LONG-OFFSET TRANSIENT ELECTROMAGNETIC MEASUREMENTS AT THE WAIRAKEI GEOTHERMAL FIELD

T. G. CALDWELL[1], S. L. BENNIE[2], D. E. KEEN[1] & D. J. GRAHAM[1]

[1]Institute of Geological and Nuclear Sciences, Lower Hutt
[2]Institute of Geological and Nuclear Sciences, Wairakei

SUMMARY – Transient electric field measurements were made in a line crossing the northwestern boundary of the Wairakei geothermal field. Three grounded bipoles were used as current sources enabling the apparent resistivity tensors to be determined at each measurement site. Source-receiver separations ranged between 7 and 10 km. Despite high levels of electrical noise instantaneous apparent resistivity tensors could be determined reliably between 0.1 and 4 s. The apparent resistivity image derived from these data shows that, beneath the southern part of the Wairakei geothermal field, higher resistivities underlie a layer of much lower resistivity about 1 km thick.

1. INTRODUCTION

In contrast to the oil industry, knowledge of the subsurface structure of the deeper parts (>1km) of geothermal reservoirs in the Taupo Volcanic Zone (TVZ) has been largely obtained from production drilling rather than from geophysical surveys. The oil industry makes wide use of seismic reflection surveys and, in most cases, obtains high quality images of the reservoir structure. Unfortunately, seismic reflection surveys that have thus far been conducted in the TVZ have produced only poor quality images. The poor quality of the images is due in large part to the difficulty of removing very strong reverberation effects produced in the near surface layers of unconsolidated volcaniclastics which cover the entire region (Bannister and Melhuish, 1994).

1.1 DC resistivity surveys

Of the geophysical techniques applied in the TVZ direct current (DC) electrical resistivity surveying (Bibby, 1988) has been by far the most effective. These surveys have successfully delineated the shallow extent of all the known geothermal fields in the TVZ (e.g. Bibby et al., 1995).

DC resistivity methods have also been used in the TVZ to study the deeper structure of the geothermal fields (Risk et al., 1970; Bibby and Risk, 1973) and more recently regional structure of the TVZ itself (Bibby et al., 1998). The majority of these surveys were made using the multiple-source bipole-dipole method (Fig. 1) that was first applied at Ohaaki in 1968 (Risk et al., 1970). Since the maximum 'detection depth' in a multiple-source bipole-dipole survey is proportional to the source-receiver separation or offset, typically 5–15km, these surveys contain information about both the near surface resistivity as well as the deeper parts of the geothermal system. In any DC resistivity survey the variation of resistivity with depth must be inferred from measurements made at different source-receiver separations. Distinguishing the effects of changes at depth from near surface resistivity variations requires careful analysis and a large number of measurement points distributed over the area of the geothermal field and its surroundings.

Using the multiple-source bipole-dipole method Bibby and Risk (1973) were able to infer the general conductivity-depth structure at the Ohaaki geothermal field. At Wairakei, multiple-source bipole-dipole surveying (Risk, 1984) clearly showed that resistivities at depth in the southern part of the field were significantly greater than in the northern part of the field (Fig 1).

While the DC method provides a map view of the data, information on the variation of resistivity with depth beneath each measurement site can also be derived from transient behavior of the same recorded-waveform used to determine the DC apparent resistivity. In this case the technique is known as the Long Offset Transient Electromagnetic (LOTEM) method.

Apparent resistivity tensor analysis techniques originally developed for multiple-source DC surveys can be applied to the transient EM data (Caldwell and Bibby, 1998). This analysis technique facilitates the visualization of data from multiple sources and enables distance-time resistivity sections or 'pseudo-sections', analogous to seismic reflection sections, to be derived directly from the measured electric field data. Results from three dimensional (3D) computer modeling suggest that apparent resistivity images made using these techniques provide good representations of the subsurface.

This paper describes an experiment at the Wairakei geothermal field designed to test the applicability of the LOTEM method for geothermal exploration in a realistic (i.e. electrically noisy) environment. The source bipoles for the LOTEM experiment were the same as those used in the DC multiple-source bipole-dipole survey described by Risk (1984). The locations of the LOTEM measurement sites, which run in a line crossing the north western (resistivity) boundary of the geothermal field, are shown as squares in Fig. 1.

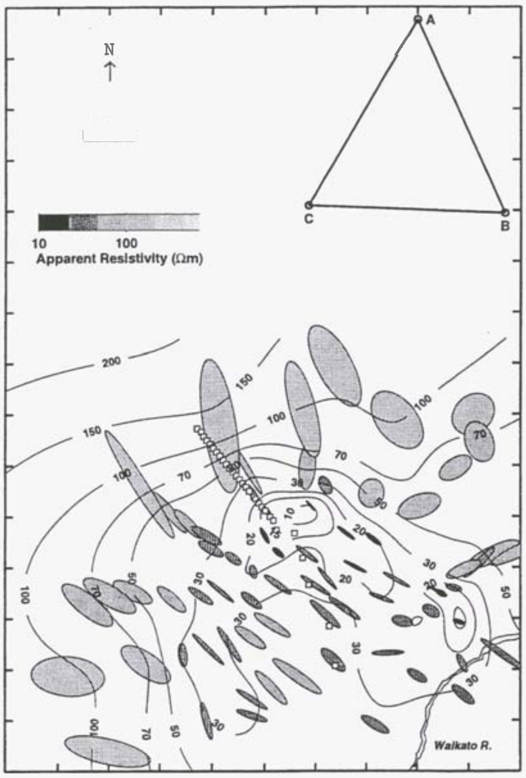

Fig. 1. DC apparent resistivity ellipses obtained by Risk et al., (1984). Contours show the P_2 invariant for the DC apparent resistivity tensor. The LOTEM measurement sites are shown by squares. The 3 positions of the current electrodes (A, **B** & C) and source-bipoles used for both the DC and LOTEM surveys are shown by the triangle.

2. LOTEMMETHOD

In order to image subsurface resistivity structures in the depth range typically used for geothermal production (1-2 km) in the TVZ, low frequency EM waves (ca. 1 **Hz)** must be used. At such frequencies propagation **of** the EM waves in the earth is described by a diffusion equation. The diffusive nature of the EM wave propagation implies that EM images will have less resolution than a comparable seismic image. In a study of different EM techniques Pellerin et al., (1996) showed that only the magnetotelluric (MT) and LOTEM methods were capable of detecting a conductive reservoir beneath a more highly conductive body, the conductive near-surface body representing a highly altered clay-rich zone above the reservoir. **A** similar study by Caldwell and Bibby (1993) showed that the DC multiple-source bipole-dipole method shared this ability.

The MT method utilizes naturally occurring EM waves and can be thought of as the EM analogue of passive **(or** earthquake) seismology. The LOTEM method is an active source technique that uses a grounded current bipole as a signal source. MT measurements are made in the frequency domain and measure the EM response at the surface due to vertically incident plane waves. In contrast, the LOTEM method measures the time domain (step function) response due to a distant current bipole. The two methods are closely related. Indeed, at short times after the current is turned on in the source bipole, the EM field at a (distant) receiver site is a close approximation to a vertically incident plane wave. It is in this 'early time' period where the controlled source MT (CSMT) method is applicable. **Thus** at high frequencies or at early **times,** the **CSMT** and LOTEM techniques are frequency and time domain versions **of** the same technique, although field measurements and analysis methods differ.

Two different polarization directions are required to completely characterize the EM response at each measurement site. This is achieved in a LOTEM survey by using current bipoles with differing orientations. By using three rather than two source bipoles, arranged as shown in Fig.1, a degree of data redundancy can be introduced into electric field measurements. This extra information can be used in the data processing to enhance signal recovery.

2.1 Measurement Equipment

In this experiment only the electric field was measured. The electric field produced by each current bipole was recorded at the receiver site using a pair of orthogonal dipoles, each 50 m long. Originally it was intended to record the LOTEM data using a Zonge GDP-32 receiver. This receiver can be phase locked with the 30kW Zonge GGT30 transmitter used to supply current to the source bipoles. Unfortunately the GDP-32 malfunctioned at the beginning of the fieldwork.

Instead voltages were recorded using a 12-bit digital data acquisition system originally developed for recording seismological data that has been adapted for resistivity data collection (Caldwell, **1990).** This system can not easily be phase locked with the transmitter and the **'1st** breaks' of EM signals were picked manually. Data were sampled at **32 Hz.** Signal amplitudes were great enough to be able to pick the origin time of the signal waveforms to within 1 or 2 sample intervals. The uncertainty in origin time is less than the 100 ms rise time of the front end Bessel filter used to condition the signals before digitization.

2.2 Signal Processing

Approximately 100 waveforms (about 30 from each bipole) were recorded at each receiver site. A least-squares procedure was used for signal processing. This procedure takes advantage of the constraint that the current-normalized signal from the 3 transmitting bipoles must sum to zero i.e.

$$\Sigma \mathbf{E}_i(t)/I_i = 0,$$

where $\mathbf{E}_i(t)$ and I_i are the measured electric field vector and current for each bipole respectively. This procedure also removes long-period telluric noise and impulsive noise spikes from the recorded time series. An example of a processed signal waveform is shown in Fig. **2.** For this site, the southern-most measurement (Fig. 1), the source-receiver separation is **10** km.

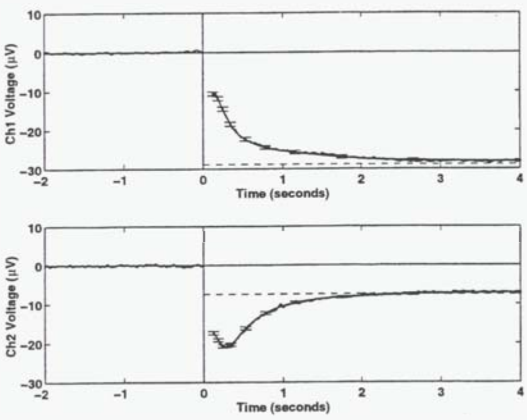

Fig. 2. Signal waveform for the transmitter bipole AB measured at the southern-most LOTEM site (locations in Fig. 1). The current, switched at on at t = 0, was **40** A giving a source moment of **16x10^4** Am. Standard errors in the voltage are shown by the error bars and are about **±1 μV** at this site.

2.3 Apparent Resistivity Tensor Analysis

The time varying electric fields are normalized for the effects of geometry and source current by using an analogue of the 3D form of Ohm's law i.e. $\mathbf{E}(t)=\rho_a(t)\mathbf{J}$. Here $\mathbf{E}(t)$ is the measured electric field vector produced by a step current switched on at t = 0, $\mathbf{J}$ is the steady state (DC) current density vector produced by a bipole source on the surface of a uniform half-space, and $\rho_a(t)$ is the 'instantaneous apparent resistivity tensor' (Caldwell and Bibby, **1998).** The current density **J** acts as a time-independent normalisation factor.

The apparent resistivity tensor provides a compact way of representing the time varying electric field data from multiple-sources. This (non-symmetric) tensor can be represented using three independent coordinate-invariant apparent resistivities and can be shown to be almost independent of the orientation of the source bipoles. Graphically the tensor can be represented as an ellipse, each ellipse representing the apparent resistivity tensor at a particular time. The lengths of the major (ρ_{max}) and minor (ρ_{min}) axes correspond to two of the three tensor invariants. Functions of the invariants are also coordinate invariant. In particular, the apparent resistivity invariant associated with the area of the ellipse (denoted by P_2) given by

$$P_2(t) = \sqrt{([\rho_{max}(t)\rho_{min}(t)])} = \sqrt{\det[\rho_a(t)]},$$

has a number of useful properties and plays a special role in the analysis.

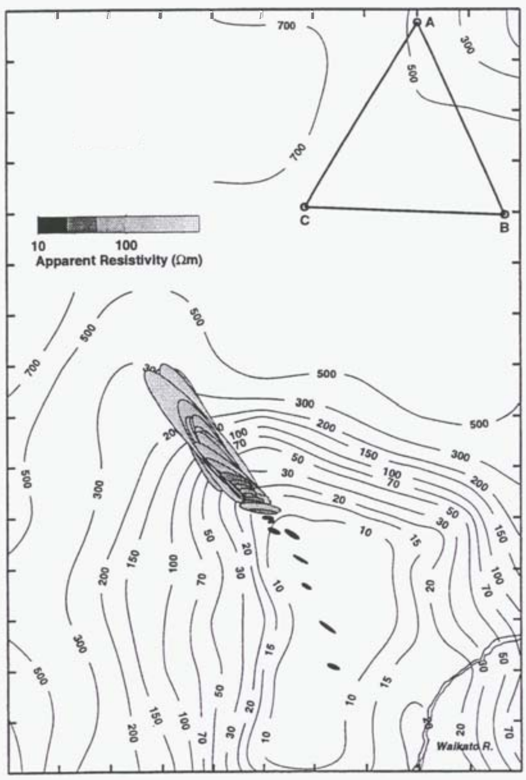

Fig. 3. LOTEM apparent resistivity ellipses at t=0.125s superimposed on to a contour map of the **AB/2=500m** Schlumberger apparent resistivity.

3. RESULTS AND DISCUSSION

Apparent resistivity ellipses are shown in Fig. **3** and Fig. **4** at t =0.125 and t =4s respectively. The LOTEM apparent resistivities at t = **4** s have reached their steady state values and are directly

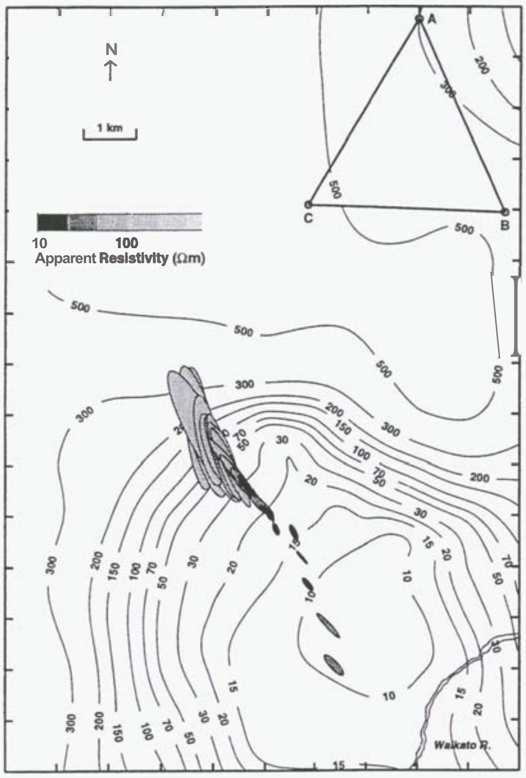

Fig. 4. LOTEM apparent resistivity ellipses at t=4s superimposed on to a contour map of the AB/2=1000m Schlumberger apparent resistivity.

comparable with the DC multiple-source bipole-dipole data shown in Fig. 1. Apparent resistivity ellipses for these two data are in good agreement. As expected theoretically, the ellipses observed at **0.125 s** (Fig. **3)** within the area of low Schlumberger apparent resistivity have a different orientation than those recorded at **4** s. Outside the low resistivity area the change in ellipse orientation with time is much less. The rate of change depends on the underlying resistivity and occurs more slowly in areas of low resistivity.

At the southern end of the LOTEM measurement line the apparent resistivities at **0.125s** are significantly smaller than those at **4s.** This change corresponds to an increase of resistivity with depth and can be seen more readily in the time-domain apparent resistivity sounding curves, an example of which is shown in Fig.5, and in the distance-time apparent resistivity pseudo-section shown in Fig. 6a. For sites near the transition from low to high Schlumberger apparent resistivities (Fig. **3** and **4)** the time-domain (P_2) apparent resistivities decrease slightly with increasing time. At this greater depth the low resistivity extends about 0.5km further north **of** the area of low Schlumberger apparent resistivity.

Outside the area of low Schlumberger apparent resistivity, in the north-western part of the survey line, the time-domain apparent resistivities are almost constant with time.

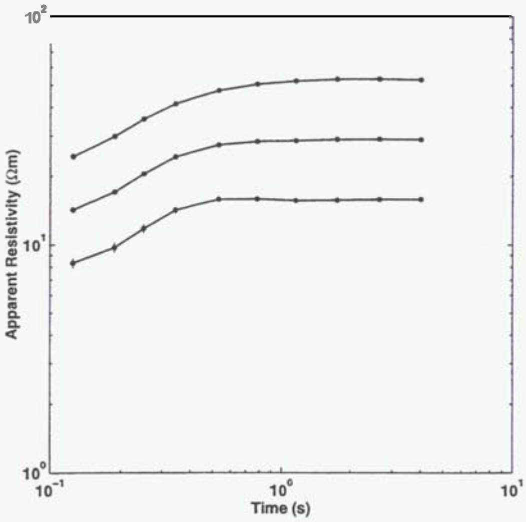

Fig. 5. Time domain sounding curves (ρ_{max}, **p—** and P_2 tensor invariants) measured at the southern most **LOTEM** station.

The pseudo-section (Fig. 6a) **also** shows the effects of a localised near surface inhomogeneity on the measurements. These effects, known as static effects in MT surveys, take the form of narrow vertical bands **of** anomalously high or low apparent resistivity. In both cases the effect is seen most strongly at one site and more weakly at an adjacent site. In the anomalously low resistivity case the most strongly **affected** measurement was made in an area of visible hydrothermal activity. The site with anomalously high apparent resistivity values was centred about 30m away fiom the well head of **WK231** and the high values here may reflect near surface changes caused the drilling and the construction of the well pad.

3.1 Apparent Depth

In a uniform half space with resistivity ρ and magnetic permeability μ_o, the 'diffusion-depth' is given by

$$\delta_{TD} = \sqrt{\frac{2\rho t}{\mu_o}}\ .$$

This parameter provides a measure of the 'detection depth' up to a limit given by the source-receiver spacing. For diffusion depths greater than the source-receiver distance the apparent resistivity is constant with time and is **equal** to the steady state or DC apparent resistivity. In a heterogeneous earth, a similar measure of detection depth is provided by the 'apparent depth' d given by the time integral of the diffusion velocity, i.e.

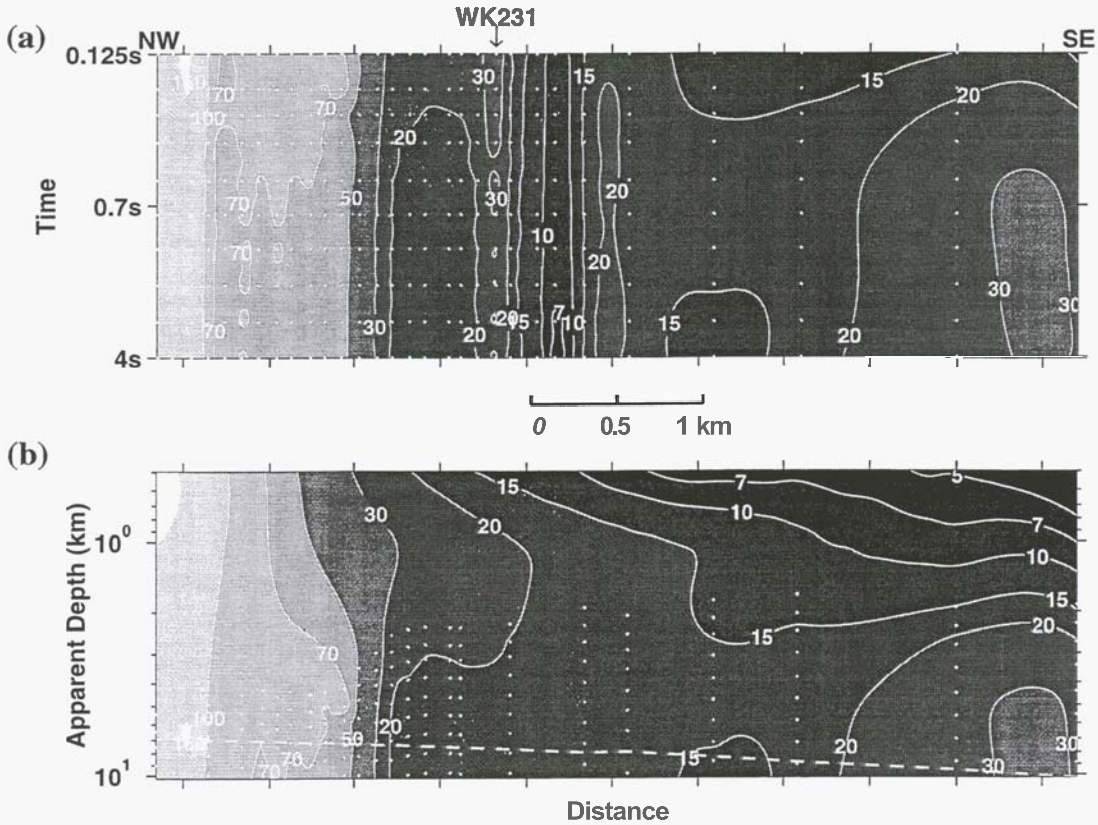

Fig. 6. (a) Distance-time apparent resistivity (P_2) pseudo-section. The vertical rows of dots show the position and times of the apparent resistivity measurements at each site. (b) Apparent-depth pseudo-section constructed from the data shown in (a) and the 500m and 1000m (AB/2) Schlumberger array data. The dots show the LOTEM apparent depth values. Measurement sites that show static effects have been removed. The dashed line shows the source-receiverdistance.

$$d = \int_0^t \left(\frac{\partial \delta_{TD}}{\partial t}\right) dt \cong \sum_i \sqrt{\frac{2P_2(t_i)\Delta t_i}{\mu_o}}$$

The corresponding measure of depth for a Schlumberger array resistivity measurement is given by half the current electrode spacing (AB/2).

Fig. 6b shows the apparent depth image constructed from (interpolated) Schlumberger array data (AB/2=0.5km and AB/2=1km) and the time-domain apparent resistivity data, $P_2(t)$, shown in Fig. 6a. In the southeast the P_2 apparent resistivity is greater than 20 Ωm for apparent depths >2 km. Schlumberger sounding data in this area (Risk, 1984) suggest that resistivities fall to <5 Ωm within 100m of the surface. Resistivity modelling near the eastern boundary of the geothermal field (Risk and Bibby, 1997) showed that, below a thin layer (thickness 100-m) of high resistivity, resistivities in the top 1 km of the field were of ca. 3 Ωm. Where the near surface resistivity is less than the shallowest (apparent depth) P_2 value the apparent depth will be an over-estimate of actual depth. Assuming a value of about 3 Ωm for the upper part of the resistivity section the EOTEM results suggest that resistivities have risen to > *20* Ωm at about 800m. The location of the high apparent resistivity (>30 Ωm in Fig. 1) area in the south correlates well with the extent of the Karapiti rhyolite. This rhyolite is known to contain cooler fluids than wells drilled further to the north in the lower resistivity area.

The preceding discussion highlights the importance of collecting the early-time data, which is needed to define the near surface resistivity structure. Without such data the apparent depth estimates may be significantly biased. The need to collect early-time data places stringent requirements on the recording instrumentation. To collect both early and late time apparent resistivity data (1ms to 10s) a wide bandwidth recording system is required. Unless the dynamic range of the recording system is very large the AC noise level will saturate the initial amplification stages of the receiver. Below the AC line frequency time-domain data can be collected quickly. In this experiment about 100 waveforms were collected from three bipoles in about 15 minutes and uncertainties in the apparent resistivities were 5% or less at most stations. At early times a frequency domain approach to recording would avoid the noise problem. Thus a combination of time and frequency domain recording techniques may prove to be most effective in practice.

4. CONCLUSIONS

The experiment reported here provides an example of the type of EM survey that will be needed to image resistivity structures within the geothermal fields in the TVZ. The high resistivity structure in the southern part of the Wairakei field, first noted by Risk (1984), can be clearly imaged beneath a cover of very low resistivity material. In the north west, at depths greater than were sensed by the Schlumberger array measurements, low resistivities continue about 0.5km further north of the resistivity boundary given in Risk (1984).

Even in an environment as electrically noisy as the Wairakei geothermal field, with its proximity to a power station and high voltage transmission lines, this experiment has shown that LOTEM signals can be easily measured in the time window between about 0.1 to 10 s. It is in this period range where the natural source signals used in MT surveys are weakest and the collection of MT data most difficult.

To determine the geometry of the resistivity structure more accurately than the apparent depth transformation used here requires the use of a 3D-inversion procedure. The apparent depth section can be considered to be the first step in this inversion and provides a starting point for an iterative inversion process. Inversion schemes capable of inverting time-domain EM data from a 3D situation are, however, still in an early stage of development.

The inability to image the internal structure of the geothermal reservoirs in the Taupo Volcanic Zone using seismic methods is a significant impediment to geothermal development. However, even the relatively low-resolution images produced by electromagnetic surveying techniques, if available at an early enough stage in a drilling program, would be of considerable value. However, it must be stressed that measurement densities more typical of a seismic reflection survey for the oil industry will be required. Without high measurement densities it will not be possible to (spatially) filter out the disturbing effects of localized near surface geological noise or to adequately image the 3D resistivity structures that characterize geothermal fields in the TVZ.

5. ACKNOWLEDGMENTS

We would especially like to thank Harry Rayner who came out of a well-earned retirement to assist us with the fieldwork. George Risk and Hugh Bibby provided many useful insights and suggestions. C.P. Wood provided encouragement and quickly approved funding for replacement parts when vital equipment broke down. Funding from Foundation for Science, Research and Technology supported this work, which is IGNS contribution No. 1776.

6. REFERENCES

Bannister, S.C. and Melhuish A., 1997. Seismic scattering and reverberation, Kaingaroa plateau, Taupo Volcanic Zone, New Zealand. *New Zealand Journal of Geology* and *Geophysics* 40,375-381.

Bibby, H.M., 1988. Electrical resistivity mapping in the Central Volcanic Region of New Zealand. N.Z. *J. Geology* and *Geophysics,* 31,259-274.

Bibby, H.M., Caldwell, T.G., Davey, F.J., and Webb, T.H., 1995. Geophysical evidence on the structure of the Taupo Volcanic Zone and its hydrothermal circulation. *Journal of Volcanology and Geothermal Research,* **68,** 29-58.

Bibby, H.M., Caldwell, T.G. and Risk, G.F., 1998. Electrical resistivity image of the upper crust within the Taupo Volcanic Zone, New Zealand. *Journal of Geophysical Research,* 103, (B5), 9665-9680.

Bibby, H.M. and Risk, G.F., 1973. Interpretation of dipole-dipole resistivity surveys using a hemispheroidal model. *Geophysics,* **38**, 719-736.

Caldwell, T.G., 1990. Digital Recording Equipment for Apparent Resistivity Measurements. Department of Scientific and Industrial Research, Geology and Geophysics, Contract Report No. 1990/9.

Caldwell, T.G. and Bibby, H.M., 1993. A comparison of electromagnetic methods with the DC resistivity multiple-source bipole-dipole method for deep geothermal exploration. *Proceedings of the 15th N.Z. Geotheml Workshop,* 187-194.

Caldwell T.G. and Bibby H.M., 1998. The instantaneous apparent resistivity tensor: a visualisation scheme for LOTEM electric field measurements. *Geophysical* Journal *International,* 135, 817-834.

Pellerin, L., Johnston, J.M. and Hohmann, G.W., 1996. A numerical evaluation of electromagnetic methods in geothermal exploration, *Geophysics* 61, 121-130.

Risk, G.F., 1984. Electrical resistivity survey of the Wairakei geothermal field. *Proceedings 6th NZ geothermal Workshop,6*, 123-128.

Risk G.F. and Bibby, H.M., 1997. Eastern boundary of Wairakei geothermal field: Resistivity measurements. *Proceedings of the 19th N.Z. Geotheml Workshop,* 141-148.

Risk, G.F., Macdonald, W.J.P. and Dawson, G.B., 1970. D.C. resistivity surveys of the Broadlands geothennal region, New Zealand, *Geothermics, Special Issue, 2*, **2**, 287-294.