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Using array MT data to image the crustal resistivity structure of the southeastern Taupo Volcanic Zone, New Zealand

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ABSTRACT

Magnetotelluric (MT) data from 169 locations covering a 700 km^2 region in the southeastern part of the Taupo Volcanic Zone, New Zealand, are used to generate a 3-D resistivity inversion model to a depth of 10 km. Isolated zones of low-resistivity (<30 Ωm) below 3 km are imaged in the basement that appear to be associated with the geothermal fields at Rotokawa, Ngatamariki and Ohaaki, but not with the geothermal fields at Orakei-Korako and Te Kopia. These low-resistivity zones are interpreted to be recent intrusions that feed heat and magmatic fluid to the overlying system of hydrothermal convection supplying the geothermal fields. At Rotokawa, hypocentres of thermally induced seismicity coincide with the top of the deep low-resistivity zone and suggest that temperatures sufficient for melting are reached at ~4.5 km depth. In most cases these deep low-resistivity zones are located at the margins of areas of low-gravity, which suggests that the intrusions lie at the margins of old caldera collapse structures. The 3-D resistivity model indicates that geological structure and magmatic intrusions play a more direct role than previously envisaged, and significantly influence the overarching system of convective heat and mass transport in the Taupo Volcanic Zone.

1. Introduction

The Taupo Volcanic Zone (TVZ), located in the central North Island of New Zealand, is an actively spreading, rifted volcanic arc, and is the on-land extension of the Tonga–Kermadec arc to the northeast (e.g. Seebeck et al., 2014). Since ~1.8 Ma ago (Eastwood et al., 2013) extension in the central TVZ has stretched and down-faulted the former surface rocks, comprised of Mesozoic meta-sediments (greywacke) and overlying (andesitic) arc-volcanics. In the past few decades, geothermal production wells drilled along the southeast margin of the TVZ have intersected these Mesozoic greywackes and arc-volcanics at progressively greater depths to the northwest, with a maximum intersection at 3.4 km depth near the Ngatamariki geothermal field (Horton et al., 2012). Extensive layers of rhyolitic ignimbrite and volcaniclastics from numerous caldera-forming eruptions since ~1.8 Ma ago now overlie this 'basement' of Mesozoic greywacke (Houghton et al., 1995; Wilson et al., 1995, 2009; Wilson and Rowland, 2011; Chambefort et al., 2014). Very few andesitic volcanics remain exposed at the surface within the central TVZ, where high-silica rhyolite lava domes, thick ignimbrite sequences, volcaniclastics and Quaternary sediments blanket the older arc-volcanoes below.

Since ~0.34 Ma ago, the central TVZ has been the most productive silicic volcanic system on Earth (e.g. Houghton et al., 1995), and has an exceptionally high heat flux of 700 mW/m^2 (Bibby et al., 1995) that is more than 10-times the global mean value (65 mW/m^2) for continental crust (Pollack et al., 1993). This heat flux is discharged at the surface through 23 localised high-temperature geothermal fields. The conceptual model of heat transport in the central part of the TVZ envisages these geothermal fields as the tops of vertical convection plumes of high-temperature fluid at near-hydrostatic pressure in the brittle part of the crust (e.g. Bibby et al., 1995 and references therein). Numerical modelling of convection in homogeneous media (McNabb, 1975; Kissling and Weir, 2005; McLellan et al., 2010) and other geological and geophysical evidence (e.g. Grindley, 1965; Browne, 1979; Simmons et al., 1993; Bibby et al., 1994) suggests that the locations of the geothermal fields, and hence the underlying plumes have been stable for periods of at least 0.2 Myr. Convection is driven by a basal layer of partial melt located below the brittle–ductile transition (Heise et al., 2007), which is associated with the cut-off of shallow seismicity at ~6–7 km depth (Bibby et al., 1995; Bryan et al., 1999). Between the geothermal fields, upward conductive heat flow is suppressed by down-flowing, cold meteoric water that recharges the hydrothermal systems. Importantly, this conceptual model does not require a direct relationship between individual geothermal fields and localised sources of heat (e.g. magmatic intrusions).

The lateral extent of the geothermal fields at shallow depth (i.e. ~300 m) is well-delineated by zones of low-resistivity, identified from ~24,000 direct-current (DC) resistivity measurements (Bibby, 1988). More recently, inversion modelling of magnetotelluric (MT)

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measurements in the southeast TVZ has imaged 'plume-like' low-resistivity zones within the basement rocks that connect to the shallow geothermal fields (Bertrand et al., 2012). These low-resistivity zones have been interpreted as high-temperature saline fluids upwelling in fractures, supplying heat (and fluids) to the geothermal fields above, consistent with the overall model of convective heat flow.

However, heterogeneity in the MT models suggests that other influences (e.g. geological structure and magmatic intrusions) also play an important role. For example, a vertical low-resistivity zone is imaged directly beneath the Rotokawa geothermal field (Bertrand et al., 2012), while at Ohaaki, a deep low-resistivity zone is offset and dips to the northwest (Bertrand et al., 2013), suggesting a pathway of preferential permeability associated with pre-existing rift-faults or a collapse feature. Although problematic from the point-of-view of a (vertical) convective driving force, an elongated low-gravity anomaly centred northwest of Ohaaki (the Mihi Volcanic Depression; Soengkono, 2012) may provide the structural (permeable) pathway that causes the low-resistivity offset.

Furthermore, while the locations of the geothermal fields appear to be stable, geological evidence suggests that the heat output of several TVZ geothermal fields have episodically waxed and waned in response to nearby magmatism (e.g. Arehart et al., 2002; Milicich et al., 2013). For example, a 0.7 Ma old intrusive complex (Chambefort et al., 2014) with an overlying halo of magmatic–hydrothermal alteration was intersected at ~2.6 km depth beneath the Ngatamariki geothermal field (Browne et al., 1992). This intrusion is much too old to represent the heat source for the present-day geothermal system, and thus implies that at least two episodes of hydrothermal activity have occurred at this location in the past 0.7 Ma.

Here we use data from 169 MT soundings to construct a 3-D resistivity inversion model of a 700 km^2 area in the southeast TVZ (Fig. 1). These MT data were collected as part of a research programme to study processes that transport heat through the brittle part of the crust, targeting depths between 3 and 7 km; below drilled depths and above the base of the seismogenic zone, respectively (Bignall, 2010). These MT measurements encompass the Rotokawa, Ngatamariki, Ohaaki, Orakei-Korako and Te Kopia geothermal fields, and several major geological features in the TVZ: the southeast rift margin, parts of the Whakamaru and Reporoa collapse calderas, the Paeroa Fault, and the Maroa Volcanic Centre (a group of young 280–16 ka rhyolite domes; Leonard et al., 2010). The 3-D inversion model images the crustal resistivity structure beneath the 700 km^2 study area to a depth of ~10 km, extending earlier DC-resistivity surveys that mapped the near-surface electrical resistivity.

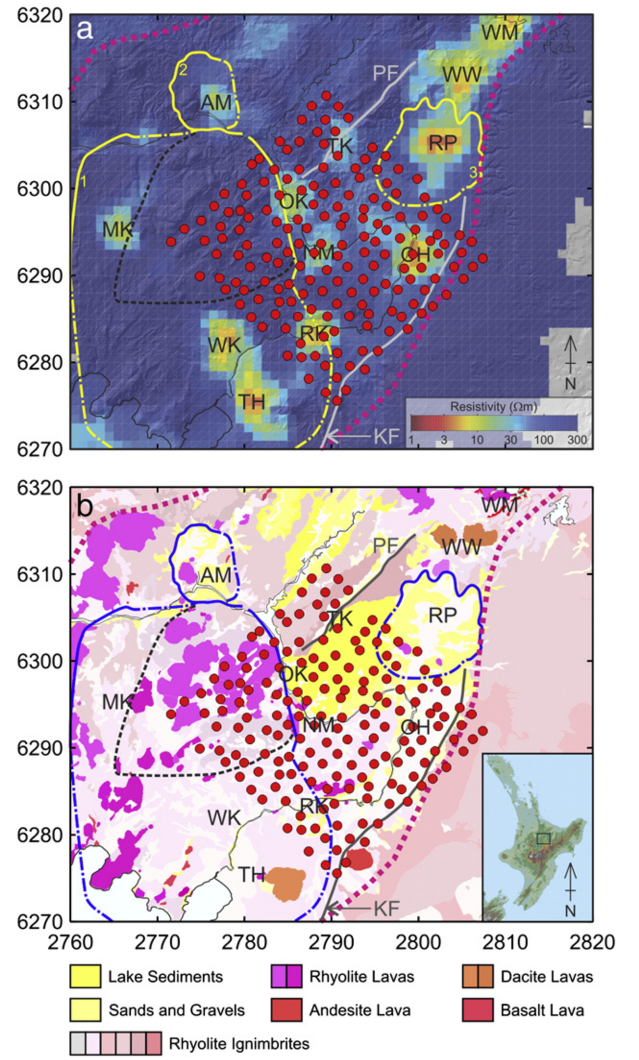

Fig. 1. a) Map of the MT array (red circles) in the southeast TVZ using the NZMG (New Zealand Map Grid) coordinate system (labels in km). The background colour shows the DC apparent resistivity map (Bibby et al., 1995) superimposed on a shaded relief map. Low resistivity zones correspond to geothermal systems: TH — Tauhara, WK — Wairakei, RK — Rotokawa, MK — Mokai, NM — Ngatamariki, OK — Orakei Korako, OH — Ohaaki, TK — Te Kopia, AM — Atiamuri, RP — Reporoa, WW — Waiotapu Waikiti, and WM — Waimungu. Yellow lines show inferred caldera collapse boundaries (1. Whakamaru, 2. Ohakuri, 3. Reporoa); solid lines show well constrained topographic collapse margins and dashed lines are inferred (Leonard et al., 2010). Dashed black line outlines the Maroa Volcanic Centre and the dashed pink lines bound the northwest and southeast extent of the TVZ (Wilson et al., 1995). Grey lines show the PF — Paeroa Fault and KF — Kaingaroa Fault. b) Similar to above with the regional geological map (Leonard et al., 2010) as the background. Black box in the inset map shows the study area within the North Island of New Zealand.

2. MT data and dimensionality

Five-component broadband (0.01 s < period < 1000 s) MT measurements were made at ~2 km spacing to form a regular array in the southeast TVZ (Fig. 1). All measurements were made using Phoenix Geophysics MT recorders and magnetic field sensors, with lead chloride electrodes forming ~70 m dipoles. Electromagnetic field time-series data were measured for a minimum of 40 h (2 night's duration) at each site. These time-series data were all processed using a remote reference (Gamble et al., 1979), located at a low-noise site ~10 km southeast of the array, on the Kaingaroa Plateau. By recording simultaneously at a remote site, the effect of uncorrelated magnetic noise at an array site can be minimised, yielding a high-quality homogeneous dataset. While parts of this dataset have been previously published (Bertrand et al., 2012, 2013), we present here a 3-D resistivity model of 169 MT measurements, including 20 new soundings made to the southeast of the Rotokawa geothermal system that were not included in the previous analysis.

Prior to inverting MT data, it is important to first assess the data dimensionality. To avoid the distorting effects of localised near-surface resistivity heterogeneities, we use properties of the magnetotelluric phase tensor Φ (Caldwell et al., 2004) for this analysis. These properties can be displayed graphically by representing the phase tensor as an ellipse: the orientation of the principal axes shown by the ellipse axes, its principal values by Φ_{max} and Φ_{min}, and the skew angle β measuring the tensor's deviation from symmetry. Maps of the observed skew angle are plotted in Fig. 2. Non-zero β values (in particular $\beta > 3°$; Booker, 2013) and variations in the orientation of the phase tensor major (Φ_{max}) and minor (Φ_{min}) axes indicate the presence of 3-D structure (Caldwell et al., 2004). Fig. 2 shows small β values for data measured at short periods (i.e. <~1 s). However, values of β are greater than 3° at longer periods, indicating that a 3-D inversion algorithm is most appropriate to model the resistivity structure that these data reflect.