcanic Region (section 2.1.4) denies the existence of a continuous greywacke basement beneath the Region and instead basement is envisaged to consist predominantly of cooled igneous material with geophysical properties resembling that of andesite. It is possibly for this reason that Hatherton (1974), when describing a residual gravity interpretation on the western side of Lake Taupo, noted that it is not necessary to place a petrological label on "basement", but merely that a density contrast of -0.5 Mg/m^3 exists between "basement" and the overlying volcanic cover.

2.2.4 Electrical work

Resistivity surveys within the Taupo Volcanic Zone have largely been directed to shallow investigations associated with geothermal exploration (Hatherton *et al.*, 1966). Geoelectromagnetic induction studies have been carried out in the southern portion of the Taupo Volcanic Zone by Hurst (1975) who found a shallow crustal conductivity anomaly of 0.1 mho/m extending to a depth of about 10 km.

Geomagnetic depth sounding work by Midha (1979) and dipole-dipole resistivity work by Bibby (reproduced by Midha, 1979) give similar results to that of Hurst. Fig. 2.11 shows Bibby's resistivity interpretation along a NW-SE profile which crosses the eastern boundary of the Taupo Volcanic Zone near Broadlands. The low-resistivity zone within the top 10 km is proposed to represent the zone of hot-water circulation associated with the hydrothermal fields (Midha, 1979).

2.2.5 Magnetic anomalies

A map of the total force magnetic field for an area that covered much of the Central Volcanic Region was first published by Gerard & Lawrie (1955). Their data were republished by Whiteford (1976) in the form of a total force anomaly field by the subtraction of a normal magnetic field from the original data. A total force magnetic anomaly map for an area that covers the central North Island has been constructed by Davey & Robinson (1978), and the part of their map that takes in the Central Volcanic Region is shown in fig. 2.12.

Preliminary interpretations of some magnetic anomalies in the vicinity of thermal areas were made by Modriniak & Studt (1959) while

Fig. 2.11 Interpretation of a dipole-dipole resistivity survey carried out by Dr H. Bibby (Geophysics Division) and reproduced by Midha (1979). The traverse crossed the eastern boundary of the Taupo Volcanic Zone near Broadlands (ie. approximately along B-B' of fig. 2.10).

Fig. 2.12 Total force magnetic anomalies for the central North Island (after Davey & Robinson, 1978). Data for the central square was obtained at a height of 1.5 km above sea level and elsewhere at a height of 3.05 km. Contour interval = 100 nT.

Malahoff (1968) modelled the more prominent magnetic anomalies, associated with the rhyolitic volcanic centres of the Taupo Volcanic Zone, in terms of three-dimensional prismatic source bodies. The models of Malahoff suggest that many of the anomalies cannot be explained by topographic effects alone and that some source bodies may have depth extents of 12-16 km.

From fig. 2.12 it can be observed that the magnetic anomaly field is relatively undisturbed away from the rhyolitic volcanic centres. Even over the Coromandel Peninsula, where large volumes of andesite and lesser amounts of diorite are exposed, the magnetic anomalies are not much different to those over the greywacke ranges to the west. The abnormally small magnetic anomalies associated with the volcanic and igneous rocks of the Coromandel area may be due to some, or all, of these rocks being reversely magnetised or the rocks may have lost a large component of their magnetisation as a result of hydrothermal alteration.

On a global scale it has been noted by Lawver & Hawkins (1978) that within the back-arc basins of the world, where spreading and associated magmatic activity is known to be occurring, that the observed magnetic anomalies are diffuse and small in amplitude. This is in contradistinction to the large and highly correlatable anomalies observed over mid-oceanic ridges where, like some back-arc basins, new lithosphere is being created. Possible explanations offered by Lawver & Hawkins for these differences are that within the back-arc basins there is a higher heat flux, thinner lithosphere, a thicker sedimentary cover, slower spreading rate, more prolonged hydrothermal alteration effects and a more diffuse pattern of magmatic intrusion. All these factors will contribute to a diminishment of magnetic anomalies. In particular, the slower cooling rate will have a strong effect on diminishing magnetic

anomalies in that it will encourage the growth of coarse grained magnetic minerals.

The magnetic method is usually a powerful method for discriminating between basement rocks of a volcanic or sedimentary affinity; the magnetisation of volcanic rocks being an order of magnitude or greater than that of sedimentary rocks. However, for the Central Volcanic Region where deep hydrothermal circulation both in the past as well as at present is inferred (Calhaem, 1973), the possibility exists that hydrothermal alteration has afflicted all the basement rocks of the Central Volcanic Region. As a consequence the magnetic method may not be an effective geophysical tool for discriminating between a sedimentary or volcanic basement for the Central Volcanic Region.

2.2.6 Heat flow

One of the most impressive geophysical features of the Taupo Volcanic Zone is its high heat output. The total natural heat output, almost all of which arrives at the surface via hydrothermal systems, has been estimated by Studt & Thompson (1969) to be 3.3×10^9 W. More recent studies give slightly higher values of $3.5\text{--}4.0 \times 10^9$ W (Allis, 1979) and $4.0\text{--}6.3 \times 10^9$ W (Hochstein, 1976).

The heat flow through the Taupo Volcanic Zone is thought to be predominantly convective (Studt & Thompson, 1969; Allis, 1979). Upflowing hot water with temperatures as high as 300°C is restricted to small areas of 20 km^2 or less called geothermal fields. The fields are separated by large recharge areas where down flowing meteoric water is interpreted to cause values of the temperature gradient to be zero (Studt & Thompson, 1969; Thompson, 1977). Studt & Thompson arrived at a figure for the average heat flow for the "volcanic belt" of

Fig. 2.13 Heat output from the geothermal areas of the Central Volcanic Region (after Allis, 1979). Areas in which the temperature of discharging water is greater than 98°C is shown in black; crosses indicate places where discharging water is less than 98°C . Bore holes with an observed zero geothermal gradient are as given by Studt & Thompson (1969) and Thompson (1977). The area bounded by straight lines is the proposed 5000 km^2 area beneath which convection is assumed to occur.

820 mW/m² by assuming 4000 km² to be the area of the belt. Allis (1979), however, considers an area of 5000 km² (fig. 2.13) to be associated with the convective heat flow, which when combined with his preferred heat output estimate of 3.5×10^9 W gives an average heat flow of 700 mW/m². This value pertains to the average heat flow for the south-eastern sector of the Central Volcanic Region shown in fig. 2.13. For the whole Region, which is an area of about 10,000 km², the heat output and therefore average heat flow are unknown.

Although the above value of 700 mW/m² is far greater than heat flow values reported from other back-arc basins (Anderson, 1975) such comparisons may not be valid. Estimates of the average heat flow from oceanic back-arc basins are notoriously difficult to obtain due to what is thought to be the effects of hydrothermal convection within the oceanic crust (Sclater, 1972). However, a volcanic spreading centre that is above sea level, and therefore has a readily available heat output estimate, is Iceland. Although Iceland represents a different plate tectonic setting to that of Central Volcanic Region, a comparison of the respective heat outputs is easy to make and is therefore useful.

Palmason & Saemundsson (1974) in a review paper of various geophysical and geological studies made on Iceland, propose that Iceland is a region of active spreading associated with the mid-Atlantic Ridge, and show that the active volcanism of Iceland, like the Central Volcanic Region, is concentrated in an elongate volcanic zone several tens of kilometres wide. Thus a comparison of the heat output per km of strike of volcanic zone will be made between Iceland and the Central Volcanic Region. The data from Iceland are:

Heat output from geothermal areas = 1.5×10^7 W/km

Heat output from volcanic extrusives = 2.0×10^7 W/km.

The heat output of volcanic extrusives erupted from the Central Volcanic Region is a little more difficult to estimate. However, taking just the rhyolitic volcanics and adopting a figure of $0.014 \text{ km}^3/\text{y}$ (Ewart *et al.*, 1975) as the rhyolite production rate for the last 10^6 years, an estimate can be made as follows: the heat available from the cooling of a volume (v) of molten rock (density = ρ) is given by Jaeger (1968) as:

$$\text{heat} = v\rho(\theta c + L) \quad \dots 2.1$$

where θ = temperature drop = $(800-20)^\circ\text{C}$ i.e. the temperature of molten rhyolite is taken as 800°C (Ewart *et al.*),

$$c = \text{specific heat} = 1.3 \times 10^3 \text{ J kg}^{-1} \text{ }^\circ\text{C}^{-1} \text{ (Lachenbruch *et al.*, 1976)}$$

$$L = \text{latent heat} = 2.7 \times 10^5 \text{ J/kg (Lachenbruch *et al.*, 1976)}$$

and the density of molten rhyolite is taken as $2.3 \times 10^3 \text{ kg/m}^3$.

Thus in a million years the heat content of the erupted rhyolites is $4.1 \times 10^{22} \text{ J}$. This is equivalent to an average heat output of $1.3 \times 10^9 \text{ W}$ or $0.8 \times 10^7 \text{ W/km}$ for 160 km strike of volcanic zone. The output from the geothermal fields is equivalent to $2.2 \times 10^7 \text{ W/km}$ and the total heat output for the Central Volcanic Region is therefore $3.0 \times 10^7 \text{ W/km}$; this is comparable to the total heat output estimate for Iceland of $3.5 \times 10^7 \text{ W/km}$.

The important point to note from the above analysis is that the average heat flow for the eastern sector of the Central Volcanic Region is many times greater than that for continental rifts (e.g. Pollack & Chapman, 1977; Zorin, 1981), yet the heat output of the Region is similar to that for an active spreading centre.

Geological evidence suggests that the heat discharge is a long-lived phenomenon; Grindley (1965) has estimated, for example, that the Wairakei geothermal field has been active for the past half million years. The problem of maintaining such activity has received much attention in

the literature (e.g. Banwell, 1963), but has so far defied a generally accepted solution. Rogan (1980) points out "such a high rate of heat transfer cannot be generated by conductive heat flow through the crust: it would require subcrustal temperatures of approximately 8000°C. Thus upward flux of magma is required to produce the surface heat flow." Rogan also performed some simple calculations, based on the numerical models of Lachenbruch *et al.* (1976), to show that putative rhyolitic magma chambers beneath some of the volcanic centres of the Taupo Volcanic Zone could supply, by conductive heat transfer, a heat flow not exceeding 200 mW/m² to these areas for a period of about 0.2 My. If convective heat transfer occurs the heat flow would be greater but at the expense of decreasing the life time of the system.

Hence, even though discrete bodies of molten rhyolite cooling within the upper crust of the Taupo Volcanic Zone could supply heat for about the required amount of time to support a geothermal system, the heat flow delivered from such bodies could not account for the observed heat flow. Clearly a much larger and pervasive heat source is required.

2.3 Scope and Outline of this Thesis

In the back-drop of the foregoing summary it is apparent that the gravity anomaly field of the Central Volcanic Region may contain the superposed gravity effects of:

- a) Rhyolitic volcanic centres and their associated deposits.
- b) The subducted Pacific plate.
- c) Crust-mantle perturbations arising from back-arc spreading.

Previous gravity work within the central North Island has been directed towards interpreting one of the first two geological structures listed above. However, there is little doubt that all three of the above are closely related. Therefore a complete gravity interpretation of the Central Volcanic Region must consider the gravity effects associated with back-arc spreading and the subducted plate as well as the effects of superficial, low-density volcanic rocks.

As the superficial low-density volcanic rocks will have a dominant gravity effect, emphasis is initially placed on establishing the thickness and velocity-density characteristics of these rocks. In chapter 3 a residual gravity anomaly field is derived for the Central Volcanic Region and a hypothesis made that the predominantly negative residuals may be explained by various thicknesses of low-density, volcanic rocks. This hypothesis is tested by analyses of bore-hole density data (chapter 4) and seismic refraction experiments (chapter 5). At two localities within the Central Volcanic Region it is found that the residual gravity effect of the seismically determined thickness of low-velocity rocks does not equal the observed residual gravity anomalies.

The work of chapter 6 is a slight digression from the central aim of this thesis. The spectral content of the aeromagnetic anomaly field is studied with a view to establishing the mean maximum depth extent of magnetic source bodies within the Taupo Volcanic Zone; this depth being interpreted to be that of the Curie Point isotherm ($\approx 500^{\circ}\text{C}$). Arguments are presented to show that the Curie depth may be used to gain an estimate for the elastic thickness of the lithosphere for the Central Volcanic Region.

In chapter 7 an attempt is made to reconcile the gravity and seismic interpretations, presented in chapter 5, by modelling the complete gravity anomaly field of the Central Volcanic Region with a simple, locally-compensated, three-dimensional model; this model is used as a basis for making profile interpretations across the Central Volcanic Region in chapter 8. Possible petrological models for the crust of the Region are then proposed and finally the study is concluded with a discussion on how the gravity and seismic work reported in this thesis bear upon previously proposed tectonic-evolutionary models for the Central Volcanic Region.

CHAPTER 3 GRAVITY DATA AND REGIONAL AND RESIDUAL GRAVITY ANOMALY FIELDS OF THE CENTRAL NORTH ISLAND

In this chapter the various sources of gravity data used in this study are described and methods of performing regional-residual separations discussed. A residual gravity anomaly map of the Central Volcanic Region is presented and the broad features of the map qualitatively analysed.

3.1 Gravity data of the central North Island

Systematic gravity surveys undertaken in New Zealand since 1949 have now secured a regional coverage with a station density of about 1 per 10 km². Since 1965 maps of the Bouguer, isostatic and isostatic first vertical derivative anomaly maps have been published at a scale of 1:250 000 (Reilly, 1972). Sheet 5 (Woodward & Ferry, 1973) and Sheet 8 (Woodward & Ferry, 1974) of this series cover most of the Central Volcanic Region where about 4000 gravity observations have been made. Although this set of observations constitutes the principal data base for this study the coverage was not uniform and it proved necessary to make further observations in three regions of interest; along and about the seismic lines at Mangakino and Otamarakau and along a profile from west of Tihoi to Taupo.

A total of 287 gravity stations were established in the following manner:

Gravity differences between field stations and various stations of the New Zealand Primary Network (Robertson & Reilly, 1960) were measured with a La Coste-Romberg gravimeter (no. G179). Repeat readings

of base stations allowed diurnal drift curves to be determined from which corrections were made to all observations. Station positions were estimated to an accuracy of 100 m from NZMS1 maps at a scale of 1:63 360 and elevations were determined with a pair of "Baromec" type aneroid barometers by measuring the difference in elevation from points of accurately known height. The effect of change in atmospheric pressure with time was corrected for by successive remeasurement at points of known height, but no corrections for horizontal pressure gradients were made. Terrain corrections for Hammer Zones B to D were made in the field and Zones E and F from topographic maps with a contour interval of 100 ft (30.4 m). Zones G to M were calculated by computer using the method of Woodward & Whitla (1980).

Reduction of the above raw data to Bouguer and isostatic anomalies was achieved by using the method of Woodward and Whitla thus giving anomalies compatible with those made by Geophysics Division. The Bouguer plate and the topographic corrections were made with a density of 2.67 Mg/m^3 and the isostatic anomalies are based on the Airy-Heiskanen model with a normal crustal thickness of 30 km and compensation density of -0.6 Mg/m^3 (Reilly, 1972). The gravity observations made for this study are held on the Geophysics Division's gravity files as:

Gravity map of New Zealand 1:250 000 Sheet 5:

Station numbers 2839 to 2930
 2996 to 3100
 3130 to 3157
 3199 to 3267

Gravity map of New Zealand 1:250 000 Sheet 8:

Station numbers 3278 to 3335

By far the greatest cause of error for any one gravity observation is the uncertainty inherent in the height determination. At the margins of the Central Volcanic Region where the land becomes mountainous errors arising from uncertainties in elevation of up to 50 $\mu\text{N/kg}$ are possible, although in general the total estimated uncertainty for any one gravity observation made within the Region is about $\pm 10 \mu\text{N/kg}$.

3.2 Regional-residual separation

3.2.1 Introduction

A gravity anomaly map commonly contains short-wavelength features which must be of shallow origin superimposed on broader anomaly trends which most likely, but not necessarily, have deep geological origins. The fundamental starting point for a gravity interpretation is often the separation of these effects. For many real earth situations this arbitrary subdivision can be troublesome and may even be fallacious; however by performing the separation the interpretation task is eased by allowing attention to be focussed on one feature at a time.

In New Zealand interpretation of gravity has previously been based on regionals obtained by interpolation of gravity observations made on "basement" rocks (Hatherton, 1952; Dibble & Suggate, 1956; Cowan & Hatherton, 1968). This method is satisfactory for investigations of superficial sedimentary basins of limited horizontal extent if one can make the assumption that the respective sources of the regional and residual gravity anomaly fields are well separated in depth.

The Central Volcanic Region is 100 km wide at its northern end, is a site of intense volcanic activity with active extension, and possibly

spreading (Chapter 2). These are not ideal conditions for a regional-residual gravity anomaly separation. For example there may be a continuum of density irregularities throughout the crust-mantle column of the Region which would effectively render a sharp wavelength distinction between regional and residual anomalies to be untenable. Nevertheless, because our knowledge of the relevant details pertaining to the deep structure of the Central Volcanic Region is incomplete at this stage, the regional-residual separation is a useful starting point for a gravity interpretation of the Region. A separation will be carried out and a hypothesis made that the resulting residual gravity anomalies represent solely the anomalous gravity effect of the low density volcanic rocks found within the near surface of the Central Volcanic Region. The corollary of this working hypothesis is that the regional field is due to density irregularities at depths large enough that they need not concern the interpretation of the residual gravity anomalies.

This hypothesis will be tested in this thesis by i) examining the variation of density with depth and ii), determining the thickness and variation of seismic velocity with depth of the volcanic rocks within the Central Volcanic Region. If a lack of consistency is found between the derived residual gravity anomalies, on one hand, and the calculated gravity effect of the seismically determined thickness of volcanic rocks, on the other, the above hypothesis will have to be rejected. A direct approach at interpreting all components of the gravity anomaly field of the Central Volcanic Region will then be required.

3.2.2 Regional and residual gravity anomalies of the central North Island

A previous Bouguer regional gravity field for the southeastern sector of the Central Volcanic Region has been presented by Modriniak & Studt (1959) (fig. 2.10). This field was constructed by manual interpolation, or "eyeballing", of greywacke-based Bouguer gravity anomaly values from one side of the volcanic region to the other. For the present study a quasi-objective method was developed whereby low-order, two-way polynomials were fitted by least squares to greywacke-based gravity observations that flank the Central Volcanic Region. The method and results of this procedure are described in Stern (1979) and a copy of this paper comprises appendix A of this thesis. Only the more important results will be discussed here.

3.2.3 Features of the residual gravity anomaly field

Regional polynomials were developed for both the Bouguer and isostatic gravity anomaly fields; residual anomalies, whether derived from either the Bouguer or the isostatic regional field were found to differ by less than 20 $\mu\text{N/kg}$.

A density of 2.67 Mg/m^3 was used for the Bouguer plate correction for all gravity anomalies, whether inside the volcanic region or out, so that the residual anomalies contain the gravity effect of rocks above sea level having densities differing from the assumed 2.67 Mg/m^3 . Fig. 3.1 shows the residual gravity anomaly field derived by subtracting the isostatic regional field from the observed isostatic anomalies. It should be noted that no account was taken for height differences of individual gravity observations when this contour map was drawn; the map is a two-dimensional representation of the residual gravity anomalies

Fig. 3.1 Residual gravity anomaly field of the Central Volcanic Region derived by subtracting a regional isostatic gravity anomaly field from the observed isostatic gravity anomalies. Contour interval = $50 \mu\text{N/kg}$. Dotted areas represent anomalies less than $-500 \mu\text{N/kg}$ and hachured areas are greater than $0 \mu\text{N/kg}$. Dots outline the approximate boundaries of the Taupo Volcanic Zone. Letters A-A' etc refer to profiles that will be discussed later.

observed on the ground surface which has a relief varying from 0 to 1000 m. The anomalies are predominantly negative which accords with the generally low densities of volcanic rocks found within the volcanic region. Moreover, by comparing fig. 3.1 with fig. 2.9 it can be noted that the majority of the most negative gravity anomalies (or gravity "lows") are associated with the rhyolitic volcanic centres of Taupo, Maroa and Okataina. However there is only a nebulous association of a gravity low with the Rotorua volcanic centre and the deepest gravity low, that found southwest of Tokoroa, is not associated with any known volcanic centre. This latter gravity low shall be referred to as the Mangakino gravity low.

Within the Central Volcanic Region several local gravity "highs" can be observed; at Otamarakau where greywacke outcrops, west of Te Puke where andesite outcrops and there is a large, circular gravity high near Tokoroa. Between Atiamuri and Broadlands (fig. 3.1) there is an elongate gravity high which corresponds with a southward plunging structural high. Modriniak & Studt (1959) called this feature the Paeroa Block and present geophysical evidence to suggest that the block is a fragment of greywacke that has subsided from the Kaingaroa Plateau to the east (fig. 3.1).

These short wavelength gravity highs are superimposed upon a broad region, correlating more or less with the boundaries of Central Volcanic Region, where the residual gravity anomalies average about $-400 \mu\text{N/kg}$. Thus, the long wavelength components of the residual gravity anomaly field tend to highlight the boundaries of the Central Volcanic Region rather than those of the Taupo Volcanic Zone.

To the east of Lake Taupo a broad residual gravity high over the Kaingaroa plateau can be noted. Drilling and seismic work (Hochstein &

Hunt, 1970) show that a thin veneer of ignimbrite overlying greywacke exists here. The residual gravity high may be an artefact of the regional polynomial being unable to accommodate the regional gravity curvature in this area or, alternatively, the anomaly may represent a high density mass distribution at a depth of about 10 km.

3.2.4 A comparison of published residual fields

No residual gravity field can be regarded as unique; for as Gupta & Ramani (1980) state "What is one mans regional is anothers residual depending on the nature of the information sought from the data." Three residual gravity anomaly fields have now been published for the Central Volcanic Region all of which essentially used a common data set and similar methods of derivation. Table 3.1 presents a comparison of these three residual fields. Several points should be noted.

Firstly, the residual fields derived separately in this study from Bouguer and isostatic regional fields are almost identical. This suggests that the isostatic correction field for the central North Island can readily be accommodated by a third-order, two-way polynomial. Secondly there are appreciable differences in the three residual field determinations contrasted in table 3.1. The residuals presented by Modriniak and Studt (1959) are more negative than those derived from this study. This difference would be even larger if the above authors had used a Bouguer reduction density of 2.67 Mg/m^3 , as used for this study, rather than a preferred density of 2.0 Mg/m^3 . Thus for say the Wairakei region, which is about 300 m above sea level, the residual anomaly of Modriniak & Studt if reduced with a Bouguer reduction density of 2.67 Mg/m^3 would be about $-480 \text{ } \mu\text{N/kg}$ instead of $-400 \text{ } \mu\text{N/kg}$; a value 130 $\mu\text{N/kg}$ lower than that derived here.

Regional		Modriniak & Studt(1959)	Stern(1979)		Rogan(1980)
Method		Manual interpolation of Bouguer anomaly field	Numerical interpolation [third degree, two-way polynomials of isostatic and Bouguer gravity anomaly fields]		Numerical interpolation [third degree, two-way polynomial of Bouguer gravity anomaly field]
Value of residual field at: [UN/kg]			Bouguer	isostatic	
Wairakei		- 400	- 345	- 350	- 250
Atiamuri		- 500	- 466	- 469	- 400
Ngakuru		- 550	- 524	- 516	- 450
Mangakino		not shown	- 649	- 657	- 500
Broadlands		- 410	- 299	- 295	- 200
Kawerau		- 300 *	- 221	- 232	- 100
Rotorua seismic line		not shown	+30 - -300	+5 - -290	+100 - -100
Otamarakau greywacke outcrop		"	+31	+ 5	+100
Whakatane Graben		- 420 *	- 440	- 446	- 250
Te Puke		not shown	- 140	- 152	0

★ Values from Studt(1958)

Table 3.1 A comparison of residual gravity anomaly values, derived from different studies, for various locations within the Central Volcanic Region. Values for Modriniak & Studt and Rogan were read from contour maps with an estimated reading accuracy of $\pm 50 \mu\text{N/kg}$.

On the other hand the residuals published by Rogan (1980) are generally more positive than those derived here (table 3.1). Hence, different residual gravity anomaly fields for the Central Volcanic Region, even when derived by similar means, may produce residual anomalies that at some locations can vary by a factor of two; this highlights some of the difficulties associated with determining a regional gravity field across a region as wide as the Central Volcanic Region.

3.3 Conclusions

The residual gravity anomaly map of the Central Volcanic Region shows a broad triangular region of negative residual anomalies which average about $-400 \mu\text{N/kg}$. This simple pattern is perturbed in places by deep gravity lows associated with some of the rhyolitic centres of the Taupo Volcanic Zone and by local gravity highs which in two cases can be correlated with outcropping rocks that are denser than the volcanic cover. However the fundamental point to be noted from fig. 3.1 is that it is the boundaries of the Central Volcanic Region that are emphasised by the shape of the residual gravity anomalies.

The general efficacy of a regional-residual separation for the central North Island gravity field has yet to be tested. The working hypothesis adopted at this stage of the investigation is that the residual gravity anomalies represent the gravity effect of varying thicknesses of low-density volcanic rocks. This hypothesis will first be tested by examining bore-hole density data from the Central Volcanic Region and, secondly by the use of seismic refraction techniques to determine the thickness and seismic velocity characteristics of the volcanic rocks.

CHAPTER 4

ROCK PROPERTIES

Summary:

An analysis of bore-hole density data is the main topic of this chapter. The variation of density with depth is examined with a view to establishing a generalised density versus depth relation for the superficial volcanic rocks of the Central Volcanic Region. Since most of the data come from geothermal areas where localised hydrothermal densification is known to occur, close attention is paid to whether or not the data can be taken as being representative of the volcanic rocks found outside of the geothermal areas.

Complementing the density versus depth data are seismic velocity versus density and seismic velocity versus depth data. Three empirical relations are developed between density (ρ), seismic velocity (V_p) and depth (Z). The sources of data used in obtaining these relations are as follows:

- ρ vs. Z - bore-hole data from Wairakei
- V_p vs. ρ - laboratory work on bore-hole samples
- V_p vs. Z - seismic refraction data from the Taupo-Ohakuri-Tihoi region.

The above three functions, which are found to be mutually consistent, lead to the following conclusions:

- i) Lateral variations of density within the top 1 km of volcanic rocks can account for gravity anomaly variations of up to 100 $\mu\text{N/kg}$ in amplitude and 10 km in wavelength; these lateral variations of density are not necessarily due to hydrothermal densification effects.

- ii) Extrapolation of the density with depth trends associated with data from the Kawerau and Wairakei bore-holes suggest that the densities of volcanic rocks approach values as high as 2.60 Mg/m^3 at depths of 2-3 km; this density is consistent with the seismic velocities of 5.0-5.5 km/s detected in the same depth range beneath the Taupo-Ohakuri-Tihoi region.

4.1 Introduction

Past gravity interpretations within the Central Volcanic Region have adopted -0.5 Mg/m^3 as the density contrast between low-density volcanics and an assumed greywacke basement (e.g. Modriniak & Studt, 1959; Rogan, 1980). With this constant density contrast, and in the absence of any seismic control, residual gravity interpretations have afforded a simplified structural model of the Central Volcanic Region. Modriniak & Studt, for example, interpreted a $-550 \text{ } \mu\text{N/kg}$ residual gravity low south of Rotorua in terms of a 4 km deep basement depression infilled with low-density volcanics (fig. 2.10).

The -0.5 Mg/m^3 density contrast was chosen by Modriniak & Studt on the basis of densities obtained from extensive surface sampling and a few drill-cores. Extrapolation of the density contrast that is applicable to near surface volcanic rocks, to depths of 4 km is clearly a gross approximation. Modriniak & Studt acknowledged this and accordingly added the proviso that if there is a decreasing density contrast with depth their interpreted depths must be regarded as minima.

Of fundamental importance when interpreting residual gravity anomalies associated with high-porosity, and hence low-density, rocks

is how the density varies with depth, rather than the mean density or the surface densities of the rocks. If the density contrast of the volcanic rocks decreases with depth, then the resolution of depth to basement, from the residual gravity anomalies, also decreases with depth. Indeed the very definition of "basement" becomes obscure in this situation.

The terms "basement" and "cover" are often used in a geological sense to distinguish two rock units of different age and sometimes different lithology. It is not at all certain that these terms are appropriate to the Central Volcanic Region where it is unknown if a rock unit of differing lithology or age underlies the acid volcanics that are observed within the near surface. However the residual gravity anomalies derived in chapter 3 purport to represent the anomalous gravity effect of rocks within the Central Volcanic Region whose densities are different to that of rocks, at an equivalent depth, that flank the Region (i.e. greywacke). Hence if a rock unit exists at some depth within the Region, regardless of its age or lithology, whose density is much the same as greywacke it will have no residual gravity effect. Therefore the term "basement" will be reserved for rocks whose densities, or seismic velocities, are similar to that of greywacke; the term "volcanic cover" will be used to describe volcanic rocks that lie above the basement rocks and whose densities are less than that of greywacke. No lithological or chronological assumptions are attached to this definition of basement. Furthermore, unless otherwise stated in this thesis, the terms density and velocity refer to wet(or saturated) density and compressional seismic velocity respectively.

4.2 Density data

4.2.1 Surface density determinations

Table 4.1 lists the densities for several rock types found at the surface throughout and adjacent to the Central Volcanic Region. Most of the data were abstracted from Whiteford & Lumb (1975) although additional density measurements were carried out on five samples of granodiorite boulders believed to be xenoliths ejected from Mt Tarawera. Ewart & Cole (1967) give textural and mineralogical descriptions of these rocks and conclude that they represent plutonic equivalents of the acid volcanic rocks found at or near the surface. These granodiorites are therefore of special interest as they possibly represent a deeper rock suite of the Central Volcanic Region that is nowhere exposed at the surface. The average wet and particle densities of the granodiorites were found to be 2.50 and 2.59 Mg/m³ respectively.

4.2.2 Bore-hole density data

Drilling for geothermal resources over the past thirty years has provided a wealth of density versus depth data from the Central Volcanic Region. Most of the density determinations were carried out by Mr A. E. Leopard (Geophysics Division, D.S.I.R.) and the data can be found tabulated either in Whiteford & Lumb (1975) or in files held at Geophysics Division.

Fig. 4.1 shows the density versus depth data and the corresponding cumulative gravity anomaly for bore 121 - the deepest hole at Wairakei. The cumulative gravity anomaly calculation assumes an infinite Bouguer plate configuration, a basement density of 2.65 Mg/m³, and is defined by:

Table 4.1 Statistics of rock densities for surface samples found within the Central Volcanic Region

Lithology	Age	Location	ρ (Mg/m ³)	Std. dev. (Mg/m ³)	No.
Rotoiti-breccia	42 000 yr	Bay of Plenty	1.91	0.18	14
Haparangi rhyolite	less than 1 My	T.V.Z.	2.23	0.21	71
Volcanic tuff	" " "	"	1.81	0.22	37
Whakamaru ignimbrite	0.33 My	Whakamaru	2.28	0.15	66
Rangitaiki ignimbrite	Quaternary	Kaingaroa Plateau	2.35	0.12	19
Besson Is. andesites	Tertiary	Coromandel	2.63	0.11	59
T.V.Z. andesites	Quaternary	Tongariro Volcanic Centre	2.53	0.17	69
Basalt	Quaternary	T.V.Z.	2.66	0.21	25
Greywacke	Mesozoic	Bordering the C.V.R.	2.65	0.06	52
Granodiorite boulders	?	Tarawera	2.50	0.03	5

T.V.Z. = Taupo Volcanic Zone

C.V.R. = Central Volcanic Region

ρ = mean density

Fig. 4.1 Density versus depth data from Wairakei bores 121 and 71. The associated cumulative residual gravity anomaly is shown at the top. A basement density of 2.65 Mg/m^3 is assumed.

$$\Delta g = 2\pi G \sum_{i=1}^N (h_i - h_{i-1}) (\rho_i - 2.65) \quad 4.1$$

where G = gravitational constant, (ρ_i, h_i) = density and depth of the i th sample, N = total number of samples and $h_0 = 0$. The density log from bore 121 does not include any samples from the top 600 m, so densities from nearby bore 71 were used for this upper interval. A marked increase of density with depth can be observed within the first 500 m, where the lithology is tuff and sandstone, whereas between 500 and 2000 m the increase of density with depth is only slight. The cumulative gravity anomaly for a depth of 2.2 km is about $-310 \mu\text{N/kg}$.

Fig. 4.2 shows the density versus depth data and cumulative gravity anomaly curve for bore 227 (see fig. 4.5 for the locations of bores 227, 121 and 71). Although bore 227 is only 5 km distant from bore 121 the densities are appreciably lower than in bore 121; indeed the densities of bore 227 were found to be the lowest of any bore-holes within the Region. No lithologies were given for bore 227 although from the cross-sections of Grindley (1965) it can be inferred that the lithology of rocks in this hole is pumice breccia of the Wairoa formation. It can be seen in fig. 4.2 that although the hole only goes to a depth of 1100 metres the cumulative gravity anomaly for this depth is $-380 \mu\text{N/kg}$. By way of comparison the cumulative gravity anomaly for bore 121 and 71 is $-250 \mu\text{N/kg}$ at a depth of 1100 m. Thus from an examination of density data from just two bore-holes it is evident that large lateral variations within the top 1 km of volcanic cover may occur. Of further note is that despite the considerable scatter in the data of fig. 4.2 there is a clear tendency for the density to increase with depth.

Fig. 4.2 Density versus depth and cumulative residual gravity anomaly for bore 227 (see fig. 4.5 for the location of the bore hole).

Fig. 4.3 shows 1011 bore-hole density determinations, plotted against depth, from Kawerau, Broadlands, Orakeikorako, Wairakei and Whakamaru. The first four localities are sites of geothermal exploration while at Whakamaru several shallow bore-holes were put down for dam investigation work. At Wairakei many holes have been put down over a wide area so preference was given to density data from those holes not directly associated with the geothermal production area. A further constraint on the data set was that only bore-holes accompanied by lithological logs were used.

Despite the large scatter for the data of Fig. 4.3 a wedge-shaped taper to higher densities with depth can be seen. The question of whether the density-depth pattern in fig. 4.3 is typical of the Central Volcanic Region in general is important, yet not easily answered. A possible bias of this data set by hydrothermal densification must first be considered.

4.2.3 Hydrothermal densification effects

Intensive gravity, seismic and magnetic investigations at Broadlands led Hochstein & Hunt (1970) to the conclusion that 80 $\mu\text{N/kg}$ of an observed 100 $\mu\text{N/kg}$ secondary residual gravity high could be attributed to effects of hydrothermal densification within the volcanic cover. Examination of cores revealed that high-rank alteration densified some of the originally porous lithologies by as much as 0.4 Mg/m^3 . Densification is thought to affect only the more porous lithologies, however. For example Cheng (1976) found that whereas densities of sandstone within the Tatun (Taiwan) geothermal field increased with hydrothermal alteration, andesite densities decreased.

Fig. 4.3 1011 bore hole density determinations from the Central Volcanic Region plotted against depth.

Rather than dismiss outright the large density data set of fig. 4.3, it was decided to examine the gravity fields around the geothermal fields, other than Broadlands, for secondary gravity highs.

4.2.4 Secondary gravity highs

Closely-spaced gravity observations have been made about the Kawerau and Wairakei geothermal fields (see Woodward & Ferry 1973, 1974). These data were used to numerically calculate 1.2 km grids of weighted residual gravity observations which were then contoured with a bicubic spline computer programme, as described by Brownrigg (1978). Contour maps of the Kawerau and Wairakei residual gravity fields are shown in figs 4.4 and 4.5 respectively.

Kawerau field: A secondary residual gravity high of approximate amplitude 20-30 $\mu\text{N/kg}$ was first noted by Studt (1958) to be associated with the Kawerau geothermal field. This secondary high can be seen in fig. 4.4. Taking the radius of the field to be 1.5 km (fig. 4.4), the volume of anomalous mass can be modelled as a vertical cylinder 1 km deep, radius 1.5 km and with a density contrast of about 0.1 Mg/m^3 (Telford *et al.*, 1976; equation 2.43e). Thus the bore-hole density data from Kawerau may contain an upward density bias from hydrothermal densification of 0.1 Mg/m^3 .

Wairakei field: Beck & Robertson (1955) first observed a local gravity high, amplitude 60-80 $\mu\text{N/kg}$, over the Wairakei region. They attributed it to a greywacke basement high. Modriniak & Studt's (fig. 2.10, chp. 2) interpretation profile across the Wairakei gravity high showed an assumed greywacke basement locally rising to a depth of 1.6 km beneath Wairakei. Grindley (1965) also subscribed to the view of

Fig. 4.4 Computer contour plot of a 1.3 km grid of weighted residual gravity anomaly values about the Kawerau area. The distribution of gravity stations used to derive the grid is given by Woodward & Ferry (1973).

★ = Borehole
 --- = Geothermal field boundary

0 5 km

contour interval = 10 UN/kg

Fig. 4.5 Computer contour plot for the Wairakei area derived in a manner similar to fig. 4.4. Positions of the bore holes used in the Wairakei density-depth analysis are shown. Insert map at top right is from Modriniak & Studt (1959) and highlights the association of the Wairakei gravity high and the Target Magnetic Anomaly (contour interval = 50 nT).

a Wairakei basement high citing stratigraphic uplift of the superficial subsurface layers as corroborating evidence for the basement uplift. However, subsequent drilling at Wairakei failed to find greywacke basement despite the deepest hole, bore-121 (fig. 4.1), terminating in andesite at a depth of 2.2 km (Smith, 1970).

Fig. 4.5 depicts the Wairakei high as a north-south ridge showing little if any correlation to the boundaries of the geothermal field. Yet there is a strongly-suggestive genetic relationship between the southern portion of the gravity high and the Target Magnetic Anomaly of Modriniak & Studt (1959). If both the Target Magnetic Anomaly and the Wairakei gravity high share a common source it is difficult to envisage this source as due either to a greywacke basement high (as greywacke is non-magnetic) or to hydrothermal alteration within the volcanic cover (as hydrothermal activity is thought to destroy magnetite (Studt, 1959)). However, it is possible to test the supposition that the source of the gravity high is due to density differences within the top 1 km of volcanic cover by making the following test.

Densities for the top 1 km of bores 71 and 121 to the west of the Waikato River were compared against densities within the top 1 km of bores 227, 225 and Th2 to the east of the river. Mean densities within the first and second 500 m of these two columns were averaged as depicted in fig. 4.6. By approximating the top 1 km of these two columns to infinite Bouguer plates it can be seen that the gravity effect of the western column is $90 \mu\text{N/kg}$ greater than that of the eastern column. The difference is due to the presence of a thin lens of andesite and a thicker sequence of Wairakei ignimbrite ($\bar{\rho} = 2.36 \text{ Mg/m}^3$) in the second 500 m of the western column (fig. 4.6) as opposed to Wairoa formation (pumice breccia) of mean density 2.01 Mg/m^3 in the second 500 m of the eastern column.

Fig. 4.6 A comparison of densities from Wairakei bore holes on either side of the Waikato River. Positions of the bore holes are shown in fig. 4.5, $\bar{d}$ refers to the mean density (Mg/m^3) of the indicated 500 m thick section and n is the number of density determinations used to estimate $\bar{d}$. The gravity difference calculation was made assuming an infinite Bouguer plate configuration.

Therefore the Wairakei gravity high may be accounted for, in total, by a lateral variation of volcanic rock type. Moreover this lateral variation of density within the volcanic cover is not necessarily due to hydrothermal densification. The mean density of Wairakei ignimbrite within the depth range 500-1000 m of bore 121 is 2.36 Mg/m^3 ; a value similar to that of the unaltered Rangitaiki ignimbrite ($\bar{\rho} = 2.35 \text{ Mg/m}^3$, table 4.1) that is found outcropping on the Kaingaroa plateau to the east, and which Martin (1961) considers as being a correlative of the Wairakei ignimbrite.

The evidence for hydrothermal densification, as outlined by Hochstein & Hunt (1970), at Broadlands is clear. At Kawerau there is possibly an upward bias of 0.1 Mg/m^3 from hydrothermal densification while at Wairakei, if hydrothermal densification is present, it is not clearly manifested in the residual gravity anomalies.

4.2.5 Statistical analysis of the density data

The data set shown in fig. 4.3 was sorted according to lithology into groups of pumice and tuff, breccia, sandstone and siltstone, ignimbrite, rhyolite, dacite and andesite. As different geologists have logged the various cores there may be inconsistencies between classifications.

Table 4.2 gives the results of four one-way analysis of variance runs between depth, density, location and lithology from which the following are noted.

- (1) Mean densities of rhyolites and ignimbrites from bore-holes are not appreciably different from those of surface samples as shown in table 4.1. In contrast, bore-hole pumice-tuff and breccias are more dense than their equivalent surface samples.

Table 4.2 Analysis of variance tables : Borehole data

a) Depth and density against location

Location	Mean depth (m)	S.D. (m)	Mean density (Mg/m ³)	S.D. (Mg/m ³)	n
Kawerau	629	290	2.31	0.22	173
Orakeikorako	553	334	2.17	0.29	106
Wairakei	654	404	2.12	0.22	322
Broadlands	633	376	2.21	0.26	344
Whakamaru	56	46	2.28	0.15	66

b) Density and depth versus lithology

Lithology	Mean depth (m)	S.D. (m)	Mean density (Mg/m ³)	S.D. (Mg/m ³)	n
Sandstone	377	257	1.95	0.21	85
Pumice-tuff	575	412	2.07	0.25	227
Breccia	701	361	2.15	0.21	216
Rhyolite	476	246	2.27	0.18	238
Ignimbrite	642	480	2.33	0.15	175
Dacite	581	324	2.34	0.17	24
Andesite	800	250	2.50	0.14	46

S.D. = standard deviation

n = number of samples used in the analysis

Mean density and depth of whole data set equals 2.20 Mg/m³ and 593 metres respectively.

Fig. 4.7 A histogram plot illustrating the higher percentage of andesite and rhyolite within the Kawerau bore holes than elsewhere. $\bar{d}$ refers to the mean density in Mg/m^3 of the respective lithologies.

- (2) Andesite is found at a greater mean depth than any of the other lithologies and represents about 5% of all the bore hole samples. This is slightly different to the volume estimates of Cole (1979), based principally on surface exposures, which gives andesite as making up 2.1% of all the volcanics within the Taupo Volcanic Zone.
- (3) Densities at Kawerau are appreciably higher than for the other localities. The difference arises from a greater proportion of andesite and rhyolite present in the Kawerau volcanic cover than at other localities. Fig. 4.7 shows a histogram plot illustrating this point.
- (4) The mean density and depth for the whole data set is 2.20 Mg/m^3 and 593 m respectively. Thus, by assuming a basement density of 2.65 Mg/m^3 one can state that the mean density contrast at a depth of about 600 m is -0.45 Mg/m^3 .

The 1011 data points of fig. 4.3 were further classified according to location and replotted in fig. 4.8. Within each plot a broad distinction is made between "pyroclastic" and "lava" type lithologies i.e. pumice, tuff, breccia; sandstone and ignimbrite take the former and rhyolite, dacite and andesite the latter classification.

A comparison of the four plots in fig. 4.8 reveals a greater scatter of data on the Broadlands plot than on the other three; this is possibly the result of the more intense hydrothermal densification at Broadlands than elsewhere. Accordingly density distributions for each lithology were recalculated, less the data from Broadlands, and are shown in histogram form by fig. 4.9. Comparison of fig. 4.9 with the data in table 4.2 shows that elimination of the Broadlands data results in a reduction of the mean densities of pumice-tuff and breccias yet little change in the other lithologies.

Fig. 4.8 Density versus depth data from four localities within the Central Volcanic Region. The lithological distinction between "lavas" and "pyroclastic" was made as described in the text. Densities in Mg/m³ and depth in metres.

Fig. 4.9 Histogram plot showing the density statistics of all bore hole data, less that from Broadlands.
 n = number of density determinations, $\bar{d}$ = mean density,
 $s.d.$ = standard deviation.

4.2.6 Numerical density - depth functions

The data and analyses of the previous sections have shown that lateral variations in density that occur within the volcanic cover may cause gravity anomalies up to 100 $\mu\text{N/kg}$ in amplitude and 10 km in wavelength. These lateral variations may be due to hydrothermal densification or just a lateral variation in unaltered rock types within the volcanic cover. Thus the density with depth behaviour of the volcanic cover is dependent on the order of stratigraphic layering as well as compaction due to geostatic pressure. Despite this it is useful to attempt to develop a generalised density-depth function from the bore-hole density data which can then be used as a standard of comparison for density-depth functions derived subsequently by other means.

The Wairakei and Kawerau data sets were chosen for the purpose of developing numerical density-depth functions. The mathematical form of the function used to simulate the density-depth behaviour is the exponentially decreasing density contrast function described by Cordell (1973):

$$\Delta\rho = \Delta\rho_0 \exp(-bz) \quad 4.2$$

where $\Delta\rho$ = density contrast, $\Delta\rho_0$ = density contrast at $z=0$ and z is depth increasing downwards. With this form of function a basement density must be assumed which will also be the limiting density that the volcanic cover is constrained to approach at large depths. Because the residual gravity anomalies derived in chapter 3 represent the gravity effect of rocks whose densities differ from that of greywacke, a basement density of 2.65 Mg/m^3 is adopted; this being the mean density of greywacke rocks found on either side of the Central Volcanic Region (table 4.1). In order to keep the model simple it is assumed that this

density is applicable to the first few kilometres of depth for the greywacke crust that flanks the Region. Finally, note that, as discussed in the introduction to this chapter, linking the basement density to that of greywacke does not necessarily imply that the basement rocks of the Central Volcanic Region are in fact greywacke.

A density-contrast function of the form given in equation 4.2 has several advantages over the previously adopted function $\Delta\rho = \text{constant}$. Firstly, a visual inspection of the data in figs 4.3 and 4.8 suggests that it is more realistic and secondly a few simple manipulations (Cordell, 1973) of equation 4.2 leads to the following results:

a) For an infinite Bouguer plate of thickness h with a density contrast given by equation 4.2 the gravity effect will be

$$g_{\text{slab}} = (2\pi G \Delta\rho_0 / b) (1 - \exp(-hb)) \quad 4.3$$

where G = gravitational constant.

b) The maximum possible gravity effect from an infinitely thick slab with a density contrast given by equation 4.2 is

$$g_{\text{max}} = 2\pi G \Delta\rho_0 / b \quad 4.4$$

The density versus depth data from Wairakei and Kawerau (fig. 4.8) were fitted to equation 4.2 and the respective values of $\Delta\rho_0$ and the exponential coefficient b were determined by least squares; the results are given in table 4.3. Fig. 4.10 shows the mean and standard deviation of successive groups of 10 density values plotted against depth for the Wairakei and Kawerau data; superimposed on these plots are the derived density contrast-depth functions.

Table 4.3 Results of fitting the Wairakei and Kawerau data to the function $\Delta\rho = \Delta\rho_0 \exp(-bz)$ where $\Delta\rho$ = density contrast, $\Delta\rho_0$ = density contrast at $z = 0$. A basement density of 2.65 Mg/m is assumed.

	Wairakei	Kawerau
Parameter values: $b(\text{km}^{-1})$	0.83	0.87
$\Delta\rho_0$ (Mg/m^3)	0.84	0.56
Std.error (Mg/m^3)	0.15	0.17
N (samples)	293	164
Gravity effect of 2 km thick slab ($\mu\text{N/kg}$)	-345	-222
Maximum obtainable gravity effect ($\mu\text{N/kg}$)	-420	-270
Density predicted at a depth of 2 km (Mg/m^3)	2.49	2.55

Fig. 4.10 Plot of mean and standard deviations for successive (with depth) groups of 10 density-depth pairs from Wairakei and Kawerau. Dashed lines are the density-depth curves given in table 4.3 and the density contrast is with respect to a basement density of 2.65 Mg/m^3 .

4.2.7 Implications of decreasing density contrast with depth for residual gravity interpretations

From fig. 4.10 it can be appreciated that the scatter of data is quite large. The standard error for the Wairakei fitting function (table 4.3) is 0.15 Mg/m^3 so that the predicted $-340 \text{ } \mu\text{N/kg}$ gravity effect of a 2 km slab of volcanic cover at Wairakei (table 4.3) should be accompanied by an uncertainty of about $\pm 130 \text{ } \mu\text{N/kg}$. However, it can also be observed from fig. 4.10 that the scatter appears to diminish with depth. Thus the predicted density of 2.49 Mg/m^3 for the volcanic cover at a depth of 2 km beneath Wairakei (table 4.3) would have an associated uncertainty of about $\pm 0.1 \text{ Mg/m}^3$.

Also given in table 4.3 are the maximum obtainable residual gravity anomalies of -270 and $-420 \text{ } \mu\text{N/kg}$ calculated from the Kawerau and Wairakei functions respectively. This calculation only has meaning if there are no other major density inversions deeper than the base of the deepest bore hole for each area, and that the density of the basement rocks found each side of the Central Volcanic Region is not increasing appreciably with depth. Without further data both of these assumptions remain open to question.

Adoption of a constant density contrast permits any magnitude of residual gravity anomaly to be explained merely by increasing the depth of the volcanic cover whereas use of the exponential density contrast function imposes an upper limit to the magnitude of residual gravity anomaly that can be attributed to the volcanic cover alone. Moreover a decreasing density contrast with depth implies a loss of basement resolution as the volcanic cover becomes thicker. Fig. 4.11 illustrates these points. Here two hypothetical models of negative mass

Fig. 4.11 Calculated gravity effects for two hypothetical models that demonstrate the difference between a constant density contrast function and an exponentially decreasing density contrast function.

anomalies are shown along with their associated gravity effects calculated for both a constant -0.4 Mg/m^3 density contrast and the Wairakei exponential density-contrast function. For model 1, which is 2 km thick, the respective gravity effects calculated with either density contrast function are similar. Thus a density contrast of -0.4 Mg/m^3 appears to be a reasonable approximation to the Wairakei density data when interpreting a 2 km thick section of volcanic cover. For model 2, which has a maximum thickness of 4 km, the respective calculated gravity effects deviate, particularly in the middle of the profile, indicating that the -0.4 Mg/m^3 density contrast is not a good approximation to the Wairakei density data when dealing with volcanic cover sequences 4 km thick.

Of further note in fig. 4.11 is that the calculated gravity effects with the exponential density contrast function for model 1 and 2 are similar despite the gross geometrical differences between the two models. This underlines that in reality gravity interpretations will be rather insensitive to the resolution of basement topography, and in particular, attempts to determine the thickness of volcanic cover within the Central Volcanic Region from the interpretation of residual gravity anomalies alone will be beset with uncertainty.

The discussion set out above once again raises the question of the meaning for the terms "cover" and "basement" as applied to the upper crustal rocks of the Central Volcanic Region. On the basis of the data and interpretations presented in this chapter, the physical boundary between basement and cover for the Central Volcanic Region is gradational, occurs somewhere between 2 and 3 km and there is no evidence as yet to suggest the boundary is a lithological one.

4.2.8 Synopsis of the density versus depth data

Three principal points emerge from this analysis:

- 1) Lateral density contrasts exist within the top 1 km of volcanic cover which may give rise to local gravity anomalies of wavelength 10 km and amplitude 100 $\mu\text{N/kg}$. These lateral variations may not necessarily be due to hydrothermal densification but merely differences in the density of, say, unaltered breccia to unaltered ignimbrite. This was the point made in the analysis of the Wairakei gravity high (section 4.2.4).
- 2) Hydrothermal densification that is manifested in density and hence gravity anomaly data from Broadlands, and to a lesser extent Kawerau, does not appear to be as prominent in the data from Wairakei.
- 3) The increase of density with depth for the volcanic rocks of the Central Volcanic Region is such that densities approaching 2.60 Mg/m^3 are predicted to occur in the depth range of 2-3 km.

The question of whether the density with depth data presented here is applicable to those parts of the volcanic region outside of the geothermal fields is still unanswered. At Kawerau it was shown that the anomalously high densities, with respect to say those at Wairakei, can in part be explained by hydrothermal densification and in part by greater amounts of rhyolite and andesite within the volcanic cover. At Wairakei the local gravity high appears to be associated with a denser volcanic unit within the volcanic cover, and in consequence the effects of hydrothermal densification at Wairakei are difficult to assess from the gravity data alone. An additional problem that besets analyses of bore-hole samples is that of sampling bias. It could be that there was an inherent bias to select the more competent, hence more dense, samples

of rock when the cores were initially recovered. Given these uncertainties it would be desirable to have some means of checking whether or not the Wairakei density versus depth data is applicable to regions outside of the geothermal fields.

At Broadlands, seismic velocities of 1.7 to 4.0 km/s have been associated with formations whose densities range from 1.8 to 2.4 Mg/m³ respectively (Hochstein & Hunt, 1970). Furthermore, these authors found that just as densities increased with rank of hydrothermal alteration so did seismic velocities. Seismic refraction techniques, therefore, offer a potentially promising method of checking the Wairakei density versus depth relationship. The quantitative link between gravity and seismic refraction interpretations is an empirical relationship between velocity and density. This is the topic to be discussed next.

4.3 Seismic velocity versus density relationships

4.3.1 Introduction

Combining gravity and seismic refraction techniques for a unified crustal structure interpretation is a strategy that has effectively been employed in the past (e.g. Woollard, 1959). The ultimate effectiveness of such a combination rests on the knowledge of an empirically-derived relationship between density and seismic velocity.

Commonly quoted is the velocity-density curve of Nafe-Drake (in Grant & West, 1965) based on about 500, *in situ* measurements of water saturated, sedimentary rocks. A functional relationship that closely approximates the Nafe-Drake curve, for sedimentary rocks, is given by Gardner *et al.* (1974) as $\rho = .23V_p^{.25}$. Birch (1961) presented

a formal analysis of the velocity-density relationship by proposing an equation of the form $\rho = A(\bar{m}) + BV_p$, where $\bar{m}$ is the mean atomic weight of the lithology under consideration. His work was based on laboratory determinations of velocity at 100 MPa to ensure that porosity had essentially vanished. Since Birch's work various workers in this field have advanced modifications to his equation. For example, Simmons (1964) introduced a correction factor for the CaO content of the rock type under investigation.

Rather than arbitrarily adopt one of the above relationships it was decided to investigate the velocity-density relationship of rock samples taken from within the Central Volcanic Region itself.

4.3.2 Laboratory work

Rock samples were cut into rectangular or cylindrical shapes with parallel faces and saturated for 24 hr under reduced pressure, following which their ultrasonic velocities were measured with a "Pundit" device. Details of the experimental work are given and the results listed in Appendix B. Typically, velocity-density relationships for surface samples measured at atmospheric pressure display a wide scatter (e.g. Golizdra, 1978a). In an effort to subdue this scatter a large percentage of samples used for this study were from boreholes. The assumption here being that relaxation of the geostatic compaction effect, since withdrawal from the subsurface, will be insignificant.

Velocities determined by the ultrasonic pulse technique should be treated with caution when compared to field determined velocities. Laboratory-determined velocities can generally be less confidently related to field measurements than density or porosity because

compressional velocity is determined by factors other than lithology; i.e. other than mineralogy and texture. It can be affected by fractures, fabric, cementation and pressure (Press, 1966). Moreover, Geertsma (1961) showed theoretically that ultrasonic velocities are somewhat frequency-dependent for water-saturated rocks with moderate porosities (13 to 29% at 82 kHz). However, the predicted differences between sonic and ultrasonic data are not large (less than 5%) and no attempt has been made to take the effect of frequency dependence into account for this study.

Relating laboratory measurements to field determinations of seismic velocity is further complicated by a sampling bias. Only competent looking rocks were selected as the more friable specimens would not stand up to cutting and polishing. This bias should, however, affect both density and velocity in the same direction. For example, two hydrothermally altered and densified tuffs from borehole BR2, at Broadlands, with densities 2.48 and 2.52 Mg/m³ gave respective ultrasonic velocities of 4.74 and 4.88 km/s.

4.3.3 Data and conclusions

Fig .4.12 shows density-velocity data measured for this study. 112 samples were measured including a few greywackes, diorites and andesites from areas bordering the Central Volcanic Region. Superimposed on fig. 4.12 is the Nafe-Drake curve and a least squares linear fit to the data less the greywacke samples. For the linear fit a correlation coefficient of 0.65 was obtained. That is, 65% of the variation of density may be accounted for by its correlation with velocity. The standard deviation of residuals to this fit is 0.16 Mg/m³. The similarity of this relationship to those of Nafe-Drake (see appendix B) and Golizdra (1978a)

Fig. 4.12 Seismic velocity versus density data for rock samples from within and bordering the Central Volcanic Region. Field determinations are taken from Hochstein & Hunt (1970) and Studt (1958). The least squares fit was made on the displayed data, less the greywacke samples (see appendix B).

is close suggesting a fundamental velocity-density relationship for moderate porosity sedimentary rocks from either water-derived or volcanic affinities.

Three further points to note are: 1) The group of andesites, diorites and greywackes (all surface samples) fall beneath the Nafe-Drake curve in fig. 4.12. This is rationalised by considering the experimental work of Birch (1960) which indicates that surface rocks under the first 100 MPa of applied pressure would show an approximate 10% increase in compressional velocity with only a minor concomitant change in density. 2) A closer examination of the andesite-diorite-greywacke grouping of fig. 4.12 reveals a random intermingling of these three rock types. Lithological distinctions, therefore, made on the basis of "basement" seismic velocities would be hazardous. 3) The Nafe-Drake curve offers a reasonable approximation to the borehole data and will therefore be adopted for this study. A conversion table based on this curve is given in appendix B.

The analysis of rock property data has reached the point of providing two relations between three parameters: i.e. density versus depth (borehole data) and velocity versus density (laboratory measurements). To complete the analysis a third relationship between velocity and depth is required.

4.4 Velocity versus depth functions

Robinson *et al.* (1981) carried out a reconnaissance refraction survey between lakes Taupo and Ohakuri. Two shots were fired in Lake Taupo and one in Lake Ohakuri with a seismograph layout as shown in fig. 4.13. The time-distance plot, irrespective of path azimuth, defines two clear

Fig. 4.13 Diagram showing the location, results and interpretation of the Taupo refraction survey as presented by Robinson *et al.* (1981). All arrivals from both shot points are projected onto one branch of the time-distance graph and crosses represent second arrivals.

lines whose gradients suggest the presence of subsurface refractors with velocities of 3.20 and 5.45 km/s. The simplest interpretation of the data is that an approximately horizontal 5.45 km/s refractor underlies the area shown in fig. 4.13 at a depth of 2 km. The crucial point to be noted is that the time-delay due to the volcanic cover for stations near Wairakei and those to the west are identical. Residual gravity anomalies for these two areas differ, however, by nearly 200 $\mu\text{N/kg}$, indicating that some of the residual gravity field is due to causes other than the volcanic cover; this aspect of the data will be discussed again in the next chapter.

From their time-distance data Robinson *et al.* derived and presented two velocity versus depth functions; one a step function made under the assumption of a subsurface consisting of discrete isovelocity layers and the other a smooth function which assumes a continuously increasing velocity with depth. These functions are shown in fig. 4.14.

Also shown in fig. 4.14 are the conversions of the two velocity versus depth functions, by the Nafe-Drake curve, to density versus depth. The Wairakei density versus depth curve is superimposed for comparison. It can be seen that there is a close correspondence between the cumulative gravity anomalies derived from the Wairakei density-depth curve (based on bore hole density data) to a depth of 2 km, the discrete layered density-depth curve to a depth of 2 km, which is based on a conversion of observed seismic velocities to densities, and the continuous density-depth curve to a depth of 3 km which is also based on a conversion of observed seismic velocities to densities.

Thus there is a three-way, mutual consistency between, 1) density versus depth data from bore holes at Wairakei, 2) velocity versus depth data from the Taupo survey, and 3) the velocity versus density function of

Fig. 4.14 Velocity and density versus depth functions for the Taupo-Ohakuri-Tihoi region. On the left are the velocity versus depth functions supplied by Robinson *et al.* (1981). The right hand graph represents a conversion, by the Nafe-Drake relationship (appendix B), to density versus depth. The Wairakei density versus depth curve, as derived from bore hole data, is shown for comparison.

Nafe-Drake. This would suggest then, that within the errors and assumptions inherent in the refraction data, that the density versus depth curve for the Wairakei region is applicable to much of the Taupo-Tihoi-Ohakuri region and not just to the immediate vicinity of the geothermal field.

Finally, a note about the possible rock-type represented by the 5.45 km/s layer detected beneath the Taupo area. Robinson *et al.* note that this velocity is "typical of greywacke". This is undoubtedly true, however, it was pointed out in the previous section that this velocity could also be associated with an andesite or a diorite, and moreover, the increase of density with depth within the volcanic cover of the Region is such that this velocity could also be interpreted as compacted volcanic rocks at a depth of 2-3 km. In this regard the results (fig. 4.15) of a seismic refraction experiment carried out on the Snake River Plain of the western United States by Pankratz & Ackermann (1982) are pertinent. Here a 3 km-deep bore hole penetrated the area of their survey to show a thick sequence of Cenozoic rhyolitic rocks whose sonic velocities varied from 1.5 km/s at shallow depths, to about 5.0 km/s at a depth of 2.5 km. Below 2.5 km a recrystallised and hydrothermally-altered rhyodacite ash flow was encountered with a sonic velocity of about 5.2 km/s, and whose bulk velocity, as determined by the seismic refraction experiment, was found to vary from 5.2 to 5.5 km/s (see fig. 4.15).

Fig. 4.15 Interpretation of a seismic refraction survey made in the Snake River Plain (U.S.A.) by Pankratz & Akerman, (1982). Note the association of the 5.2-5.5 km/s velocity layer with the rhyolitic rocks found in bore INEL-1. Lithological log is that given by Sparlin *et al.* (1982).

4.5 Conclusions

The prime point to emerge from this chapter is that the volcanic rocks of the Central Volcanic Region are capable of attaining densities approaching 2.60 Mg/m^3 at a depth of about 2-3 km. Whether or not these same rocks can be associated with the 5.0-5.5 km/s velocities detected at the same depth beneath the Taupo region is uncertain. However, it is clear that if there is a lithological transition, deeper than the 2.2 km depth of the deepest bore hole, to an igneous or sedimentary basement, topography on such a basement would have little effect on the residual gravity anomaly field.

In order to account for the residual gravity anomalies within the Central Volcanic Region that are less than say $-500 \text{ } \mu\text{N/kg}$ the following possibilities are considered:

- 1) There are some localities within the Region where large thicknesses of very low-density pumice-tuff or breccia are found i.e. sequences such as that in bore 227.
- 2) There are density inversions at depths greater than 2 km due to discrete bodies of molten rhyolite within the upper crust (e.g. Rogan, 1980).
- 3) The basement rocks of the Central Volcanic Region possess a small but systematic negative density contrast with respect to the basement rocks, at an equivalent depth, that flank the Region. If such a density contrast occurred down to depths of say 10 km or greater, then the associated anomalous gravity effect would make contributions to both the residual and regional gravity anomaly field.

To test or differentiate between the above hypotheses more refraction data from other parts of the Central Volcanic Region are required. This is the subject matter of the next chapter.

CHAPTER 5 SEISMIC REFRACTION SURVEYS CARRIED OUT WITHIN THE CENTRAL VOLCANIC REGION

Summary:

Seismic refraction surveys carried out along lines near Mangakino and between Lake Rotoehu and Otamarakau are reported in this Chapter. Interpretations of the data in terms of plane isovelocity layers show a shallow-dipping, 4.8 km/s refractor at a depth of 2.2 km beneath Mangakino and for the area just north of Lake Rotoehu a 5.5 km/s refractor at a depth of 1.8 km.

5.1 Introduction

The principal objectives of the seismic refraction experiments described in this chapter were to: i) gain constraints for gravity interpretations by determining the thickness and velocity structure of low-velocity volcanic rocks within the Central Volcanic Region; and ii) make integrated gravity and seismic interpretations at two locations within the Central Volcanic Region that were considered geophysically interesting - the Mangakino gravity low and the Otamarakau greywacke outcrop.

Only the salient features of the refraction interpretations are presented here. Appendix C contains details of equipment, short-spread interpretations, and a quantitative discussion of errors that may occur in refraction interpretations within volcanic provinces.

5.1.1 Previous seismic work within the Central Volcanic Region

Some previous seismic refraction and reflection surveys made within the Central Volcanic Region have encountered difficulty due to ground noise and high attenuation of seismic energy within the volcanic cover (Studt, 1958; Modriniak & Studt, 1959; Hochstein & Hunt, 1970; Rogan, 1980). Recent refraction surveys southeast of Mt Ruapehu (Sissons & Dibble, 1981) and the Taupo survey of Robinson *et al.* (1981), discussed in Chapter 4 have, however, successfully recorded seismic energy over shot-detector distances of up to 40 km. Deep water, or distributed shallow-water shots, and microearthquake instruments were used. Microearthquake recorders have filter settings allowing peak responses at 2-5 Hz whereas most refraction seismograph sets have a peak response at about 14 Hz. Given that the attenuation coefficient of seismic energy is directly proportional to frequency (Dobrin, 1976), the results of these surveys suggest that problems of energy transmission may be overcome by working with low-frequency seismic waves. On the other hand the use of microearthquake equipment has the disadvantage that the timing and reading accuracy of microearthquake instruments is somewhat less than that of standard refraction equipment. Moreover, the number of microearthquake instruments, and hence geophones per spread, were limited. Therefore the surveys described here are strictly reconnaissance in nature and are only capable of delineating the broad scale structure.

Results of the Taupo survey and many smaller surveys carried out within the Central Volcanic Region are summarised in Table 5.1. Lithological correlations given in Table 5.1 are those of their respective authors and only at Broadlands (Hochstein and Hunt, 1970) have these correlations been corroborated by drill-hole data.

Table 5.1 Some results of previous seismic work from or adjacent to the Central Volcanic Region

Authors	Locality	Velocity range (km/s)	Depth range (m)	Lithological correlation*
Studt (1958)	Kawerau	1.8-2.2 3.0	0-100 >100	pumice breccia rhyolite
Modriniak & Studt (1959)	Wairakei	1.4-4.0	0-500	pumice, ash, rhyolite
Hochstein & Hunt (1970)	Broadlands	0.4-1.2 1.7-1.9 3.4-4.0	0-20 20-400 >400	pumice pumice-tuff rhyolite
Hochstein & Hunt (1970)	Kaingaroa plateau	3.0-3.4 4.3-4.8	0-250 >250	ignimbrite greywacke
Robinson et al. (1981)	Taupo-Ohakuri	1.5-3.2 3.2 5.45	0-250 250-2000 >2000	ash, pumice ignimbrite, rhyolite ?
Sissons & Dibble (1981)	S.E. of Mt Ruapehu	2.0 4.95 5.4	0-500 500-1700 >1700	lahar deposits and Tertiary rocks greywacke greywacke
Hochstein & Nixon (1979)	Hauraki Rift	0.3-0.8 1.5 3.5-3.7	0-60 60-100 >100	Recent sediments Quaternary sediments greywacke
Ferguson et al. (1980)	Northern Hauraki Rift	1.6 5.0 5.95	0-300 300-2500 >2500	Quaternary sediments greywacke greywacke

*lithological correlations are those made by the respective authors.

5.2 Field and Interpretation Methods

5.2.1 Field methods

Both the Rotoehu and Mangakino surveys involved the positioning of independent microearthquake recorders along an approximately straight line between two principal shot points. Shallow velocity information for both surveys was obtained by shooting short spreads at suitable locations with an RS44 refraction seismograph unit; this information was then utilised in the calculation of datum and weathering corrections for arrivals from the main shots.

Three of the main shots, MS1, MS2 and RS1, consisted of 50 kg of explosives detonated in a water depth of 10 metres. RS2 involved a 75 kg shot in three metres of water and all shots were fired during the New Zealand standard time signals, which were also simultaneously recorded against the shot instant on a radio cassette tape recorder. Most of the microearthquake recorders automatically recorded time signals against their internal time marks although timing for three of the seismographs was achieved by recording V.N.G. minute marks against their internal timing marks. Measurements of many time signals on the seismographs indicated that the maximum timing error of these instruments was less than 0.05 sec. Station positions were determined to an estimated precision of 100 metres on topographical maps of scale 1:63 360 and, for the stations in the Mangakino region, 1:20 000. Heights were determined to ± 5 metres by barometric levelling.

5.2.2 Interpretation methods

Seismic refraction observations constitute a set of number pairs in distance and time. The first decision faced by the interpreter is

whether to fit a series of contiguous straight-line segments or a continuous smooth curve to the data. Implicit in the straight-line segment choice is that the subsurface consists of a series of plane isovelocity layers whereas, on the other hand, fitting a smooth curve through the data implies a subsurface model where velocity increases continuously with depth. Fig. 5.1 illustrates the differences between these two interpretation models. Both methods rely on there being an increase of velocity with depth; the presence of a low-velocity layer within the subsurface will lead to erroneous depth estimates.

Although the assumption of a continuous or discrete increase of velocity with depth can be made with some confidence for say fine grained sediments deposited in a marine environment, it is doubtful if the same confidence can be shared when it comes to the volcanic cover of the Central Volcanic Region. It can be inferred from the density log of bore 121 (fig. 4.1) that low-velocity and thin high-velocity layers occur within the volcanic cover. Some time will be spent in this section on examining the possible effects of such layers, as their presence are considered to be one of the principal sources of error for any refraction interpretation made within the Central Volcanic Region.

Fig. 5.2 shows three situations that will lead to errors in depth estimates. Case 1 is that of the low-velocity zone where no first arrivals from the top of the V_2 layer will be present on the time-distance graph; the computed depth to the V_3 layer will be over-estimated. Case 2 is the blind zone problem (Dobrin, 1976, p.301) where the velocity and thickness of the V_2 layer are such that refractions from its top surface will all be second arrivals and depth estimates to the top surface of the V_3 layer will be too small.

$$Z_1 = T_1/2 \quad V_2 V_1 / (V_2^2 - V_1^2)^{1/2}$$

$$Z_2 = 0.5 (T_2 - 2Z_1 (V_3^2 - V_1^2)^{1/2} / V_3 V_1) \quad V_3 V_2 / (V_3^2 - V_2^2)^{1/2}$$

$$Z(V) = 1/n \int_0^x \cosh^{-1}(Vp(x_i)) dx$$

Fig. 5.1 Interpretational models used for seismic refraction data. On the left is the plane, non-dipping layer model with the time intercept method described by Dobrin (1976). On the right is the continuous velocity with depth model and the Wiechert-Herglotz-Bateman integral (Grant & West, 1965, p.139).

Fig. 5.2 Three subsurface structures that will give rise to erroneous depth estimates in refraction interpretation.

Case 3 is not often discussed but has particular relevance to this study where only a limited number of seismographs were available to cover the seismic lines. Here the discrete spacing of the seismographs is such that the V_2 layer does not contribute any first arrivals and the layer is a blind zone as in case 2; with a continuous coverage of seismographs the layer would be detected.

In an attempt to assess the possible effects of low-velocity and blind zones within the volcanic cover an analysis of the density log of bore 121 and 71 (fig. 4.1) was made in the following manner. Fig. 5.3 shows an approximate conversion of the density log to isovelocity layers, with each layer being required to be at least 100 metres thick. Also shown in fig. 5.3 are two theoretical travel-time graphs, for station spacings of 0 and 2 km, of critically refracted ray paths out to offsets of 20 km, along with the velocity sections interpreted from these travel-time graphs. Straight-line segments were selected by eye.

The 0 km spacing is an ideal situation where there is a continuous line of seismograph stations, whereas the 2 km spacing represents a situation more akin to the conditions of the surveys to be described in this chapter. For the 0 km station spacing it can be seen that the 2.0 km/s layer constitutes a blind zone and the 3.0 and 3.7 km/s layers constitute low-velocity zones. The low-velocity zones have a dominant effect on the error in the depth estimate and the depth to the 5.0 km/s layer is over estimated by nearly 40%.

For the 2.0 km station-spacing case the 2.0 and 3.1 km/s layers are blind zones, the 3.0 and 3.7 km/s layers constitute low-velocity zones and the effect, with respect to the error in the depth estimate, of the blind zones nearly counteracts that of the low-velocity zones. The depth to the 5.0 km/s refractor is thus overestimated by about 10%.

Fig. 5.3

(a) Density vs depth data from bores 121 and 71 that have been converted to approximate isovelocity layers. Each layer was required to be ≥ 100 m and velocities in km/s.

(b) Theoretically derived interpretation section from time-distance graph at top right.

(c) As for (b) but derived from time-distance graph at bottom right.

Bore 121 lies close to the line connecting the two shotpoints of the Taupo refraction survey; hence the velocity sections interpreted by Robinson *et al.* (1981), as shown in fig. 4.14, should bear some resemblance to those predicted in fig. 5.3. This is of course assuming that bore 121 is representative of the whole area covered by the Taupo refraction survey. The 0 km spacing solution in fig. 5.3 closely resembles the solution obtained for the continuously increasing velocity with depth interpretation shown in fig. 4.14. In particular, note that the predicted 3 km depth to rocks with a velocity of about 5 km/s is almost identical in both cases. This would suggest that by adopting a continuously increasing velocity with depth model to interpret refraction surveys within the Central Volcanic Region, depth to "basement" may be overestimated. On the other hand the plane-layer interpretation given in fig. 4.14 predicts a lower overall velocity within the volcanic cover and a shallower depth to the bottom refractor than for the model solution (fig. 5.3) with a 2 km station spacing. This suggests that the station spacing of the Taupo survey, which was about 5 km, was such that the effect of blind zones on the depth estimate negated that of the low-velocity zones.

The principal conclusion drawn from this analysis is that providing the sequences of low-velocity and hidden zones within bore 121 are typical of those elsewhere within the Central Volcanic Region, then a plane layer interpretation of refraction data should give depth estimates that are within about 10% of the true depths, although this is dependent on the seismograph spacing, and a continuously increasing velocity with depth model may give depth estimates that are up to 40% too large. In any event, when the interpreted velocity-depth sections are converted to density-depth sections and their integrated residual

gravity effects calculated, the effects of both the continuous and the discrete increase of density with depth models should be approximately equal. This is borne out by the interpretation of the Taupo refraction data shown in fig. 4.14.

For this study plane-layer models are initially adopted for the purpose of obtaining a first-order structural solution. Gravity data are then used to locate surface projections of major structural discontinuities detected by the seismic data. Finally trial and error ray-tracing methods are employed to refine the first-order model. At the end of the chapter an attempt is made to interpret the data from both surveys in terms of a continuously increasing velocity with depth model.

5.3 Rotoehu Seismic Survey

5.3.1 Introduction

The line between Lake Rotoehu and the greywacke outcrop near Otamarakau (fig. 2.8) was selected for a detailed seismic and gravity investigation for the following reasons.

(i) The absence of short wavelength magnetic or topographic expression, volcanic centres or manifestations of geothermal activity was taken to suggest that the subsurface structure is likely to be less complex than for areas further to the south.

(ii) The greywacke outcrop near Otamarakau is viewed as anomalous for reasons discussed previously. It was hoped that a combined gravity and seismic interpretation would provide some insight into the subsurface structure of this outcrop.

(iii) Logistical advantages offered by this line were that deep water shot-points were available at each end and the line, for a large part, passed through the Rotoehu State Forest; an area with good roads yet of limited public access. The latter factor is crucial when using microearthquake recorders of the continuously rotating drum type. The extreme sensitivity of these instruments can result in pre- or post-shot vehicular or cultural noise obliterating the recorded shot arrival.

Fig. 5.4 gives the position of the seismic line with respect to the residual gravity anomaly field. Apart from three minor greywacke outcrops near Otamarakau, surface geology along and about the line is dominated by a blanket of Rotoiti Pumice Breccia; a hot avalanche deposit erupted from Lake Rotorua 40,000 years ago (Nathan, 1976).

5.3.2 Data

Some difficulty was encountered in finding good geophone sites along the Rotoehu seismic line. The Rotoiti Breccia consists of unconsolidated volcanic ash and breccia and in general the line did not appear to be particularly promising. However, as indicated by the records shown in figs 5.5 and 5.6 noise levels were low and the arrivals surprisingly impulsive. Fig. 5.5 gives two records from station 5 in the middle of the spread for both shots. Fig. 5.6 shows the record made at the far end of the spread and a record from one of the high speed Kinematic recorders.

Table 5.2 lists, for each station, observed first arrival time, distance from shot point, height, datum height, weathering and elevation correction and the corrected arrival time. Elevation and weathering corrections were made as outlined in the scheme depicted in fig. 5.7.

Fig. 5.4 Map of residual gravity anomalies and gravity and seismograph stations for the Lake Rotoehu-Otamarakau area. The position of this seismic line with respect to the whole Central Volcanic Region is shown in fig. 2.8.

Table 5.2 Data for the Rotoehu Seismic Survey

Shot: R.S.1: Shot elevation = 290 metres A.S.L.

Location: South end of the Rotoehu seismic line (Lake Rotoehu)

Stn	Obs (s)	Distance (km)	Height (m)	Dat. Ht. (m)	Correction (s)	Corr. time (s)
1	0.48	0.55	305	280	0.06	0.42
2A	1.07	1.82	274	261	0.06	1.01
2B	1.30	2.66	274	248	0.06	1.24
2C	1.38	3.00	274	241	0.07	1.31
2D	1.64	3.82	260	225	0.07	1.57
3	1.85	4.35	244	215	0.07	1.78
4	2.23	5.53	244	193	0.08	2.15
5	2.51	7.17	183	164	0.06	2.45
6	2.72	8.44	174	140	0.07	2.65
7	2.94	9.54	137	120	0.06	2.88
8	3.01	10.17	63	109	0.03	2.98
9	3.09	11.60	61	83	0.04	3.05
10	3.39	13.32	31	51	0.04	3.35
11	4.04	15.66	15	10	0.05	3.99

Shot: R.S.2: Shot elevation = 10 metres A.S.L.

Location: North end of Rotoehu seismic line (near Otamarakau)

Stn	Obs. (s)	Distance (km)	Height (m)	Dat. Ht. (m)	Correction (s)	Corr. time (s)
12A	0.32	0.59	17	21	0	0.32
12B	0.51	0.92	17	24	0	0.51
10	0.95	2.28	31	52	0.05	0.90
11A	1.26	3.39	91	72	0.06	1.20
9	1.36	4.00	61	83	0.04	1.32
8	1.85	5.43	63	109	0.03	1.82
7	2.21	6.07	137	120	0.06	2.15
6	2.38	7.16	174	140	0.07	2.31
5	2.65	8.45	183	164	0.06	2.59
4	3.00	10.07	244	193	0.08	2.91
3	3.16	11.26	244	215	0.06	3.10
2D	3.22	11.79	260	225	0.07	3.15
2C	3.44	12.70	274	241	0.07	3.37
2B	3.50	12.94	274	248	0.07	3.43
2A	3.64	13.89	274	261	0.06	3.58
1A	3.98	15.44	294	291	0.07	3.91

Corrected time = observed time - correction.

Least squares solutions for selected straight-line segments:

	n	V_{app} (km/s)	ΔV_{app} (km/s)	T_O (s)	ΔT_O (s)	T_E (s)	ΔT_E (s)
Shot: R.S.1							
line 1	2	1.3					
line 2	7	3.0	± 0.1	0.31	± 0.03	5.53	± 0.06
line 3	5	5.6	± 0.2	1.16	± 0.05	3.95	± 0.05
Shot: R.S.2							
line 1	13	1.8	± 0.04	0.01	± 0.01		
line 2	4	3.8	± 0.2	0.28	± 0.04	4.33	± 0.07
line 3	9	5.4	± 0.1	1.00	± 0.05	3.91	± 0.03

n = number of points used in the least squares determination.

The origin is used in line 1 for R.S. 1 and 2.

Fig. 5.5a. Seismograph record for the south shot (R.S.1). Note the time signals immediately preceding the arrival.

Fig. 5.5b Seismograph record for the north shot (R.S.2) at station 5. Time signals preceding the arrival are slightly obscured by noise. Set no. S.1 is a Sprengnether 600 microearthquake instrument.

From the interpretation of shallow velocity spreads (appendix C) the weathered layer was taken as a 20 metre thick layer of velocity 0.3 km/s. This layer is possibly just the aerated layer above the water table and stripping of this layer was done purely to ease the interpretation process.

Stations 12a and 12b from shot RS2 were situated beside a stream where the aerated layer was observed to be of negligible thickness; accordingly no weathering correction was made for these stations. Elevation corrections were made with respect to a sloping datum that passed through both shot points as the surface topography was such that this datum level closely paralleled the ground surface and therefore minimised the maximum elevation correction.

Elevation differences between stations were assumed to be due to differences in the thickness of the 2.2 km/s layer found in the shallow velocity spread on Hannon Rd (appendix C); thus V_1 was taken to be 2.2 km/s and approximate estimates of V_2 were obtained from an initial examination of the uncorrected time-distance data. The largest source of error in making the combined weathering-elevation correction is in the assumed value of t (thickness of aerated layer). It was not possible to determine t all along the line so that departures in t from the assumed 20 metres value may cause errors of up to 0.05 sec in the corrected travel times.

5.3.3 First order interpretation

A subsurface structural model consisting of a discrete number of plane isovelocity layers is assumed in the following interpretation. Adoption of the plane-layered model requires fitting of straight-line

Correction of refraction times for weathering and elevation.

Elevation corrections were made using a similar method to that described by Dobrin (1976). The method differs from that of Dobrin in that a sloping datum is allowed for; necessitating the specification of the datum level beneath both shot and detector. The combined elevation and weathering correction is :

$$\text{Correction} = (e + E - h - t - d_s - d_d)(V_2^2 - V_1^2)^{\frac{1}{2}}/V_2V_1 + t(V_2^2 - V_0^2)^{\frac{1}{2}}/V_2V_0$$

where: e = elevation of shot point

E = " " detector

h = depth of shot point beneath shot point site

t = thickness of weathered or aerated (low velocity) layer

d_s = elevation of datum at shot point

d_d = " " " " detector

V_0 = velocity of weathered layer

V_1 = velocity of layer where the elevation differences between the various stations is assumed to occur.

V_2 = velocity of deepest refractor along which the particular arrival has travelled.

Fig. 5.7 Correction scheme for weathering and elevation used in this study.

segments and a first order interpretation is shown in fig. 5.5; least square determinations of apparent velocities, time intercepts and their associated uncertainties at the 90% confidence level are given in table 5.2. Offsets in the travel-time data from both shots can be observed between stations 7 and 9. The offsets are in opposite directions for each shot indicating a structural discontinuity beneath these stations; therefore arrivals at Stations 9, 10 and 11 from shot RS1 and the arrival at station 8 from RS2 are possibly diffractions from the discontinuity and no straight-line segments were fitted between these points.

Line segments 3 from both RS1 and RS2 are taken to represent arrivals from a common refractor as their end times (T_E) match within the 90% confidence limits (table 5.2). Moreover, near equality of apparent velocities implies that these line segments represent arrivals from a 5.5 km/s refractor lying parallel to the datum surface. The simplest explanation for the 3.0 km/s line segment from RS1 is that it also represents a refractor parallel to the datum. Accordingly, a non-dipping plane layered interpretation was applied to line segments 1, 2 and 3 from RS1 to obtain the structural solution shown in fig. 5.8, i.e. using the time-intercept method (Dobrin, 1976):

$$Z_1 = (T_{i1}/2) V_0 V_1 / (V_1^2 - V_0^2)^{1/2}$$

$$Z_2 = 0.5(T_{i2} - 2Z_1(V_2^2 - V_0^2) / V_2 V_0) V_2 V_1 / (V_2^2 - V_1^2)^{1/2}$$

where $V_0 = 1.3$ km/s

$V_1 = 3.0$ km/s

$V_2 = 5.5$ km/s

$T_{i1} = 0.31$ s

$T_{i2} = 1.15$ s

gives:

Fig. 5.8 Corrected time-distance data for the Rotoehu seismic survey and the first-order structural solution based on the indicated straight-line segments. Apparent velocities of the straight-line segments are shown and station numbers are given at the top of the time-distance plot.

Z_1 = thickness of V_0 layer = 225 metres

Z_2 = " " V_1 " = 1459 metres

and the depth to the 5.5 km/s refractor is $Z_1 + Z_2 = 1684$ metres.

Given that the 1.3 km/s layer possibly represents a smoothed average of the 0.8 and 2.2 km/s layers found in the shallow velocity spread (appendix C) and taking into account the uncertainties in the time intercepts (Table 5.2) the estimate of $Z_1 + Z_2$ would have an associated uncertainty of about ± 150 metres.

5.3.4 Gravity interpretation

Gravity observations with an average spacing of 430 metres were made along the seismic line. The large structural offset in the first order interpretation (Fig. 5.8) was assumed to be associated with the subsurface structure of the greywacke outcrop and for the purposes of making a gravity interpretation, the gravity effect of this body was regarded as being anomalous with respect to the 1700 metres of volcanic cover to the south. This transformation of viewpoint was simply effected by subtracting $-300 \mu\text{N/kg}$, the lowest value of residual gravity over the volcanic cover to the south, from all residual gravity anomalies along the profile. The greywacke block could then be represented by a mass anomaly divided into three horizontal sections of constant density contrast; density contrast decreasing with depth with respect to the volcanic cover. These density contrasts and the geometry of the greywacke body were then progressively modified until a satisfactory fit of observed and calculated gravity was obtained. Fig. 5.10 summarises this analysis. The horizontal dimensions of the block can be inferred from fig. 5.4 to be at least four times the depth to the centre of the mass anomaly so that the two-dimensional approximation used in this analysis is reasonable.

Fig. 5.9 Density-depth functions used for the two and three-dimensional gravity interpretations of the Otamarakau greywacke body. Also shown for comparison is the Wairakei density-depth curve as developed from bore hole data (Chapter 4).

Fig. 5.10 Two-dimensional interpretation of residual gravity anomalies along the Rotoehu seismic line. Density contrasts within the greywacke block are with respect to the low-density volcanics. The lower boundary represents the level at which the velocity, and therefore density, contrasts are considered to vanish. Datum level is the same as the seismic datum (ie. 280 m at R.S.1 and 10 m a.s.l. at R.S.2).

5.3.5 Ray tracing solution

Firstly, a combined structural model featuring aspects of both the first order and gravity models was constructed. Then, geometrical ray paths obeying Snell's law at interfaces of velocity contrast were drawn and travel-times calculated. These were then compared with the observed travel-times.

Finally the model was modified progressively until a satisfactory set of residuals - observed minus calculated times - was obtained. Fig. 5.11 shows the best fit model with a standard error of residuals equal to 0.05 sec.

In arriving at this model the shallow velocity structure determined beneath the Hannon Road spread (appendix C) was considered fixed and the depth of the 3.0-5.5 km/s interface along with the velocity structure within the greywacke block perturbed until the best fit was found. Major modifications to the first order solution are the introduction of topography on top of the greywacke block, as determined by gravity, a lowering of the 3.0-5.5 km/s interface by 100 metres and a division of the velocity structure within the greywacke block.

The one aspect of the model that is of most interest is the structural relationship between the 5.5 km/s refractor and the greywacke block; this cannot be determined by these seismic data alone. For example an alternative model which would give a satisfactory fit to the data is one where the base of the greywacke block formed a continuous unit with the 5.5 km/s refractor to the south.

5.3.6 Geological interpretation

One of the most striking features associated with the Otamarakau greywacke outcrop is its associated, circular-shaped gravity anomaly.

Fig. 5.11 Ray tracing solution for the Rotoehu seismic data. The observed travel-times minus the calculated travel-times of the theoretical ray-paths (dotted lines) are shown at the top. Velocities in km/s and the shaded region represents the interpreted subsurface structure of the greywacke outcrop.

In order to gain an appreciation of the subsurface shape of the block a simple three-dimensional gravity model was constructed using the polygonal lamina method described by Talwani & Ewing (1960). Five lamina were used with the density contrast of each lamina decreasing with depth in accordance with the Wairakei density-depth function as shown in fig. 5.9. Subsurface contours of the model along with the calculated and observed anomalies are shown in fig. 5.12. The three-dimensional model is terminated offshore although the true northward extent of the block is unknown. An examination of many unpublished marine gravity profiles from the Bay of Plenty (held by Geophysics Division, D.S.I.R.) gave no indication of an offshore extent for this gravity anomaly with the closest profile passing within 15 km of the coast near Otamarakau.

The Otamarakau greywacke outcrop is viewed as the surface expression of a flat topped, approximately circular structure with a 65° dip on the edge coincident with the refraction line. From the outset, then, the idea of a continuous erosional surface from the outcrop to a greywacke basement to the south is not favoured; the steep sides suggest a tectonic rather than an erosional origin. Yet, on the other hand, it is difficult to explain the outcrop within the context of a volcanic collapse model. That is a model where melting of a pre-existing sialic crust resulted in the formation of rhyolitic magma, and the subsequent eruption of these magmas led to a tectonic collapse of the crust (e.g. Healy, 1964). One would have to advance reasons why the collapse occurred in a circular fashion about the outcrop at a location many kilometres from any known volcanic centre.

An alternative explanation for the greywacke outcrop has been offered by Calhaem (1973). His proposal for lithospheric spreading within the Central Volcanic Region (chapter 2) accounts for this outcrop as being

Fig. 5.12 Three-dimensional gravity model for the Otamarakau greywacke block (left) and the calculated and observed gravity anomalies.

the surface expression of an allochthonous block, spalled sometime in the past from the trailing edge of the Hawkes Bay proto-plate; the block is thus considered to be "floating" within a medium of intruded igneous rocks and low-density volcanic rocks. In order to account for the isostatic support of the block it would be necessary to propose that either the block has sufficient vertical extent to extend some distance into a higher density crustal or upper mantle layer or the flexural rigidity of the lithosphere is great enough to support this 10-12 km wide mass anomaly.

To test the allochthonous block hypothesis against that of the continuous greywacke basement would require a geophysical method that could determine either the base of the outcrop or the rock-type of basement that surrounds the outcrop. Both the gravity and seismic data reported here are insufficient to do this because of the similarity in seismic velocities and densities of greywacke, on one hand, and diorite, andesite, or compacted volcanic rocks on the other. The magnetic method is usually the most decisive method for discriminating between basement rocks of either a sedimentary or volcanic-igneous affinity. However, as discussed in section 2.2.5 this may not apply in a back-arc situation where magnetic anomalies typically show a diffuse and subdued pattern; for instance, the magnetic anomalies observed over the Coromandel Peninsula (fig. 2.12), where considerable volumes of andesite can be found, are not markedly different from the magnetic anomalies observed around the Otamarakau-Lake Rotoehu area. Thus, on the basis of the unusual subsurface geometry of the Otamarakau greywacke outcrop, its preferred interpretation is the allochthonous block model discussed above.

The occurrence of blocks and fragments of old continental crust enclosed within younger rocks of a back-arc basin has been noted elsewhere.

For example, isolated blocks of Mesozoic continental crust within the younger oceanic crust of the Japan Sea are discussed by Hilde & Wageman (1973) and similar occurrences of continental crust have been found within a Cretaceous back-arc basin of southern Chile, which has now been exposed by uplift and erosion (Dalziel *et al.*, 1974; Bruhn & Dalziel, 1977; De Wit & Stern, 1981). Furthermore, in a general overview of back-arc basins that have formed throughout geological time, Tarney & Windley (1981) note, when discussing back-arc basins that have formed at continental margins, that "...back-arc spreading may lead to the detachment and dispersal of continental fragments."

Reverting briefly back to the Central Volcanic Region, it is possible that the Otamarakau greywacke block is not the only tectonically-rifted greywacke block within the Region. For example, writing before the concepts of plate tectonics and back-arc basins had been developed, Healy (1962) noted that Modriniak & Studt (1959) had interpreted the Paeroa fault block as "...a part of the (greywacke) Kaingaroa Plateau which has slipped and subsided towards the volcanic zone" and then Healy went further to conject "This has in fact occurred on a much larger scale, and the entire Kaingaroa Plateau seems to be part of the eastern ranges which have sunk and slid towards the Taupo Volcanic Zone".

The interpretation of the Otamarakau greywacke outcrop being the surface expression of an allochthonous block is based on circumstantial geophysical evidence alone. More work about the outcrop is clearly warranted, in particular marine gravity observations close to the coast would be useful in order to determine the northern extent of the block.

5.4 Mangakino Seismic Survey

5.4.1 Introduction

One of the most impressive geophysical features of the Central Volcanic Region is the Mangakino gravity low shown in fig. 5.13. Five major gravity lows were identified on the residual gravity anomaly map in chapter 3, four of which are more or less confined to the axial strip of the Taupo Volcanic Zone and correlate with rhyolitic volcanic centres. The Mangakino gravity low stands apart from these four gravity lows in that it lies further to the west, has no associated magnetic anomaly or volcanic centre and is greater in magnitude (-650 compared to $-500 \mu\text{N/kg}$).

Hochstein (1976) suggested that the gravity low at Mangakino might be partly due to molten rhyolite at depths below 6 km. Alternatively, the Mangakino gravity low may be explained by an unusually large thickness of low-density volcanic rocks. For example Rogan (1980) modelled the Mangakino gravity low in terms of a 5 km depression filled with rocks of density contrast -0.5 Mg/m^3 . There is no strong thermal anomaly in the Mangakino region which might corroborate the molten rhyolite hypothesis yet, on the other hand, the model of a 5 km thickness of volcanic cover with a density contrast of -0.5 Mg/m^3 conflicts with the density-depth analysis given in chapter 4. The prime objective of this refraction survey was to determine at what depth basement-type seismic velocities are reached and therefore possibly differentiate between the two models outlined above.

Surface geology along the greater portion of the line consists of Whakamaru Ignimbrite, beneath which Martin (1965) has proposed exists a graben structure that is aligned with the Hauraki Rift. Factors which guided the selection of the seismic line were that it had deep water shot-points at each end, passed near the centre of the gravity low, crossed the Waipapa fault and ended on outcropping greywacke at its northern end

(fig. 5.13). Moreover, for a large part, the line passed through the Kinleith forest; a private forest with limited public access.

5.4.2 Data

Despite good geophone sites the arrivals recorded at Mangakino tended to be less impulsive than those recorded at Rotoehu. Fig. 5.14 shows two seismograph records from the Mangakino survey. Table 5.3 lists for each seismograph station the observed first arrival time, height, datum height, elevation and weathering correction as well as the corrected arrival time. Elevation and weathering corrections were made as described in section 5.3.2 for the Rotoehu survey. Values of t , V_0 , V_1 and V_2 , were obtained from interpretations of the shallow velocity spreads reported in appendix C and estimates of V_2 from a preliminary inspection of an uncorrected travel-time graph. The thickness, t and velocity V_0 of the aerated layer were taken as 20 metres and 0.3 km/s respectively. This layer was considered to be present at all stations except 14(a) which was placed on a greywacke outcrop. The layer for which elevation differences were assumed to occur in (V_1 -layer) was taken to be the 2 km/s layer. V_2 varied from 3.2 km/s to 5.2 km/s and was determined from an examination of the uncorrected travel-time data.

5.4.3 First order structural solution

The basic assumption in the interpretation of these data is that the subsurface structure can be approximated by a series of isovelocity layers that are bounded by plane interfaces, and whose dip directions are parallel to the direction of the profile. However, the gravity anomaly pattern shown in fig. 5.13 suggests the limitations inherent in the two-dimensional design and interpretation of this seismic survey. In particular, the three-dimensional character of the subsurface structure, as may be

Fig. 5.13 Gravity and seismograph stations and residual gravity anomalies in the Mangakino area. The position of this line with respect to the rest of the Central Volcanic Region is shown in fig. 2.8.

STATION; 6
 Set No; 5
 Record No; shot 1.

GAIN 66db

SHOT DISTANCE 8.55 km

0 20 sec

Fig. 5.14a Seismograph record from a Sprengnether 800 instrument at Station 6. Note time signals preceding the first arrival.

STATION; 14a
 Set No; K.3.
 Record No; shot 1.

ATTENUATION - 30db FILTER 1

SHOT DISTANCE 18.7 km

0 20 sec

Fig. 5.14b Seismograph record at the most distant station from M.S.1. Station 14(a) was the only station located on outcropping greywacke.

inferred from the gravity data, suggests the dip direction of the subsurface layers may deviate somewhat from the direction of the profile. However, the distribution of shot-points and seismograph stations was dictated by the availability of deep-water lakes and roads, respectively, and the interpretation must be recognised for these limitations.

Table 5.3 lists the least squares solutions to straight-line segments indicated in fig. 5.15. An offset in time between stations 4 and 2 from shot MS1 is suggestive of a subsurface structural discontinuity. A similar time offset, but in the opposite direction, is also evident between stations 3 and 6 from MS2. The position of this offset correlates closely with the position of maximum gravity gradient and the geologically interpreted position of the Waipapa fault as shown by Healy *et al.* (1964, fig. 5.13).

After making allowance for this time offset, reference to values of reciprocal end times (T_E) in table 5.3 suggests that the only reversed straight line pair that represent arrivals from a common refractor are lines 2 and 3 from shots MS2 and 1 respectively. No line segment from MS2 can satisfactorily be matched to line segment 2 from MS1 (apparent velocity = 3.4 km/s) and the true velocity cannot be calculated from these data alone.

A refractor of velocity 3.0 km/s at a depth of 200 metres was detected beneath the Murry Road shallow velocity spread (appendix C). No such refractor was determined beneath the Christiansen Road shallow velocity spread. Hence the most straightforward interpretation is that the 3.4 km/s apparent velocity of line segment 2 represents an updip velocity for a refractor which shallows to a depth of 200 metres beneath the Murry Road spread.

The quality of data from the Murry Road spread was not good and only from shot MS4, at the north end of Murry Road, did seismic energy

Fig. 5.15 Corrected time-distance data and the first-order structural solution for the Mangakino seismic data. The Christiansen and Murry Rd spreads are discussed in appendix C. Numbers at the top of the time-distance graph indicate the station numbers and numbers on the time-distance graph are apparent velocities in km/s of the indicated straight line segments.

Table 5.3 Data for the Mangakino Seismic Survey

Shot: M.S.1: Shot elevation = 195 metres A.S.L.

Location: South end of the Mangakino seismic line (Lake Maraetai)

Stn	Obs. (s)	Distance (km)	Height (m)	Dat. Ht. (m)	Correction (s)	Corr. time (s)
13	0.73	1.24	326	280	0.07	0.66
12	1.76	3.24	366	275	0.07	1.69
11	2.08	4.44	326	273	0.06	2.02
10	2.55	6.22	359	268	0.07	2.48
9	3.00	7.51	305	265	0.05	2.95
8	3.18	8.55	323	263	0.06	3.12
6	3.78	12.05	280	255	0.04	3.74
5	4.23	13.79	293	251	0.05	4.18
4	4.35	14.59	311	249	0.06	4.29
3	5.15	15.31	326	247	0.07	5.08
2	4.25	16.26	305	245	0.06	4.19
1	4.45	16.99	427	243	0.12	4.33
14	4.73	18.64	183	240	0.04	4.77

A technical malfunction resulted in stations 7A and 7B not recording time signals.

Shot: M.S.2: Shot elevation = 111 metres A.S.L.

Location: North end of Mangakino seismic line (Lake Arapuni)

Stn	Obs. (s)	Distance (km)	Height (m)	Dat. Ht. (m)	Correction (s)	Corr. time (s)
1	0.45	1.69	427	243	0.06	0.39
2	0.59	2.42	305	245	0.06	0.53
3	0.83	3.37	326	247	0.06	0.77
4	1.12	4.09	311	249	0.08	1.04
5	1.41	4.89	293	251	0.07	1.34
6	1.99	6.63	280	255	0.06	1.93
7A	2.39	8.36	311	259	0.07	2.32
7B	2.63	9.40	347	261	0.09	2.54
8	2.89	10.14	323	263	0.08	2.81
9	2.99	11.17	305	265	0.07	2.92
10	3.41	12.46	359	268	0.09	3.32
11	3.85	14.24	326	273	0.08	3.77
12	4.10	15.44	366	275	0.09	4.01
13	4.39	17.44	326	280	0.07	4.32
14A	4.72	18.68	201	283	0.02	4.70

Corrected time = observed time - correction.

Least squares solutions for selected straight-line segments:

	n	V_{app} (km/s)	ΔV_{app} (km/s)	T_O (s)	ΔT_O (s)	T_E (s)	ΔT_E (s)
Shot: M.S.1							
line 1	3	1.9	± 0.1	0.0			
line 2	4	3.4	± 0.2	0.73	± 0.16	6.13	± 0.14
line 3	5	5.2	± 0.3	1.48	± 0.13	5.06	± 0.10
Shot: M.S.2							
line 1	4	4.4	± 0.01	0.0	± 0.02	4.18	± 0.19
line 2	10	4.4	± 0.1	0.43	± 0.09	4.71	± 0.05

n = number of points used in the least squares determination.

The origin is used in line 1 for M.S. 1 and 2.

propagate the full 1100 metres of the spread. Hence, the 3.0 km/s velocity recorded in this spread and the 3.4 km/s velocity of line segment 2 from MS1 are matched as down and updip velocities, respectively, from the same refractor.

At the Christiansen Road spread a 100 metres thick layer of velocity 1.2 km/s was detected and a similar layer of slightly lower velocity and thickness was detected on the Murry Road spread. Due to the large spacing between the microearthquake seismographs, this layer could not be identified from the main shots. Nevertheless, for the purpose of this first order interpretation it will be assumed that it is continuous between MS1 and station 3. With both this assumption, and that regarding the up and down-dip velocities of the 3.0-3.4 km/s refractor, there is enough data to solve the two dipping layer problem.

Fig. 5.16 shows the model and geometrical relations given by Mota (1954) that will be used to solve the two-dipping layer problem. The model will initially be used to obtain information about the 3.0-3.4 km/s refractor; this being the V_2 layer in the model shown in fig. 5.16. The V_0 and V_1 layers are considered to be the 1.2 and 2.0 km/s layers respectively. Information pertaining to the deeper refractor with up and down-dip velocities of 5.2 and 4.4 km/s, respectively, is then obtained by numerically stripping the 1.2 km/s layer and using the two-dipping layer model again with the 2.0 and 3.0-3.4 km/s layers representing V_0 and V_1 .

a) Depth, dip and true velocity of the 3.0-3.4 km/s layer:

Substituting the following in the equations of fig. 5.16

$$V_0 = 1.2 \text{ km/s}$$

$$V_1 = 2.0 \text{ km/s}$$

$$Z_1 = 0.1 \text{ km}$$

$$\theta_1 = 0^\circ$$

$$V_{2u} = 3.4 \text{ km/s}$$

$$V_{2d} = 3.0 \text{ km/s}$$

$$T_{1u} = 0.73 \text{ s (see table 5.3)}$$

the following are derived:

$$i_2 = 38.8^\circ$$

$$\theta_2 = 3.1^\circ$$

$$Z_2 = 0.74 \text{ km}$$

$$H_2 = 0.84 \text{ km (depth beneath MSL)}$$

$$V_2 = 3.2 \text{ km/s}$$

b) Depth, dip and true velocity of the 4.4-5.2 km/s layer:

In order to determine the parameters pertaining to the lower refractor, the 1.2 km/s layer is stripped off by subtracting from the intercept time (T_{2u}), the two-way travel-time of a near vertical ray path through the 1.2 km/s layer. Thus the new input data for the two-dipping layer model are:

$$V_0 = 2.0 \text{ km/s}$$

$$V_1 = 3.2 \text{ km/s}$$

$$\theta_1 = 3.1^\circ$$

$$V_{2u} = 5.2 \text{ km/s}$$

$$V_{2d} = 4.4 \text{ km/s}$$

$$T_{2u} = 1.48 \text{ s -when corrected for the stripped 100 metre}$$

thick, 1.2 km/s layer $T_{2u} = 1.32 \text{ s}$.

With the above parameters substituted into the equations of fig. 5.16 the following are derived:

$$i_2 = 42.2^\circ$$

$$\theta_2 = 1.9^\circ$$

$$V_2 = 4.8 \text{ km/s}$$

$$Z_2 = 1.4 \text{ km}$$

$$H_2 = 2.14 \text{ km (2.24 km when the 1.2 km/s layer is allowed for)}$$

The travel-time data and the first-order structural solution are shown in fig. 5.15. Table 5.3 lists the uncertainty in T_0 for line segment 3 (i.e. T_{2u}) to be 0.13 s at the 90% confidence level. This would give an uncertainty in the thickness estimate for the 1400 metre thick 3.2 km/s layer of 350 metres. Thus the error in the above depth estimates is about 25%. The principal causes of this rather large error are attributed to the wide geophone spacing and the deviation of the real structure from plane isovelocity layers.

5.4.4 Ray tracing

The best-fit ray-tracing solution is shown in fig. 5.17. This solution was arrived at by holding the 3.2-4.8 km/s interface at the position indicated by the first-order structural solution and adopting a 5.0 km/s velocity for the interpreted greywacke abutment.

The arrival at station 3 from MS1 was up to 0.8 s late. Trial ray-tracing solutions indicated that this station is placed almost directly above the interpreted surface projection of the greywacke fault and was effectively in a shadow zone with respect to arrivals from the 4.8 km/s refractor. Furthermore, the line up of this arrival with the projection of the 3.4 km/s line segment (fig. 5.17) from MS1 suggests that this arrival represents seismic energy propagating up the proposed 3.2 km/s refractor.

5.4.5 Gravity interpretation

Seismic velocities in the model of fig. 5.1 were converted to densities with the Nafe-Drake relationship given in appendix B. Using these converted densities, with respect to a basement density of 2.65 Mg/m³, and depths obtained from the refraction interpretation it was found that

Fig. 5.17 Ray tracing solution for the Mangakino refraction data. Velocities in km/s. Depth is with respect to sea level.

only $-440 \mu\text{N/kg}$ of the observed $-650 \mu\text{N/kg}$ residual anomaly could be accounted for. Hence, there is a "secondary residual gravity anomaly" of $-210 \mu\text{N/kg}$ and it would appear that a low-density mass distribution deeper than 2.2 km is required.

A trial solution to the residual gravity profile along the seismic line is shown in fig. 5.18. This solution, just one of many possible, involves a rectangular volume of low density material (density contrast -0.45 Mg/m^3) in the depth range 3 to 7 km.

An alternative trial solution would be to consider that the 4.8 km/s layer represents a thin, high-velocity layer which overlies a deeper sequence of low-density volcanic rocks. However, the density data discussed in the previous chapter indicate that the increase of density with depth for the volcanic rocks of the Central Volcanic Region is such that their densities are approaching $2.50\text{--}2.60 \text{ Mg/m}^3$ in the depth range of 2–3 km. A trial model showed that a mass anomaly of density contrast -0.15 Mg/m^3 (i.e. absolute density of 2.50 Mg/m^3), and with its top surface at a depth of 3 km, would be required to extend to depths of 14 km just to account for amplitude of the observed residuals, with a consequence that the gradients could not be matched. This trial model was therefore rejected.

5.4.6 Geological Interpretation

Experimental and theoretical studies by Ewart *et al.* (1975), Murase & McBirney (1973) and Heming (1979) point to a density range of molten rhyolite from 2.05 to 2.41 Mg/m^3 ; the range being due to differing chemical compositions and water content. Hence the density contrast of -0.45 Mg/m^3 used for the mass anomaly beneath seismic basement in fig. 5.18 is compatible with a model of molten rhyolite emplaced in basement rocks of density 2.65 to 2.75 Mg/m^3 .

Fig. 5.18 A trial gravity model for a profile along the Mangakino seismic line. The seismic velocity and density structure for the volcanic cover is shown on the right. The gravity effects are calculated with respect to a basement density of 2.65 Mg/m^3 . Two-dimensionality is assumed for the volcanic cover (mass anomaly 1) but the lower mass anomaly (2) has a strike length of 5 km both into and out of the plane of the profile. Calculated gravity effect obtained using the $2\frac{1}{2}$ -dimensional gravity interpretation method outlined in appendix D.

Further geophysical and geological observations that lend both direct and indirect support to the model of molten rhyolite beneath the Mangakino region are: 1) The geomagnetic depth sounding work of Midha (1979) detected a suppressed Z-response of the natural variation in the geomagnetic field over an area which is approximately coincident with the Mangakino gravity low (see fig. 5.19). Although no modelling of this anomaly was performed, Midha noted that the anomalous Z-response "lends qualitative support" to Hochstein's proposal of molten rock at a shallow depth beneath Mangakino.

2) The geological map of Healy *et al.* (1964) shows a small geothermal field in the vicinity of Lake Maraetai. Visible surface manifestations of this field have now been eradicated by the man-made lake and the heat output of this field is unknown. However, a local resistivity low, which are characteristic of many of the geothermal fields in the Central Volcanic Region (Hatherton *et al.*, 1966), has been detected in the vicinity of Lake Maraetai (W.J.P. McDonald, pers. comm.).

3) A stratigraphic study of North Island ignimbrites by Blank (1965) show that a large ignimbrite sheet found to the west of the Central Volcanic Region has an estimated volume of 200 km^3 and "...appeared to have a source somewhere in the Mangakino Basin". This ignimbrite sheet, named the Ongatiti ignimbrite by Blank, is possibly the largest and oldest (fission track dated at 0.73 My by Kohn (1973)) to have been erupted from the Central Volcanic Region.

If the proposed magma body beneath Mangakino is a relic from an ignimbrite eruption that occurred 0.7 My ago a study of the cooling curves presented by Lachenbruch *et al.* (1976) would suggest that any remanent magma would certainly have cooled and be frozen by now. Alternatively, magma may have been reinjected more recently but in this case one might

Fig. 5.19 The association of a geomagnetic induction anomaly found by Midha (1979) with the Bouguer gravity anomaly in the Mangakino area.

anticipate a more prominent thermal anomaly to be observed in the Mangakino area.

Heat-flow measurements were attempted in bore holes around the Maraetai dam, close to the southern shot-point (fig. 5.13). These holes were put down about 30 years ago for dam investigation work but have now collapsed at or near the water table mark. All attempts at heat flow measurements proved unsuccessful. However in view of the widespread hydrothermal circulation that is suspected to occur within the upper crust of the Central Volcanic Region (Studt & Thompson, 1969) it is feasible that even if there is a large body of molten rhyolite at a shallow depth beneath Mangakino, heat from the molten rock could be carried by the circulating water and discharged elsewhere.

At this stage the interpretational model in fig. 5.18 is still considered tentative. Although there is converging evidence to suggest that there is molten, or partly molten, rhyolite beneath Mangakino, the sheer size of the body depicted in fig. 5.18 seems doubtful. The volume of the proposed body is about three times larger than that of any known ignimbrite sheets erupted from the Central Volcanic Region and the differential stresses within the upper crust implied by the body of low-density molten rock are large. Before proceeding with further discussion on the Mangakino gravity low, seismic and gravity data from the Taupo region will be examined.

5.5 Seismic and gravity data from the Tihoi-Taupo area

In chapter 4 results and an interpretation of the Taupo refraction survey were introduced (see figs 4.13 and 4.14). Although this survey was only for reconnaissance purposes, and consequently the data is somewhat

Fig. 5.20 Residual gravity interpretation along a line between Tihoi and Taupo (see fig. 4.13). The lower mass anomaly (b) has a strike direction of 5 km into and out of the plane of the profile. Positions and depths of bores 223, 121 and 227 are shown. Density contrasts in Mg/m^3 .

sparse, it nevertheless is sufficient to highlight a fundamental conflict between the seismic interpretation offered by Robinson *et al.* (1981) and the observed residual gravity data from the same region. Namely, the seismic interpretation suggests the presence of a flat refractor at a depth of 2 km beneath the Wairakei-Tihoi-Lake Ohakuri region, whereas residual gravity anomalies show a $-200 \mu\text{N/kg}$ variation over the same area. In other words the delay times due to low-velocity volcanic rocks are identical, irrespective of whether the observed residual gravity anomaly is -350 or $-550 \mu\text{N/kg}$. One possible explanation for this conflict is that a portion of the residual gravity anomaly is due to a low-density source deeper than 2 km.

An interpretation of a closely-spaced west-east gravity profile across this region is shown in fig. 5.20. The velocities supplied by Robinson *et al.* (1981) were converted to densities and the residual gravity effect of the low-velocity rocks, with respect to a density of 2.65 Mg/m^3 , was calculated to be $-340 \mu\text{N/kg}$. For a 2 km-thick section this is equivalent to an average density contrast of -0.4 Mg/m^3 ; however allowance for this seismically-determined section still leaves a secondary residual anomaly of about $-200 \mu\text{N/kg}$ unaccounted for. The interpretation offered in fig. 5.20 features a three-dimensional body of density contrast -0.4 Mg/m^3 , in the depth range 3 to 6 km, to explain this secondary residual anomaly.

The best explanation for such a low-density mass is molten rhyolite when the same arguments as those developed for the Mangakino secondary residual gravity interpretation are applied. Few objections can be raised to this model on the basis of the heat flow considerations, as heat discharges from the Wairakei area at the rate of about 530 megawatts (fig. 2.13). However, the model does conflict with an interpretation of microearthquake

data from this region. Evison *et al.* (1976) found that microearthquake activity west of Wairakei was high in the depth range 2-6 km, while Robinson *et al.* (1981) found no appreciable P-wave delays for stations located in the same region. The presence of a large mass of molten rhyolite, as indicated in fig. 5.20, would reduce strength, and hence earthquake activity, and possibly cause a P-wave delay.

The secondary residual gravity anomaly of about $-200 \mu\text{N/kg}$ present in the Tihoi-Taupo region, after the gravity effect of the seismically-determined thickness of low-velocity rocks is removed, is quite similar to that at Mangakino. An alternative means of explaining these secondary residuals is to consider that there is a continuum of density irregularities right down through the crust-mantle column of the Central Volcanic Region. For such a situation it is not possible to make a geologically meaningful separation of the total gravity anomaly field into regional and residual gravity anomalies (see chapter 3). This point is illustrated by a comparison of the two possible "deep structure" models for the Central Volcanic Region shown in fig. 5.21.

In model a) of fig. 5.21 a model for the subducted plate, similar to that proposed by Hatherton (1970) (see fig. 2.4a), is considered as the source of the regional gravity anomaly field, and the negative residuals of about $-340 \mu\text{N/kg}$ are due entirely to the 2 km thickness of superficial volcanics with a density contrast of -0.4 Mg/m^3 . Because the respective sources of the regional and residual gravity anomaly fields are well separated in depth, with respect to the width of the Region, the regional-residual gravity anomaly separation can be performed unambiguously and thus has some geological meaning. Model b), on the other hand, contains the same 2 km thick section of superficial volcanics as model a), but also includes a negative mass anomaly in the depth range 2-10 km of density contrast

Fig. 5.21 Two possible "extreme-case" models to explain the regional and residual gravity anomalies of the Central Volcanic Region. Both models incorporate the same 2 km thick, near-surface mass anomaly that represents the low-density volcanics. Density contrasts in Mg/m^3 .

-0.15 Mg/m^3 ; these two negative mass anomalies are isostatically compensated by a positive mass anomaly, which may be interpreted to represent a thin crust for the Region, in the depth range of 15 to 22.5 km. Model b) is similar to the isostatically compensated structures found to be associated with some continental rifts; for example the Baikal Rift as described by Zorin (1981). Because the three mass anomalies in model b are not well separated in depth, with respect to the width of the Region, the derived residual anomaly of $-420 \text{ } \mu\text{N/kg}$ bears little relationship to the gravity effect of any one of the three mass anomalies in the model; in this situation the separation of the gravity anomaly field into regional and residual components has little geological meaning.

The use of a -0.15 Mg/m^3 density contrast (model b, fig. 5.21) between the crustal rocks of the Central Volcanic Region, in the depth range of 2-10 km, and the greywacke rocks in the same depth range that flank the Region, requires some justification. Although seismic velocities at a depth of 2-3 km within the Region (4.8-5.5 km/s) appear to be only slightly lower than those for greywacke rocks (5.4-5.9 km/s) at an equivalent depth found elsewhere within the North Island (Garrick, 1968; Ferguson *et al.*, 1980; Sissons & Dibble, 1981) the uncertainty in the empirical relationship between density and seismic velocity is such that two rock types with similar velocities may be associated with densities that differ by as much as 0.15 Mg/m^3 (section 4.3.3).

The density of greywacke rocks deep beneath the ranges that flank the Central Volcanic Region is unknown. However, an indication of the upper limit for the density that these rocks are capable of attaining can be gained by considering the particle density of the low-grade metamorphic equivalent of these greywackes - the alpine and Otago schists found in the South Island (Wellman, 1956). Hatherton & Leopard (1964)

give a value of 2.75 Mg/m^3 as the mean particle density of these schists. Thus if the crustal rocks of the Central Volcanic Region in the depth range of, say, 2 to 10 km have an average density of 2.60 Mg/m^3 , a density not incompatible with the observed seismic velocities of 4.8-5.5 km/s (appendix B), then the proposed density contrast of -0.15 Mg/m^3 between the crustal rocks of the Region, on one hand, and the greywacke ranges, on the other, is not unreasonable.

A three-dimensional version of a model similar to that of model b) in fig. 5.21 will be used in chapter 7 as one component of a gravity interpretational model for the central North Island, and eventually preferred interpretations of the gravity lows at Taupo and Mangakino will be presented.

5.6 Velocity Versus Depth Relationships

Interpretation of the refraction data from the Rotoehu and Mangakino surveys has so far been made under the assumption of a subsurface consisting of plane isovelocity layers. Accordingly, the data were approximated by a series of straight line segments. The data from MS1, up to and including station 4 (fig. 5.15), and RS1, up to and including station 8 (fig. 5.8), could equally well be described by smooth curves; the necessary assumption then being that the subsurface consists of rocks whose velocity variation is only in the vertical direction and that this variation is a continuous increase with depth. Smooth curves were constructed that provided a fit to the data from MS1 and RS1 with standard errors less than 0.1 s. The gradients of these curves were then calculated at intervals of 0.5 km and these data were then applied to a numerical integration of the Weichert-Herglotz-Bateman function (fig. 5.1).

Fig. 5.22 shows the velocity-depth functions so derived, for both Mangakino

Fig. 5.22 Seismic velocity and density versus depth functions for the Mangakino and Rotoehu seismic data. The continuous velocity with depth functions were calculated with the W-H-B integral (fig. 5.1) as described in the text.

and Rotoehu, compared with the respective plane-layered velocity with depth functions.

For the data from Rotoehu, Mangakino and Taupo (fig. 4.14) the continuously increasing velocity with depth functions give larger estimates for the depth to the 4.8 to 5.5 km/s rocks than do the plane-layer interpretations. As discussed in section 5.2.2 this can be explained by there being a presence of both low-velocity and blind zones (fig. 5.2) within the subsurface. Blind zones cause an underestimation of the true depths, and only affect the plane-layer interpretation method, whereas low-velocity zones cause an overestimate of the true depths for both methods of interpreting the data. Therefore, given a set of refraction data from an area which has both low-velocity and blind zones within the subsurface, a plane-layered interpretation of the data would be expected to yield a smaller depth to the highest velocity rocks and this depth should be closer to the true depth.

5.7 Conclusions

The deployment of microearthquake instruments to record seismic energy from artificial explosions has proved an effective means of carrying out seismic refraction experiments within the Central Volcanic Region. Judicious selection of seismograph sites and deep-water shot points is, however, desirable.

Fig. 5.23 summarises the plane layer interpretations of the three refraction surveys discussed in this chapter. The most notable points of these data are summarised by the following four points:

1. The thickness of low-velocity volcanic rocks is much the same at the two seismic refraction sites that are outside of the boundaries of the Taupo Volcanic Zone (Rotoehu and Mangakino) as at the one site within the Zone (Taupo). Thus structural depression

Fig. 5.23 A summary of the three refraction interpretations discussed in this chapter. The locations of the three surveys are shown superimposed on the residual gravity map in the lower figure.

is not just confined to the youthful and currently active Taupo Volcanic Zone.

2. The thickness of volcanic rocks with velocities less than 3.2 km/s is greater at Mangakino than at the other two refraction sites. This, in part, accounts for the larger negative residual gravity anomalies that are observed at Mangakino than at Taupo or Rotoehu.
3. At all three localities no appreciable dip was found on the deepest, or basement, refractors. If the Central Volcanic Region is underlain by a Mesozoic sedimentary basement that has slumped, and has then been locally ruptured as a result of subcrustal melting and concomitant rhyolitic volcanism, then it is anticipated that such a basement would display considerable structural relief on its top surface, rather than being flat as suggested by the refraction data. The preferred interpretation of the 4.8-5.5 km/s seismic velocities that occur at a depth of 2-3 km is that they do not represent a lithological boundary, but merely represents igneous or compacted volcanic rocks at this depth; this interpretation is consistent with the rock property data presented in chapter 4. However, no firm conclusions regarding the petrology of basement rocks for the Central Volcanic Region can be drawn from these seismic data alone. The question of what type of rock constitutes basement for the Central Volcanic Region need not concern the gravity interpretation at this stage, but it will be discussed again in chapter 8 where the heat output for the Region is examined.
4. The interpreted velocity sections at Rotoehu and Taupo are similar yet the residual gravity anomalies at these two localities differ by about 200 $\mu\text{N/kg}$. The residual anomalies along the refraction profile at Rotoehu can satisfactorily be explained by the seismically determined thickness of low-velocity, and therefore low-density, volcanic rocks. In contrast, at both Taupo and Mangakino, secondary

residuals of about $-200 \mu\text{N/kg}$ remain after the calculated gravity effect of the seismically determined thickness of volcanic rocks is allowed for. Interpretations of these secondary residuals in terms of molten rhyolite emplaced beneath the seismic basement were considered tentative for reasons discussed in the text. In particular the sheer volumes ($\approx 700 \text{ km}^3$) of the required magma bodies is doubtful, and at Taupo there is geophysical evidence which would not support such a model. For the purpose of obtaining a physically plausible explanation for these secondary residual gravity anomalies it is proposed that the upper crustal rocks of the Central Volcanic Region are themselves less dense than the upper crustal rocks, at an equivalent depth, elsewhere within the North Island. In such a situation the anomalous gravity effect of the upper crustal rocks of the Region will possess both regional and residual components and the distinction between regional and residual gravity anomalies for the Central Volcanic Region will become obscure.

To demonstrate this proposal a deep structure model for the Central Volcanic Region is required whose calculated gravity effect will account for the complete gravity anomaly field. This will be the topic presented in chapter 7 and in chapter 8 preferred interpretations of the gravity lows at Taupo and Mangakino will be made.

The following chapter is a slight digression from the central aim of this thesis in that it has only an indirect bearing on a gravity interpretation for the Central Volcanic Region. The magnetic anomaly field is studied in the frequency domain for the purpose of finding the average maximum depth extent of magnetic source bodies within the Taupo Volcanic Zone; this depth being interpreted as that of the Curie point isotherm.

CHAPTER 6 DETERMINATION OF DEPTH TO THE CURIE POINT ISOTHERM
FROM MAGNETIC ANOMALY DATA

Summary:

The magnetic anomalies of the Central Volcanic Region are analysed in the frequency domain with a view to determining the maximum depth extent of the magnetic source bodies. This is estimated to be 6.2 ± 2.0 km and is interpreted as the average depth of the Curie point isotherm.

6.1 Introduction

A smoothed version of the aeromagnetic map of Whiteford (1976) for the Central Volcanic Region is shown in fig. 6.1. Viewed alone, the anomalies appear impressive. The magnetic signature of the Region is, however, seen in its broader context by inspecting fig. 6.2, where North Island magnetic anomalies both on and off shore are shown. It can be observed from this perspective that the Central Volcanic Region is situated within an area of generally negative anomalies upon which small-amplitude, short-wavelength, positive anomalies are superimposed. It would appear, therefore, that the magnetic source bodies within the Central Volcanic Region are shallow relative to those situated offshore.

The level at which the deepest, discrete, magnetic source exists is here termed the base of the magnetic crust. This depth may depend on one of two factors. Temperatures may be such that rock magnetism vanishes - because the Curie point isotherm, or Curie depth, is reached. Alternatively, the base of the magnetic crust may correspond to the termination of a lithological unit which possesses a strong magnetisation contrast with respect to surrounding rocks.

Fig. 6.1 Total force magnetic anomaly map of the Central Volcanic Region; simplified from Whiteford (1976), contour interval = 50 nT, flight altitude = 1525 metres. Dashed lines enclose the area that was digitised for spectral analysis.

- Magnetic anomaly map (total force) of the North Island, New Zealand. Measurements over land areas were made at 3.05 km above sea level except in the Central Volcanic Region where they were made at 1.5 km, and in North Auckland at 0.91 km above sea level. Measurements over sea areas along the west coasts of the North and South Islands were made at 0.5 km above sea level; in Bay of Plenty at sea level. Data from Reilly (1969; 1970) Hunt and Syms (1976), Hunt and Whiteford (in press), Roberts (1967), and J.T. Lumb (pers. comm). A revised version of the anomalies in the Central Volcanic Region is given by Whiteford (1975). Anomaly values in nanotesla (nT); contour interval 100 nT.

To test for the alternative proposition would require corroborating geophysical evidence in the form of say deep seismic sounding, to establish a correspondence with some seismic velocity boundary. In the absence of the necessary data this proposal cannot be rejected out of hand. For the Central Volcanic Region, however, where the average heat-flow of at least the eastern half of the Region is 12 times greater than normal (section 2.2.6) and where high temperatures are likely to occur at abnormally shallow depths, the determination of the base of the magnetic crust will be considered equivalent to a determination of the average depth to the Curie point isotherm.

The Curie Point for pure magnetite (Fe_3O_4), is 580°C , while for titanomagnetite - the most common magnetite-bearing mineral in igneous rocks - it is less than 570°C (Byerly & Stolt, 1977). An increase in the titanium content of titanomagnetite causes a reduction in the Curie temperature; such an increase is more common in mafic igneous rocks (Nagata, 1961). Byerly & Stolt argue for a 500°C Curie temperature for felsic plutons, and since acid volcanic rocks dominate at least the top 2 km of the Central Volcanic Region this figure will be adopted here.

6.2 Magnetic anomaly analysis in the spatial domain

It is instructive to compare the results of previous magnetic modelling within the Central Volcanic Region. Common to the studies of Modriniak & Studt (1959) and Malahoff (1968) are models for the magnetic anomaly at Waiotapu and the "Target Magnetic Anomaly" near Wairakei. Using similar magnetisation values, there is a close correspondence in the depth to the top (ZT) and horizontal extent, of the respective models, but a considerable discrepancy in the depth to the bottom (ZB). The greatest difference in ZB determinations comes from a comparison of the models for

the magnetic anomaly over the Okataina volcanic centre given by Rogan (1980) (2 km) and that given by Malahoff (1968) (12 km). The ZB irresolution problem can be ascribed to the swamping by adjacent anomalies of the broad, low-amplitude negative flanks of positive anomalies. It is these long-wavelength negative flanks which contain the crucial information pertaining to ZB.

Shuey *et al.* (1977) tried to infer depth to the bottom of magnetic sources within the Utah high plateaus from aeromagnetic data. They found that in a statistical sense, for a set of eight isolated anomalies, "the data...fit equally well for any postulated depth to the bottom."

These difficulties in estimating ZB provided the necessary incentive to turn to a statistical frequency-domain approach of analysing magnetic data from the Central Volcanic Region. The prime theoretical advantage of this approach is that signals emitted from the top and bottom of a group of source bodies, although largely superimposed on each other in the spatial domain, have differing wavelength characteristics and will therefore separate in the frequency domain.

6.3 Spectral analysis of two-dimensional arrays of spatial data

The aeromagnetic map can be considered as a description of the total magnetic anomaly field on a horizontal plane at a fixed elevation above the surface of the earth. This is called a "spatial description" of the data and is generally represented as a contoured map of the field $T(x,y)$. An alternative to the "spatial description" is a presentation of the same data in the frequency domain. For a given magnetic map digitised at intervals Δx and Δy km and extending over the range $0 \leq x \leq M-1(\Delta x)$ and $0 \leq y \leq N-1(\Delta y)$ it is possible to represent the data in the frequency domain by using the following discrete Fourier transform:

$$T(\Delta f_x, \Delta f_y) = \sum_{\Delta x=0}^{M-1} \sum_{\Delta y=0}^{N-1} T(\Delta x, \Delta y) \exp(-2\pi i (\Delta x \Delta f_x / M + \Delta y \Delta f_y / N))$$

the inverse is similar being:

6.1

$$T(\Delta x, \Delta y) = 1/MN \sum_{k=0}^{M-1} \sum_{l=0}^{N-1} T(\Delta f_x, \Delta f_y) \exp(2\pi i (\Delta x k / M + \Delta y l / N))$$

The frequencies resolved are:

$$\Delta f_x = k/M\Delta x \text{ cycles/km} \quad k = 0, \pm 1, \pm 2 \dots M/2$$

$$\Delta f_y = l/N\Delta y \text{ cycles/km} \quad l = 0, \pm 1, \pm 2 \dots N/2$$

and the series is bounded by the upper frequency limit, or Nyquist frequency, of $1/2\Delta x$ ($1/2\Delta y$) cycles/km. The frequency increments, or frequency resolution, is $1/M\Delta x$ ($1/N\Delta y$) cycles/km.

As both the map dimensions $M\Delta x$ and $N\Delta y$ are made infinite the frequency increments become infinitesimal. And in addition if the sampling intervals Δx , Δy are reduced to zero the upper frequency bounds will become infinite. In such a case the discrete Fourier transform pair (equation 6.1) reduces to the Fourier integral:

$$T(f_x, f_y) = \iint_{-\infty}^{\infty} T(x, y) \exp(-2\pi i (x f_x + y f_y)) dx dy$$

6.2

$$T(x, y) = \iint_{-\infty}^{\infty} T(f_x, f_y) \exp(2\pi i (x f_x + y f_y)) df_x df_y$$

A definition of the power or energy spectrum may be obtained from equation 6.1 as follows:

$$E(f_x, f_y) = |T(f_x, f_y)|^2$$

and the square root of this quantity is called the amplitude spectrum. The energy spectrum is a real-symmetric function and therefore only two quadrants of the frequency plane need be evaluated. Thus energy spectra

are most easily viewed in profile form by taking radial averages over 180° of f_x, f_y space and will be designated $E(f)$ where $f = (f_x^2 + f_y^2)^{1/2}$.

For a finite data set sampled at constant intervals one must use the discrete Fourier transform (equation 6.1) to obtain an estimate of the true spectrum for the data. In a comprehensive review of the spectral estimation process, Spector (1968) notes that the useable portion of the spectrum of a discrete data set will be band-width limited. Imperfect removal of regional trends from the original spatial data and finite data effects will distort the low frequency end of the spectrum while the spectral content of random noise may climb above the signal level at the high frequency end. Appendix E contains a discussion of the steps taken to minimise these effects as well as a description and listing of the fast Fourier transform computer programme used for this study.

The energy spectrum is related to the aeromagnetic map through the Fourier transformation. Therefore at each position of the frequency plane the energy density conveys information pertaining to a particular characteristic of the whole map sheet. It is because of this integration effect that analyses of spectra may be used to obtain parameter averages, such as average depth or size, of the magnetised bodies within the subsurface of the region being studied as a whole. Thus the method is particularly well suited to the study of geologically distinct areas, such as the Central Volcanic Region, where magnetic anomalies overlap, the overall pattern of anomalies is complex and it is difficult to obtain consistent results from the modelling of individual anomalies in the spatial domain.

6.4 Recovering source parameters from the spectrum

The physical basis of applying the spectrum to the resolution of source body parameters lies in the remarkable manner in which these parameters separate in the mathematical expression for the power spectrum. Fig. 6.3 illustrates this point. Here the gravity effect of an infinite horizontal cylinder and its one-dimensional frequency response are calculated. Note that the exponential decay rate of the spectrum is related to only one source body parameter - the depth to the centre of the cylinder. This then is the basis of source parameter resolution in the frequency domain; the DC level of the spectrum is a function of density or magnetisation while the shape and decay rate are solely functions of source body geometry.

Mathematical expressions for magnetic source bodies are slightly more complex than the above gravity example, yet in essence the procedure is much the same. Foremost in this field of study is the work of Bhattacharyya (1966) who gave the frequency domain expressions for simple magnetic models such as poles, dipoles and a single rectangular prism. In each case the transform separates into multiplicative factors: for a radially averaged spectrum $f = (f_x^2 + f_y^2)^{1/2}$.

$$E(f) = (4\pi^2 k^2) \cdot R_T(I_T, D_T) \cdot R_k(I_k, D_k) \cdot S(f) \cdot C(f) \cdot H(f) \quad 6.3$$

where: k = magnetic moment

R_T and R_k are functions of the inclination (I) and declination (D) of the geomagnetic field vector and magnetic moment vector respectively.

The first three factors of equation 6.3 are functions of pole strength, geomagnetic field vector and the magnetic moment factor respectively. They are not functions of frequency. Hence these three

Gravity effect of a horizontal cylinder

Fig. 6.3 Theoretical gravity effect of a horizontal cylinder (top) and its associated amplitude spectrum calculated with the fast Fourier transform described in appendix D.

factors do not affect the behaviour of the spectrum, but only have a bearing on the relative level of the spectrum. Henceforth they will be contained in one constant. The last three factors in equation 6.3 are of critical interest as these are functions of frequency:

a) $S(f)$ - shape factor

for a prism of sides $2a \times 2b$: $S(f) = [\sin 2\pi f_x a \sin 2\pi f_y b / 2\pi f_a 2\pi f_b]^2$

for a vertical cylinder of radius R : $S(f) = (J_1(2\pi fR) / \pi fR)^2$

b) $H(f)$ - depth factor = $\exp(-4\pi fZT)$ 6.4

ZT = depth to top of cylinder or prism

c) $C(f)$ - thickness factor = $(1 - \exp(-2\pi ft))^2$ 6.5

t = thickness of cylinder or prism

A modified form of equation 6.3 may now be written.

$$E(f) = \text{Const.} \cdot C(f) \cdot S(f) \cdot H(f)$$

$$\text{or } \ln E(f) = \text{Const.} + \ln C(f) + \ln S(f) + \ln H(f) \quad 6.6$$

By varying the parameters ZT , t , and R (or a and b) equation 6.6 can accommodate a wide range of geometrical forms with the behaviour of the spectrum being influenced only by the gross geometrical characteristics of the source model.

The above analysis applies to the spectrum for a single magnetic body; to analyse a map which contains many anomalies a statistical approach is required.

6.5 Statistical treatment of the magnetic map spectrum

6.5.1 Introduction

At each harmonic of the frequency plot the spectrum conveys information regarding a particular characteristic of the entire map sheet.

This is to say the information given is a statistical estimate of parameter averages of the whole sheet.

The theory of the statistical approach, termed ensemble averaging, has been discussed in detail by Spector & Grant (1970) and Nadiu (1968, 1970a) and only the salient points and derivations will be outlined here. The theory relies on sufficient data being contained within the map for the statistics of the data to be reliably sampled. For example, Spector & Grant suggest that at least five or six anomalies are required to satisfy this criterion. Another important consideration is that the statistical properties of the process do not change over space. This has been demonstrated to be a reasonable assumption when measurements of the field are restricted to one geological province (Nadiu, 1970a).

The aim of the statistical approach is to separate the ensemble source parameters from one or two ensembles of bodies giving rise to the aeromagnetic map and then interpret them. It is not intended to furnish a detailed analysis of specific anomalies.

6.5.2 Modifications to $E(f)$ for an ensemble

The analyses for various ensemble distributions defining $H(f)$, $S(f)$ and $C(f)$ have been performed by Spector & Grant (1970) for the case of the rectangular prism, and by Pedersen (1978) for a cylinder. Spector & Grant considered rectangular distributions for the various source parameters between 0 and $2\bar{a}$ where $\bar{a}$ represents the mean value of the parameter in question. Equation 6.6 when modified for an ensemble of vertical cylinders with source parameters distributed as described above is given by:

$$\ln E(f) = \text{constant} + \ln\langle H(f) \rangle + \ln\langle C(f) \rangle + \ln\langle S(f) \rangle \quad 6.7$$

where $\langle \rangle$ denotes ensemble average, and

$$\langle H(f) \rangle = \exp(-4\pi f \bar{Z}\bar{T}) \text{ for } \Delta ZT/\bar{Z}\bar{T} < 0.5$$

$$\langle C(f) \rangle = 1 - (3 - \exp(-4\pi f \bar{t})) (1 - \exp(-4\pi f \bar{t})) / 8\pi f \bar{t}$$

$$\langle S(f) \rangle = 1/2\bar{R} \int_0^{2\bar{R}} (J_1(2\pi f \bar{R}) / \pi f \bar{R})^2 dR$$

where $\bar{R}$, $\bar{t}$, $\bar{Z}\bar{T}$ are mean values of radius, thickness and depth to top of the cylinders respectively.

In fig. 6.4(a) $\text{Ln}\langle S(f) \rangle$ is evaluated for various values of $\bar{R}$. These curves will be used as corrections for the spectra. After correction for the $\text{Ln}\langle S(f) \rangle$ factor, recovery of the source parameters may follow one of two methods:

- (a) Location of the spectral peak: Possible solutions of $\bar{t}$ and $\bar{Z}\bar{T}$ that produce a peak in the spectrum for a particular value of f can be deduced iteratively from solutions of $d/df\{\text{Ln}\langle H(f) \rangle + \text{Ln}\langle C(f) \rangle\} = 0$. These are shown in fig. 6.4(b). It was noted by both Shuey *et al.* (1977) and Green (1972), and it is evident from fig. 6.4(b), that the resolution of $\bar{Z}\bar{B}$ is poor when the spectral peak occurs at frequencies less than say 0.02 cycles/km.
- (b) Linear gradient method: Spector & Grant (1970) noted that it is common for spectra of magnetic anomaly data to possess two sections of the spectrum showing different decay rates. They interpret this as being due to two ensembles of magnetic sources at distinctly different depths. For instance, consider the expression for the amplitude spectrum of a single vertical cylinder (Shuey *et al.*, 1977):

$$T(f) = \text{constant} \cdot (\exp(-2\pi f ZT) - \exp(-2\pi f ZB)).$$

If ZT and ZB are sufficiently different the above equation suggests that the top surface of the cylinder will act as a shallow ensemble of magnetic sources, the bottom surface as a deeper source and the logarithmic amplitude spectrum will be dominated by a linear decay

Solutions to $d/df [\ln\langle H(f) \rangle + \ln\langle C(f) \rangle] = 0$

proportional to ZB at low frequencies and a linear decay proportional to ZT at the higher frequency end of the spectrum.

This will also be true for an ensemble of vertical cylinders as long as $\bar{ZT}$ and $\bar{ZB}$ fill well defined and separate rectangular volumes of parameter space (Spector & Grant, 1970).

Thus the predominant emphasis in past studies has been to identify at least one or two, and in some cases three (Nadiu, 1970b), linear segments on the $\ln E(f)$ plot and interpret these gradients in terms of the depth to a source zone. Other studies, for example Pal *et al.* (1978), utilise the position of the spectral peak to obtain a thickness estimate for the ensemble distribution.

Fig. 6.5 shows three published examples of parameter determinations based on magnetic anomaly spectra. Example a), from Smith *et al.* (1974), is a radially averaged amplitude spectra of two-dimensional magnetic data from the Yellowstone caldera, Wyoming, United States. Dashed lines on the spectrum represent logarithmic slopes of $-2\pi D$ where D is the indicated depth. The low frequency peak at 0.01 cycles/km is taken to signify that the magnetic map was large enough to resolve the depth to the Curie point isotherm and the depth to the isotherm is interpreted to be 10 km.

Example b) is from Spector & Grant (1970) and is a radially averaged spectrum of two-dimensional magnetic data from Mantonipi, Canada. The spectrum is interpreted to represent a two-ensemble structure with the base of the lower ensemble not being resolved but its top surface being at a depth of about 2 km; overlying this is a shallower ensemble of magnetic source bodies with an interpreted thickness of about 0.5 km.

Example c), after Green (1976), is a simulated one-dimensional energy spectrum for a double ensemble of two-dimensional magnetised prisms. Also shown is a cross-section of the modelled prisms and the

Fig. 6.5 Three examples of published spectra as discussed in text.

calculated magnetic profile. The portion of the spectrum NMOP was obtained from the simulated profile using only the deeper ensemble. The peak in the low frequency portion of the spectrum, due to the interaction of the depth and thickness terms, is clearly seen. An average depth of 5.4 km is obtained from the section of the spectrum MOP which compares favourably with the true average depth to the top surface for the deeper ensemble of 5.6 km. The portion of the spectrum NMOQ was obtained by using both the deep and shallow ensembles in the simulation. The portion marked NMP is not affected by the introduction of the shallow ensemble and the section MO yields a depth estimate of 6.2 km (true value 5.6 km). The linear trend OQ yields an average depth of 0.54 km which compares well with the true average depth of 0.6 km for the shallow ensemble.

6.6 Central Volcanic Region data

A 125×125 km area of the Central Volcanic Region (fig. 6.1) was digitised at a grid interval of 2.5 km giving a 50×50 data array, and then processed as described in appendix E. Pre-transformation treatment of the data involved detrending to remove any spurious long wavelength effects, windowing of the data in order to reduce the variance of the spectral estimates, extrapolation of a 10 grid-unit wide taper around the edge of the data so as to minimise sharp edge effects and finally, padding of the data with zeros out to a 128×128 array. The padding of the data with zeros is necessary for two reasons. Firstly, for purposes of carrying out filtering in the frequency domain (see section 6.7) it is desirable to have a padded array of spatial data - this aspect is elaborated on in appendix E - and secondly, the fast Fourier transform algorithm used in this study required the dimension of

the data array to be a power of two. In order to meet these two requirements it was necessary to pad the 60×60 extrapolated data array out to a 128×128 array.

6.6.1 Detrending and finite map size effects

Fig. 6.6 shows three radially-averaged spectra a, b, c which have, prior to transformation, been detrended with polynomials of order 0, 1, and 2 respectively. For harmonics greater than 10 (0.0313 cycles/km) there is little marked difference in the spectra, yet the low frequency peak at the second harmonic in (a) moves to higher harmonics in spectra (b) and (c). These low frequency peaks occur at frequencies ranging from 0.00625 to 0.01875 cycles/km, which correspond to wavelengths of 160 to 53 km; wavelengths which exceed that of the individual anomalies shown on the magnetic anomaly map of the Central Volcanic Region (fig. 6.1).

The low-frequency peak is ascribed to a finite data length effect, or more specifically to the zero-padding procedure. A typical row or column of pre-transformed data consists of 34 zeros, five extrapolated and tapered data points, a string of 50 data points, five extrapolated values then a further 34 zeros. Removing the mean of any row or column results in the lowest frequency present being equal to the reciprocal of the central 50-60 point data string; that is, a frequency of $1/50$ to $1/60$ cycles per grid unit or 0.008 to 0.0167 cycles per km. As higher order detrending polynomials are applied to the data the low-frequency content is progressively removed resulting in a migration of the low-frequency spectral peak to higher frequencies. It is difficult to make an objective decision regarding which polynomial represents the true

Fig. 6.6 Radially averaged spectra for the Central Volcanic Region data. Spectra a), b) and c) represent spectra with pre-transformation detrending with zero, first and second order polynomials, respectively.

anomalous background or regional effect, and instead all three spectra will be considered in the following analysis.

The interpretable portion of the spectrum is also limited at the high frequency end where the contribution of random noise may rise above the signal level. The procedure adopted here follows that of Spector (1968) and only considers that part of the spectrum where the energy level is $\geq 1\%$ of the maximum level. Thus the band-width of interest in the spectra of fig. 6.6 runs from the 10th to the 35th or 40th harmonic (0.0281 to 0.125 cycles/km).

6.6.2 Spectral models

All three spectra in fig. 6.6 have, in the band-width of interest, three features in common.

1. A spectral peak in the frequency range 0.0313 to 0.0375 cycles/km (10th to 12th harmonic);
2. A roll off from the peak followed by a linear gradient between the 14th and 16th harmonics. Preliminary depth estimates from these gradients range from 9.5 to 8.6 km.
3. A linear decay between the 16th and 40th harmonic furnishing preliminary depth estimates of 1.8 to 2.0 km. These depth estimates are preliminary as the spectra have not been corrected for the size effect.

The presence of two linear segments in each spectra suggests two dominant levels, or two depth ensembles, of magnetic sources; however the spectral overlap from the shallow ensemble is such that the linear segment pertaining to the deeper ensemble is only defined by three harmonics, and is therefore poorly resolved by this method. The existence of peaks at the low frequency end of the three spectra suggests that the

source bodies are of a finite thickness (Spector & Grant, 1970). Hence, the position of this peak together with an estimate for $\bar{ZT}$ from the high frequency end of the spectrum will be used in Section 6.6.4 to gain an estimate of $\bar{ZB}$ using fig. 6.4.

The model proposed is that of an ensemble of vertical cylinders; their tops at or near the ground surface behaving as positive poles and their bases at a depth $\bar{ZB}$ acting as negative poles. There is good geological evidence to suppose that an additional source of high-frequency energy is contributed by magnetised surface topography in the form of rhyolite domes. These sources would correspond to a series of narrow poles or shallow dipoles.

6.6.3 Size correction factor

In order to obtain an estimate for $\bar{R}$ (mean radius of ensemble) the horizontal half-widths of Malahoff's model solutions were analysed. For 130 half-widths measured the mean and standard deviation were calculated to be 5.3 and 3.2 km respectively. Malahoff only modelled the more prominent anomalies so that $\bar{R} = 6$ km was adopted for the deeper ensemble, and in view of the observation made previously that some of the high-frequency energy possibly emanates from topographic features, $\bar{R} = 2$ km was adopted as the preferred value for the shallow ensemble.

Corrections for the size factor were made by selecting the appropriate curve in fig. 6.4(a) then evaluating the mean gradient within the frequency range of interest. Division of this value by -4π gives a depth correction to be subtracted from the preliminary depth estimates in fig. 6.6. Although the effect of this correction on the preliminary depth estimates is appreciable it was found that for a wide range of values of $\bar{R}$ the spectral peak remained stable at the 12th harmonic.

6.6.4 Parameter estimates

Parameter estimates were made by both the spectral peak and the linear gradient methods as outlined in section 6.5.2. Common to both methods is the necessity for an initial estimate of $\bar{Z}\bar{T}$ from the high-frequency end of the spectrum. After correction for the size factor with $\bar{R} = 2$ km the average value of $\bar{Z}\bar{T}$ from the three spectra in fig. 6.6 was estimated to be 1.5 km. This appears a reasonable value if one considers that the mean ground clearance of the flight lines was 1.2 km. Results of both the spectral peak and the linear gradient analyses are presented in fig. 6.7. For a spectral peak at the 12th harmonic and a value of $\bar{Z}\bar{T} = 1.5$ km, $\bar{t}$ is estimated to be 6.1 km (fig. 6.7(a)). Surrounding this estimate is a parallelogram of uncertainty based on a ± 1 harmonic uncertainty in the spectral peak position and a range of values for $\bar{Z}\bar{T}$ from 1.4 to 2.0 km. Thus the preferred solution from the analysis of the spectral peak position is $\bar{t} = 6.1 \pm 2$ km, or, taking into account that $t = ZB - ZT$ (see fig. 6.4) and the average flight line clearance of about 1.2 km, the Curie depth estimate is $6.1 + (1.5 - 1.2) = 6.4 \pm 2$ km.

Estimates of $\bar{Z}\bar{B}$ based on the linear gradient method are given in fig. 6.7(b). The vertical bars illustrate the variation in $\bar{Z}\bar{B}$ when the size correction is made for extreme values of $\bar{R}$. For example the top point on the vertical bar represents a size correction for $\bar{R}=0$, the mid-point for $\bar{R}=6$ km and the lowest point for $\bar{R}=12$ km. The preferred value of $\bar{R}=6$ km furnishes an average estimate of $\bar{Z}\bar{B}=7.2$ km, which when corrected for the flight line clearance, gives a Curie depth of 6.0 km beneath ground level.

Both methods of determining $\bar{Z}\bar{B}$ suffer from uncertainty. The spectral peak method is sensitive to the value of $\bar{Z}\bar{T}$, particularly when $\bar{Z}\bar{T}$ approaches say half the value of $\bar{Z}\bar{B}$, whereas the linear gradient method suffers from there being only three points available to determine

the linear gradient pertaining to $\bar{Z}\bar{B}$. Moreover, the padding of the pre-transformed spatial data with zeros introduced an artificial long-wavelength trend to the data resulting in a low frequency peak appearing in the derived spectra; the long wavelength trend may also have introduced spurious frequency components to other parts of the spectra (Black & Scollar, 1969). However, some confidence for the validity of the model and depth estimates is gained from the similar values that are obtained for $\bar{Z}\bar{B}$ by using two essentially independent methods.

The average of the two Curie depth estimates is 6.2 km and the uncertainty in this determination, based on the uncertainty in the position of the spectral peak, is estimated as ± 2 km. Although this Curie depth estimate is considered to be that for the Central Volcanic Region, it should be noted that the area that was subjected to digitisation (fig. 6.1) did not include the whole of the Region, and the magnetic anomalies that were included in the analysis are dominated by those associated with the rhyolitic volcanic centres of the Taupo Volcanic Zone. Therefore this Curie depth estimate may be weighted more to reflect the subsurface thermal conditions pertaining to the Taupo Volcanic Zone rather than that for the Central Volcanic Region as a whole.

6.7 Filtering in the frequency domain

The application of conventional filters (continuation, derivatives and band-pass filters) and more specialised filters (Sydberg, 1972) is more accurate and convenient in the frequency domain (Black & Schollar, 1969). In order to bring out the dominant trends in the data of fig. 6.1 low-pass filters were designed then applied to the transformed magnetic data. The two low-pass filters used in this study were the upward

continuation and pseudo-gravity filters described in appendix E and the resulting filtered maps are shown in fig. 6.8.

The upward continuation filter purports to represent the data as though it were viewed from a higher plane, whereas the pseudo-gravity filter integrates the magnetic anomaly field, then multiplies by a constant factor to give the filtered anomaly values in gravity units. A pseudo-gravity anomaly field similar to that of fig. 6.8 was derived then used by Rogan (1980) as a constraint in the interpretation of residual gravity anomalies of the Taupo Volcanic Zone. The necessary assumption required for the application of pseudo-gravity anomalies to the interpretation of residual gravity anomalies, is that there is a common geological source for the magnetic and residual gravity anomalies, and further, that the ratio of density contrast to magnetisation contrast is constant throughout the source body.

No quantitative analysis of the filtered magnetic maps (fig. 6.8) will be made for this study. However, a qualitative comparison of the filtered magnetic maps with the residual gravity anomalies of fig. 3.1 and the geological map of fig. 2.9 reveals a close correspondence between the positive magnetic anomalies, some of the negative residual gravity anomalies and the rhyolitic volcanic centres of the Taupo Volcanic Zone.

6.8 Lithospheric thickness, elastic thickness and the Curie depth

While the Curie depth is an important geophysical parameter in its own right, it also has an indirect, yet important bearing on gravity models for the Central Volcanic Region.

The term "lithosphere" has received various definitions based on seismic, thermal, compositional and mechanical properties. From a historical viewpoint, the first definition was a mechanical one.

Fig. 6.8 Low-pass filtered maps of the magnetic data from the Central Volcanic Region. The pseudo-gravity map was derived for (density/magnetisation) ratio of $(-0.4 \times 10^3 \text{ kg/m}^3) / 2.0 \text{ A/m}$.

Barrell (1914a,b) initially coined the terms lithosphere and asthenosphere when he summarised the evidence for a strong outer layer overlying a weak layer in the earth, and for this study the term lithosphere will also be used in a mechanical sense to mean that region of the crust and upper mantle with long term strength; underlying the lithosphere is the asthenosphere which is assumed to possess the properties of an incompressible fluid on geological time scales. Recently there has been a tendency to use the terms thermal, seismic or elastic thickness of the lithosphere, depending on how the thickness is determined (Hanks & Raleigh, 1980). These three thicknesses are not the same.

The thermal thickness of the lithosphere is based on the depth to the isotherm that corresponds to the solidus for mantle rocks; generally the 1200-1300°C isotherm and depths to this isotherm are proposed by Turcotte (1979) to be 100 km for oceanic basins and 200 km for stable continental areas. The seismic thickness of the lithosphere is defined by surface wave studies, and these studies point to a seismic thickness being approximately equivalent to the thermal thickness of the lithosphere (Kanamori & Press, 1970). The elastic thickness of the lithosphere is determined by studying the observed flexure of the lithosphere produced by such surface loads as mountains, seamounts and isolated loads that cause sedimentary basins. By applying a mathematical model pertaining to the bending of thin elastic plates that overlie a weak fluid, an effective elastic thickness can be found (e.g. Walcott, 1970; Watts & Talwani, 1974).

Some examples of elastic thickness determinations are:

- a) 110 km for stable continental lithosphere (Walcott, 1970).
- b) 30 km for 80 My old oceanic lithosphere (Watts *et al.*, 1980).

c) 20 km for the tectonically active Basin & Range province
(Walcott, 1970).

d) 2-13 km (average 6 km) for the mid-oceanic ridges (Cochran, 1979).

Watts *et al.* (1980) propose from the basis of loading studies that the elastic thickness of oceanic lithosphere is two to three times smaller than the seismic thickness as determined by surface wave studies. Furthermore, by applying theoretical, age-dependent cooling models of oceanic lithosphere to their study they find a correlation between elastic thickness and temperature. More specifically, they find an association between the lower boundary of the elastic portion of the lithosphere and the depth to the $500 \pm 150^\circ\text{C}$ isotherm. Similar studies by other workers are reported by Turcotte (1979) to find that this lower boundary is defined by the $600 \pm 100^\circ\text{C}$ isotherm. The $500\text{--}600^\circ\text{C}$ temperature range spans the range of the Curie point for most crustal rocks (section 6.1) and hence suggests that a determination of the average Curie depth for some region is tantamount to a determination of the elastic thickness of the lithosphere for that region.

It is therefore proposed that the base of the elastic portion of the lithosphere for the Central Volcanic Region coincides with the average depth of the Curie point isotherm at a depth of about 6 km. For the northwestern North Island - that is, the area that takes in the regions of Northland, Waikato and northern Taranaki (fig. 1.2), no Curie depth analysis (or loading studies) have been performed. However, Pandey (1981b), from an analysis of heat flow and radiogenic heat production data, has calculated temperature profiles for the Taranaki and Waikato-Northland regions (fig. 6.9). From these profiles it can be observed that the 500°C isotherm occurs at a depth of about 20 km; this will be adopted as the elastic thickness of the lithosphere for the northwestern North Island.

Fig. 6.9 Temperature-depth profiles for the Northland-Waikato and Taranaki regions of the North Island as given by Pandey (1981b). The shaded areas represent the melting ranges of Basalt-eclogite. The association of the Curie point and elastic thickness of the lithosphere assumes a temperature range of 500-600°C for both (see Text).

A useful comparison may be drawn between the northwestern North Island and the Basin and Range Province of the Western United States. Firstly, heat flow for both these regions averages about 70-90 mW/m² (Pandey, 1981b; Pollack & Chapman, 1977). Secondly, the elastic thickness of the lithosphere for both regions is estimated to be about 20 km. For the northwestern North Island the thickness was estimated above from the depth of the 500°C isotherm and for the Basin and Range it has been estimated by Walcott (1970) from an analysis of the isostatic uplift of Lake Bonneville. These similarities in elastic thickness, and heat flow are consistent with the proposal of Watts *et al.* (1980) that the elastic thickness is a function of temperature.

The manner in which the lithosphere will deform under a load of given amplitude and wavelength is dependent upon the flexural rigidity of the elastic layer of the lithosphere; the flexural rigidity is in turn a function of the elastic thickness raised to the third power (Walcott, 1970). More specifically, for a given load and elastic thickness, it can be determined whether isostatic compensation for the load will tend to be regional or local in character (Gunn, 1943; McNutt, 1980). Thus the estimates of elastic thickness made here will be utilised in the following chapter where a regional gravity interpretation of the central North Island will be presented.

6.9 Conclusions

The spectral method has provided a means by which a statistical estimate for the mean depth extent of magnetic source bodies within the Central Volcanic Region can be obtained. The determination of $\bar{Z}_B$ in the frequency domain depends on two variables less than the equivalent

problem in the spatial domain. Those variables which have no influence on the shape of the radially averaged spectrum are the intensity and direction of magnetisation. Although some uncertainty is inherent in the depth estimates made here they are a substantial improvement over those based on analyses of single anomalies in the spatial domain.

The three principal points to be noted from this analysis are:

- 1) The estimated Curie point isotherm level of 6.2 km is unusually shallow; only at Yellowstone (U.S.A.) has a similarly shallow Curie depth been interpreted (Bhattacharyya & Leu, 1975).
- 2) There is nothing in the spectra to suggest a magnetisation contrast at a depth of 2-3 km. Seismic refraction results, presented in the last chapter, indicate a seismic velocity contrast at a depth of 2.0 ± 0.2 km at three locations within the Central Volcanic Region. If this seismic velocity boundary represents a magnetisation boundary a peak in the spectra within the vicinity of the 21st harmonic (0.066 cycles/km) would be expected (i.e. refer to fig. 6.4(b) for a model of $\bar{Z}\bar{T}=1.5$ and $\bar{t}=2$ km).
- 3) The Curie depth determination has an indirect application to gravity interpretations in that the Curie temperature for crustal rocks is approximately equal to the value of the isotherm that defines the base of the elastic portion of the lithosphere.

CHAPTER 7 A REGIONAL GRAVITY INTERPRETATION OF THE
CENTRAL NORTH ISLAND

Summary

The North Island presents a favourable opportunity to study the gravity effect of a subducted lithospheric plate. Although the seismically defined Benioff zone continues south to the north end of the South Island it is not accompanied throughout its length by a back-arc gravity high; the back-arc high terminates at the southern extremity of the Central Volcanic Region. Furthermore strong gravity trends are developed at angles to the strike of the subducted plate suggesting that the subducted plate may not have an overriding influence on the gravity anomaly field of the central North Island. An attempt is made here to model the major gravity anomalies of the central North Island with a superposition of gravity effects from:

- (i) A boundary joining Mt Egmont, Mt Ruapehu and Opotiki which divides thin crust and low-density mantle to the north and northwest from a normal crustal thickness and mantle density to the south and southeast;
- (ii) a thinner crust and a lower-density mantle beneath the Central Volcanic Region; and
- (iii) the subducted Pacific Plate.

The model cannot be considered definitive as it lacks control from long-range seismic refraction data. It is, however, an attempt to explain the complete gravity anomaly field of the Central Volcanic Region and to provide a structural model for the central North Island which may be tested by future work.

7.1 Introduction

In Chapter 5 it was found that at both Mangakino and Taupo the estimated gravity effect for the seismically determined thickness of low-velocity volcanic rocks is less than the observed residual gravity anomaly. Trial interpretation models were presented which explained these secondary residual gravity anomalies as being due to large volumes ($\approx 700 \text{ km}^3$) of molten rhyolite within the seismic basement (see Chap. 5). Such large volumes of molten rhyolite, although possible, were not favoured. Instead it was proposed that part of the secondary negative residuals may come from the basement rocks themselves.

In order to pursue this argument it is necessary to present a plausible "deep-structure" gravity model for the Central Volcanic Region and compare the calculated gravity effect of the model against the total observed gravity anomaly field. Due to the overlapping gravity effects of deeper density irregularities, the structure of the Central Volcanic Region cannot be treated in isolation. The anomalous gravity effect of the subducted Pacific plate and other structural inhomogeneities within the central North Island must be considered simultaneously with the anomalous gravity effect of the Central Volcanic Region.

The isostatic gravity anomaly map of the North Island possesses some strong gravity trends (fig. 7.1(a)) that do not parallel the strike of the subducted plate. Furthermore, it was pointed out in section 2.1.2 that the back-arc gravity high over the northwestern North Island terminates at about 39° S whereas subduction of the Pacific plate persists to about 41° S . There is not, therefore, a clear association between the North Island back-arc gravity high and the present position of the subducted Pacific plate.

The procedure to be followed in this modelling exercise is first to explain the gravity trends that do not parallel the strike of the subducted plate, in terms of crustal thickness and mantle density variations, and secondly, to assign the remaining gravity anomalies to the gravity effect of the subducted plate. Deep seismic sounding is generally required as a control on estimates of crustal thickness before a regional interpretation can be performed. Only a limited amount of such data are available for the North Island, so that the main constraint on the model will be the principle of isostasy. Implicit in the adoption of this constraint is that mass anomalies within the lithosphere of the North Island are not supported by subduction-induced dynamic flow in the underlying asthenosphere. Thus the model to be presented in this chapter is a first-order static one that is kept simple enough to be tested by future geophysical work.

7.2 Isostasy and isostatic gravity anomalies

7.2.1 Definitions

The definition of isostasy used here is taken from Heiskanen and Vening Meinesz (1958, p.133) and reads "...that at a depth of compensation D hydrostatic equilibrium prevails, so that every area unit at the depth D is under the same pressure regardless of whether the unit is under mountain, low-land or ocean". This definition encompasses both local and regional compensation. Topographic loads are said to be locally compensated if the total mass in any vertical column above the compensation depth is constant. In this scheme elevated regions are pointwise compensated whereas a regional mechanism distributes compensation laterally around the feature as well as vertically beneath it. The discussion of isostatic compensation need not be restricted to that for topographic loads alone. Loads exist

7.1a Isostatic gravity anomalies at 1 km above sea level (after Reilly *et al.*, 1977). Black triangles represent recently active andesitic and dacitic volcanoes.

Fig. 7.1b Two Bouguer and isostatic gravity anomaly profiles across the North Island (after Reilly, 1978). Note that, apart from regions near the trench, the Bouguer and isostatic gravity anomalies are broadly similar.

in the form of both positive and negative lithospheric mass anomalies; these loads will also be compensated either locally or regionally depending on the wavelength of the load as compared to the elastic thickness of the lithosphere.

7.2.2 Elastic thickness and isostatic compensation

The elastic thickness of the lithosphere refers to that part which possesses long-term strength (see section 6.8). The elastic thickness is an important parameter for gravity studies, as lithospheric loads whose wavelengths are long in relation to the elastic thickness will, except near the edges, be locally compensated. Loads with wavelengths less than the elastic thickness will be supported by stresses within the elastic layer and in general there will be a departure from local isostasy (McNutt, 1980; Beaumont, 1978; Artemjev & Artyushkov, 1971). A more quantitative account of the relationship between the degree of isostatic compensation for a given load on an elastic plate, the wavelength of the load, and the thickness of the plate is given by Gunn (1943).

This relationship between wavelength of load and elastic thickness can be applied, in a qualitative manner, to two regions of the North Island.

(i) The Central Volcanic Region: The seismic refraction interpretations of chapter 5 show that at three localities within the Region the thickness of low-velocity, and therefore low-density, volcanic rocks is about 2 km. Gravity profile interpretations presented in the next chapter will indicate that this thickness of low-density volcanic rocks covers much of the Region apart from localised areas, such as the greywacke outcrop at Otamarakau and the andesite outcrop near Te Puke, where higher density rocks come near or to the surface. For the most part, the width of the Central Volcanic Region is many times greater than

the 6 km elastic thickness of the lithosphere for the Region as estimated in section 6.8. Hence it will be assumed that the low-density volcanic rocks that are found within the top 2 km of the Region are locally compensated with respect to the structure found on each side of the Region.

(ii) The northwestern North Island: A large positive gravity anomaly can be observed over the northwestern North Island (fig. 7.1(a)). This is the back-arc gravity high discussed in Chapter 2, and for reasons enumerated in that chapter it is considered unlikely that this positive gravity anomaly can be entirely due to an excess mass within the subducted plate. An alternative explanation for the back-arc gravity high will be sought.

The back-arc gravity high over the northwestern North Island has a wavelength of about 100 km yet the elastic thickness of the lithosphere for this region is estimated to be about 20 km (section 6.8). It is therefore considered doubtful that the gravity high can be explained by an uncompensated lithospheric load; the width of the required load would be such that stress-differences equal to about 0.6 times the amplitude of the load will be present at depths considerably greater than 20 km (see Jeffreys, 1976; p.267). For instance, if it were proposed that the gravity high was due to a load in the form of a thinner (or more dense) crust for the northwestern North Island than elsewhere, then consideration would have to be given to how the thinner, or more dense, crust is compensated.

7.2.3 Gravity "edge-effects"

For a region with thin crust that is locally compensated by say a low-density mantle, the integrated gravity effect of the thin crust and

low-density mantle will be zero, the net gravity effect at the centre of the region will approach zero but around the boundaries of the region large dipolar gravity anomalies will be observed. Following Sibuet & Le Pichon (1974), these will be called edge-effect gravity anomalies.

Fig. 7.2 shows a gravity edge-effect that arises from the juxtaposition of two lithospheric blocks that are equal in mass but differ in their respective mass distributions. The block on the left represents a normal continental crust situation whereas the block on the right represents a crust thinned to 22.5 km which is pointwise, or locally, compensated by an anomalous low-density mantle; the density of the anomalous mantle has been adjusted so that compensation is achieved at a depth of 100 km. Stresses are everywhere hydrostatic at and below 100 km, the distance between peaks of the dipolar gravity anomaly is about 100 km and the anomaly will clearly integrate to zero. Note that the amplitude and wavelength of this edge-effect gravity anomaly are dependent on the amount of crustal thinning and depth of compensation respectively.

A three dimensional form of the model shown in fig. 7.2 will eventually be used to model the sharp gravity gradient across a boundary between Mt Egmont and Mt Ruapehu.

7.2.4 Isostatic gravity anomalies of the North Island

The main point to note from the model in fig. 7.2 is that the existence of non-zero isostatic gravity anomalies over some region does not necessarily imply departures from isostasy, as defined in section 7.2.1, but merely departures from the Airy-Heiskanen model. The Airy-Heiskanen model, such as that used in the construction of the isostatic gravity anomaly map of fig. 7.1(a), only makes corrections for

Fig. 7.2 Diagram showing the principle of an edge-effect gravity anomaly as discussed in the text. Densities and density contrasts in Mg/m^3 and the compensation depth is taken to be 100 km. Note that the density of the anomalous mantle on the right must be taken to four decimal places in the modelling process in order to get a symmetrical edge-effect.

topographic loads and their assumed compensating masses. For a region that is in isostatic equilibrium isostatic gravity anomalies can arise from, i) gravity edge effects associated with subsurface, compensated loads (fig. 7.2), or ii) anomalies of density of sufficiently small width that all stress differences are confined to be above the depth of compensation (Jeffreys, 1976; p.276).

For broad regions one, admittedly non-unique, indicator of isostatic equilibrium is the condition:

$$\iint_s \Delta g ds = \text{constant} \quad (\text{Golizdra, 1978b}) \quad \dots 7.1$$

where Δg = free air or isostatic gravity anomaly, s is a suitably large enough area of integration, with respect to the depth of compensation, and the constant is the background gravity anomaly field for the region as given by say a low-order, spherical harmonic representation of the earth's gravity anomaly field; these low-order or long wavelength gravity anomalies can be taken as originating at depths far greater than the base of the lithosphere (McKenzie, 1967; Lambeck, 1972).

The North Island isostatic gravity anomaly map (fig. 7.1(a)) exhibits approximately equal areas of positive and negative anomalies and a statistical analysis of the whole map (fig. 7.3) shows that the mean value of the isostatic gravity anomaly field, averaged over 20 x 20 km squares, is 6 $\mu\text{N/kg}$. By way of comparison the low-order, spherical harmonic representations for the gravity anomaly field of the earth (e.g. Jeffreys, 1976, p.242) show that the background field over the North Island lies between 0 and 100 $\mu\text{N/kg}$; this is in approximate agreement with the mean value of the isostatic anomaly over the North Island (fig. 7.3) and therefore a possible model to explain the North Island gravity anomaly field is one that conforms to the condition:

Fig. 7.3 Histogram of averaged 20×20 km squares for the isostatic gravity anomaly over the North Island.

$\sum_{i=1}^N M_i = 0$ where M_i = lithospheric mass anomaly associated with either the subducted Pacific plate or the overriding Indian plate.

7.2.5 Modelling procedure

In studying regional structures from gravity data it is important to note that gravity data on its own is insufficient to establish the subsurface structure. Gravity data can be used to 1) give indications, from a study of gravity gradients, of a maximum depth to which lateral density variations take place, and 2) permit the extrapolation and interpolation of seismically deduced structures to unknown areas. The only published estimate for crustal thickness within the North Island, as deduced by seismic refraction methods, is that for the Wellington profile (Garrick, 1968). His estimate, which shows the crust to be about 35 km thick, is based on an unreversed profile and the arrivals which were interpreted as refractions from the mantle were all second arrivals. However, these are the only data available and this crustal thickness estimate, combined with the constraint of isostatic balance for the North Island as a whole, will be the starting point for the interpretation. Because these constraints are insufficient to lead to a unique interpretation the model will be kept simple; in particular a single value of 2.75 Mg/m^3 will be used for the density of the crust, whether it be upper or lower continental crust or oceanic crust. Crustal densities may in reality vary from 2.60 to 3.0 Mg/m^3 but these complexities will not alter the main results of this model; as more information becomes available crustal densities may be varied, or intermediate layers introduced, and accommodated by a corresponding adjustment of the crustal boundaries.

Fig. 7.1(b) shows two Bouguer and isostatic gravity anomaly profiles across the central North Island. Apart from differences where the smoothed topography departs from 500 m of sea level, the broad scale features of the Bouguer and isostatic gravity anomalies are similar. However, in section 2.1.2 (figs 2.2 and 2.3) it was noted that in two-dimensions, or plan view, the isostatic gravity anomalies tend to highlight several of the geological and geophysical boundaries of the central North Island, and in general present a clearer gravity anomaly pattern than do the Bouguer anomalies. Thus for discussion purposes the principal gravity anomalies of the North Island will be considered to be the isostatic gravity anomalies shown in fig. 7.1(a). Moreover, because of the simple anomaly pattern afforded by the isostatic anomalies, as opposed to that of the Bouguer anomalies, it was found that the three-dimensional modelling process is eased considerably if models are initially constructed whose gravity effect matches that of the isostatic anomalies. However, these models can only be considered as first-order ones, because the isostatic correction has already removed the gravity effect of an assumed compensation mechanism for the topography of the North Island. In order to obtain an interpretational model which contains the true amount of anomalous mass beneath sealevel, the Bouguer anomalies must be modelled.

At the end of this chapter a Bouguer gravity anomaly profile across the North Island will be modelled which will demonstrate that the differences between structural models for the central portion of the North Island, derived by modelling either the Bouguer or isostatic gravity anomalies, are minor.

The gravity model will attempt to explain the observed gravity anomalies of the North Island by the anomalous gravity effect of the subducted

Pacific plate and by gravity edge-effects associated with the following two boundaries:

1) A boundary which follows the path of the zero isostatic gravity anomaly between Mt Egmont, Mt Ruapehu and Opotiki; to the north and northwest of this boundary the crust is proposed to be thinner and the mantle less dense than elsewhere.

2) A "vee-shaped" boundary that corresponds to that of the Central Volcanic Region. Within the vee the crust is proposed to be thinner and the upper mantle less dense than outside of the vee.

For mass anomalies associated with the subducted Pacific plate a form of distributed, or regional, compensation will be assumed whereas for mass anomalies within the overriding Indian plate local isostatic compensation will be assumed. These three structures, that is the Mt Egmont-Mt Ruapehu-Opotiki boundary, the Central Volcanic Region boundary and the subducted plate will be modelled separately then their associated gravity effects superimposed and compared with the observed isostatic gravity anomalies.

7.3 Egmont-Ruapehu-Opotiki Boundary

7.3.1 Introduction

In an analysis of manifestations of upper mantle inhomogeneities of the North Island, Hatherton (1970) first drew attention to the association of volcanism, heat-flow, and gravity anomalies with the seismic attenuating/transmitting boundary described by Mooney (1970). This boundary closely parallels the zero isostatic gravity anomaly contour (Fig. 7.1a) and the path of this contour is here termed the Egmont-Ruapehu-Opotiki boundary. Fig. 7.4 summarises the geophysical properties defining the boundary. To the north and northwest heatflow is high, gravity anomalies are positive and volcanism, both

Fig. 7.4 Diagram summarising various geophysical parameters discussed in the text. Heat flow values from Pandey (1981a). Note the association of positive gravity anomalies with areas of volcanism and high heat flow.

past and present, is abundant. The only parameters that remain unchanged across the boundary are isobaths of deep seismic activity (fig. 7.4) and the Stokes Magnetic Anomaly (fig. 6.2) which both cross the Egmont-Ruapehu portion of the boundary. Temperature profiles (fig. 6.9) of Pandey (1981b) indicate that the Curie point for crustal rocks is reached at a depth of about 15 km beneath Taranaki, suggesting that the source of the Stokes Magnetic Anomaly must be shallower than 15 km. Hence there appears to be geological continuity across the Egmont-Ruapehu boundary at depths less than 15 km (Stokes Magnetic Anomaly) and greater than 100 km (the subducted plate); accordingly it is proposed that the source of the intense gravity gradient across this boundary lies in the depth range 15-100 km.

Not shown in fig. 7.4 are the upper-mantle compressional velocity (P_n) determinations of Haines (1979). Anomalously low values are found beneath much of the northwestern North Island and markedly low values (7.4 km/s) are detected beneath an area which includes much of the Central Volcanic Region. Haines suggests that the very low P_n velocities could be due to a partially molten mantle.

It is difficult to place any rigorous constraints on a gravity interpretation from the above associations. The data summarised in fig. 7.4 is, however, taken as qualitative support for the proposition that the upper mantle is both hotter and less dense to the north and northwest of the Egmont-Ruapehu-Opotiki boundary than elsewhere.

7.3.2 Gravity model of the Egmont-Ruapehu-Opotiki boundary

A simple locally compensated model that will produce the required gravity relief across the Egmont-Ruapehu-Opotiki boundary was discussed in section 7.2.3 and is shown in fig. 7.2. The crust-mantle section to the

south is taken as standard; the crust is modelled to be 35 km thick, with a mean density of 2.75 Mg/m^3 , and the crust is underlain by a mantle of density 3.4 Mg/m^3 . To the north the crust is thinned to 22.5 km and the mantle density lowered accordingly so that compensation is achieved at a depth of 100 km. A similar gravity edge-effect to that of fig. 7.2 could also be simulated by holding the crustal thicknesses the same across this boundary but postulating a denser crust to the north. For example, trial models indicate that if a 35 km crust to the north of the boundary was 0.2 Mg/m^3 more dense than that to the south, and the denser crust is compensated by a low density mantle similar to that shown in fig. 7.2, much the same edge-effect gravity anomaly would be produced. Alternatively, some combination of say a crust 6 km thinner and 0.1 Mg/m^3 more dense to the north, and compensated by a low density mantle, would also give the desired edge-effect gravity anomaly.

Fig. 7.2 displays the edge-effect anomaly on an enlarged scale for a north-south profile through Mt Egmont. Although there is a reasonable match of gradients with the observed anomaly the model can only be considered an approximation to reality as no allowance has been made for the sediments along the profile, the gravity effect of the subducted plate or the overall three-dimensional character of the Egmont-Ruapehu-Opotiki boundary.

The three-dimensional gravity effect for the edge of fig. 7.2 placed along the Egmont-Ruapehu-Opotiki boundary is shown in fig. 7.5; a "corner-effect" is created producing a local gravity high over Lake Taupo.

7.4 Central Volcanic Region density model

The density model to represent the Central Volcanic Region will consist of the following three components:

Fig. 7.5 Three-dimensional gravity effect of the model in fig. 7.2 when the vertical edge, that separates normal continental crust and mantle from thinned crust and low-density mantle, is placed along the Egmont-Ruapehu-Opotiki boundary (ie. ABC in figure). Gravity effect in $\mu\text{N/kg}$.

a) Near surface: the seismic refraction data presented in chapter 5 show that at three localities within the Central Volcanic Region there is about 2 km of low-velocity volcanic rocks. The residual gravity effect of these rocks is equivalent to a 2 km layer with an average density contrast of -0.4 Mg/m^3 , with respect to the greywacke rocks that are found either side of the Region. For the purposes of a regional interpretation this low-velocity, low-density layer will be assumed to be present right over the Region.

b) Upper crust: beneath 2 km the rock type and density structure for the crust of the Central Volcanic Region are unknown. However in order to obtain a coherent explanation for the secondary negative residual anomalies detected at Mangakino and Taupo it will be proposed that the upper crustal rocks in the depth range of 2-10 km are less dense than those crustal rocks, at an equivalent depth, that flank the Region.

c) Upper mantle: the most definitive geophysical data that bears upon the upper mantle of the Central Volcanic Region are the Pn velocity determinations of Haines (1979) (see chapter 2). The Pn velocity for the Region is $7.4 \pm 0.1 \text{ km/s}$, which is anomalously low. By adopting the velocity-density relationships of Woollard (1959) for rocks at a depth of 8 km or greater, a 7.4 km/s velocity is associated with a density of about 3.1 Mg/m^3 , whereas rocks with a normal mantle velocity of say 8.2 km/s are associated with a density of about 3.4 Mg/m^3 . Deviations from curves such as that presented by Woollard (1959) may occur, but any reasonable variations from his curve will not alter the general results of this modelling exercise. Rocks below the 7.4 km/s boundary will be denoted as the "upper mantle" for the Central Volcanic Region whereas all rocks above this boundary will be referred to as the "crust".

Fig. 7.6 Diagram showing the density model and associated gravity effect used to represent the Central Volcanic Region; in two-dimensions (at top) and three-dimensions (bottom). Compensation at 22.5 km, densities in Mg/m^3 and gravity effect in $\mu\text{N/kg}$.

The principal constraint for the gravity model of the Central Volcanic Region is that it be locally compensated with respect to the structure each side of the Region. In the previous section it was proposed that the structure on the concave side of the Egmont-Ruapehu-Opotiki boundary consists of a crust 22.5 km thick underlain by a mantle of density 3.30 Mg/m^3 . The model for the Central Volcanic Region will represent deviations from this structure with isostatic compensation being achieved at a depth of 22.5 km; this depth of compensation is chosen because it corresponds to the proposed position for the base of the crust each side of the Region, hence simplifying the computational procedures, and also because it approximately coincides with the estimated depth for the base of the elastic portion of the lithosphere of the northwestern North Island (section 6.8).

A density model that satisfies the above conditions along with the calculated gravity effects of the model in two and three dimensions is shown in fig. 7.6. The model is non-unique and in particular the depth of the proposed crust-upper mantle interface is poorly constrained. For example, trial models indicated that providing isostatic compensation is achieved at a depth of 22.5 km, with respect to the structure each side of the Region, the upper mantle density and crustal thickness could be simultaneously varied from 3.02 to 3.30 Mg/m^3 and 15 to 19 km, respectively, without any appreciable changes in the calculated gravity anomaly. However, the model shown is not physically unreasonable and accounts for several important features on the observed isostatic gravity anomaly map (fig. 7.1(a)); namely, the gravity highs that flank the Region, the deep gravity low where the Region becomes narrow near Lake Taupo and a broad, centrally placed, relative gravity high at the northern extremity of the Region.

There are three features of the model in fig. 7.6 that are purely artefacts of computational convenience. Firstly, the use of 2.75 Mg/m^3 as

the density for the lower crust of the Central Volcanic Region is not meant to imply that these rocks are the same as those 2.75 Mg/m^3 crustal rocks that are shown to flank the Region. Secondly, the use of sharp horizontal boundaries between the various isodensity layers may be misleading; particularly the crust-upper mantle interface for which earthquake studies (Haines, 1979) specifically suggests to be an indistinct boundary. Lastly, the choice of vertical boundaries to separate the isodensity layers within the Region from those without is a consequence of the requirement that the model be a simple, locally compensated one. In reality these boundaries may not be vertical, but this cannot be determined from gravity data alone.

The two components of the central North Island gravity model so far discussed deal with a proposed variation in crustal thickness and mantle densities; these are summarised in fig. 7.7. The final component of the gravity model is the model of the subducted Pacific plate.

7.5 The gravity effect of subducted lithospheric plates

7.5.1 Introduction

The gravity effect of subducted lithospheric plates remains uncertain despite the topic having received a great deal of attention (e.g. Hatherton, 1970; Griggs, 1972; Grow, 1973). These authors attempt to determine the gravity effect of subducted plates directly from gravity observations. A theoretical approach to the problem has been made by Minear & Toksöz (1970) who calculated the temperature distribution as a means to gain an estimate for the anomalous density and hence gravity effect of a subducted plate. Their models show a broad positive anomaly of $850 \text{ } \mu\text{N/kg}$ and a broad negative anomaly of $-1050 \text{ } \mu\text{N/kg}$ for subduction rates of 8 and 1 cm/y respectively. Such an extreme range of gravity anomalies are rarely observed over subduction zones.

Fig. 7.7 A plan view summarising the proposed crust-upper mantle structure associated with the Central Volcanic Region and the Egmont-Ruapehu-Opotiki boundary. Densities in Mg/m^3 .

The process of subduction at active margins is a dynamical phenomenon and the question naturally arises of how the dynamics may affect the gravity anomaly field. The problem of calculating dynamic fluid flow and induced stresses in back-arc regions has been considered by McKenzie (1969), Sclater (1972), Andrews & Sleep (1974) and Toksöz & Bird (1977) among others. The problem is a difficult one for it depends on poorly known parameters such as asthenospheric viscosity, which is both stress and temperature dependent. Coupled to this question of induced dynamic flow is that of how the excess mass of the subducted plate is compensated.

Davies (1981) presents some exploratory models for subduction that appear to be particularly promising in this regard. He begins by asserting that since subducted plates represent some of the largest mass anomalies within the earth it is reasonable that they be at least partially, if not wholly, compensated. He considers that compensation could be achieved by the earth's surface being pulled downwards by one, or a combination of, the following two mechanisms:

(i) By stresses transmitted up the slab to the earth's surface thus causing the often observed trench. This is called trench flexure.

(ii) By dynamically induced stresses transmitted through the fluid asthenospheric wedge that overlies the subducted plate. This is called wedge flexure.

Wedge flexure has previously been predicted by McKenzie (1969) and is discussed more fully by Torsvik *et al.* (1978).

Fig. 7.8 gives some of Davies results calculated for a two-dimensional, 45° dipping plate that is 1000 km in length, subducting at a velocity of 80 mm/y, represents an excess mass of $3.5 \times 10^{12} \text{ kg m}^{-1}$ and is overlain by an asthenospheric wedge of viscosity $3 \times 10^{20} \text{ N s m}^{-1}$. The cartoon at the

bottom of fig. 7.8 is a schematic earth model showing the various surface displacements associated with subduction that will give rise to gravity anomalies. Anomalies due to trench flexure can arise from either a depression in the earth's surface at, or about, the point above where the plate begins its descent, or from the downward displacement to mantle depths of oceanic crust that caps the upper surface of the subducted plate. These models are highly simplified and in particular note that the displacement curve for trench flexure is calculated for the vertical component of the plate load acting at a symmetrical trench; the effects of the horizontal component of the plate load and of moments at the trench are neglected.

The top graph of fig. 7.8 shows the calculated gravity effect of the excess mass of the subducted plate and the maximum gravity effects of wedge and trench flexure. The solid line in this graph represents the calculated gravity effect of the excess mass of the subducted plate that is 30% compensated by trench flexure and 50% by wedge flexure; further examples with various other combinations of wedge and trench flexure are given by Davies (1981).

The principal point of these theoretical models is that a wide variety of gravity profiles across subduction zones is possible depending on how compensation for the subducted plate is achieved. Of particular note is that if compensation for the plate is wholly due to wedge flexure the respective gravity effects of wedge flexure and the excess mass of the plate are nearly equal and opposite (fig. 7.8). It is therefore possible that subduction could occur beneath some region without any noticeable gravity anomalies at all. In reality, however, trenches or deep sedimentary basins associated with the point on the earth's surface above where a subducted plate makes its principal downturn are observed, so that it is reasonable

Earth model

Fig. 7.8 Calculated surface displacements and gravity effects of a subducted plate (after Davies, 1981). The earth model is schematic only and shows the various displacements that give rise to wedge (W) and trench (T) associated gravity anomalies.

to conject that in most cases at least part of the compensation for a subducted plate is realised by trench flexure.

Whether or not wedge flexure can be predicted to be an important contributing factor to the compensation of a subducted plate is largely dependent on the viscosity within the asthenospheric wedge that overlies the plate; the higher the viscosity the better the coupling between induced convection in the fluid asthenosphere and the overlying lithosphere. McKenzie's (1969) mathematical formulation of wedge flexure shows that the amplitude of the flexure is dependent on the length scale: $L(\text{km}) = (2\eta v/\Delta\rho g)^{\frac{1}{2}}$ where η is the viscosity in the asthenosphere, v is the subduction velocity, $\Delta\rho$ is the density contrast between the asthenosphere and the material that overlies the depressed lithosphere and g is the acceleration of gravity.

Of all the above parameters that define L the viscosity is the least confidently known. Furthermore, being an exponential function of temperature, the viscosity of the asthenosphere may be highly variable and at best can only be estimated to an order of magnitude (McKenzie, 1969). Davies and McKenzie use a viscosity of $3 \times 10^{20} \text{ N s m}^{-1}$ in their models, but if this value were to be reduced by an order of magnitude the amplitude of wedge flexure predicted by their models would be lessened accordingly. In this regard it is worth while noting that an asthenospheric viscosity of $10^{18} \text{ N s m}^{-1}$ has been proposed by Artyushkov (1974) for tectonically active areas and viscosities of 10^{18} - $10^{19} \text{ N s m}^{-1}$ have been suggested by Mavko & Nur (1975) and Spence (1977) specifically for the asthenosphere in the wedge above subducted plates. Moreover, Artyushkov goes further to argue that dynamic motion within the asthenosphere of tectonically active areas cannot be responsible for observed isostatic anomalies which commonly amount to a few hundred $\mu\text{N/kg}$. He calculates that for an

asthenospheric viscosity of $10^{18} \text{ N s m}^{-1}$ "improbably high rates of asthenospheric flow ($>10 \text{ m/y}$) would be necessary to create viscous stresses that could sustain uncompensated loads $>10^2$ bars (10 MPa) corresponding to the observed isostatic anomalies."

South of the Egmont-Ruapehu line where the heat flow is lower than to the north, and no attenuating zone within the upper mantle is inferred to be present (fig. 2.1), it appears possible that the asthenospheric wedge is cooler and therefore more viscous than the asthenospheric wedge north of the Egmont-Ruapehu line. Thus conditions to the south of the line are more favourable than to the north for wedge flexure to be a contributing factor to the compensation of the subducted plate. However, for the part of the North Island that this model is primarily concerned with, that is the region north of the latitude coincident with the Egmont-Ruapehu line, it will be assumed that compensation for the subducted plate is achieved wholly by trench flexure.

7.5.2 Mass anomaly models of subduction

The trench flexure model of Davies (1981) is physically identical to the flexural or elastic plate model of subduction as described by Watts & Talwani (1974). Watts & Talwani consider the simple analogy for subduction of an applied force acting on the free edge of a semi-infinite elastic plate which overlies a weak fluid. They note "The total gravity anomaly associated with a flexural model is obtained from the combined gravity effects of the surface load and the induced deflections. Since the deflections are a form of compensation for the applied surface load the total gravity effect of the model integrated to great distances will be zero."

The elastic plate model predicts three principal sources for the gravity anomalies observed over subduction zones:

(i) The gravity effect of the applied load - in this case the equivalent applied load is the excess mass of the subducted plate.

(ii) The gravity effect of trench flexure. As discussed in section 7.5.1 this gravity effect may arise from a depression in the surface of the earth at, or about, the point above the principal downturn of the subducted plate, or from the downward displacement of crustal material, that caps the plate, to mantle depths.

(iii) The gravity effect of a topographic high which is predicted, and is indeed commonly observed (Watts & Talwani, 1974), to occur oceanward of the point of downturn for the subducted plate. This outer gravity and topographic high arises from an upward elastic flexure that provides partial compensation for the trench flexure described in (ii) above. The mass balance associated with subduction may therefore be approximated by:

$$M_p^+ + M_d^- + M_{OH}^+ = 0 \quad \dots 7.2$$

where M_p^+ = mass anomaly of the plate (positive sign)

M_d^- = mass anomaly associated with the principal downturn
(negative sign)

M_{OH}^+ = mass anomaly associated with the first upward deflection
(outer topographic or gravity high) (positive sign).

Due to a paucity of marine gravity data off the East Coast of the North Island the outer gravity-topographic high for the North Island subduction zone cannot be readily identified. However, the theoretical models presented by Watts & Talwani indicate that the integrated gravity effect of the outer gravity high will be about 10% of the integrated

gravity effect associated with the principal downturn (M_d^-). Accordingly, equation 7.1 is modified to:

$$M_p^+ + (0.9) \times M_d^- = 0 \quad \dots 7.3$$

Seismic studies carried out by Reyners (1978) provide the following geometrical constraints for the gravity model:

1) Between the Hikurangi Trench and the 80 km isobath of deep seismic activity the plate dips northwest with the dip increasing from 5° to about 30° ; beneath 80 km the plate dips northwest at an angle of about 50° .

2) The 80 km isobath approximately corresponds to the position of the active volcanic front.

The model adopted here is essentially a modification of Hatherton's model (b) of fig. 2.4. It is based on petrological arguments presented by Grow (1973) and Anderson *et al.* (1976) and applied by Sayer (1980) to the Mariana subduction zone. The negative mass anomaly is attributed to a layer of oceanic crust capping the plate and being carried down to mantle depths. At a depth of about 80 km the oceanic crust is considered to have transformed to eclogite of density 3.56 Mg/m^3 (Anderson *et al.* 1976). At a depth of 150 km temperatures are assumed to have risen sufficiently for the eclogite layer to have melted (Cole, 1978), and being denser than the standard mantle, the eclogite layer constitutes part of the positive mass anomaly in M_p^+ (equation 7.3). The remainder of M_p^+ is modelled as being due to that part of a 80 km thick, relatively cool subducted plate, less the eclogite layer, in the depth range of 80 to 250 km.

It is assumed here that the Wanganui and East Coast sedimentary basins arise from subduction related processes. That is the sedimentary basins

are, in part, a form of distributed compensation for the subducted plate and accordingly the negative mass anomalies associated with the sedimentary infill are considered to be part of M_d^- in equation 7.3. This assumption may not be strictly valid, particularly for the northern portion of the Wanganui Sedimentary Basin, where sediments may be compensated by mechanisms quite unrelated to the subduction process. Nevertheless, even though the near surface sediments have a pronounced gravitational effect their contribution to M_d^- is only minor.

The positions of the mass anomalies described above and the mass balance of equation 7.3 effectively define the wavelength of the subduction-related gravity anomaly; what remains to be determined is the amplitude of the anomaly. A southeast-northwest profile was taken across the North Island (A-A' in fig. 7.4) and a model constructed incorporating the Central Volcanic Region, the Ruapehu-Opotiki boundary and the subduction effect. The free parameters of the subduction model were then varied until a satisfactory fit between the calculated and observed isostatic gravity anomalies was obtained. The free parameters of the subduction model were considered to be the thickness of the oceanic crust/eclogite layer and the positive density contrast in the main body of the subducted plate at depths greater than 80 km. The best fit model is shown in fig. 7.9. It can be seen that after allowance for the gravity effect of the Central Volcanic Region and the Ruapehu-Opotiki boundaries the effect of the subduction process is required to be quite small; 120 $\mu\text{N/kg}$ in the back-arc region and, if the contribution from the near surface sediments is neglected, -300 $\mu\text{N/kg}$ over the principal downturn of the plate.

The 0.015 Mg/m^3 density contrast proposed for the bulk of the subducted plate appears small when compared say to Hatherton's models (fig. 2.4). However, the total mass anomaly of the slab is equal to

Fig. 7.9 The proposed model and gravity effect, in profile form (left) and plan view (right), for the subducted plate beneath the North Island.

$3.4 \times 10^8 \text{ Mg m}^{-1}$ which if distributed uniformly throughout an 80 km thick, 222 km long plate is equivalent to a mean density contrast of 0.02 Mg/m^3 or for a 50 km thick plate, as used by Hatherton in his models, a density contrast of 0.03 Mg/m^3 .

Watts & Talwani (1974) found that outer gravity highs seaward of the Aleutian and other trenches may be explained in total by an upward topographic flexure of the seafloor without any contribution from the subducted plate. They concluded that if a 80 km thick plate, in the depth range of 80 to 250 km, is to make a negligible contribution to the outer gravity high then the mean density contrast throughout the plate must be 0.04 Mg/m^3 or less. In light of this result the proposal of an average 0.02 to 0.03 Mg/m^3 density contrast for the subducted plate beneath the North Island does not appear unreasonable.

In any event, the important point of this analysis is that it is difficult to isolate the gravity contribution of a subducted plate from the observed gravity data alone; possible gravity effects of crust-mantle perturbations associated with extension in the back-arc regions may have a more dominant gravity effect and need to be isolated first.

7.6 Combined gravity effect of the three structures

The calculated three-dimensional effects of the Ruapehu-Egmont-Opotiki boundary, the Central Volcanic Region boundary and the subduction model, excluding the effects of near surface sediments, are shown superimposed on a $20 \times 20 \text{ km}$ grid of the North Island in fig. 7.10(a). The estimated gravity effects of sediments within the Wanganui and East Coast basins are shown in fig. 7.10(b). For the Wanganui Basin the gravity effects are a smoothed version of the residual anomalies given by Hunt (1980) whereas for the East Coast Basin the gravity effects can only be considered as

Fig. 7.10 Upper grid shows the combined effects of the Egmont-Ruapehu-Opotiki boundary (fig. 7.5), the C.V.R. boundary (fig. 7.7) and the subduction model (fig. 7.9). The lower grid represents the estimated gravity anomaly effect of low-density sediments.

Fig. 7.11 Top map shows the combined effect of the two grids shown in fig. 7.10. The bottom map is the observed isostatic anomalies.

crude estimates. However, as for the Wanganui Basin, the gravity effects of the East Coast sediments are considered to have a minimum value of $-500 \mu\text{N/kg}$ and the effects of the Miocene sediments are taken to be substantially less than that of Pleistocene-Pliocene sediments (Hunt, 1980).

Fig. 7.11 shows a contoured presentation of the calculated gravity anomalies compared with the observed isostatic gravity anomaly field.

Applying the goodness of fit criterion:

$$\% \text{ of anomaly explained} = (\Sigma(\text{obs})^2 - \Sigma(\text{residual})^2 / \Sigma(\text{obs})^2 \times 100\%$$

where residual = calculated-observed, it was found that 90% of the observed gravity anomaly pattern could be explained by the model with a standard error of residuals equal to $150 \mu\text{N/kg}$.

The model provides a satisfactory fit to the observed gravity for regions either side of and within the Central Volcanic Region; thus fulfilling the original objective of this exercise. The model obviously fails, however, by some $-500 \mu\text{N/kg}$ to satisfy the intense negative anomaly near Wanganui.

7.7 The Wanganui gravity anomaly

The Wanganui gravity anomaly has attracted the attention of various New Zealand geophysicists for some time (Robertson & Reilly, 1958; Eiby, 1964; Reyners, 1978). However, no generally accepted solution has yet been proposed. Although it is going beyond the scope of this study to attempt a quantitative solution for the Wanganui anomaly, the following modifications to the regional gravity model described in this chapter may be useful:

i) A bending of the subducted plate model to the southwest once south of Mt Ruapehu. The isobaths of deep seismic activity as given by Adams & Ware (1977) show a slight hint of such a bend, although possibly

not enough to account for the southwest bend in the axis of the Rangitikei-Waiapu negative gravity anomaly.

(ii) A change in the mode of compensation for the subducted plate from trench to wedge flexure once south of Mt Ruapehu. This is considered the most promising avenue of inquiry for the following reasons. Firstly, it would account for the deviation of the axis of the negative gravity anomaly from the strike direction of Benioff zone; from fig. 7.8 it can be seen that the gravity minimum for wedge flexure will occur further landward, or in the case of the North Island further to the northwest, than the minimum produced by trench flexure. Secondly, wedge flexure may go some way in providing an explanation for the "crustal downwarp" (Suggate, 1978) that exists between the Wanganui Basin in the north to the Marlborough area in the south. Lastly, it is physically reasonable to propose two forms of compensation for the subducted plate within the North Island, because whether or not wedge or trench flexure will occur is largely dependent on the viscosity of the asthenospheric wedge that overlies the plate (section 7.5.1). The geophysical observations discussed in section 7.4 point to a colder and therefore more viscous asthenosphere south of the Egmont-Ruapehu boundary than to the north, thus making conditions more favourable for wedge flexure to the south and trench flexure to the north of the boundary.

The practical, or computational, consequence of allowing for wedge flexure beneath the Wanganui Basin, would be that the gravity effect of a downwarp of the crust-mantle interface, beneath the Basin, be included in the model.

At this stage the above discussion must be regarded as conjecture; more data and detailed modelling is required before the exact cause of the impressive Wanganui gravity anomaly can be identified.

7.8 2½-Dimensional solution to the Bouguer gravity anomaly field

The solution for the isostatic gravity anomaly field can only be considered to represent a first-order structural solution as no allowance has been made for compensation of topography. For the model to take account of topography, crustal thicknesses of the model solution to the isostatic gravity anomaly field may be increased under the mountainous regions and decreased off-shore until the gravity effect of the modified model satisfies the Bouguer gravity anomaly field.

Fig. 7.12 shows a possible solution to the Bouguer gravity anomaly field along profile A-A' of fig. 2.2. The 2½-dimensional modelling technique (Appendix D) was used for this interpretation because the profile represents a cut through a three-dimensional structure. Crustal thicknesses are now interpreted to be 23-25 km beneath the northwestern North Island and 27 and 35 km beneath the Kaingaroa plateau and Ruahine range respectively.

However, it is stressed that these crustal thicknesses can only be regarded as being approximate for the following reasons. Firstly, it is assumed that compensation for topography of the North Island is achieved by a simple pointwise thickening of the crust and secondly, the starting point for this interpretation was a single determination of crustal thickness for the southern portion of the North Island; and this determination was gained from the results of an unreversed seismic refraction profile. The distribution of the low-density upper mantle shown in fig. 7.12 is also poorly constrained. For example, a distribution more in accord with Karig's (1974) suggestion of a "low-density mantle diapir" rising up beneath the back-arc basin would require that the low-density mantle material were concentrated in a narrow region directly

Fig. 7.12 A $2\frac{1}{2}$ -dimensional (appendix D) interpretational model for the Bouguer gravity anomaly field along A-A' of fig. 7.4. The density contrasts of mass anomalies, other than those associated with the subducted plate beneath 80 km, are calculated with respect to the standard density model shown on the right. Mass anomalies associated with the thin crust and low-density mantle beneath the C.V.R. and western North Island are calculated with a southwest strike-length of 50 km. Other mass anomalies have infinite strike-lengths.

beneath the basin and extending down to the top surface of the subducted plate. These are aspects of the model that cannot be resolved from gravity data alone. Nevertheless, the important features that the model does illustrate are:

- (i) The relative variation in crustal thicknesses and upper mantle densities of the central North Island that are required to explain the observed gravity anomalies.
- (ii) The abrupt change in the vertical mass distribution at the boundaries of the Central Volcanic Region that is required for a compensated model to simulate the observed, steep gravity gradients.

Fig. 7.13 summarises the regional gravity interpretation model for the North Island. The isostatic gravity anomaly map is divided into the following four regions:

Region 1: Northland-Waikato-Taranaki region - gravity anomalies are almost entirely positive here and are interpreted to arise from gravity edge-effects associated with the Egmont-Ruapehu-Opotiki and Central Volcanic Region boundaries. Structurally, the region is interpreted to have a thin (or more dense) crust which is compensated by a low-density mantle.

Region 2: Central Volcanic Region - an area of very thin crust (half the normal continental thickness) which is underlain by a low-density mantle. The relative gravity low in the centre of the Region is interpreted to be due to superficial, low-density volcanics and the flanking gravity highs as being due to a shallow mantle beneath the Region which provides compensation for the low-density volcanic rocks above.

Region 3: The elongated gravity low and deep sedimentary pile that dominates the Hawke Bay area are interpreted as being due to a trench flexure which provides compensation for the excess mass of the subducted

Fig. 7.13 Diagram showing the North Island isostatic gravity anomaly field divided into four distinct regions. A summary description for each region is given in the text. The area shown beneath which the subducted plate is inferred to be present is taken from Walcott (1978) and the arrows represent the direction of the present day motion between the Pacific and Indian plates.

plate that underlies regions 1 and 2. In addition, a component of the negative gravity anomaly is interpreted to come from the gravity edge-effect associated with the Ruapehu-Opotiki boundary.

Region 4: The broad and intense gravity low centred near Wanganui is proposed to be due to both a gravity edge effect associated with the Egmont-Ruapehu boundary and a wedge flexure which provides compensation for the subducted plate in this region.

7.9 Conclusions

The gravity interpretation presented here is the first attempt to consider both shallow crustal and deep mantle structure of the central North Island. Because the model is a static, locally compensated one it must be considered limiting. Nevertheless the model is useful in that it is broadly compatible with other geophysical data pertaining to the central North Island and simple enough to be tested by deep seismic sounding. Therefore the model should be regarded as a testable hypothesis.

Up to this stage no explanation has been offered which might account for the proposed variation of crust and lithospheric thicknesses of the North Island. In a very general way an explanation may be secured from any one of the various theoretical accounts of back-arc asthenospheric convection (e.g. McKenzie, 1969; Sclater, 1972; Andrews & Sleep, 1974; Toksöz & Bird, 1977). These accounts predict a thermally weakened and thinned lithosphere in back-arc regions as a result of convection within the asthenosphere induced by the downgoing plate; as the lithosphere thins, low-density asthenosphere rises causing crustal extension, uplift and supracrustal erosion and, ultimately, back-arc spreading with the creation of new lithosphere. Yet this generalised account still begs the question of why the Egmont-Ruapehu boundary passes nearly orthogonally

over the isobaths of deep seismic activity. To answer this, the history of subduction beneath the North Island over the last 20 My should be considered. Hatherton (1969b), Calhaem (1973), and Cole & Lewis (1981) have all entertained the idea that a NW-SE trending subduction zone extended beneath the northwestern North Island during Tertiary times and has rotated and migrated southeastwards since. If this is so then it would appear that the Northland and Waikato regions of the North Island have been subjected to possibly as much as 20 My of back-arc convection and lithospheric and crustal thinning. On the other hand, subduction at the south end of the North Island is thought to be at a comparatively juvenile stage (Walcott, 1978). Thus it is possible that south of the Egmont-Ruapehu boundary insufficient time has elapsed for back-arc asthenospheric convection to thermally weaken and thin the lithosphere.

Because the North Island back-arc gravity high is coincident with areas of extensional tectonics, higher than normal heat-flow, volcanism and anomalous seismic properties of the upper mantle, yet not coincident with the isobaths of deep seismic activity, it is proposed that it is crustal extension and back-arc spreading that is the principal cause of the North Island back-arc gravity high. This is not a new proposal. Both Karig (1971) and Watts & Talwani (1974) have presented arguments based on studies from other asymmetric active regions of the western Pacific to support this view. The case of the Mariana subduction zone and associated back-arc basin, as discussed by Karig, is particularly convincing. Here the Benioff zone is vertical yet a positive back-arc gravity high of about 500-600 $\mu\text{N/kg}$ peaks over the centre of the back-arc basin, and not above the subducted plate. Karig presents a simple locally-compensated model, similar in principle to that used here, to explain this back-arc gravity high. Moreover, a recent study of geoid and gravity

anomalies associated with some subduction zones, by Chapman and Talwani (1982), arrive at conclusions similar to those arrived at in this study. Namely, they find that to explain both the geoid and gravity anomalies, observed within the vicinity of subduction zones with an associated extensional back-arc basin, a "dipolar layered mass distribution" in the back-arc regions and a fairly small density contrast for the subducted plate is required. The dipolar layered mass distributions in the models of Chapman & Talwani consists of a shallow sea compensated by a low-density upper mantle. This is a similar mass distribution to that used here for the back-arc regions of the North Island, except that it is thin crust, rather than a shallow sea, that is compensated by a low-density mantle.

Of prime interest to this study is that the gravity model provides a coherent explanation for the observed gravity anomaly field of the Central Volcanic Region. In doing so the model permits a reassessment of some of the problems encountered with interpretations of the residual gravity anomaly field presented in chapter 5. For example, the observation that the residual gravity anomaly field west of Wairakei is about 200 $\mu\text{N/kg}$ less than the residual field at Rotoehu, yet the seismic sections at these two locations are similar, can be accounted for, in part, by the gravity effect of a shallow mantle which becomes more dominant as the Region widens (fig. 7.6).

In the following chapter gravity profiles across the Central Volcanic Region will be examined with this point in mind.

Note: Subsequent to the development of this gravity model a long-range seismic refraction experiment has been carried out within the Central Volcanic Region by Geophysics Division of the DSIR. The writer participated

in this experiment and a preliminary interpretation of the results is given in appendix F. The interpretation is in good agreement with the structure predicted for the Central Volcanic Region from the gravity interpretation. Namely, a 7.3-7.7 km/s "upper mantle" at a depth of about 15 km.

CHAPTER 8 GRAVITY AND HEATFLOW MODELS OF THE CENTRAL
VOLCANIC REGION

Summary

By way of a summary of the data and interpretations presented in this thesis, a quantitative, integrated geophysical model for the Central Volcanic Region is presented in this chapter.

Three west-east gravity interpretation profiles across the Central Volcanic Region are examined in detail. Comparisons are drawn between interpretations based on the analytically derived residual gravity anomaly field and the isostatically compensated model of the Central Volcanic Region presented in chapter 7.

A discussion is then presented relating to the possible petrological models for the Central Volcanic Region that are implied by the crustal structure model developed in this study. Finally, consideration is given to the question of whether the Central Volcanic Region is the site of active spreading or merely passive extension.

8.1 Gravity interpretation profiles across the Central Volcanic Region

8.1.1 Introduction

In the following section emphasis will be placed on two different methods of interpreting gravity profiles across the Central Volcanic Region. In brief the procedures and accompanying assumptions for the two methods are:

(i) Interpretations of the analytically derived residual gravity anomaly field: This is the method of gravity interpretation that has been applied to the Central Volcanic Region in the past; examples were

given previously in chapter 5 (figs 5.18 and 5.20). Separation of the residual gravity components of the observed gravity anomaly field is achieved by determining a regional field by means of some analytical criteria such as wavelength filtering. These residuals will be called the analytical residuals. Two assumptions are necessary when interpreting the analytical residuals. The first is that there is a clear separation in depth between the source bodies responsible for the residual gravity anomalies, on one hand, and the source bodies responsible for the regional gravity anomalies on the other. The second assumption, which is implicit rather than stated, is either that the low-density volcanic rocks within the Region are uncompensated, and therefore supported by stress-differences within the elastic layer of the lithosphere, or that they are compensated and the gravity effect of the compensation mechanism is contained wholly within the regional gravity anomaly field.

(ii) Interpretations of the modelled residual gravity anomalies: With this method the residuals to be explained are deviations of the observed gravity anomalies from the calculated gravity effect of the regional model presented in the previous chapter. In practice this requires that the modelled gravity effects of the Egmont-Ruapehu-Opotiki boundary and the subduction model (figs 7.5 and 7.9) are taken as the background regional gravity field for the Central Volcanic Region and what is left, here termed the modelled residuals, are the gravity effects considered to represent density irregularities solely associated with the Central Volcanic Region. The modelled residuals have both positive and negative components: sharp negative anomalies within the Central Volcanic Region and broader positive anomalies on the flanks and adjacent to the Region. With this method all components of the gravity anomaly field are accounted for and the ultimate fidelity of the interpretational

models rests on the validity of the regional interpretational model presented in the previous chapter.

The locations of the three profiles to be discussed are shown in fig. 8.1.

8.1.2 Tihoi-Wairakei-Kaingaroa profile

An interpretation of the analytic residuals along this profile (fig. 5.20) was presented in chapter 5. Fig. 8.2(a) shows that interpretation, and the interpretation of the modelled residuals along the same profile. Both models of fig. 8.2 are constrained to conform with the seismic results of Robinson *et al.* (1981) (section 4.4) which show that for a large part of this profile the volcanic cover can be regarded as a 2 km thick section of rocks (for which the residual gravity effect is about $-340 \mu\text{N/kg}$) and that the seismic basement beneath the volcanic cover is essentially horizontal.

Comparison of the two sets of residuals in fig. 8.2 shows that the modelled residuals are similar in shape to the analytical residuals, but about $120 \mu\text{N/kg}$ greater in amplitude. The first-order fit to the modelled residuals (dashed line in fig. 8.2(b)) is made with a locally compensated structure similar to that proposed for the regional model discussed in chapter 7. Negative secondary residuals are still present west of Wairakei, but they are only half the amplitude of the secondary residuals present in the model of fig. 8.2(a). In fig. 8.2(b) the negative secondary residuals are explained by the additional gravity effect of a rectangular body of density contrast -0.4 Mg/m^3 in the depth range 7–9 km and is interpreted to represent a body of molten rhyolite. Such an interpretation is consistent with the numerical-petrochemical model of Ewart *et al.* (1975) which suggests that the

Fig. 8.1 Residual gravity anomaly map of the Central Volcanic Region showing the location of the three profiles A-A', B-B' and C-C'-C'' to be analysed in this chapter.

Fig. 8.2 Tihoi-Wairakei-Kaingaroa profile. The top model is a solution to the analytical residual gravity anomaly field and the bottom model is a solution to the modelled residuals. Bracketed number pairs refer to the strike lengths (in km) of the indicated mass anomalies into and out of the plane of the profile. If no strike lengths are given the mass anomaly is taken to be two-dimensional. All density contrasts and densities in Mg/m^3 and datum level for the profile is 300 m A.S.L.

molten rhyolite required to produce the large scale ash-flows, that are associated with the Central Volcanic Region, would have equilibrated at an average depth of 7-8 km in volumes with a characteristic radius of about 5 km. Moreover, because the secondary residuals are only half the amplitude of those in model a), the proposed volume of molten rhyolite (density contrast -0.4 Mg/m^3) can be modelled deeper than the 6 km depth limit of maximum microearthquake activity found by Evison *et al.* (1976) (section 5.5).

Other interpretations of these negative secondary residuals are possible. For example, it could be that the low-density volcanic cover is less dense or thicker west of Wairakei than elsewhere and that the resolution of the refraction survey of Robinson *et al.* (1981) was insufficient to detect this. More detailed seismic work in this area would be desirable.

The local gravity high over Wairakei is explained in model b) by a mass anomaly of density contrast $+0.2 \text{ Mg/m}^3$, with respect to the volcanic cover, in the depth range 500 to 1500 m. On the other hand, the interpretation of fig. 8.2(a) implies that the Wairakei gravity high is "normal" and that the region to the east of Wairakei is the site of a local gravity low. The best interpretation of the Wairakei high possibly lies somewhere between these two alternatives, in that the densities of bore-121 are higher than "normal" and the densities of bore-227 are lower than "normal". "Normal" refers here to the Wairakei density with depth curve developed from all the Wairakei data as discussed in chapter 4.

8.1.3 Mangakino-Maroa-Kaingaroa profile

Fig. 8.3(a) and (b) show interpretations of the analytical and modelled residual gravity anomalies, respectively, along B-B' of fig. 8.1.

This profile crosses the Mangakino seismic line of fig. 5.13 almost at right angles. Thus the volcanic cover beneath Mangakino in both models of fig. 8.3 conforms to the density structure previously deduced from the observed seismic velocities. Elsewhere along the profile the volcanic cover is taken to be the same as that used for the Tihoi-Wairakei-Kaingaroa profile; that is, 2 km thick, with an average density contrast of -0.4 Mg/m^3 .

Comparison of the two models in fig. 8.3 reveals that the magnitude of the mass anomaly beneath seismic basement that is required to explain the Mangakino gravity low is substantially less for the modelled residuals (model b) than for analytical residuals (model a); anomalous masses of $-90 \times 10^3 \text{ kg}$ and $-224 \times 10^3 \text{ kg}$ are required respectively. Furthermore in model b) the Mangakino low can be accounted for by a mass anomaly of density 2.48 Mg/m^3 , or density contrast -0.12 Mg/m^3 , extending from seismic basement level to a depth of 8 km. A rock of this density satisfactorily accounts for the 4.8 km/s velocity detected at depth of 2.2 km beneath the Mangakino seismic line; this velocity being lower than those found at a similar depth for the Taupo and Rotoehu surveys (chapter 5).

Whether or not the proposed 2.48 Mg/m^3 mass anomaly represents solidified, partially molten or even molten magma is uncertain. Rhyolite densities vary from 2.05 Mg/m^3 for rhyolite in the molten state (section 5.4.6), to $2.50\text{--}2.59 \text{ Mg/m}^3$ for the cooled plutonic equivalent of the rhyolitic volcanic rocks found within the Central Volcanic Region (section 4.2.1). A further possibility is that the 2.48 Mg/m^3 density represents molten andesite; Murase & McBirney (1973) show from experimental work that andesite glass with a density of 2.59 Mg/m^3 at room temperature has a density of 2.45 Mg/m^3 at a temperature of 1000°C .

Fig. 8.3 Mangakino-Maroa-Kaingaroa profile (B-B' of fig. 8.1). As for fig. 8.2 and the numbers in brackets representing the strike lengths out of and into the plane of the profile respectively. The section of low-density, low-velocity volcanic rocks at Mangakino is constrained to conform to the density model as interpreted from the Mangakino refraction survey. Datum level is 300 m A.S.L. All densities and density contrasts in Mg/m³.

Clearly the non-uniqueness of gravity inversions and our incomplete knowledge of densities for various molten rocks militates against any firm conclusions being drawn from the gravity interpretation models alone. However, other data relevant to the cause of the Mangakino gravity low were discussed in section 5.4.6 and may be briefly reiterated.

a) An electrical conductivity anomaly (Midha, 1979) (fig. 5.19) is coincident with the Mangakino gravity low; Midha suggests that the existence of this conductivity anomaly lends qualitative support to there being molten magma beneath Mangakino.

b) The stratigraphic mapping of Blank (1965) suggested that the source vent for some of the earliest and largest ignimbrite sheets erupted from the Central Volcanic Region is situated near Mangakino. Dating of these ignimbrites reveals an age of 0.7-1.1 My. Furthermore there is a topographic depression in the Mangakino area despite an unusually large thickness of Whakamaru ignimbrite. This ignimbrite was erupted about 0.33 My BP from the Taupo Volcanic Centre to the southeast of Mangakino (Briggs, 1976); Cole (1981) suggests that the topographic depression could represent a caldera-collapse structure associated with an old rhyolitic centre.

The preferred interpretation of the Mangakino gravity low is that it is an expression of a now partially cooled acidic magma body which erupted during the period 0.7 to 1.1 My BP to produce the extensive ignimbrite deposits found on and adjacent to the western margin of the Central Volcanic Region. However, it may be recalled from the discussion in section 5.4.5 that the cooling models of Lachenbruch *et al.* (1976) show that for a magma body which was similar in dimensions to that shown in figs 5.18 and 8.3 to be still molten would require that there has been reinjection of heat within the last 0.2-0.4 My. Therefore, in view

of the apparent southeastward migration of both andesitic and rhyolitic volcanism across the Central Volcanic Region (Kear, 1959; Cole, 1979), the proposed presence of subsurface magmatic activity at the western margin of the Region would seem to be highly anomalous.

Immediately east of the Mangakino low the traverse cuts across the southern end of the circular Tokoroa high. Rogan (1980) ascribed this gravity high to a greywacke basement high. An equally plausible explanation, particularly as the gravity high is associated with minor magnetic anomalies (Whiteford, 1976), is that the Tokoroa high represents a buried andesite massif. About 60 km to the northeast of Tokoroa is the Te Puke gravity high (fig. 8.1). This gravity high does correlate with outcropping andesite, yet like the Tokoroa high it has only minor magnetic expression.

A density of 2.5 Mg/m^3 was chosen to represent the proposed andesite massif; this is somewhat less than the mean density of andesites found within the adjacent Coromandel region (Chapter 4). However, the lower density used here takes account of the fact that andesite volcanoes like Mt Ruapehu occur in a stratovolcanic formation with alternating layers of low-density breccia and higher-density lava.

To the east of the Tokoroa high is a local gravity low associated with the Maroa Volcanic Centre. No seismic or deep-borehole data is available from this area. If it is assumed that the top 2 km of volcanic cover is much the same here as that beneath the Wairakei-Tihoi region then a deeper low-density mass anomaly, similar to that postulated at Mangakino, is required. However, there appears little justification for extrapolating the Taupo refraction results northwards, and it is possible that the local gravity low is due to either a deeper or a less dense sequence of volcanic cover. More seismic work is required here.

8.1.4 Karapiro-Te Puke-Whakatane profile

Fig. 8.4(a) is a first-order interpretation of the modelled residuals along the profile C-C'-C" of fig. 8.1. Despite the obvious short-wavelength differences between the calculated and observed anomalies, the overall fit is considered reasonable. Note that the calculated gravity in the middle of the profile is about $-150 \mu\text{N/kg}$ whereas in the middle of the Tihoi-Wairakei-Kaingaroa profile the calculated gravity is $-300 \mu\text{N/kg}$. This difference can be understood by considering that as the Central Volcanic Region widens the gravity effect of the proposed shallow, low-density mantle becomes more prominent. Indeed, because the model is a locally compensated one, as the density model of fig. 8.4(a) becomes infinitely wide the calculated gravity in the centre of the profile will approach zero. This effect explains, in part, why the observed residual gravity anomalies at Taupo and Rotoehu differ by $250 \mu\text{N/kg}$ yet the seismic velocity sections at these two localities are similar (Chapter 5).

Model b) of fig. 8.4 is an attempt to modify the first-order interpretation for the observed variation in the surface geology along the profile. The dashed line at a depth of 2 km along most of the profile signifies the depth where density contrasts are assumed to vanish; it is not intended to imply that the respective lithological units terminate at this depth.

At the western margin the profile passes across the southern extension of the Hauraki Rift. Here Hochstein & Nixon (1979) have carried out geophysical studies and the density contrasts used in model b) (fig. 8.4) are adopted from their models. The profile then crosses the Kaimai range which has only a minor gravitational expression. A large rhyolite dome outcrops here; hence the density of 2.35 Mg/m^3 used in the

Fig. 8.4 Karapiro-Te Puke-Whakatane profile (C-C'-C'' of fig. 8.1). The top model is a first order solution to the modelled residuals. The lower model is similar but modified for the various geological structures observed along the profile. The dashed line at a depth of about 2 km signifies the depth at which the density contrasts are assumed to vanish. All densities in Mg/m³ and datum level is sea level.

model. Further east is the broad Mamaku Plateau. The geological cross-section of Healy *et al.* (1964) suggests a 500 m thick section of ignimbrite here, but it is evident from the gravity data that the thickness of low-density volcanics required is about 2 km.

At Te Puke an elliptically-shaped gravity high is coincident with outcropping andesite and a weak magnetic anomaly as discussed previously. As for the Tokoroa gravity high a density of 2.5 Mg/m^3 is used to model the andesite. Between Te Puke and the greywacke outcrop near Otamarakau a 2 km thickness of low-density rocks is required and this depth is in approximate accordance with the results of the Rotoehu seismic survey. A 1.8 km thickness of low-velocity rock was interpreted along the Rotoehu seismic line, but the line ran due south from the greywacke outcrop, orthogonal to, but not crossing the gravity interpretation profile.

East of the Otamarakau greywacke outcrop, but west of the Whakatane Graben, the surface geology is varied, consisting largely of ignimbrite, marine sandstones with a small outcrop of andesite. The density of 2.4 Mg/m^3 that is used here implies that the andesite may have a larger sub-surface extent than the small surface outcrop would suggest.

On the eastern margin of the profile lies the Whakatane Graben (Healy *et al.* 1964). Bounded on the west by normal faults and on the east by a complex set of intersecting north-south transcurrent and north-east normal faults, the Whakatane Graben is considered by Healy *et al.* to represent the northward continuation of the Taupo Volcanic Zone. In order to account for both the magnitude and gradients of the gravity anomaly observed here a lower average density of 1.95 Mg/m^3 is postulated for the top 1 km of low-density rocks. Other explanations are, however, possible; in particular it could be that the low-density rocks are thicker than the 2 km indicated.

The essential point to be noted from this profile analysis is that, apart from local outcrops of higher density rocks at Te Puke and Otamarakau, the 2 km thick section of low-density rocks is almost continuous right across the Central Volcanic Region. Furthermore the observation that the analytical residuals decrease towards the north-end of the Central Volcanic Region, yet the seismic velocity sections at Rotoehu and Taupo are similar, is largely explained by the gravity effect of the proposed low-density, shallow mantle becoming more prominent as the Region widens.

8.1.5 Summary of the profiles

In this section two methods of interpreting gravity profiles across the Central Volcanic Region have been demonstrated. The difference between the two methods lies in the manner in which the background regional field was determined; either an analytically determined regional field or a regional field which is modelled directly. The statement made in Chapter 3 that no regional field is unique still stands, and there is little merit in placing a right or wrong label on either of the two methods just discussed. However, the interpretations based on the modelled residuals are preferred, primarily because all components of the gravity anomaly field are explained and features such as the secondary residual gravity lows at Mangakino and Taupo can be accounted for with physically tenable models.

8.2 Petrological models for the Central Volcanic Region

8.2.1 Introduction

The density-velocity model for the Central Volcanic Region, as derived from the data and interpretations presented in this study, is summarised in fig. 8.5. This is a simple representation of the Region in terms of isodensity and isovelocity layers and, apart from the top 2 km of

Fig. 8.5 A summary of the crustal structure model for the Central Volcanic Region as derived from the seismic and gravity investigations of this study. The two proposed petrological models shown on the right are discussed in section 8.2.

the Region which has been investigated by bore holes, the petrological characteristics of the various layers are unknown. In particular the nature of the 2.60 Mg/m^3 "basement" layer still remains unknown, yet is of great interest. For example in the continental rift situation where thin crust, higher than normal heat-flow and extensional tectonics are observed, a mechanical model often favoured is one of plastically attenuated lower crust and brittle fractured upper crust (Artemjev & Artyushkov, 1971; McKenzie, 1978). If this model were applied to the Central Volcanic Region, basement rocks would be greywacke that has been extended and thinned without actually rupturing. Alternatively, basement rocks could be intruded igneous rocks as suggested by Calhaem (1973).

Either rock type is consistent with the gravity models and seismic data presented so far in this study. Magnetic anomalies are of limited use as a discriminator because of the demagnetisation effects of hydrothermal activity, the shallow Curie Point isotherm (chapter 6) and the dominating effect of near surface rhyolite domes. Similarly, electrical sounding methods made within the Central Volcanic Region have proved incapable of unequivocally differentiating between basement rocks of an igneous or sedimentary affinity (Hurst, 1975).

The one geophysical parameter that has not been considered as yet is the anomalously high heat output of the Region. In this section petrological models implied by two previously proposed heat flow models for the Region will be discussed, and their compatibility, or otherwise, with the density-velocity model of fig. 8.5 will be assessed.

8.2.2 Model 1 - Banwell's model

Banwell (1963) gave consideration to the idea of circulating magma at a depth of about 9 km beneath the Region supplying heat via a thin

conducting layer to convecting meteoric water within the upper crust.

He found that the observed natural heat flow of the Taupo-Wairakei area could be explained by a circulating magma chamber, at a depth of about 9 km, if the top side of the magma chamber has an area of 227 km^2 and is separated from the convecting meteoric water by a 0.5 km thick conducting shell. The crucial assumption in this model being that the upper 8.5 km of crust is sufficiently permeable to allow for the free convection of meteoric water between the surface and the top side of the conducting plate.

The gravity models of chapter 7 and section 8.1, and the seismic data of appendix F, suggest that a low-velocity, and therefore low-density, upper mantle exists at a depth of about 15 km beneath the Central Volcanic Region. A possible explanation for this anomalous upper mantle is that it is slowly convecting asthenospheric material at a temperature of about 1100°C . For example the numerical models of Andrews & Sleep (1974) suggest that convecting asthenosphere at this temperature will be present beneath a back-arc basin, at a depth of 10-20 km, after a period of about 10 My since the inception of subduction. If this is so, and assuming that the circulating asthenosphere is delivering heat to a conducting plate of area 5000 km^2 (the proposed area of hydrothermal convection - section 2.2.6), the required thickness of the plate can be calculated. Taking the temperature drop across the plate as being 800°C (i.e. $1100-300^\circ\text{C}$), and the thermal conductivity of the plate as being $1.5 \text{ W/m}^\circ\text{C}$, then a 1.7 km thick conducting layer would be required to maintain the average heat flow of 0.7 W/m^2 . However, for this heat flow model to be acceptable it must be assumed that the upper 13 km of crust for the Region is sufficiently permeable to allow for the free convection of meteoric water to act as an efficient heat transfer mechanism.

Banwell was also troubled by this aspect of the circulating magma beneath a hot plate model. He noted "...it would be quite insufficient for surface water to percolate down faults to a few parts of this area and return to the surface up the same or other faults." Thus, although there is an element of uncertainty about this model, it cannot be dismissed out of hand until more data is available pertaining to the possibility, or otherwise, of the movement of meteoric water through the upper crust of the Central Volcanic Region.

With this model the 2.60 Mg/m^3 "basement" layer would be greywacke, of a slightly lower density than normal for the 2-10 km depth range, and the genesis of the rhyolitic rocks can be explained by partial melting at the base of the 15 km thick crust.

8.2.3 Model 2 - Calhaem's model

The simple idea of supplying heat to the geothermal areas by a series of intrusions was rejected by Banwell (1963) on the grounds that the required volume of intrusives would be too large; it being assumed that the Region was underlain by sedimentary rocks at a depth of 2 km. However, Calhaem's (1973) plate tectonic model for the development of the Central Volcanic Region implies that it is the cooling of andesitic or basaltic rock, intruded into the Region as an accompaniment to spreading, that provides the necessary heat. He suggests that the spreading is one sided, or asymmetric, with the present day intrusion occurring along the eastern boundary of the Taupo Volcanic Zone. Under this scheme molten rock will be at a comparatively shallow level at the eastern boundary and the cooled magmatic crust will thicken to the west; a scheme which has since received support from an interpretation of microearthquake data (Evison *et al.*, 1976). Calhaem accepts the prevailing view, exemplified by the work of Ewart & Stipp (1968), that the rhyolitic rocks have resulted from a partial melting of Mesozoic greywacke rocks.

Because Calhaem's model precludes the existence of a continuous greywacke basement beneath the Region, he suggests that the rhyolites result from a partial melting of greywacke blocks that have become detached from the Hawke Bay proto-plate (see section 2.1.4) at the eastern margin of the Central Volcanic Region.

Although Ewart & Stipp (1968) specifically propose that the parent rock for the rhyolites be the marginal facies greywacke, then thought to be only found on the western side of the Region, recent geochemical work (Reid, 1982) has identified marginal facies greywacke on the eastern side of the Region. Hence, Calhaem's model does not appear to conflict with the geophysical or geochemical evidence, although no attempt was made by Calhaem to quantify his proposal. In particular, it should be noted that his proposal involves the partial melting of greywacke blocks at upper crustal levels, and therefore the intruded magma must be of a sufficient volume to supply enough heat to maintain the heat output over geological time scales, and in addition, sufficient heat to partially melt the greywacke blocks so as to produce the observed volume of erupted rhyolitic rocks.

Spreading within active oceanic back-arc basins is accompanied by the intrusion of basalt to form new oceanic crust (Karig, 1971; Lawver & Hawkins, 1978); the basalt is thought to be derived from the hydration and partial melting of upper mantle material (Ringwood, 1977). For the Central Volcanic Region where the solid surface is above sea level, and the characteristic gravity anomaly pattern consists of a deep central low with flanking gravity highs, the idea of vast volumes of high density, cooled basalts within the top 10 km of the crust appears doubtful. For example, Daly *et al.* (1966) give the mean density of 40 typical basalts to be 2.94 Mg/m^3 , and basalts from the Taupo Volcanic Zone have mean wet

and particle densities (table 8.1) of 2.62 and 2.94 Mg/m³ respectively. The large difference between these latter two figures, and hence the implied high porosity, can be explained by the highly scoriaceous character of these basalts; a characteristic which is thought to be acquired as the rising basaltic magma comes into contact with the water saturated volcanic rocks near the surface (Cole, 1979).

Table 8.1 Densities of some New Zealand volcanic and igneous rocks

<u>Rock type</u>	<u>Mean Density (Mg/m³)</u>		<u>n</u>	<u>Data source</u>
	wet	particle		
T.V.Z. andesite	2.54	2.62	57	Whiteford & Lumb (1975)
T.V.Z. K-trig basalt	2.62	2.94	8	"
T.V.Z. granodiorite	2.50	2.59	5	Chapter 4
Fiordland granites	2.62	2.66	15	Hatherton & Leopard (1964)

T.V.Z. = Taupo Volcanic Zone

n = number of samples

Unless basalt can exist in a highly scoriaceous, and therefore highly porous, form at depths greater than 2 km, it appears necessary to propose, within the context of Calhaem's plate tectonic model, that the intruded igneous rocks are of a medium density. The densities listed in table 8.1 suggest that an andesitic rock or a granite-granodiorite would suffice. Note that the rock listed as Taupo Volcanic Zone andesite in table 8.1 would, when cooled, presumably constitute a diorite. Dioritic rocks are generally more dense than andesites (Daly *et al.*, 1966), however, intruded andesitic rock that has been extensively leached and altered by hot hydrothermal waters may well become less dense than the original andesite (Cheng, 1976; see section 4.2.3).

Under this intrusion hypothesis the 2.60 Mg/m^3 "basement" layer can be ascribed to intruded andesitic rock derived by either partial melting and fractionation of upper mantle material (Ringwood, 1977), or from a partial melting of oceanic crust and assimilated sediments on the subducted plate (Cole, 1979). The rhyolitic rocks are, as suggested by Calhaem (1973), considered to be derived from a partial melting, at upper crustal levels, of detached greywacke blocks.

The 15 km depth to a transition from crust to low-density upper mantle, as inferred from the gravity model in Chapter 7, was corroborated by the seismic work discussed in appendix F. However, no uniqueness is claimed for the densities of 2.75 and 3.02 Mg/m^3 (fig. 8.5) used on either side of this boundary. Nevertheless, a higher density than 2.60 Mg/m^3 within the lower crust is required to satisfy the gravity data, and an explanation for this lower crustal layer may be obtained by giving consideration to the partial melt process that the greywacke rocks are thought to undergo in order to produce the rhyolitic volcanics. Recent work on greywacke melts by Reid (1982) suggests that a 35% partial melt, by volume, will produce, in a chemical sense, the best fit rhyolite; a higher density, refractory residue will remain. Hence, to a first approximation, for every unit volume of rhyolite produced in the partial melt process, two unit volumes of residue will also be produced. Accordingly, the 5 km thick 2.75 Mg/m^3 layer in the density model shown in fig. 8.5 is attributed to the refractory residue produced in the partial melt process.

8.2.4 Summary of the petrological models

Two distinctly different petrological models for the Central Volcanic Region have been discussed in this section. The first is based upon a

heat source in the form of convecting asthenospheric material, at a depth of 15 km, supplying heat to the Region via a thin conducting shell to convecting meteoric water within the upper crust. With this model the crustal rocks may be considered to be predominantly greywacke that have been extended and thinned by back-arc tensional forces. The second model involves a heat source in the form of mantle-derived andesitic magma that has been intruded at the eastern margin of the Region as an accompaniment to back-arc spreading.

Neither of these two models is entirely free of doubtful or unattractive features. The first model is dependent on greywacke rocks, to a depth of about 13 km, being sufficiently permeable to allow for the free convection of meteoric water to act as an efficient heat transfer mechanism. On the other hand, the second model is dependent on there being sufficient heat available from a continuous 8 km thick layer of intruded rock beneath the Region to i) maintain the heat output of the geothermal systems over geological time scales, and ii) partially melt a sufficient volume of greywacke rocks so as to produce the required volume of rhyolitic volcanics.

It would appear, therefore, that as the classical geophysical methods are limited in their ability to resolve a satisfactory petrological model for the Central Volcanic Region, a more effective way of obtaining constraints on the problem would be to carry out thermal modelling to test one, or both, of the models discussed above. This is a topic that goes beyond the scope of this study and consequently will not be attempted here.

8.3 Discussion: passive extension or active spreading?

It was the contention of Calhaem (1973) that the Central Volcanic Region is a site of active, asymmetric spreading accompanied by igneous intrusion at the eastern boundary of the Region. His proposal is based largely on the following evidence:

- i) The extraordinarily high heat output of the Region which cannot be explained by conductive heat transfer through the crust alone.
- ii) The asymmetric distribution of geothermal fields within the Region (fig. 8.6) and the tendency for the fields to young to the east.
- iii) The apparent, rapid migration of the andesite axis across the Central Volcanic Region during the period 4 My BP to the present (fig. 8.6).
- iv) The marked change in surface geology across the boundaries of the Central Volcanic Region (see section 2.1.4).

The principal contribution of the present study to this question is that the position of Calhaem's 4 and 0 My boundaries (fig. 8.6) coincides with a subsurface boundary, which from gravity (chapter 7) and seismic (appendix F) evidence is inferred to separate thin crust and low-density upper mantle within the Central Volcanic Region from a higher density upper mantle and thicker crust in the areas that flank the Region. The occurrence of a shallow low-density upper mantle beneath the Region, which could be equated with a low-density "thermal mantle diapir" that Karig (1974) proposes to be the driving force behind the formation of back-arc basins, strengthens the proposal of Calhaem - although cannot be taken as evidence that unequivocally confirms it.

Anomalously thin crust underlain by a low-velocity upper mantle is a feature also associated with continental rifts and ensialic back-arc basins where extension, but not necessarily active spreading, is thought to be occurring. This point is demonstrated by the comparison presented

Fig. 8.6 Map showing the asymmetrical distribution of geothermal fields and low-potash andesite ages within the Central Volcanic Region. The down flow region shown here is similar to that proposed by Allis (1979) (section 2.2.6) and is based on the three measurements of zero geothermal gradient located within the Central Volcanic Region but west of the active geothermal areas. The 4 and 0 My boundaries are those suggested by Calhaem (1973) and coincide with those of the Central Volcanic Region (section 2.2.1). The 2 My boundary is suggested here on the basis of the data displayed.

in table 8.2. Here various geophysical characteristics of the Central Volcanic Region are compared with those of three other ensialic back-arc basins for which the relevant geophysical data are available; these other basins are interpreted (Hovarth *et al.*, 1975, 1981; McKenzie, 1978) to be extensional in origin yet without actually being in a state of active spreading accompanied by the intrusion and creation of new lithosphere.

From table 8.2 it can be seen that although the characteristics of anomalously thin crust and a low-velocity upper mantle is common to all the back-arc basins listed, the average heat flow estimate for the Central Volcanic Region is substantially higher than that for the other basins. Some caution should be exercised, however, when it comes to making comparisons between these heat flow estimates, because i) the ages of the various basins differ, and ii) the average heat flow estimate for the Central Volcanic Region is based on the measured convective heat flow, whereas estimates for the other three basins are based on measurements of conducted heat. Nevertheless, it is notable that although the average heat flow estimate for the Central Volcanic Region is several times larger than that for the other back-arc basins listed in table 8.2, the heat output per km of strike length of volcanic zone is on a par with that for the active spreading centre of Iceland (section 2.2.6). Thus on the basis of this observation, coupled with those of Calhaem (1973) outlined at the beginning of this section, the most promising evolutionary model for the Central Volcanic Region, considered here, is one of active asymmetric spreading accompanied by the intrusion of mantle derived igneous rocks at the eastern boundary of the Region. The quantitative testing of this spreading proposal, however, must await an attempt at some detailed thermal modelling.

Table 8.2 A comparison of geophysical characteristics
 for some ensialic back-arc basins

	Crustal thickness (continental norm = 30-40 km)	Pn (continental norm = 8.1 km/s)	Heatflow (continental norm = 60 mW/m ²)	Age of oldest volcanism (My)
Basin & Range* (U.S.A.)	25-30	7.7-7.9	90	40
Pannonian Basin** (Mediterranean)	25	-	90-110	15
Aegean** (Mediterranean)	20-30	7.6-7.8	50-90	2
Central Volcanic Region (New Zealand)	15	7.4	350-700***	2-4

* Data from Scholz *et al.* (1971) and Pollack & Chapman (1977)

** Data from Hovarth *et al.* (1975, 1981)

*** The upper bound of 700 mW/m² for the Central Volcanic Region is based on assuming that the average heat flow for the eastern half of the Region is the average for the whole Region. The lower bound of 350 mW/m² assumes that the heat output of 3.5×10^9 W from the eastern half of the Region represents the total heat output of the Region i.e. 3.5×10^9 W/10000 km² = 350 mW/m² (see section 2.2.6).

CHAPTER 9 SUMMARY AND CONCLUSIONS

The conclusions of this thesis are presented in two parts. Firstly, work concerning the structure of the Central Volcanic Region is summarised. Secondly, some implications of the regional gravity analysis are discussed and suggestions made for future work.

9.1 Near surface structure of the Central Volcanic Region

The bore hole data analysed in chapter 4 provided important information on several counts. The data showed that appreciable lateral variations in density within the top 1 km of volcanic cover can occur, and these variations are capable of producing gravity anomalies that might otherwise be interpreted in terms of "basement relief". The bore hole data also showed a significant increase of density with depth such that the average density of the volcanic cover is predicted to approach 2.60 Mg/m^3 in the depth range of 2-3 km. This density, and the depth range where it is predicted to occur, are consistent with the seismic velocities of 4.8-5.5 km/s detected at depths of 1.8-2.2 km by the seismic refraction experiments discussed in chapter 5. Hence a variety of interpretations for these "basement refractors" are possible. For example, it cannot be determined from the seismic data alone if the 4.8-5.5 km/s velocities represent a distinct lithological boundary, compacted volcanics or intruded igneous rocks. The only bore hole that is coincident with any of the refraction lines, and also went deeper than the calculated depth of these basement refractors, was bore 121 at Wairakei; this hole terminated at a depth of 2.2 km in andesite of density 2.65 Mg/m^3 .

It has been hitherto assumed that the Central Volcanic Region is underlain by a continuous greywacke basement. Because of the similarity in density and seismic velocity for greywacke, on one hand, and andesite, compacted volcanics or intruded igneous rocks on the other, gravity and seismic methods are of limited use in the testing of this assumption. The question of what is the predominant rock-type for the crust of the Central Volcanic Region is possibly best answered by starting from the viewpoint that whatever petrological model is proposed, it must be consistent with, i) the anomalously high heat output of the Region and ii) the crustal thickness of about 15 km, as deduced from the gravity and seismic work reported in this study. On this basis, two extreme-case models were suggested in chapter 8 that could be tested by thermal modelling.

At the more detailed level the seismic refraction data presented and discussed in chapter 5 has furnished some important structural information as well as providing important constraints for gravity interpretations. Investigations about the Otamarakau greywacke outcrop show it to be the surface expression of an approximately circular, flat topped and steep-sided block that is surrounded by about 1.8 km of low-velocity volcanic rocks. Although the base of the block cannot be determined by seismic methods, the unusual subsurface geometry of the block lends support to the proposal of Calhaem (1973) that the block is allochthonous.

To the south of the outcrop is 1.8 km of low-velocity, and therefore low-density volcanic rocks. This is an important finding as a previous residual gravity interpretation (Rogan, 1980) had shown only 0.25 km of low-density volcanic rocks here. This discrepancy, in addition to the overall large variations between the three published residual

gravity anomaly fields for the Region (Chapter 3) emphasises the difficulty in obtaining a unique and a geologically meaningful residual gravity anomaly field for the Region by analytical means. The analytical approach to the separation of gravity anomaly fields into regional and residual components is useful when dealing with narrow features such as superficial sedimentary basins. For the Central Volcanic Region, however, which is 100 km wide at its northern end, and where a continuum of density irregularities is inferred to exist from the surface down to large depths, a clear wavelength separation between residual and regional gravity anomalies may not be possible. In order to obtain a coherent structural model for the Central Volcanic Region it appears necessary to consider all components of the observed gravity anomaly field.

At both Mangakino and Taupo second-order gravity lows of about $-100 \mu\text{N/kg}$ remained unexplained after the seismically determined thickness of low-velocity, low-density volcanic rocks, and the gravity effect of the proposed structural model for the central North Island, were allowed for (Chapter 8). These second-order, negative gravity anomalies were explained by discrete volumes of negative density contrast emplaced within the seismic basement. The geological interpretation of these negative mass anomalies is that at Taupo it represents a deep (7-9 km) body of molten rhyolite, whereas at Mangakino the low-density mass anomaly represents a partially molten to cooled body of rhyolite in the depth range of 2.2-8 km. Given that some 14000 km^3 of rhyolitic rocks have been erupted from the Region in the last 1 My, the existence of molten rhyolite at shallow crustal depths poses no serious geological problems. However, in view of the generally held belief that magmatic activity within the Region has migrated southward and eastward with

time (Kear, 1959; Calhaem, 1973) the possible occurrence of molten magma beneath Mangakino, at the western margin of the Central Volcanic Region, remains an enigma.

9.2 Regional gravity analysis of the central North Island

The regional model presented in chapter 7 represents the first quantitative attempt to explain the three-dimensional gravity anomaly pattern associated with the central North Island. The model is a first-order one in that the principal constraint on the model is that it be an equilibrium one. However, despite the gross simplicity of the model it is useful in that, a) the calculated gravity effect of the model provides a reasonable fit to the observed gravity anomalies and, b) various structural aspects of the model provide qualitative explanations for the variations in the observed heat flow and anomalous seismic properties for the upper mantle of the central North Island.

Important features of the model are:

(i) It predicts a crustal thickness of about 15 km for the Central Volcanic Region and 22-25 km for the northwestern North Island; these areas of anomalously thin crust are inferred to be locally compensated by low-densities within the upper mantle.

(ii) The principal gravity anomalies of the North Island, in particular the 650 $\mu\text{N/kg}$ back-arc gravity high over the northwestern North Island, can simply be explained by gravity edge-effects associated with the Egmont-Ruapehu-Opotiki and Central Volcanic Region boundaries.

(iii) The gravity effect of an excess mass associated with the subducted plate is deduced to make only a minor contribution to the observed back-arc gravity high - about 120 of the observed 650 $\mu\text{N/kg}$.

(iv) The tectonic interpretation of the regional model is that the western North Island, north of the Egmont-Ruapehu boundary, has been subjected to back-arc asthenospheric convection and associated lithospheric and crustal thinning since as far back as the early Miocene. The extremely thin crust and lithosphere of the Central Volcanic Region is possibly the direct result of active spreading and the creation of new crust and lithosphere. South of the Egmont-Ruapehu boundary subduction is at a comparatively juvenile stage and it is accordingly inferred that back-arc convection has not had sufficient time to thermally weaken and thin the lithosphere.

Hence, the unusual situation of highly three-dimensional gravity anomaly, heat flow and upper mantle seismic attenuation patterns observed above a two-dimensional subduction zone is considered to be a reflection of the time-lag between the onset of subduction, on one hand, and the initiation of back-arc asthenospheric convection and associated lithospheric and crustal thinning on the other.

The gravity anomalies of the central North Island are explained in this study by the anomalous gravity effects associated with the following three structures: the thin, but compensated, crust on the concave, or northwestern, side of the Egmont-Ruapehu-Opotiki boundary; the even thinner crust of the Central Volcanic Region and the subducted Pacific plate. If either of the first two structures could be determined unambiguously by deep seismic sounding then the remaining two could be resolved by the gravity interpretation methods outlined in chapter 7. In this regard the preliminary results of the long-range refraction experiment reported in appendix F appear highly promising, and moreover, further such seismic work within and west of the Central Volcanic Region is logistically feasible providing suitable deep-water shot points can be found.

Thus, the North Island of New Zealand represents an unusually good location to carry out further geophysical studies that could ultimately lead to a precise determination for the gravity effect of a subducted lithospheric plate.

APPENDIX A THE ANALYTICAL DERIVATION OF A RESIDUAL GRAVITY
ANOMALY FIELD FOR THE CENTRAL VOLCANIC REGION

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Regional and residual gravity fields, central North Island, New Zealand

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ABSTRACT

Regional components of the central North Island Bouguer and isostatic gravity anomaly fields are presented as third-degree, two-way polynomials. The Bouguer regional contours strike north-east, except in the Bay of Plenty region, where contours parallel the coast. The isostatic regional contours strike uniformly north-east and go through a maximum which coincides with the axis of epicentres of intermediate depth earthquakes.

The residual isostatic field shows four prominent negative anomalies in Central Volcanic Zone; three of these coincide with centres of rhyolitic volcanism.

KEYWORDS Gravity interpretation, Central Volcanic Zone.

INTRODUCTION

Maps of North Island isostatic and Bouguer gravity anomalies (1:1 000 000) published by Reilly *et al.* (1977), display a strong north-west-south-east gradient in, and to the east of, Central Volcanic Zone. Superimposed on this gradient are many short-wavelength anomalies presumed to be due to structures in the shallower layers of the lithosphere. In order to interpret these anomalies it is usual to first subtract the regional field.

A significant regional component in Bouguer anomalies is removed by the application of a topographic-isostatic correction; this removes the gravitational effect of the transition from continental to oceanic lithosphere (Reilly 1972). The effect is illustrated by a comparison of the 1:4 000 000 Bouguer and isostatic anomaly maps (Reilly 1965) in which the strong gradient of the Bouguer anomaly from positive values at the coast to negative values inland, particularly in the South Island, is largely absent from the isostatic anomaly. A further example of this effect is shown by Hatherton (1966). He found in Southland that the broadest significant contribution to the Bouguer regional is the effect due to isostatic compensation, and he therefore adopted the isostatic anomaly map itself as a map of primary residuals.

Departures from the assumed isostatic model may still cause regional trends to be present on the isostatic anomaly map. These may be removed by graphical or analytical means.

Various New Zealand studies (e.g., Hatherton 1952; Dibble & Suggate 1956; Modriniak & Stude 1959) adopted regionals constructed graphically by interpolation of hand-drawn curves between gravity values measured on Torlesse Group greywacke. The Torlesse Group is a suitable suite of rocks on which to base a gravity regional as it is widespread throughout New Zealand and exhibits only small variations in density (Hatherton & Leopard 1964).

Nettleton (1976) provides an extensive discussion on the relative merits of graphical and analytical methods of determining the regional. Regionals obtained analytically by interpolation of gravity data

from a contour map to a regular grid, such as second-derivative and upward-continuation methods, have been found to be highly dependent on the mode of contouring and the grid spacing. On the other hand, as graphical methods leave scope for personal judgment, a unique regional is difficult to obtain. The purpose of this study is to illustrate how, in an area such as Central Volcanic Zone, the regional-residual separation may be performed in a manner which bridges the gap between subjective graphical methods and purely analytical methods.

THE REGIONAL POLYNOMIAL

An interpreter who constructs a smooth gravity field from anomaly values obtained on basement outcrops either side of a sedimentary basin, and calls this smooth surface the regional, is in effect fitting a gravity field to those areas he believes to be least effected by local anomalies. The method adopted here follows the suggestion of Skeels (1967): a polynomial of low order is fitted by least squares to gravity values obtained at stations situated on a geologically uniform basement.

Johnson (1974) used a terminated Fourier series, orthogonalised with respect to irregularly spaced data, to represent the regional Bouguer gravity field of Tasmania. Crain (1970), in a review of computer interpolation techniques of two-dimensional data, found the simplest mathematical form to express a surface generated by non-equispaced data to be the non-orthogonal polynomial. For surfaces of order greater than five, however, there are advantages in using the computationally more complex orthogonal polynomials.

SELECTION OF THE DATA SET

Figure 1 shows the locations of 114 gravity stations, situated on Torlesse Group greywacke, the anomaly values of which were used for the computation of the regional field. In order to adequately sample any curvature of the regional gradient, data points were incorporated across as wide a region as possible, particularly east of Central Volcanic Zone, where it was suspected that curvature is high.

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FIG. 1—Locality map of Central Volcanic Zone showing positions of gravity stations used in the regional computation. The boundary of Torlesse Group greywacke is indicated by the dashed line.

Greywacke crops out near Otamarakau on the Bay of Plenty coast. Gravity readings from here were excluded from the data set, however, as there was some uncertainty as to the origin and nature of this outcrop. Extensive drill-hole data at Kawerau, near the eastern boundary of Central Volcanic Zone, has afforded a detailed geological section down to greywacke (Browne 1978). A north-west-south-east section was taken across the Kawerau geothermal area. The gravity deficit, due to low density volcanic rocks overlying greywacke, was calculated by assuming this section to continue indefinitely in the north-east-south-west direction and by using measured densities of drill-core samples (Macdonald *et al.* 1970). At three gravity stations the deficit was added to surface gravity

anomalies, and the resulting estimates of regional gravity were included in the computation.

SELECTING THE POLYNOMIAL

A programme was used for fitting a two-way polynomial to the data by least squares. Independent variables X, Y are the eastings and northings, respectively, of the gravity stations in North Island National Yard Grid co-ordinates; $G(X_i, Y_i)$ is the set of gravity anomaly values used in the regional computation ($i=1 \dots N$); and dependent variable $P_m(X, Y)$ is a polynomial of order m , with p coefficients, representing the regional gravity field in $\mu\text{N/kg}$.

APPENDIX A THE ANALYTICAL DERIVATION OF A RESIDUAL GRAVITY ANOMALY FIELD FOR THE CENTRAL VOLCANIC REGION

STERN - GRAVITY FIELDS, NORTH ISLAND

The following are calculated:

$$P_m(X, Y) = \sum_{i=0}^m \sum_{j=0}^i (a_{i-j, j}) X^{i-j} Y^j \quad (1)$$

$a_{i-j, j}$ are coefficients derived in the least squares procedure.

The residual at station i ;

$$R_i = G(X_i, Y_i) - P_m(X_i, Y_i) \quad (2)$$

An estimate of the error variance;

$$V_m = \sum_{i=1}^N (R_i)^2 / (N-p) \quad (3)$$

and $\sqrt{V_m}$ is the r.m.s. residual.

In selecting the order of polynomial the following procedure was carried out.

1. Polynomials from order zero to five were estimated.
2. Values of R_i were examined. Random errors in $G(X_i, Y_i)$ can arise from observational error, incorrect terrain correction, or error in the calculated elevation of the particular gravity station. If a station on greywacke, particularly near the margin of the volcanic zone, is close to a steep contact with, or faulted against, low density volcanic sediments, its gravity value would be expected to be below that of the estimated regional value. Thus a negative value of R_i would result.

Any R_i greater than $3\sqrt{V_m}$ was attributed to one of the above causes, hence the associated gravity station was rejected from the regional computation. In all, ten stations of the initial data set failed to meet this criterion and were replaced by stations chosen so that a uniform spatial distribution of stations was maintained. The locations of the ten rejected stations tended to be in the more isolated regions of the greywacke ranges where elevation and terrain control are the most difficult.

3. An F-test (Dixon & Massey 1969) was applied to determine the significance of coefficients. The application of this test allows the order of polynomial that most adequately incorporates the regional trend to be determined from the data and not be pre-assigned arbitrarily.

RESULTS

For both Bouguer and isostatic data sets it was found that going from a third to a fourth order polynomial did not produce a change in the variances which was significant at the 95% level. Figure 2 shows a comparison of the variance of residuals with polynomial order.

Coefficients in expression (1) for the third order polynomial estimates of the Bouguer and isostatic regional fields respectively are shown in Table 1.

FIG. 2—Normalised error variance versus order of polynomial. Calculated for polynomial estimates, order zero to five, of the isostatic regional field.

TABLE 1—Coefficients for third order polynomial estimates of the Bouguer and isostatic regional fields.

Coefficient	Bouguer	Isostatic
a00	0.5547×10^3	0.6271×10^3
a10	-0.1287×10^{-2}	-0.2146×10^{-2}
a01	0.4898×10^{-2}	0.1553×10^{-2}
a20	-0.2452×10^{-7}	-0.3286×10^{-7}
a11	0.6189×10^{-7}	0.4827×10^{-7}
a02	-0.1773×10^{-7}	-0.2362×10^{-7}
a30	-0.1512×10^{-12}	-0.7327×10^{-13}
a21	0.1774×10^{-12}	0.1534×10^{-12}
a12	-0.9093×10^{-13}	-0.2212×10^{-13}
a03	0.3361×10^{-13}	0.4969×10^{-13}

The origin was chosen for each polynomial to be approximately central with respect to the data set, i.e. 375 000 yards east, 510 000 yards north. Figures 3 and 4 show plots of the isostatic and Bouguer regional fields respectively. The residual isostatic gravity anomaly map (Fig. 5) was derived by hand-contouring values of the residual anomaly calculated from approximately 5000 gravity stations whose positions are shown on sheet 5 (Woodward & Ferry 1973) and sheet 8 (Woodward & Ferry 1974) of the 1:250 000 New Zealand Gravity Map series.

FIG. 3—Isostatic regional gravity field. Contour interval = 50 $\mu\text{N/kg}$.

FIG. 4—Bouguer regional gravity field. Contour interval = 50 $\mu\text{N/kg}$.

FIG. 5—Isostatic residual gravity field. Contour interval = 50 $\mu\text{N/kg}$. Dotted areas are less than -500 $\mu\text{N/kg}$, hachured areas are greater than 0 $\mu\text{N/kg}$.

INTERPRETATION

A comparison of Fig. 4 and Fig. 3 reveals that while both Bouguer and isostatic regional fields are dominated by a north-east trend, the Bouguer regional field incorporates an additional trend, most prominent at, and parallel to, the Bay of Plenty coast. This Bouguer anomaly pattern, which has values positive at the coast and decreasing inland as previously noted for the South Island, reflects the transition of continental to oceanic lithosphere.

The envelope of earthquake epicentres for earthquakes occurring between 100 km and 150 km beneath the North Island (Adams & Ware 1977) coincides with the 600 $\mu\text{N/kg}$ regional isostatic high in Fig. 3. The axis of this gravity high is also close to Mooney's (1970) boundary between regions of high and low attenuation of high frequency seismic energy. These correlations further support associations of gravity and seismic phenomena found in asymmetric active regions, as discussed by Hatherton (1970).

The closing of contours in Fig. 3, off-shore Bay of Plenty, may be a consequence of lack of data in this region. Indeed, if these contours were left open they would be more consistent with the limited data available from the Lau-Havre trough (Talwani *et al.* 1961), which indicates a positive isostatic field in this region.

Although no detailed analysis of the residual map is attempted here, some general features may be noted:

1. There are four regions of residual anomaly below $-500 \mu\text{N/kg}$: Mangakino, northern end of Lake Taupo, southern end of Lake Rotorua, and Okataina Volcanic Centre. The latter three coincide with the rhyolitic volcanic centres of Maroa, Rotorua, and Okataina respectively. These three regions are associated with total force magnetic anomalies greater than 300 nT taken from an aeromagnetic survey flown at 1525 m altitude (Whiteford 1976). On the other hand, while the Mangakino low is an especially impressive feature of $-600 \mu\text{N/kg}$, it has no associated magnetic anomaly or known volcanic centre.

2. At Otamarakau, where greywacke crops out, the residual anomaly is zero. This implies that this outcrop is part of an extensive greywacke basement, or else that it is underlain by rocks with effectively the same density as greywacke.

3. There is a broad region of relatively high gravity north-east of the Mangakino low. This is probably genetically related to a magnetic high (Whiteford 1976) that extends over the same region.

CONCLUSIONS

The technique of fitting a polynomial of low order to data from gravity stations on basement appears to be a plausible, non-subjective, and reproducible method for constructing the regional gravity field of the central North Island. The regional field is expressed in closed mathematical form so that evaluation at a particular location may be carried out uniquely and quickly.

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APPENDIX B

Measurements and results of ultrasonic velocity determinations:

All samples selected for this study were homogeneous and lacked obvious flaws and fractures. Borehole samples were generally about 50 mm long and cylindrical in shape with a diameter of about 50 mm. Each velocity determination was made across a pair of faces cut parallel and polished smooth. Velocities were determined using the travel time of a ultrasonic pulse of frequency 82 kHz. Pulse generation and timing were by means of a "PUNDIT" (portable ultrasonic non-destructive digital indicating tester). The transducers used were barium titanate in stainless steel cases, the energising pulse being 800 V peak for 2 μ sec. Calibration was by means of a calibration rod supplied by the manufacturers with the stated travel time of 26.3 μ sec.

Good contact between the transducer faces and the sample were necessary to give reliable, consistent readings. Vaseline was used as a coupling between the faces then moderate pressure applied by hand until a steady reproducible reading was attained. Transducers were placed exactly opposite each other; the distance measured by calipers. A series of readings were taken with the least time, corresponding to the highest velocity, being the value accepted. Samples were saturated for 24 hours under reduced pressure so that the velocities measured could be plotted against saturated densities.

Reliability of results:

The source of greatest error in the ultrasonic velocity determinations is in measuring the travel time of the pulse. Most

specimens were of a size to give travel times of 5-30 μsec . Travel times were measured to 0.1 μsec ; the stated accuracy of the instrument. A maximum error of about 2% would arise from this source. Distance measurements are a much smaller source of error; they can consistently be made to 0.1 mm, which, for intervals as small as 20 mm represents an error of only 0.5%. Thus the maximum possible error for any velocity determination is estimated to be about 5%.

Table B.1 lists ultrasonic velocity, density (wet), depth of the sample if it is from a bore hole and code for its lithological classification.

Table B.1 Ultrasonic seismic velocity versus density data for
rocks from the Central Volcanic Region.

V_p (km/s)	ρ (Mg/m ³)	Depth (m)	Code	V_p (km/s)	ρ (Mg/m ³)	Depth (m)	Code
2.72	2.42	506	3	3.05	2.29	24	2
3.54	2.18	445	4	2.24	2.16	0	2
2.84	2.29	778	3	4.16	2.46	0	5
3.16	2.42	843	3	2.24	2.13	1051	3
3.30	2.47	973	2	2.38	2.35	0	4
2.08	2.16	823	3	2.12	1.70	91	3
3.30	2.40	0	6	4.18	2.43	149	2
4.04	2.67	0	6	3.21	2.21	1132	3
4.30	2.46	0	6	0.80	1.96	0	2
1.29	1.79	0	2	1.28	2.25	0	2
2.79	1.97	0	2	4.82	2.73	0	6
3.32	2.18	0	2	2.00	1.87	7	4
5.15	2.73	50	7	1.90	2.18	929	3
5.07	2.75	60	7	2.80	2.31	644	6
4.57	2.65	20	7	1.85	2.11	838	3
1.30	2.01	23	2	3.21	2.25	0	4
1.58	2.09	11	2	2.68	2.27	12	2
1.51	2.05	11	2	4.55	2.66	0	7
3.06	2.23	651	3	1.75	1.70	97	1
4.65	2.72	0	7	0.75	1.78	146	1
5.05	2.72	35	7	1.79	1.66	119	3
3.94	2.23	222	3	2.97	2.45	1.46	2
1.63	1.84	152	3	4.76	2.63	0	5
3.97	2.42	134	4	1.11	1.79	0	3
2.61	2.18	556	4	1.30	1.71	76	3
3.81	2.38	61	2	4.14	2.41	0	6
3.02	2.30	41	2	0.93	2.08	15	1
5.36	2.69	0	7	4.74	2.65	0	7
1.88	2.20	15	2	3.05	2.40	119	2
3.15	2.46	0	6	2.26	2.33	0	3
3.52	2.23	91	4	3.75	2.34	30	4
2.50	2.29	0	4	2.44	1.75	37	1
1.92	2.14	853	3	2.93	2.30	250	3
3.20	2.35	686	3	2.40	1.77	98	1
5.00	2.64	0	7	1.82	1.69	0	4
1.37	1.84	174	3	3.06	2.29	1003	1
2.02	2.26	9	2	5.07	2.74	0	6
4.87	2.64	0	6	0.76	1.67	0	1
2.30	2.21	0	5	2.42	2.21	609	4
2.32	2.08	164	8	3.97	2.70	0	6
3.98	2.46	0	6	1.67	2.05	748	4
3.46	2.51	0	4	2.54	2.19	930	3
4.25	2.64	0	6	2.84	2.29	730	8
3.22	2.51	0	5	2.58	2.09	274	4
2.14	1.79	0	4	1.92	2.19	792	3
3.18	1.91	0	3	2.66	2.21	883	2
1.48	1.95	308	1	2.58	2.21	851	2
1.12	1.88	0	2	1.99	2.06	652	3
5.26	2.72	0	7	4.74	2.48	550	3

/continued

Table B.1 continued

V_p (km/s)	ρ (Mg/m ³)	Depth (m)	Code	V_p (km/s)	ρ (Mg/m ³)	Depth (m)	Code
1.06	1.51	0	3	2.03	1.61	0	3
4.42	2.48	0	6	3.09	2.14	364	4
3.07	1.91	124	3	3.00	2.17	290	4
5.26	2.65	0	6	2.23	2.36	459	4
1.71	1.86	608	3	1.70	2.11	686	3
4.88	2.52	576	3	5.24	2.69	0	7
2.98	2.22	336	4	5.39	2.73	0	7
3.36	2.14	1005	2	5.35	2.71	0	7
3.43	2.25	260	4	4.98	2.75	0	6
1.96	2.19	851	1	5.35	2.75	0	6
1.76	2.04	14	8	1.34	1.73	0	2

Key to Code:

- 1 Pumice and ash
- 2 Ignimbrite
- 3 Volcanic breccia
- 4 Rhyolite
- 5 Dacite
- 6 Andesite
- 7 Greywacke
- 8 Mudstone and sandstone

Least squares fit to the data, less the greywacke samples is

$$\rho = 1.62 + 0.21 V_p$$

$$r^2 = 0.65$$

and standard deviation of the residuals is 0.16 Mg/m³.

$$d\rho/dV_p = 0.21 \text{ Mg/m}^3/\text{km/s}$$

Table B.2 The empirical seismic velocity-density relationship
of Nafe-Drake (see Grant and West p.200) for the
velocity range of interest to this study.

V_p (km/s)	ρ (Mg/m ³)	V_p (km/s)	ρ (Mg/m ³)
1.6	1.70	3.6	2.33
1.7	1.79	3.7	2.34
1.8	1.85	3.8	2.37
1.9	1.90	3.9	2.38
2.0	1.95	4.0	2.39
2.1	2.00	4.1	2.40
2.2	2.02	4.2	2.42
2.3	2.06	4.3	2.43
2.4	2.09	4.4	2.44
2.5	2.12	4.5	2.46
2.6	2.14	4.6	2.47
2.7	2.17	4.7	2.48
2.8	2.20	4.8	2.50
2.9	2.21	4.9	2.52
3.0	2.23	5.0	2.53
3.1	2.24	5.1	2.56
3.2	2.26	5.2	2.58
3.3	2.28	5.3	2.60
3.4	2.30	5.4	2.62
3.5	2.31	5.5	2.64

Least squares fit to the data: $\rho = 1.58 + 0.20 V_p$

$$r^2 = 0.95$$

$$d\rho/dV_p = 0.2 \text{ Mg/m}^3/\text{km/s}$$

Appendix C - Equipment, statistical uncertainties and data from
shallow velocity investigations

C.1 Equipment

The following seismographs were used for the refraction surveys described in Chapter 5.

- 1) A 12 channel R.S.44 refraction seismograph with a spread length of 330 m, extendable to 1.1 km with 400 m cables at each end.
- 2) Four model M.E.Q.600, Sprengnether microearthquake recorders which record unattended for 26 hours at a drum speed of 60 mm/minute. These instruments have a normal peak displacement magnification of 5.2×10^7 at 23 Hz. In the field the recorders are usually operated at 36 db below full gain resulting in a displacement magnification of 8.5×10^5 . The remaining M.E.Q.800 recorder is capable of 50 hours unattended recording at a drum speed of 60 mm/minute. Average displacement magnifications in the field with this instrument were also about 8.0×10^5 , peaking at 14 Hz. For all the shots fired in this experiment, both the M.E.Q.600 and 800 had their drum speeds increased to 120 mm/minute.

All the Sprengnether recorders incorporate a quartz clock which when used in conjunction with radio-time signals provides a timing uncertainty no greater than ± 0.1 sec. These instruments automatically record time-signals broadcast on the hour by New Zealand time service.

3) Two model PS-1 and three model PS-1A Kinematics micro-earthquake recorders were also used in this study. Both types of instruments had displacement magnifications of approximately 7.5×10^5 and similar filter settings. Filter 2 (5 Hz peak) was used for sets within 8 km of an explosion and for distances greater than 8 km Filter 1 (2 Hz peak) was preferred. The internal clocks of these sets were calibrated by the operator, who recorded either the V.N.G. or Radio New Zealand time signals against the internal clock minute marks. This generally was done some hours prior to or after the main explosions. However two of the Kinematic instruments were manned at the time of the explosion so that the New Zealand time signals could be recorded against the hour mark immediately prior to the explosion.

Most of the Kinematics had drum speeds of 60 mm/minute although at the time of the Rotoehu survey two of the Kinematics had been modified so that they had drum speeds of 600 mm/min.

All microearthquake records were read with a travelling microscope which provided a reading accuracy of about ± 0.0025 mm; for the slowest drum speed of 60 mm/min this is equivalent to a reading accuracy better than ± 0.01 sec.

All Sprengnether and Kinematic recorders were operated with identical 1 Hz vertical seismometers (Mark products, model L4).

4) Two portable slow motion magnetic tape recorders used in the study are similar and have been described by Dibble (1964). A Willmore mark 1 vertical seismometer, $T=1$ sec, was used with each recorder. A tape speed of 0.4 mm/sec enables recording frequencies in the range 0.1-30 Hz to be played back with flat response. The magnetic tapes were played back at 196 mm/sec on a four-channel tape

recorder. The channels were then displayed two at a time on a dual-beam digital storage oscilloscope from which time mark corrections and seismic arrival times were read off. The overall timing accuracy of the recording and playback system is about 0.03 s.

C.2 Errors and minimum statistical uncertainty in refraction interpretation

Errors in refraction interpretation are here classified as model or data consequent.

- a) Model errors: These arise when the assumed subsurface model of either plane isovelocity layers or continuous velocity with depth is inapplicable. As discussed in section 5.2.2 both low-velocity and blind zones are departures from the above models and will cause errors in depth estimates. For this reason a complete statistical treatment of refraction errors is placed out of reach.
- b) Data consequent errors: Under this classification errors arising from a failure of the chosen model to fit exactly the observed data is considered. For example if a layered isovelocity model is assumed and straight-lines fitted to the data, there will be a scatter of data about these lines and the problem is then a statistical model fitting exercise. The uncertainty estimated will be a minimum in the sense that for the given model the uncertainty is at least this large. The numerical method used in this study is taken from Steinhart & Meyer (1961) and only the salient points will be reproduced here.

For N number pairs in distance and time (X_i, T_i) a straight-line segment of the form: $T = T_i + X_i/V_{app}$, $i=1, \dots, N$ is found. Parameters of interpretative interest are V_{app} (apparent velocity),

T_0 (time intercept at zero distance) and for the case of reversed profiles T_E (reciprocal or end time).

a) Estimate of uncertainty in $(V_{app})^{-1}$: The best estimate, in a maximum likelihood sense, of the reciprocal apparent velocity is obtained by: $S_m = (Stt/Sxx - (Sxt^2/Sxx^2)/(N-2))^{1/2}$

where: S_m = estimate of the standard deviation of the reciprocal apparent velocity $(V_{app})^{-1}$

$$S_{xx} = \sum X_i^2 - (\sum X_i)^2 / N$$

$$S_{xt} = \text{covariance } (X_i, T_i) \quad i = 1, \dots, N$$

If M is the true value of the slope m , $(M-m)/S_m$ is distributed according to a student's t -distribution. Thus $M = m \pm S_m \tau_{N-2}$ where τ_{N-2} is the numerical value of the t -distribution with $N-2$ degrees of freedom at some pre-selected level of probability. A close estimate for the uncertainty in V_{app} can then be obtained from $(V_{app}^{-1} \pm \Delta V_{app}^{-1})^{-1}$.

b) Uncertainty in the prediction of T_n : It can be shown (Steinhardt & Meyer, 1961), that:

$$\Delta T_x = \tau_{N-2} S_m (S_{xx})^{1/2} \left[\frac{1}{N} + \left(X - \frac{\sum x_i}{N} \right)^2 / S_{xx} \right]^{1/2}$$

where: ΔT_x = estimate of uncertainty for time intercept at position X .

The estimates of uncertainties for both T_0 and T_E can therefore be obtained by inserting different values of X into the equation for ΔT_x .

The procedure adopted for this study is to tabulate for each straight line determination the uncertainty in V_{app} , T_0 and T_E . The uncertainties for determinations with only two time-distance pairs is essentially infinite and the uncertainty in T_E is only given for line segments with a reversed counterpart. Included in these measures of uncertainty will be perturbations in the data from random reading errors, horizontal velocity variations and deviations of refracting interfaces

from horizontal planes. The uncertainties are useful for deciding whether to further complicate a postulated model (i.e. choose a dipping rather than a horizontal layered model) and to make estimates for the minimum uncertainty in depth determinations when using the various time intercept formulae given in Chapter 5.

C.3 Shallow velocity investigations along the Mangakino and Rotoehu seismic lines

For both the Rotoehu and Mangakino seismic lines, shallow velocity information was obtained from short spreads shot with the R.S.44 refraction seismograph. This information was then utilised for elevation and weathering corrections to arrivals from the main shots.

Equipment organisation: At most sites the 12 geophone, 330 metre long spread was end shot from both directions with shot offsets of 15-30 metres. In some cases offset shots of 400 metres were made to gain deeper structural information. For these long spreads end geophones were extended, with the aid of 400 metre extension cables in either direction, to give an end-to-end spread length of approximately 1.1 km.

Shots were fired in hand auger dug holes approximately 2 metres deep. Shot weights were 0.8 and 5 kg for spreads of 350 and 1100 metres, respectively.

Seismic recordings: Examples of some recordings made on the R.S.44 seismograph are shown in fig. C.1. Arrival times were obtained by reading the first onset, or in the cases where this was difficult, by reading the first peak and subtracting the average time interval between well-recorded onsets and peaks on that record.

6

Fig. C.1 Three examples of records from the 12 channel RS44 seismograph.

Interpretation methods: The main object in the interpretation of these lines was the identification of characteristic near-surface velocity units and their corresponding depths. For each set of corrected time-distance pairs from a particular shot, straight line segments were chosen which could be matched with straight line segments from its reverse shot with equal reciprocal end times.

Satisfactory sets of lines were interpreted to find depth, and true velocity using the time intercept method of Dobrin (1976) (see Section 5.3.3). For two of the short spreads reported here there was sufficient data to use the plus-minus method of Hagedoorn (1959) over the central portion of the spreads. Interpretations made with this method are very similar to those made with the plane layer interpretation method.

Tables C.1 to C.6 list observed and where appropriate, corrected travel times for all shallow velocity spreads on the Rotoehu and Mangakino seismic lines. Beneath each table is a list of the derived parameters from least square solutions for selected straight line segments. These parameters are:

V_{app} = apparent velocity

T_o = time intercept at zero distance

T_E = for reversed spreads, T_E = time intercept at position of the reverse shot.

ΔV_{app} , ΔT_o , ΔT_E = respective minimum statistical uncertainties at the 90% confidence level.

Mangakino shallow velocity spreads: Two sites, Murry Rd and Christiansen Rd sites, were found to be suitable for shallow-velocity investigations along the Mangakino seismic line. At both sites two

reversed spreads of lengths 330 and 1100 metres were shot.

(a) Murry Road spreads: Refraction arrivals at this site tended to be weak and for the long spread of shot M.S.3 energy failed to propagate the full length of the profile (1100 metres). The end time for this shot was adopted from its reversed shot M.S.4. Travel-time data and a plane layer interpretation of the data are shown in fig. C.2

(b) Christiansen Road spreads: Data for these spreads were good and the interpretation is simple and reliable. The long-spread shot, M.S.9, and geophone 12 were at a lower elevation than the rest of the spread, necessitating a small time correction to all arrivals; effectively placing shot and geophone on the same datum as the rest of the spread. Fig. C.3 shows the travel-time data and structural solutions obtained by both the horizontal plane-layer and the plus-minus method.

Rotoehu shallow velocity spreads: Two principal surface lithological units along the length of the Rotoehu profile were evident. Fluvial deposits within the vicinity of the two main shots and poorly-compacted Rotoiti pumice breccias elsewhere.

Accordingly, two shallow investigation sites were selected; a reversed 760 metre spread on Hannon Road and two unreversed profiles with respective shot offsets of 90 and 590 metres along Cambell Road. The latter Cambell Road profile served a dual purpose in that it was shot from the same location as R.S.2; providing extra data points for the time-distance graph associated with R.S.2.

(a) Hannon Road spreads: The pattern of geophone and shot layout at this site gave poor control on the top two layers but good control on the third. Fig. C.4 shows the travel time data and structural solutions

obtained by both the plane-layer and the plus-minus method.

The polarity of arrivals at the first geophone for each shot were examined for normality as 0.3 km/s is close to the speed of sound. They were normal in both cases. Poor control on the 0.8 km/s layer could lead to a more simple two layer interpretation for this spread. Such an interpretation would have a 0.3 km/s layer, 60 metres thick, directly overlying the 2.2 km/s layer. This interpretation was rejected in favour of the three layer model for the following reasons. Firstly, the influence of the water table and lithostatic compaction would make it unlikely that the 0.3 km/s layer would extend for 60 metres and secondly, the proposed three layer model is commensurate with that found at Mangakino where shallow velocity control was superior.

(b) Cambell Road spreads: Single straight line segments could be satisfactorily fitted to the data from these spreads. Velocities of 1.55 and 1.76 km/s were indicated by spreads with shot offsets of 90 and 590 metres respectively. Ray paths for the latter spread would be sampling to deeper levels, hence the difference in velocities is ascribed to an increase in velocity with depth.

Murry Rd. spread

Fig. C.2 Short spread data from Murry Rd and a simple plane-layer interpretation of the data. Note that there is no arrival from the south shot for the longer of the two spreads.

Christiansen Rd. spread

Fig. C.3 Data from the Christiansen Rd. spread. A plane-layer interpretation is shown and an interpretation by the plus-minus method of Hagedoorn (1959) is shown for the central portion of the spread.

Fig. C.4 Short-spread data from Hannon Rd (Rotoehu survey). A plane-layer and a "plus-minus" interpretation for the central portion of the spread are shown.

Table C.1 Seismic data for the Murry Rd short spread

Shot: M.S.5

Location: North end of short spread on Murry Road

Geophone No.	Distance (m)	Travel time (s)
1	22	0.065
12	52	0.085
11	82	0.120
10	112	0.150
9	142	0.180
8	172	0.230
7	202	0.235
6	232	0.260
5	262	0.270
4	292	0.285
3	322	0.295
2	352	0.307

Least squares solutions:

	n	V_{app} (km/s)	ΔV_{app} (km/s)	T_O (s)	ΔT_O (s)	T_E (s)	ΔT_E (s)
line 1	2*	0.3					
line 2	5	1.0	± 0.1	0.04	± 0.02	0.44	± 0.05
line 3	7	2.2	± 0.2	0.15	± 0.01	0.31	± 0.01

Shot: M.S.6

Location: South end of short spread on Murry Road

Geophone No.	Distance (m)	Travel time (s)
2	13	0.050
3	43	0.095
4	73	0.135
5	103	0.165
6	133	0.202
7	163	0.220
8	193	0.235
9	223	0.265
10	253	0.275
11	283	0.280
12	313	0.295
1	343	0.305

Least squares solutions:

	n	V_{app} (km/s)	ΔV_{app} (km/s)	T_O (s)	ΔT_O (s)	T_E (s)	ΔT_E (s)
line 1	2*	0.3					
line 2	4	0.8	± 0.1	0.04	± 0.02	0.49	± 0.04
line 3	8	2.1	± 0.2	0.15	± 0.01	0.32	± 0.01

Shots and geophones for above spreads all at 327 metres A.S.L.

* Origin used in these straight-line determinations.

Table C.2 Seismic data for the Murry Rd long spread

Shot: M.S.3

Shot height: 345 metres A.S.L.

Location: South end of long spread on Murry Road

Geo. No.	Distance (m)	Height (m)	Corrn. (s)	Obs. Time (s)	Corrected Time (s)
12	32	347	0	0.100	0.100
1	408	327	-0.02	0.365	0.345
11	464	327	-0.02	0.390	0.370
10	500	327	-0.02	0.410	0.390
9	530	327	-0.02	0.425	0.405
8	560	327	-0.02	0.430	0.415
7	590	327	-0.02	0.440	0.420
6	620	327	-0.02	0.455	0.435
5	650	327	-0.02	0.460	0.440
4	680	327	-0.02	0.475	0.455
2	740	327	-0.02	0.495	0.475
3	1100	311	Not Recorded 0.59 s adopted from M.S.4		

Least squares solutions:

	n	V _{app} (km/s)	ΔV_{app} (km/s)	T _O (s)	ΔT_O (s)	T _E (s)	ΔT_E (s)
line 1	4	2.0	± 0.2	0.142	± 0.03	0.69	± 0.03
line 2	7	3.0	± 0.2	0.230	± 0.01	0.59	± 0.01

Shot: M.S.4

Shot height: 311 metres A.S.L.

Location: North end of long spread on Murry Road

Geo. No.	Distance (m)	Height (m)	Corrn. (s)	Obs. Time (s)	Corrected Time (s)
3	26.5	311	0.00	0.050	0.05
2	390	327	0.02	0.310	0.330
4	450	327	0.02	0.336	0.356
5	480	327	0.02	0.350	0.370
6	510	327	0.02	0.365	0.385
7	540	327	0.02	0.375	0.395
8	570	327	0.02	0.390	0.410
9	600	327	0.02	0.405	0.425
10	630	327	0.02	0.415	0.435
11	660	327	0.02	0.425	0.445
1	720	327	0.02	0.440	0.460
12	1100	347	0.00	0.590	0.590

Least squares solutions:

	n	V _{app} (km/s)	ΔV_{app} (km/s)	T _O (s)	ΔT_O (s)	T _E (s)	ΔT_E (s)
line 1	7	2.2	± 0.1	0.15	± 0.01	0.65	± 0.01
line 2	5	3.0	± 0.2	0.23	± 0.01	0.59	± 0.01

Table C.3 Seismic data for the Christiansen Rd short spread

Shot: M.S.7

Shot and station heights all 347 metres A.S.L.

Location: South end of short spread on Christiansen Road

Geophone No.	Distance (m)	Travel Time (s)
11	14	0.040
12	44	0.135
10	74	0.175
9	104	0.190
8	144	0.235
7	174	0.240
6	204	0.260
5	234	0.280
4	264	0.310
3	294	0.330
1	324	0.360
2	354	0.380

Least squares solutions:

	n	V_{app} (km/s)	ΔV_{app} (km/s)	T_O (s)	ΔT_O (s)	T_E (s)	ΔT_E (s)
line 1	3*	0.3	± 0.1	0	± 0.01	1.09	± 0.05
line 2	10	1.4	± 0.1	0.12	± 0.01	0.38	± 0.01

Shot: M.S.8

Shot and geophones all at 347 metres A.S.L.

Location: North end of short spread on Christiansen Road

Geophone No.	Distance (m)	Travel Time (s)
2	17	0.055
1	47	0.125
3	77	0.160
4	107	0.190
5	147	0.210
6	177	0.240
7	207	0.280
8	237	0.305
9	267	0.330
10	297	0.350
12	327	0.370
11	357	0.390

Least squares solutions:

	n	V_{app} (km/s)	ΔV_{app} (km/s)	T_O (s)	ΔT_O (s)	T_E (s)	ΔT_E (s)
line 1	3*	0.4	± 0.1	0	± 0.01	0.9	± 0.05
line 2	10	1.2	± 0.1	0.10	± 0.01	0.40	± 0.01

* Origin included in these line segments.

Table C.4 Seismic data for the Christiansen Rd long spread

Shot: M.S.9

Shot and geophones all at 347 metres (A.S.L.), except geophone No.12

Location: Long spread from north end of Christiansen Road

Geophone No.	Distance (m)	Travel Time (s)
1	30	0.095
2	448	0.440
3	508	0.460
4	538	0.480
5	568	0.485
6	598	0.510
7	628	0.535
8	658	0.550
9	688	0.565
10	708	0.575
11	778	0.605
12	1200	0.780

Least squares solutions:

	n	V_{app} (km/s)	ΔV_{app} (km/s)	T_O (s)	ΔT_O (s)	T_E (s)	ΔT_E (s)
line 1	2	1.2	n/a	0.07	n/a	1.06	n/a
line 2	11	2.2	± 0.07	0.23	± 0.01	0.79	± 0.01

Shot: M.S.10

Shot height of 329 metres. Corr'n. of 0.02 sec added to all arrivals so shot and detector all lie on 347 metres datum.

Location: South end of long spread on Christiansen Road

Geophone No.	Distance (m)	Travel Time (s)
12	32	0.070
11	453	0.430
10	513	0.450
9	543	0.480
8	573	0.495
7	603	0.505
6	633	0.520
5	663	0.525
4	693	0.530
3	723	0.545
2	783	0.585
1	1200	0.800

Least squares solutions:

	n	V_{app} (km/s)	ΔV_{app} (km/s)	T_O (s)	ΔT_O (s)	T_E (s)	ΔT_E (s)
line 1	2	1.2	n/a	0.04	n/a	1.07	n/a
line 2	11	2.0	± 0.1	0.20	± 0.03	0.79	± 0.02

Table C.5 Seismic data for the Hannon Rd spread

Shot: R.S.3

Location: South end of Hannon Road

Geophone No.	Distance (m)	Travel Time (s)
11	60	0.178
12	260	0.435
10	310	0.460
9	340	0.480
8	370	0.483
7	400	0.498
6	430	0.513
5	460	0.517
4	490	0.538
3	520	0.548
2	580	0.570
1	760	0.667

Least squares solutions:

	n	V_{app} (km/s)	ΔV_{app} (km/s)	T_O (s)	ΔT_O (s)	T_E (s)	ΔT_E (s)
line 1	2*	0.3	n/a	0	n/a	2.25	n/a
line 2	2	0.8	n/a	0.10	n/a	1.08	n/a
line 3	11	2.2	± 0.1	0.32	± 0.01	0.66	± 0.06

Shot: R.S.4

Location: North end of Hannon Road

Geophone No.	Distance (m)	Travel Time (s)
2	60	0.177
1	260	0.427
3	310	0.452
4	340	0.465
5	370	0.470
6	400	0.490
7	430	0.510
8	460	0.513
9	490	0.540
10	520	0.547
12	580	0.578
11	760	0.665

Least squares solutions:

	n	V_{app} (km/s)	ΔV_{app} (km/s)	T_O (s)	ΔT_O (s)	T_E (s)	ΔT_E (s)
line 1	2*	0.3	n/a	0	n/a	2.24	n/a
line 2	2	0.8	n/a	0.10	n/a	1.05	n/a
line 3	11	2.1	± 0.1	0.30	± 0.01	0.66	± 0.01

* line segments which use the origin in the least squares determination.

Table C.6 Seismic data for the Cambell Rd spread

Shot: R.S.6
 Location: Cambell Road
 Shot Offset: 90 metres

Geophone No.	Distance (m)	Travel Time (s)
1	90	0.055
2	120	0.075
3	150	0.092
4	180	0.112
5	210	0.134
6	240	0.152
7	270	0.170
8	300	0.192
9	330	0.212
10	360	0.230
11	390	0.250

Least squares solutions:

	n	V_{app} (km/s)	ΔV_{app} (km/s)	T_o (s)	ΔT_o (s)
line 1	12	1.55	± 0.01	0	± 0.01

Shot: R.S.5
 Location: Cambell Road
 Shot Offset: 590 metres

Geophone No.	Distance (m)	Travel Time (s)
1	590	0.324
2	620	0.340
3	650	0.350
4	680	0.372
5	710	0.392
6	730	0.408
7	760	0.425
8	790	0.448
9	820	0.470
10	850	0.481
11	880	0.502
12	910	0.510

Least squares solutions:

	n	V_{app} (km/s)	ΔV_{app} (km/s)	T_o (s)	ΔT_o (s)
line 1	13	1.76	± 0.04	0	± 0.01

Plane layer solution from R.S. 5 and 6: layer 1 = 1.5 $\rightarrow$ 1.8 km/s.

APPENDIX D $2\frac{1}{2}$ -DIMENSIONAL GRAVITY INTERPRETATIONS

In the quantitative interpretation of gravity and magnetic data, two-dimensional calculations along profiles perpendicular to the axis of an infinitely long body have been popular. Reasons for this popularity are that structures which approach two-dimensionality are common in geology and data are often collected in profiles perpendicular to the strike direction. Polygonal cross-sections of two-dimensional bodies are conveniently represented on paper and input to the computer, whereas three-dimensional calculations are slower and data input and display are more cumbersome.

Shuey & Pasquale (1973) derived equations for the magnetic field of a compromise " $2\frac{1}{2}$ -dimensional body", invariant in cross section but terminated a finite distance along strike. Cady (1977, 1980) has derived a similar expression for the gravity effect of a $2\frac{1}{2}$ -dimensional body which also incorporated the facility for different strike lengths into and out of the plane of the profile. Moreover by specifying one strike direction to be negative the calculation is performed for a profile which does not pass directly over the body; this is demonstrated in the following example.

Fig. D.1 shows a hypothetical model with various strike directions and their calculated gravity effects. For L_1 and L_2 infinite the gravity effect is merely that of a two-dimensional body. For $L_1 = -3$ and $L_2 = 10$ km the profile does not cross the body but is 3 km distant from one end of the body and 10 km distant from the other i.e. the length of the body is 7 km.

This method has particular relevance for gravity interpretations made in volcanic provinces where geological forms tend to be three-dimensional in character. The $2\frac{1}{2}$ -dimensional method has the convenience and speed of the two-dimensional approach, yet still retains much of the generality of the three-dimensional approach.

Table D.1 contains a Fortran listing of the $2\frac{1}{2}$ -D gravity interpretation programme.

Table D.1 Listing of the 2½-dimensional gravity interpretation programme.

```

C*****2 1/2-D GRAVITY INTERPRETATION PRGM (AFTER CADY(1977))
C***** T.A. STERN 1979
FILE 95(TITLE='2DEXP',KIND=DISK,FILETYPE=7)
80(KIND=DISK,MAXRECSIZE=22,BLOCKSIZE=330,PROTECTION=SAVE)
DIMENSION STRK1(10),STRK2(10),XSTAT(150),ZSTAT(150),TT(150)
DIMENSION GTE(10,150),GTT(150),NSIDES(10),RES(150),PC(10)
DIMENSION GOBS(150)
DIMENSION DELM(10,25,150),XSTART(10,100),ZSTART(10,100)
C*****ENTER NUMBER OF COMPUTING POINTS AND NO. OF POLYGONS
READ(95,/) NUMX,NBODS
C***** ENTER X-COORDINATE OF CPT. PT. AND OBSERVED ANOMALY
C***** UNITS OF METRES,MILLIGALS
DO 4 J=1,NUMX
  READ(95,15)XSTAT(J),GOBS(J)
15 FORMAT( F8.2,F8.2)
  WRITE(80,15) XSTAT(J),GOBS(J)
4 CONTINUE
DO 50 I=1,NUMX
  TT(I)=0.0
DO 50 J=1,NBODS
  GTT(I)=0.0
50 GTE(J,I)=0.0
  CNST=6.666667E-03
C*****READ PARAMETERS FOR EACH BODY: STRIKE LGTHS ARE +VE IF PROFILE
C*****CROSSES BODY
C***** DENSITY CONTRAST(PC(L)) IN MG/M3
DO 500 L=1,NBODS
  READ(95,/)NSIDES(L),STRK1(L),STRK2(L),PC(L)
  WRITE(80,23)L,NSIDES(L),STRK1(L),STRK2(L)
23 FORMAT(1X,"POLYGON",I3,"HAS",I3,"SIDES & HALF STRIKE
1 LENGTHS",2F10.1)
  WRITE(80,24)PC(L)
24 FORMAT(1X,"DENSITY CONTRAST FOR POLYGON:",F8.6)
  M=NSIDES(L)+1
DO 90 LL=1,M
  READ(95,16) XSTART(L,LL),ZSTART(L,LL)
16 FORMAT( 2F10.2)
  WRITE(80,21)XSTART(L,LL),ZSTART(L,LL),L
21 FORMAT(1X,I6,"(X,Z)=",2F12.2,"BODY NO.",I3)
90 CONTINUE
DO 600 K=1,NUMX
  MM=NSIDES(L)
DO 610 J=1,MM
  Z1=ZSTART(L,J)
  Z2=ZSTART(L,J+1)
  X1=XSTART(L,J)-XSTAT(K)
  X2=XSTART(L,J+1)-XSTAT(K)
  Y1=STRK1(L)
  Y2=STRK2(L)
  IF(Z1.LT.0.0) Z1=0.00001
  IF(Z2.LT.0.0) Z2=0.00001
  DEL = PC(L)
  CALL GRAV(Z1,Z2,X1,X2,Y1,Y2,DEL,T)
  DELM(L,J,K)=T
  GTE(L,K)=GTE(L,K)+T*CNST
610 CONTINUE
600 CONTINUE
500 CONTINUE
DO 260 I=1,NUMX
DO 261 J=1,NBODS
  TT(I)=TT(I)+GTE(J,I)
261 CONTINUE
  RES(I)=GOBS(I)-TT(I)
260 WRITE(80,17) XSTAT(I),TT(I),GOBS(I),RES(I)
17 FORMAT(1X,"X=",F8.0,2X,"CALC=",F5.1,5X,"OBS=",F5.1,5X,
1"RESIDUAL=",F5.1)
  SSR=0.0
DO 270 I=1,NUMX
  SSR=SSR+RES(I)*RES(I)
270 CONTINUE
  WRITE(80,271)NUMX,SSR/NUMX,SQRT(SSR/NUMX)
271 FORMAT(1X,I6,"STATIONS",4X,"MEAN RES**2 =",F15.3,"SQRT=",F15.3)
  LOCK 80
  CALL EXIT
END
C*****PERFORM LINE INTEGRAL FOR EACH LINE INCREMENT
SUBROUTINE GRAV(Z1,Z2,X1,X2,Y1,Y2,DEL,TERMS)
  TERMS=0.
  IF(ABS(X1-X2).LT.0.001) GO TO 50
  IF(ABS(X1).LT.0.00001) X1=0.00001
  IF(ABS(X2).LT.0.00001) X2=0.00001
  IF(ABS(Z1).LT.0.00001) Z1=0.00000001
  IF(ABS(Z2).LT.0.00001) Z2=0.00000001
  TWOPI=2.*3.1415927

```

Table D.1 - continued

```

DX=X2-X1
IF (ABS(DX).LT.0.001) DX=0.0001
DZ=Z2-Z1
IF (ABS(DZ).LT.0.001) DZ=0.0000001
EM=DZ/DX
CSQ=1.+EM**2
C=(SQRT(DZ*DZ+DX*DX))/DX
ZO=Z1-EM*X1
A=ZO/C
ASQ=A*A
AK=ZO/CSQ
XO=EM*AK
BX=X2+XO
IF (ABS(BX).LT.0.0001) BX=0.0001
BI=X1+XO
IF (ABS(BI).LT.0.0001) BI=0.0001
CBX=C*BX
CBI=C*BI
JEND=1
IF (ABS(Y1).LT.0.0001.AND.ABS(Y2).LT.0.0001) JEND=0
RSQ1=X1**2+Z1**2
RSQ2=X2**2+Z2**2
RY1=SQRT(Y1*Y1+RSQ1)
RY11=SQRT(Y1*Y1+RSQ2)
RY2=SQRT(Y2*Y2+RSQ1)
RY22=SQRT(Y2*Y2+RSQ2)
T1=0.
T2=0.
IF (RSQ2.GT.0.) T1=BX*ALOG(RSQ2)
IF (RSQ1.GT.0.) T2=BI*ALOG(RSQ1)
T3=ATAN2(BX,AK)
IF (BX.GT.0..AND.AK.LT.0.) T3=T3-TWOPI
T4=ATAN2(BI,AK)
IF (BI.GT.0..AND.AK.LT.0.) T4=T4-TWOPI
IF (JEND.EQ.0) GO TO 48
T5=BI*ALOG((Y1+RY1)*(Y2+RY2))
T6=BX*ALOG((Y1+RY11)*(Y2+RY22))
T7=Y1/C*ALOG((CBI+RY1)/(CBX+RY11))
T8=Y2/C*ALOG((CBI+RY2)/(CBX+RY22))
FNUM=ASQ+Y1**2+Y1*RY11
DEN=ZO*BX
T9=ATAN2(FNUM,DEN)
IF (FNUM.LT.0..AND.DEN.LT.0.) T9=T9+TWOPI
FNUM=ASQ+Y2**2+Y2*RY22
T10=ATAN2(FNUM,DEN)
IF (FNUM.LT.0..AND.DEN.LT.0.) T10=T10+TWOPI
FNUM=ASQ+Y1**2+Y1*RY1
DEN=ZO*BI
T11=ATAN2(FNUM,DEN)
IF (FNUM.LT.0..AND.DEN.LT.0.) T11=T11+TWOPI
FNUM=ASQ+Y2**2+Y2*RY2
T12=ATAN2(FNUM,DEN)
IF (FNUM.LT.0..AND.DEN.LT.0.) T12=T12+TWOPI
TERMS=T1-T2+2.*AK*(T3-T4)+T5-T6+T7+T8+AK*(T9+T10-T11-
1T12)
GO TO 49
48 TERMS=T1-T2+2.*AK*(T3-T4)
49 TERMS=-1.*TERMS*DEL
50 RETURN
END

```

APPENDIX E SPECTRAL ANALYSIS FOR TWO-DIMENSIONAL ARRAYS
OF SPATIAL DATA

This appendix gives a description of the computational procedure to obtain energy-spectra from a two-dimensional array of spatial data. When a discrete approximation to the infinite Fourier integral is made a certain amount of distortion of the true spectrum is unavoidable. Close attention is paid to these sources of distortion and methods of minimising them.

E.1 Pre-transformation procedures

a) Digitisation of the data: The aeromagnetic map of Whiteford (1976) was selected for digitisation. The map was digitised on a north-south, east-west grid at an interval of 2.5 km. The region chosen was a 125x125 km square and incorporated all of the major rhyolitic centres (fig. 6.1) giving 2500 data points in total. The digitisation interval of 2.5 km results in the highest frequency available for study, the Nyquist frequency, being 0.2 cycles/km. Anomalies of wavelength 5 km or less will be folded back (aliased) about the Nyquist frequency. Within the Central Volcanic Region there are anomalies with wavelengths less than 5 km. However, as the original flight-line spacing is of the order of 5 km, these anomalies would not be reliably represented in the hand-contoured map. Furthermore, the purpose of this analysis is to determine the base of the magnetic crust so that the degrading of information pertaining to short wavelength (hence shallow) magnetic sources does not seriously prejudice this objective. Fig. E.1 shows a comparison of a section of the original magnetic map with a contour map, computer drawn, from the digitised data. Apart from a degree of

Fig. E.1 Comparison of the original magnetic anomaly data with the digitised then computer contoured data.

smoothing, inherent in digitising and computer contouring, the main anomaly features are well preserved.

b) Scan for errors: Spikes due to reading or card-punching errors will propagate through the spectrum causing spurious output at all frequencies (Black and Scollar, 1969). A simple scanning routine was used to check for any unacceptable variations between consecutive field values; i.e. the rate of change across a row or down any column is examined such that any violation of: $T_{j+1,i} - T_{j,i} < \Delta T_{\max}$ produces an error message.

c) Detrending the data: Trends in the data of a wavelength similar to or larger than the horizontal dimensions of the map will contaminate the low-frequency end of the spectrum. These trends could arise from an incomplete removal of the geomagnetic normal field, magnetic edge effects of deep sources either side of or under the Central Volcanic Region or systematic errors introduced in the digitisation process.

Using the digitised data set, polynomials of order 0, 1 and 2 were constructed then subtracted from the original data to yield three detrended data sets for transformation.

d) Windowing the data: A full discussion of "window carpentry" is given by Bath (1974, p155) and only the basic principles will be discussed here. Direct transformation of data is equivalent to applying a rectangular spatial window, or box-car window, the frequency response of which is a sinc function. This function has a narrow spectral window, thus giving high spectral resolution, but large side lobes resulting in a large variance. To circumvent this the Hanning window is often used. This window tapers the spectral data to zero at the edges and its transform has a spectral window width twice that of the rectangular window but with smaller side lobes. Hence, resolution is not as good but variance is decreased.

A third window (Bath, 1974, pl61) is a combination of the previous two and offers a reasonable trade off between variance and resolution. This data window, amounting to approximately a 17 percent cosine taper, was applied to the data sets in this study.

e) Padding of the data: A further source of variance in the spectral estimation is the partial inclusion of anomalies around the map border. To offset this effect an extrapolated data border around the map, five grid units wide, was added to the digitised grid. The mean, or D.C. component, was subtracted from the extrapolated grid, then the cosine-tapered rectangular window described above was applied to the extrapolated border.

Historically, the filtering of gravity and magnetic data was carried out by a linear convolution process with a selected set of constant coefficients in the spatial domain (e.g. Peters (1949)). With the advent of the fast Fourier transform this process can be performed more accurately and conveniently by multiplication in the frequency domain (Black & Scollar, 1969).

However, multiplication of discrete Fourier transforms is not exactly equivalent to a linear convolution. A sampling of the transform implies that the map is periodically extended in space and multiplication of the discrete transform is therefore equivalent to circular or periodic convolution of the extended map with a set of weights the full size of the map. In order to obtain the equivalent linear convolution carried out in the spatial domain, the pre-transformed data was padded with zeros out to a convenient power of two grid units in each direction; the choice of a power of two being necessary for the fast Fourier transform programme used in this study. A more detailed account of the techniques to eliminate circular convolution is given by Oppenheim & Schaffer (1975).

E.2 Description and listing of the fast Fourier transform- filter package

Theoretical aspects of the programme were outlined in chapter 6 and the programme used for this study is summarised in the flow diagram of fig. E.2. In this section more detailed aspects of the programme are outlined in terms of input and output variables then concluded with a Fortran listing, compatible for a Burroughs 6700, of a sample programme.

The programme consists of four independent units comprising a main driver routine to window the data prior to transform, transformation routines, a series of filter routines, and various output routines. The author is indebted to M. Rogan (pers. comm.) of Auckland University for kindly supplying the fast Fourier transform routines. The filtering subroutines used in this study come from Carter & Cook (1978) and the main driver part of the programme was written by the author. A detailed description of the programme follows.

1) Main driver routine: This routine initially removes a regional trend which, in the case of this study, consisted of a polynomial in terms of grid units. The routine then extrapolates the end values of each row and column out to a distance of EX (variable) grid units. The mean value is then removed from the grid. Note that the regional polynomial removed on the previous step may have already removed the mean.

The data grid is then windowed with a cosine-taper window as outlined in section E.1. If the user wishes he may now print out the modified data grid for inspection. The grid is then padded out with zeros to a dimension of 2^m by 2^m where m is some integer.

TWO-DIMENSIONAL FAST FOURIER TRANSFORM (F.F.T.-2D)

Fig. E.2 Flow diagram illustrating the working of the two-dimensional Fast Fourier Transform programme.

Input variables for main driver:

first data card: A = digitisation interval in X-direction (km)

B = " " " Y- " "

LX = data length in X-direction (grid units)

LY = " " " Y- " "

M = power of two for which grid is to be padded to

EX = length of extrapolated border (grid units)

REG = card containing polynomial expression for regional field.

F(LX,LY) = original data grid either read from cards or from computer file.

Output variables: LY = LX = dimensions of modified data grid.

F(LX,LY) = modified data grid.

Subroutines called in main driver:

(a) Subroutine FMEAN (LX,LY,F,SUM,EX)

(b) Subroutine TAPER (I2,X,N,I)

Input variables: I2 = last grid location in the row or column of
the cosine taper.

X = extrapolated gravity value at column or row
location I

N = switch 1 if last value to be tapered = 1
-1 if last value to be tapered = 0

I = grid location of value to be tapered.

(c) Subroutine EBORD(X,J,L,J2,L2)

This subroutine makes a two-dimensional half cosine bell taper
to the corners of the extrapolated grid. Input variables:

X = digitised data value at grid location (L,J)

L = column grid location

J = row grid location

J2 = last row grid location of the corner of the
non-extrapolated data

L2 = last column grid location of the corner of the
non-extrapolated data.

Output: X is tapered digitised value.

Note: This routine is called four times; once for each corner of the array.

2) Transformation subroutines

Subroutine SINTAB(LX,SX)

Input variables: $LX = 2^m$

Comments: This routine sets up a sin table of values to be stored in memory for use in subroutine TRNS2D. It is called twice: once for LX, once for LY.

Subroutine TRNS2D(LX,LY,F,G,SIGNI)

Input variables: $F(LX,LY)$ = real part of array to be transformed.

$G(LX,LY)$ = imaginary part of array for transformation.

SIGNI = 1 for forward transformation

= -1 for reverse transformation

Output: $F(LX,LY)$ = real part of transformed array.

$G(LX,LY)$ = imaginary part of transformed array.

Required subroutines that are called by TRNS2D: FURRY, VTOM, MTOV.

Comments: This subroutine calculates the two-dimensional Fourier transform by repeated application of a one-dimensional transform to the complex array $F(L,J) + iG(L,J)$.

3) Filter subroutines: Subroutine FILTER acts as controlling element for subsequent calls to various filter subroutines.

Subroutine FILTER(N1,N2,TVR,TVI)

Input variables: N1 = number of rows in transformed array
 N2 = number of columns in transformed array
 TVR = real transformed array values
 TVI = imaginary transformed array values.

Other internal variables: f2 and f1 are variables which convert column and row grid locations into spatial frequencies (cycles/grid unit).

Output variables: TVR are the filtered real array values.

Subroutine REDPOL(FE,FN,TR,TI): This filter produces a magnetic anomaly field that would result if the source bodies, which are originally assumed to be magnetised in the direction of the earth's field, were vertically magnetised. The real and imaginary parts are multiplied by the factor:

$$(\sin(I) + i \cos(I) (f_x \cos(D) + f_y \sin(D) / (f_x^2 + f_y^2)^{\frac{1}{2}})^{-2}$$

where I and D are the inclination and declination, respectively, of the earth's magnetic field for the survey area; f_x and f_y are the spatial frequencies and i is the square root of -1.

For this study I and D were taken as -60° and 0° respectively.

Subroutine STRTUM(F1,F2,RR,RI): This filter calculates the vertical integral factor by multiplying the transformed data by:

$$\text{grid interval (km)} / 2\pi (f_x^2 + f_y^2)^{\frac{1}{2}}$$

To calculate the pseudogravity field the transformed data is further multiplied by $G\Delta\rho/\Delta M$; where G is the gravitational constant, $\Delta\rho$ is the assumed density contrast and ΔM the magnetisation contrast throughout the source body. When calculating the pseudogravity field a call to REDPOL must precede the call to STRTUM.

Subroutine CONTIN(H,FE,FN,TR,TI): This filter calculates the upward (H-negative) or downward (H-positive) continuation field by multiplying the transformed data by $\exp(2\pi H(f_x^2 + f_y^2))$. H - in grid units.

4) Output routines

Subroutine AMP2(LX,LY,f,G,D): Calculates the natural log of the power spectrum, i.e. $D = \ln((F+iG) \times (F-iG))$. Writes out for inspection the real and imaginary parts of the transformed array (optional).

Subroutine RADAVG(LX,LY,SU,D): Calculates and writes out the radial average of $D(L,J)$ averaged over 180° . A call to AMP2 must precede this subroutine.

Subroutine OUTPUT(LX,LY,NS,A,B,H,f): A general output routine for writing out the real and imaginary parts of the transformed field if called after one call to TRNS2D. If called after the second call to TRNS2D will output the filtered spatial field.

Table E.1 Listing of the Fast Fourier Transform programme used in this study.

```

S      SET S=0 LINEINFO
FILE 11 TITLE="MATFIT",KIND=DISK,MAXRECSIZE=60,BLOCKSIZE=600)
      DIMENSION F(128,128),G(128,128),C(128,128),U(128,128)
      COMMON SX(128),SY(128)
      READ(5,1) A,B,LX,LY,M,EX
MAIN DRIVER
1  FORMAT(2F4.1,4I4)
   DO 10 J=1,LY
   READ(7,2)(F(L+EX,J+EX),L=1,LX)

36  CONTINUE
2  FORMAT(13F6.1)
C  REMOVE REGIONAL TREND - PLANE SURFACE: REG=A+B*X+C*Y
   DO 35 L=1,LX
   DO 36 J=1,LY
   REG=-119.68+.7934*L+.6483*J
   F(L+EX,J+EX)=F(L+EX,J+EX)-REG
35  CONTINUE
3  EXTRAPOLATE 5 GRID UNITS EACH END OF ROWS AND COLS
   DO 3 L=EX+1,EX+LX
   DO 4 J=1,EX
   F(L,J)=F(L,1+EX)
4  CONTINUE
   DO 3 J=1+EX+LY,LY+2*EX
   F(L,J)=F(L,LY+EX)
3  CONTINUE
   DO 5 J=1,LY+2*EX
   DO 6 L=1,EX
   F(L,J)=F(1+EX,J)
6  CONTINUE
   DO 5 L=1+LX+EX,LX+2*EX
   F(L,J)=F(LX+EX,J)
5  CONTINUE
C  REMOVE DC VALUE FROM GRID
CALL FMEAN(LX,LY,F,SUM,EX)
C  COSINE TAPER ALL ROWS IN DATA GRID
   DO 7 J=1+EX,LY+EX
   DO 8 L=1,EX
   CALL TAPER(EX,F(L,J),1,L)
8  CONTINUE
   DO 7 L=LX+EX+1,LX+2*EX
   CALL TAPER(LY+2*EX,F(L,J),-1,L)
7  CONTINUE
C  COSINE TAPER ALL COLS IN DATA GRID
   DO 9 L=1+EX,LX+EX
   DO 20 J=1,EX
   CALL TAPER(EX,F(L,J),1,J)
20  CONTINUE
   DO 9 J=LY+EX+1,LY+2*EX
   CALL TAPER(LX+2*EX,F(L,J),-1,J)
9  CONTINUE
C  PUT COSINE TAPER IN EXTRAPOLATED CORNERS OF GRID
   DO 11 J=1,EX
   DO 12 L=1,EX
   CALL EBURD(F(L,J),J,L,1+EX,1+EX)
12  CONTINUE
   DO 11 L=LX+EX+1,LX+2*EX
   CALL EBURD(F(L,J),J,L,1+EX,LX+EX)

11  CONTINUE
   DO 13 J=LY+EX+1,LY+2*EX
   DO 14 L=1,EX
   CALL EBURD(F(L,J),J,L,LY+EX,EX+1)
14  CONTINUE
   DO 13 L=LX+EX+1,LX+2*EX
   CALL EBURD(F(L,J),J,L,LX+EX,LX+EX)
13  CONTINUE
   WRITE(6,31) A,B,LX,LY,M,EX

31  FORMAT(1X,'GRID PARAMETERS ARE',6F4.1)
   LX=2**M
   LY=2**M
   WRITE(6,101) LX,LY
101  FORMAT(2I4)
C  GRID HAS NOW BEEN PADDED OUT WITH ZEROS TO 2**M BY 2**M
   CALL SINTAB(LX,SX)
   CALL SINTAB(LY,SY)
   CALL TRNS2D(LX,LY,F,G,1.0)
   CALL FILTER(LX,LY,F,G)
   CALL AMP2(LX,LY,F,G,0)
   CALL PADAVG(LX,LY,SU,0)
   CALL TRNS2D(LX,LY,F,G,-1.0)
   CALL OUTPUT(60,60,0,A,B,0,F)
   DO 22 IY=1,60
   WRITE(11,2)(F(IX,IY),IX=1,60)
22  CONTINUE
   LUCK 11
   STOP
   END

SUBROUTINE SINTAB(LX,S)
  DIMENSION S(128)
  DOUBLE PRECISION ARG,DELARG
  N1=LX/4+1
  ARG=0.0
  DELARG=0.6283185307179586001/LX
  DO 10 I=1,N1
  S(I)=DSIN(ARG)
  ARG=ARG+DELARG
10  CONTINUE
  RETURN
  END

SUBROUTINE TRNS2D(LX,LY,F,G,SIGMI)
  DIMENSION F(128,128),G(128,128),X(128),Y(128)
  COMMON SX(128),SY(128)
  DO 10 J=1,LY
  CALL MTGV(LX,LY,J,1,X,Y,F,G)
  CALL FURRY(LX,X,Y,SIGMI,SX)
  CALL VTUM(LX,LY,J,1,X,Y,F,G)
10  CONTINUE
   DO 20 L=1,LX
   CALL MTGV(LX,LY,L,2,X,Y,F,G)
   CALL FURRY(LY,X,Y,SIGMI,SY)
   CALL VTUM(LX,LY,L,2,X,Y,F,G)
20  CONTINUE
  RETURN
  END

```

Table E.2 - continued

```

SUBROUTINE FURRY(LX,X,Y,SIGNI,S)
  DIMENSION X(128),Y(128),S(128)
  SC=1.0
  IF(SIGNI.LT.0.0) SC=1.0/LX
  NN=LX/2
  N=LX/4
  J=1
  DO 30 I=1,LX
    IF(I.GT.J) GO TO 10
    TEMPX=X(J)*SC
    TEMPY=Y(J)*SC
    X(J)=X(I)*SC
    Y(J)=Y(I)*SC
    X(I)=TEMPX
    Y(I)=TEMPY
  10 M=NN
  20 IF(J.LE.M) GO TO 30
  J=-M
  M=M/2
  IF(M.GE.1) GO TO 20
  J=J+M
  L=1
  NL=NN
  40 ISTEP=2*L
  IND=1
  DO 50 M=1,L
    INDN=IND+N-1
    IF(INDN.41,42,43)
  41 MY=S(INDN)*SIGNI
    GO TO 45
  42 MY=SIGNI
    GO TO 45
  43 MY=-S(INDN+1)
    MY=S(N+1-INDN)*SIGNI
  45 DO 50 I=M,LX,ISTEP
    TEMPX=MX*X(I+L)-MY*Y(I+L)
    TEMPY=MX*Y(I+L)+MY*X(I+L)
    X(I+L)=TEMPX
    Y(I+L)=TEMPY
    X(I)=X(I)+TEMPX
    Y(I)=Y(I)+TEMPY
  50 IND=IND+NL
  55 L=ISTEP
  NL=NL/2
  IF(L.LT.LX) GO TO 40
  RETURN
END

SUBROUTINE MTOV(LX,LY,J,NI,X,Y,F,G)
  DIMENSION X(128),Y(128),F(128,128),G(128,128)
  IF(NI.EQ.2) GO TO 15
  DO 10 L=1,LX
    X(L)=F(L,J)
    Y(L)=G(L,J)
  10 CONTINUE
  RETURN
  15 CONTINUE
  DO 20 L=1,LY
    X(L)=F(L,J)
    Y(L)=G(L,J)
  20 CONTINUE
  RETURN
END

SUBROUTINE VTOM(LX,LY,J,NI,X,Y,F,G)
  DIMENSION X(128),Y(128),F(128,128),G(128,128)
  IF(NI.EQ.2) GO TO 15
  DO 10 L=1,LX
    F(L,J)=X(L)
    G(L,J)=Y(L)
  10 CONTINUE
  RETURN
  15 CONTINUE
  DO 20 L=1,LY
    F(L,J)=X(L)
    G(L,J)=Y(L)
  20 CONTINUE
  RETURN
END

SUBROUTINE FMEAN(LX,LY,F,SUM,EX)
  DIMENSION F(128,128)
  SUM=0
  DO 5 L=1,LX+2*EX
    DO 5 J=1,LY+2*EX
      SUM=SUM+F(L,J)
  5 CONTINUE
  SUM=SUM/FLOAT((LX+2*EX)*(LY+2*EX))
  DO 1 L=1,LX+2*EX
    DO 1 J=1,LY+2*EX
      F(L,J)=F(L,J)-SUM
  1 CONTINUE
  35 WRITE(6,35) SUM
  FORMAT(1X,'DC VALUE OF DATA IS',F7.2)
  RETURN
END

SUBROUTINE TAPER(I2,X,N,I)
  PI=3.141592654
  FI=FLOAT(I2-I)
  X=X*0.5*(1.+N*COS((FI/4.)*PI))
  RETURN
END

SUBROUTINE FBORD(X,I,J,I2,J2)
  PI=3.141592654
  FI=FLOAT(I-I2)
  FJ=FLOAT(J-J2)
  F=SQRT(FI*FI+FJ*FJ)
  F2=SQRT(2.)
  FC=SQRT(26.)
  IF(F.GT.FC) GO TO 2
  DIF=FC-F2
  X=X*0.5*(1.+COS((F-F2)/DIF)*PI))
  GO TO 3
  2 X=0.
  3 RETURN
END

SUBROUTINE FILTER(N1,N2,IVR,TVI)
  DIMENSION TVP(1),TVI(1)
  NYQUIST FREQUENCY = 0.5
  DO 50 I=1,N2
    F2=FLOAT(I-1)/FLOAT(N2)
    IF(F2.GT.0.5) F2=F2-1
    DO 50 J=1,N1
      F1=FLOAT(J-1)/FLOAT(N1)
      IF(F1.GT.0.5) F1=F1-1
      K=(I-1)*N1+J
      DR=TVP(K)
      DI=TVI(K)
      CALL REDPOL(F2,F1,DR,DI)
      TVP(K)=DR
      TVI(K)=DI
    50 CONTINUE
  RETURN
END

SUBROUTINE CONTIN(H,FE,FN,TR,TI)
  *H FOR UPWARDS CONTIN, -H FOR DOWWARDS CONTIN
  H IN UNITS OF GRID INTERVAL
  PI=3.141592654
  FAC=EXP(2.*PI*SQRT(FE**2+FN**2)*H)
  TR=TR*FAC
  TI=TI*FAC
  RETURN
END

SUBROUTINE STRTUM(F1,F2,RR,RI)
  DATA GRID/2.5/
  PI=3.14159265
  RI=0.0
  F=SQRT(F1**2+F2**2)
  IF(F.LT.1.E-5) GO TO 1
  RR=GRID/(2.0*PI*F)
  RETURN
  1 RR=0.0
  RETURN
END

SUBROUTINE REDPOL(FE,FN,TR,TI)
  FE=EAST FREQ, FN=NORTH FREQ
  COMPLEX Z,ZD,T
  PI=3.14159265
  AI=0.0
  D=0.0
  DEGRAD=PI/180.0
  SINI=SIN(AT*DEGRAD)
  COSI=COS(AT*DEGRAD)
  SIND=SIN(D*DEGRAD)
  COSD=COS(D*DEGRAD)
  F=SQRT(FE**2+FN**2)
  IF(F.LT.1.0E-05) GO TO 2
  DI=COSI*(COSD*FN/F+SIND*FE/F)
  ZD=CMPLX(SINI,DI)
  Z=1.0/(ZD)**2
  T=CMPLX(TR,TI)
  TR=REAL(T)
  TI=AIMAG(T)
  2 RETURN
END

SUBROUTINE AMP2(LX,LY,F,G,D)
  DIMENSION F(128,128),G(128,128),D(128,128)
  DO 22 J=1,16
    WRITE(6,21)(F(L,J),L=1,16)
  22 CONTINUE
  21 FORMAT(1X,16F8.0)
  DO 23 J=1,16
    WRITE(6,21)(G(L,J),L=1,16)
  23 CONTINUE
  DO 10 I=1,LX
    DO 10 J=1,LY
      IF(F(I,J)+G(I,J)) 2,10,2
  2 D(I,J)=ALOG(F(I,J)**2+G(I,J)**2)
  10 CONTINUE
  RETURN
END

```

Table E.2 - continued

```

SUBROUTINE RADAVG(LX,LY,SU,D)
  DIMENSION D(128,128)
  PI=3.141592654
  DO 87 R=1,(LX/2)-1
    SU=0
    DO 86 B=1,9
      L=IFIX(R*SIN((PI/16)*(B-1)))
      J=IFIX(R*COS((PI/16)*(B-1)))
      SU=SU+D(L+1,J+1)
    CONTINUE
  86 DO 88 B=1,9
    L=IFIX(R*SIN((PI/16)*(B-1)))
    J=IFIX(LY-R*SIN((PI/16)*(B-1)))
    SU=SU+D(L+1,J)
  CONTINUE
  88 WRITE(6,90) R/(LX*2.5),SU/18
  90 FORMAT(1X,'RADIAL AVERAGE AT',F7.5,1X,'CY/KM IS',F8.2)
  CONTINUE
  87 RETURN
END

```

```

SUBROUTINE OUTPUT(LX,LY,NS,A,B,H,F)
  DIMENSION F(128,128),BB(100)
  WRITE(6,110) H
  IF (NS.EQ.1) WRITE(6,113)
  IF (NS.EQ.2) WRITE(6,114)
  110 FORMAT('1','UPWARD CONT TO HEIGHT',F10.2)
  111 FORMAT('1',19X,'Y',10F10.1,19X,'X/')
  112 FORMAT('/',1,F20.2,10F10.1)
  113 FORMAT('REAL PART OF FIELD')
  114 FORMAT('IMAGINARY PART OF FIELD')
  KK=LY/10
  DO 40 K=1,KK
    J1=(K-1)*10+1
    J2=J1+9
    DO 25 J=J1,J2
      25 BB(J)=J*B
      WRITE(6,111) (BB(J),J=J1,J2)
      DO 30 L=1,LX
        AL=L*A
        WRITE(6,112) AL,(F(L,J),J=J1,J2)
      30 CONTINUE
    40 CONTINUE
    IF (LY.LE.J2) GO TO 50
    J1=J2+1
    DO 42 J=J1,LY
      42 BB(J)=J*B
      WRITE(6,111) (BB(J),J=J1,LY)
      DO 45 L=1,LX
        AL=L*A
        WRITE(6,112) AL,(F(L,J),J=J1,LY)
      45 CONTINUE
    50 CONTINUE
    RETURN
  END

```

APPENDIX F DATA AND INTERPRETATION OF A LONG-RANGE, SEISMIC
REFRACTION EXPERIMENT WITHIN THE CENTRAL VOLCANIC
REGION

F.1 Introduction

In February 1982 the writer participated in a seismic refraction experiment carried out by Geophysics Division of the D.S.I.R. The experiment was reconnaissance in nature, the prime aim of the experiment being to ascertain the feasibility of propagating seismic energy over large distances through the Central Volcanic Region. The experiment consisted of three artificial explosions (one 136 kg and two 410 kg charges) fired from the ship R.N.Z.N. "Tui" at positions west of White Island as shown in fig. F.1.

Microearthquake recorders were laid out on a line from the greywacke outcrop at Otamarakau, south to Lake Taupo with a spacing between seismographs of about 10 km. An initial examination of the records showed that for shot-detector distances greater than 80 km there was a severe attenuation of seismic energy and many of the arrivals could not be reliably picked because of a high level of background noise. However, energy from Shots 2 and 3 was recorded on seismographs of the New Zealand Seismograph Network at Karapiro (KRP), Great Barrier Island (GBZ), Whakatane (WTZ) and Gisborne (GNZ). The interpretation to be presented here is a preliminary structural one, and is based on arrivals at the above stations of the New Zealand Network as well as arrivals at three temporary stations that were placed along the Rotoehu seismic line (Stations OTA, TYY, and ROT in Fig. F.1). Data from these three portable stations were used because i) the stations were within 80 km of the shot, and therefore gave good impulsive arrivals, and

Fig. F.1 Diagram showing the shot and seismograph station distribution for the long-range seismic refraction experiment.

ii) corrections for the superficial low-velocity volcanic rocks could be made for these stations based on the interpretation gained from the Rotoehu seismic data (see Chapter 5).

F.2 Data

Table F.1 gives the observed and corrected travel-time data from the three shots. The shot and station corrections given in Table F.1 were made so as to place shots and detectors on a datum 650 m below sea level. For stations TYY and ROT an initial correction, which is incorporated in the station correction, was made that replaces a 1.8 km thick section of low-velocity volcanic rocks (average velocity of 3 km/s) by rocks of velocity 5.0 km/s. A similar correction for the Tertiary-Quaternary sediments beneath the Gisborne station was also made. Grindley (1960) provides a geological cross-section for a SE-NW line 50 km to the south of Gisborne showing 9 km of Tertiary-Quaternary sediments. The velocity variation with depth within this sedimentary pile is unknown although Garrick (1969) finds that near surface Tertiary sediments from the East Coast have seismic velocities of 2.0-3.0 km/s. It is unlikely, however, that these near surface velocities will persist to depths of 9 km. Accordingly it is assumed, for the purpose of making a correction to the arrival times at Gisborne, that the sedimentary pile at Gisborne is equivalent to a 5 km thick section of average velocity 3.5 km/s.

A preliminary time-distance plot showed that data from Shots 1 and 3, for stations within a shot-detector range of 80 km, can be described by a straight line of the form: $\text{time(s)} = (\text{distance (km)}/6.17) + 1.24$ whereas data from the same stations for Shot 2 are described by: $\text{time} = (\text{distance}/6.19) + 0.77$. Thus there was a delay of about -0.5 s for

Station and elevation —> OTA (122m) TYY (152m) ROT (560m) KRP (60m) WTP (43m) GNZ (30m) GBZ (70m) Sonobuoy (0)

<u>Shot 1:</u>									
Arrival time (hr s)	1920 25.15	1920 26.53	--	--	1920 25.03	--	--	--	--
Shot correction (s)	-0.33	-0.33			-0.33				
Station correction (s)	-0.1	-0.40			-0.1				
Corrected travel time (s)	10.78	11.82			10.66				
Distance (km)	58.6	65.2			58.0				
<u>Shot 2:</u>									
Arrival time (hr s)	0701 16.50	0701 17.90	0701 19.49	0701 27.68	0701 16.50	0701 34.65	0701 33.91		
Shot correction (s)	-0.34	-0.34	-0.34	-0.34	-0.34	-0.34	-0.34	-0.34	-0.34
Station correction (s)	-0.1	-0.40	-0.45	-0.1	-0.1	-0.72	-0.1	-0.32	
Corrected travel time (s)	11.46	12.57	13.80	22.64	11.45	28.99	28.87	2.96	
Distance (km)	62.8	69.9	76.5	141.0	63.6	161.0	192.2	10.4	
<u>Shot 3:</u>									
Arrival time (hr s)	1900 21.03	1900 22.28	1900 23.46	1900 32.98	1900 21.27	1900 40.03	--		
Shot correction (s)	-0.25	-0.25	-0.25	-0.25	-0.25	-0.25		-0.25	
Station correction (s)	-0.1	-0.40	-0.45	-0.1	-0.1	-0.72		-0.22	
Corrected travel time (s)	9.10	10.03	11.20	21.07	9.36	27.50		2.63	
Distance (km)	48.0	54.1	61.5	130.0	50.0	153.5		8.45	

Shot instants:

Shot 1 1400(hr) 13.94(s) N.Z.D.T. (16/2/82)

Shot 2* 0701(hr) 4.6(s) N.Z.D.T. (17/2/82)

Shot 3 1900(hr) 11.56(s) N.Z.D.T. (17/2/82)

* -Shot 2 time was deduced from those of Shot 1 and 3 as explained in the text.

Shot depths all at 150m below sea level.

Water depths beneath shots of 650, 688 and 468m below sea level for shots 1, 2, and 3 respectively.

Shot 1 consisted of 136kg of explosives

Shots 2 and 3 consisted of 410kg of explosives

Shot correction places shots at an elevation of 650m below sea level.

Station correction places stations at an elevation of 650m below sea level

and for those stations not situated on greywacke makes an appropriate correction

for any low-velocity rocks that may be present. For TYY and ROT 1.8 km of low-velocity

(3.0 km/s) volcanics were assumed and for GNZ 5 km of Tertiary-Quaternary sediments

(average velocity 3.5 km/s) were assumed.

Table F.1 Travel-time data for the Bay of Plenty seismic refraction survey.

arrivals from Shot 2 with respect to arrivals from Shots 1 and 3 at a shot-detector range of 50-80 km. Shot 2 was situated the furthest offshore and contributions to the delay could be due to an excessive amount of sediments beneath Shot 2 or an error in either determining the position of the shot or the timing of the shot instant. In order to make the arrivals from Shot 2 compatible with those from Shots 1 and 3, 0.5 s was added to all arrivals from Shot 2.

F.3 Interpretation

Fig. F.2 shows all first arrival data plotted on a single time-distance graph. A fundamental assumption in the interpretation of these data is that the subsurface velocity structure can be approximated by a series of isovelocity layers with the velocity of successive layers increasing with depth; accordingly straight line segments were fitted to the data as shown. The inverse gradients of these straight line segments give apparent velocities; to obtain true velocities reversed time-distance data would be required. Because no reversed data are available it is necessary to assume that the inverse gradients of these straight line segments represent the true velocities of a series of horizontal, subsurface isovelocity layers. The time intercept method (Dobrin, 1976) was used to obtain estimates for depths to the various isovelocity layers. Uncertainties for the velocities were calculated at the 90% confidence level by the method of Steinhart and Meyer (1961). Uncertainties were also calculated for the intercept times, and the uncertainties in the depth calculations were estimated on the basis of the uncertainty in the intercept time.

Fig. F.2 Time-distance data and interpretations of the seismic refraction data. An explanation for each of the four interpretations is given in the text.

The time intercept of 1.27 s for the line segment with an inverse gradient of about 6.2 km/s is due to the sum of delay times at the shot and at the detectors; these delays being caused by an unknown thickness of rocks with velocities less than 6.2 km/s beneath both the shot and detectors. By interpreting the data in terms of uniform isovelocity layers we are effectively distributing the delay time (or time intercept) of 1.27 s evenly between the shot and detectors. In reality, however, the respective delay times at the shot and the detectors may not be equal. At the shot end there is an unknown thickness of sediments whereas beneath the detectors there are no sediments (for the two stations that were situated on low-velocity volcanic rocks appropriate corrections were made (section F.2)) but the rocks that the detectors were placed on have near-surface velocities less than 6.2 km/s. Hence the 2.4 km thick, 3.2 km/s layer has been introduced to simply account for the total delay-time but can only be considered an approximation to the real near-surface velocity structure. However, this approximation in the models will not appreciably alter the principal conclusions of the following interpretations.

Fig. F.2 shows four plane-layer interpretations.

Solution 1: This interpretation projects all arrivals from ray paths that have traversed beneath the Central Volcanic Region onto one series of straight line segments. Implicit in doing this is the assumption that the velocity structure of the Central Volcanic Region does not vary laterally. However, the ray path to station GBZ lies, in part, beyond the seaward projection for the western boundary of the Central Volcanic Region. Therefore Solution 1 may not be a structure that pertains strictly to the Central Volcanic Region.

Solution 2: This interpretation uses the same stations as used for Solution 1 less GBZ. Ray paths from the various shots to the stations used in this interpretation would have traversed, for the most part, within the Central Volcanic Region. The velocity of the deepest layer is 7.3 km/s and this is in good agreement with the 7.4 km/s velocity of Pn determined by Haines (1979) for the Region. However, the 7.3 km/s velocity has a large uncertainty of ± 0.8 km/s because the determination was made on just three widely-separated points.

Solution 3: This solution is arrived at by adopting the velocity for Pn of 7.4 km/s and then using the arrival times at Karapiro to determine the depth at which the top surface of the 7.4 km/s layer lies. That is, a line of gradient $(7.4 \text{ km/s})^{-1}$ was drawn through a point midway between the Karapiro arrivals and projected back to the distance axis so as to obtain a time intercept.

Solution 4: This is an interpretation made for a ray path between the shot points and Gisborne. Using the arrivals at the sonobuoys and Gisborne from Shot 3, a straight line segment of inverse gradient 5.81 km/s is derived. Solution 4 is then arrived at by adopting 8.1 km/s as the velocity of Pn within the Whakatane-Gisborne area (Haines, 1979) and assuming that the critical distance for a refraction from the 8.1 km/s layer is at, or after, 161 km (the distance of Shot 2 to Gisborne). With this assumption, and that of horizontal isovelocity layers, it can be calculated that the crust is at least 35 km thick at a point midway between the shot points and Gisborne.

It should be noted that a large gravity gradient exists between Whakatane and Gisborne and therefore the assumption of horizontal isovelocity layers for this area may not be realistic; Solution 4 should be viewed with this borne in mind.

F.4 Conclusions

The results of this one-way seismic refraction experiment are interpreted to show that the top surface of a 7.3-7.7 km/s layer exists at a depth of about 15 km beneath the Central Volcanic Region. Because there were only a few seismograph stations beyond the critical distance, the velocity and depth to this layer are not well constrained by the seismic refraction data alone. However, Solution 3, which is based on Haine's determination of the velocity for Pn, is considered more reliable than Solution 1 or 2 because: i) the uncertainty in his velocity determination is less than those derived from the refraction data and, ii) his velocity determination is based on more or less reversed travel-time data; i.e. for ray paths that passed beneath CNZ to KRP and WTZ to CNZ (see Fig. F.1).

The relative "speed-up" for the ray path to Karapiro, with respect to that to Gisborne, is about 3.5 s at a distance of 150 km (Fig. F.2). A similar phenomenon was noted by Latter (1979) for arrivals recorded at Mangahao and Karapiro associated with the 1969 eruption of Mt Ruapehu. He found that "...there is a relative speed-up on the path north to Karapiro of 2.5 s compared with the path south to Mangahao." These two stations are both situated almost exactly 150 km from Mt Ruapehu (Fig. F.1). This relative speed-up is similar to that observed for this experiment, and can be explained by a ray path from Mt Ruapehu to Karapiro which, for the most part, traversed along the top of a 15 km deep, 7.4 km/s layer at the western margin of the Central Volcanic Region.

The principal conclusions arrived at from this refraction experiment are:

(i) The 7.3-7.7 km/s layer is here interpreted as representing an anomalous low-velocity, and therefore low-density, upper mantle for the Central Volcanic Region. The thin crust for the Region, as deduced by this seismic experiment, provides a suitable mechanism by which the low-density mantle and superficial low-density volcanic rocks are isostatically compensated.

(ii) It was noted in section 2.1.3 (fig. 2.5) that Woollard (1970) had found a systematic relationship between depth to the mantle (H_m) and seismic velocity of the upper mantle (V_m). His relationship, which is based on over 100 data points from studies made in North America, is:

$$H_m(\text{km}) = (V_m (\text{km/s}) - 7.95) 33.2 + 34.$$

Inserting a value of $V_m = 7.4$ km/s in the above equation gives $H_m = 15.7$ km which is a result that is in close agreement with what is found here for the Central Volcanic Region. Thus the relationship of Woollard may have an applicability which extends beyond the continental structure of North America.

(iii) The geometry of the survey was such that the ray paths between the shots and the seismograph stations (except Gisborne) traversed beneath an area which, for the most part, was within the Central Volcanic Region but outside of the Taupo Volcanic Zone (Fig. F.1). However the very low values for the velocity of P_n were also found by Haines (1979) for a ray path that passed beneath the Chateau and Whakatane Stations - i.e. a ray path that would have traversed entirely beneath the Taupo Volcanic Zone. Hence it appears that it is the whole of the vee-shaped Central Volcanic Region (Fig. F.1),

which includes the Taupo Volcanic Zone, that is characterised by a low-velocity, low-density upper mantle, and therefore by deduction, a thin crust. This is in accordance with the gravity models presented previously.

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