Imaging the Roots of High-Temperature Geothermal Systems Using MT: Results From the Taupo Volcanic Zone, New Zealand

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ABSTRACT

High-temperature, economic geothermal resources are currently restricted to drillable depths (typically less than 3 km). However, the roots of these geothermal resources originate at much greater depths, where fluids extract heat from high temperature sources (i.e. magma) and are transported to the surface. This convective process is not well understood; nor is the associated deep-seated permeability required. Besides sourcing hydrothermal systems from which electricity can now be generated, these deep-rooted geothermal fluids also represent untapped energy resources themselves, which in the coming years may well be exploited with advances in deep drilling technology. Thus a very good case can be made to expand our knowledge on the roots of geothermal systems to better extract the existing resource base as well as tap into new sources of geothermal energy, including sources beyond conventional extraction depths in the coming years. In this paper, we probe mechanisms of heat exchange and fluid transport beneath four high-temperature geothermal systems in the Taupo Volcanic Zone (TVZ) using one of the largest 3D magnetotelluric (MT) arrays ever assembled (259 MT soundings, 1250 km²). We demonstrate the reliability of full tensor 3D MT modeling by applying a different data processing and 3D inversion algorithm to data analyzed by Bertrand et al. (2012), retrieving many of the same dominant resistivity features. We expand upon their study by using 90 additional MT soundings that cover a new area to the NW, and considering two geothermal systems not discussed in the previous work. According to our resistivity model, sustained convection cells extract energy and volatiles from quasi-plastic rock at ~5 – 7 km, elevating fluid density. Hot saline fluids migrate upward from this depth, potentially controlled by heavily fractured zones (i.e., fault accommodation zones), depressing resistivity as they are transported to the geothermal systems. Intrusive volcanics may also contribute to the low resistivity features imaged between 3 and 7 km.

1. Introduction

The TVZ is characterized by active extension (8 mm/yr) and rhyolitic volcanism driven by oblique subduction of the Pacific
Plate under the Australian Plate. 23 high-temperature (>250°C) geothermal systems hosted within this rift signal an extraordinarily high heat flux, ten times the average continental heat flow (Rowland & Sibson, 2004).

It is currently believed that high temperature convection plumes extending down to depths of 7 km provide the fluids for the TVZ geothermal systems (Bibby et al., 1995). However, there is a high degree of uncertainty regarding the structure of basement rocks (greywacke and meta-sediments), mechanisms of heat transport and rock permeability below the present maximum-drilled depth of 3 km. While shallow geothermal systems have been discovered and imaged by DC or time-domain EM techniques in the past, magnetotellurics (MT) has a unique ability to characterize deep resistivity. In recent analysis of broadband MT data, Bertrand and his colleagues illuminated deep-seated electrically-conductive plumes extending down to 8 km depth beneath two geothermal systems (Ohaaki and Rotokawa), which manifest in near-surface expressions of hydrothermal activity and clear permeable pathways for transporting geothermal fluids to the Earth’s surface. In this paper we expand upon their analysis by imaging an expanded set of these data, now covering 1250 km² with a station spacing of two kilometers; the additional data extend the survey coverage to the NW beyond the Atiamuri geothermal system.

2. MT Method

The MT method exploits naturally occurring electromagnetic (EM) wave fields over the Earth’s surface as sources to image subsurface resistivity structure. The EM fields arise from regional and worldwide storm activity and also from interaction of the solar wind with the Earth’s magnetic field (magnetosphere). These fields radiate through the air and enter the Earth with vertical plane-wave incidence, and are arbitrarily polarized over a 3D Earth, requiring a tensor formulation (i.e., vector measurement of the EM fields) to completely represent the subsurface geoelectric structure.

The measured horizontal EM field spectra (Ex, Ey, Hx, Hy) are interrelated by:

\[ E = [Z] H \]

where \( Z \) is a complex 2x2 tensor, called impedance, obtained for each MT recording station as a function of frequency. An equivalent parameterization of impedance yields apparent resistivity and impedance phase quantities, which are more intuitive to inspect and interpret (Vozoff, 1991).

3. MT Data Analysis

Two hundred fifty-nine (259) MT broadband soundings (100 – 0.001 Hz) have been recorded in the south-eastern TVZ between 2009—2013. These stations form a rectangular array (25 x 50 km) extending from the south-eastern rift margin, southeast of Ohaaki and Rotokawa geothermal systems, to the northwest, beyond the Atiamuri geothermal system (see Figure 1). All data were robustly processed by GNS Science using a remote reference to eliminate local and cultural noise (Bertrand et al., 2012).

Our experience modeling MT data shows that it is necessary to treat the full MT impedance tensor in a three-dimensional (3D) resistivity imaging process for this case. Analysis of deeper and increasingly complex resistivity structure using only principle component tensor data (the off-diagonal components that are dominant in 2D analysis) can produce artifacts in the imaging process. While it may be possible to align the imaging grid to preferred structural trends in the data (Kiyan et al., 2013; Tietze & Ritter, 2013), soundings that show electrical strike changing with frequency can make this difficult if not impossible to achieve (e.g., Bedrosian et al., 2004). Nevertheless, full MT tensor analysis also has its own issues, especially how one treats data noise for all components in the impedance tensor, where individual tensor elements can vary by several orders of magnitude at a fixed period.

Assigning data weights for full tensor inversion is non-trivial, because on-diagonal impedance tensor elements are of relatively low amplitude and higher variance. We have developed a data weighting by which we down-weight data with high variance and use conservative error floors based on impedance element amplitude: 15% for Zxx and Zyy; 3% for Zxy and Zyx.

![Figure 2: Examples of LBNL forward model fits (lines) to recorded data (circles) at 4 stations in 4 different geothermal systems covered by the array. The colors indicate impedance tensor element: Zxy=red; Zyx=blue; Zxx=green; Zyy=turquoise.](image)
4. 3D MT Inversion

The inversion process solves Maxwell’s equations for 3D resistivity variations and plane wave source excitation at a discrete set of frequencies. We invert only the low frequency data (1—0.001 Hz) on a coarse modeling mesh (i.e., finite difference node separations of 1000 m across interior). We will refine this model in future work using the high frequency data to add resolution above 2 km depth. To stabilize the inversion, additional constraints were added such as spatial smoothing of the resistivity model. This inversion workflow has been shown to retrieve 3D images of geothermal systems in an accurate and efficient manner (Lindsey & Newman, in review). The 3D MT preconditioned inversion code (Newman & Alumbaugh, 2000; Newman & Boggs, 2004) was run on approximately 4000 cores of NERSC Cray XT4 Hopper system.

3D MT inversion achieved a reduction in misfit to a root-mean squared value (i.e., summation of the square of the data-model residuals, normalized by their weights) of 2.5 after 219 iterations. Figure 2 compares forward model predictions of apparent resistivity and phase with recorded data. Off-diagonal data (Zxy and Zyx) are better fit than on-diagonal data; however, an acceptable fit of the Zxx and Zyy components was achieved when the data are of comparable amplitude.

5. Interpretation

Although resolution of our 3D model is coarse in the upper 2 km (it will be refined in future work), we image a near-uniform resistive surface layer down to 1.5 km depth consistent with Bertrand et al. (2012). The interpretation for this surface layer is based upon continuous (Quaternary–present) subsidence and periodic caldera collapse across the TVZ. It has been infilled by ignimbrite and rhyolitic lava, interbedded with lacustrine and fluvial sediments to create this resistive near-surface (Wood et al., 2001).

We image several low-resistivity anomalies in the upper few hundred meters, and interpret these as surface manifestations of the geothermal systems contained by the MT array (Ohaaki, Rotokawa, Ngatamariki, Orakei Korako, Te Kopia and Atiamuri) following Bibby et al (1995). Low electrical resistivity (2—5 Ohm-m) is a common geophysical indicator of clay mineralogy (e.g., smectite clays) below 200°C, which results from high-temperature convection in the underlying core of the geothermal reservoir. Above this temperature, the clays (illite and chlorite) are more resistive (50 – 200 Ohm-m).

We image at least four vertical low-resistivity (3—6 Ohm-m) zones that extend from 3 – 8 km depth in proximity to Ohaaki, Rotokawa, Ngatamariki and Atiamuri. Bertrand et al. (2012) showed that low-resistivity (3 – 5 Ohm-m) plumes rise from 10 km up to the Ohaaki and Rotokawa geothermal fields. Their modeling found that the deep low-resistivity plume at Ohaaki was offset to the NW relative to the surface manifestation by a few kilometers, while at Rotokawa the plume rose vertically under the geothermal field. Our model shows both deep plumes beneath Ohaaki and Rotokawa offset to the NW by ~3 km (Figure 3). We image similar offset plumes beneath Ngatamariki (Figure 4) and Atiamuri (Figure 5).

At Ohaaki and Rotokawa, we find the ascending features are connected at 6 – 12 km depth by an elongated low-resistivity (2 – 5 Ohm-m) branch that traverses the MT array SW—NE (Figure 6). This deep branch is characterized by two local minima.

Figure 3. Section views through the 3D resistivity model showing (a) Rotokawa and (b) Ohaaki geothermal systems (dashed lines). Representative data fits for Rotokawa and Ohaaki shown in Figure 2 are located at RK and OH.

Figure 4. Section views through the 3D resistivity model at Ngatamariki geothermal area (dashed circle). “NM” indicates position of MT sounding in Figure 2.
of approximately 2 Ohm-m at which points the vertical low-resistivity features begin to rise (Figure 6).

Immediately beside the vertical low-resistive zones of each geothermal system, we note the presence of vertical high-resistivity (>300 Ohm-m) zones. In the cases of Ohaaki and Rotokawa, these resistors are interpreted as the greywacke Kaingaroa Plateau, the eastern flank of the TVZ, which appears to be a barrier to hydrothermal fluid flow.

6. Discussion

Could the deep, vertically-oriented, laterally-offset low-resistivity features below Ohaaki, Rotokawa, Ngatamariki and Atiamuri be upwelling high-temperature geothermal fluids?

Bertrand et al. (2012) argue that upwelling plumes of hot (~330°C at Rotokawa) saline fluids migrate from 7 km depth at near hydrostatic pressure, chemically-altering the surrounding country rock and depressing electrical resistivity to below 5 Ohm-m (see Figure 6). This hypothesis directly supports the long-lived existence of convection cells, where low-resistivity features represent upward-propagating hot fluid pathways (Bibby et al., 1995). Our resistivity model indicates the surface geothermal manifestations are laterally-offset from the deep vertical low-resistivity features, indicating geothermal fluid upflow may be structurally-controlled (e.g., Rowland and Simmons, 2012). However, further analysis of vertical magnetic field data as well as high frequency MT recordings will add greater structural resolution across the upper 2 km of our model, perhaps clarifying this preliminary observation.

Could the vertically-oriented, laterally-offset low-resistivity features below Ohaaki, Rotokawa, Ngatamariki and Atiamuri be intrusive volcanics (i.e., lateral dikes)?

Dikes, sills and other intrusive volcanism can serve as structural controls on flow (e.g., Rowland & Simmons, 2012), as well as sources of geothermal heat for ~10,000 years (Norton & Knight, 1977). Naturally high TVZ heat flow indicates invasive volcanics at depth, although there is no exposed geological evidence of this fact. Thus, an important rheological boundary may exist at the base of the seismogenic zone (~6 – 7 km), potentially trapping basaltic dikes (Rowland and Sibson, 2004; Nairn et al., 2004). The SW-NE elongated low-resistivity branch imaged below Ohaaki and Rotokawa at 6 – 12 km depth could be evidence of a ponding zone of volcanic material. Yet sufficient buoyancy of rhyolite and perhaps andesite may allow dikes to intrude to shallower depths (Henley & Ellis, 1983; Johnson et al., 2011; Rowland and Sibson, 2004). For example, a few kilometers below Ngatamariki a dioritic pluton was intersected during drilling; however, this pluton was found to be too old (550-700 ka) to heat the current geothermal system (Browne et al., 1992). Deep drilling at Krafla geothermal field in Iceland, an analogous rift-hosted geothermal play, found an intrusive magma heat source at ~2 km depth, which 3D MT analysis indicates was transported to the surface in a vertical low resistivity zone extending down to 6 km depth (Gasperikova et al., 2011). Furthermore, observations of widely-varying epithermal ore vein deposits across the TVZ demonstrate the variable chemical-composition and geometry of emplaced magma below several geothermal systems (Rowland & Simmons, 2012). Hence it is possible that igneous bodies intrude above the TVZ brittle-ductile transition at ~6–7 km along preferential directions in the stress field, elevating the local geothermal gradient at 3 – 7 km.
How is Permeability Controlled?

Down-going cold, meteoric water filters through the interconnected pore networks of unfractured media (Rowland and Sibson, 2004). Across the southeast flank of the TVZ, the electrically resistive Kaingaroa Plateau is an example of this type of broad fluid source.

Within the rift, accommodation zones develop between ~20 km-long normal-faulted rift segments due to oblique tectonic forces, which act to enhance vertical permeability and enable upflow of buoyant high-temperature fluids (Rowland & Sibson, 2004). Accommodation zones have been shown to correlate with geothermal occurrence worldwide (Faulds et al., 2010); 60% of TVZ geothermal systems are located in accommodation zones (Rowland & Sibson, 2004). Thus, upwelling TVZ geothermal brines likely exploit structural permeability from 3 to 7 kilometers depth. For example, the Kaingaroa Fault zone appears to channel high-temperature upflow to the Ohaaki and Rotokawa geothermal features.

Within the geothermal reservoir permeability is fracture-dominated leading to intermediate resistivity values of several 10’s Ohm-m (Pellerin et al, 1996). Heise et al. (2008) found the core of the Rotokawa geothermal reservoir was ~100 Ohm-m, interpreting this as fracture permeability. We find similar intermediate resistivity values below the Ohaaki, Rotokawa, Ngatamariki and Atiamuri surface features.

7. Conclusions

Increasing our understanding of how heat is extracted from deeply-rooted geothermal systems carries obvious implications for resource exploration. Using a large (259 station) 3D MT array, we have probed mechanisms of heat exchange and fluid transport beneath four high-temperature geothermal systems in the TVZ. We demonstrated repeatability of MT modeling by applying a different data processing and inversion methodology to the same dataset, retrieving many of the same dominant resistivity features. Resistivity imaging cannot distinguish between different mechanisms that both cause low resistivity (i.e., saline brine versus partial melt), therefore combining MT information with seismic tomography or other data represents an important future direction.

References


