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Characterisation of induced fracture networks within an enhanced geothermal system using anisotropic electromagnetic modelling

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ABSTRACT

As opinions regarding the future of energy production shift towards renewable sources, enhanced geothermal systems (EGS) are becoming an attractive prospect. The characterisation of fracture permeability at depth is central to the success of EGS. Recent magnetotelluric (MT) studies of the Paralana geothermal system (PGS), an EGS in South Australia, have measured changes in MT responses which were attributed to fracture networks generated during fluid injection experiments. However, extracting permeabilities from these measurements remains problematic as conventional isotropic MT modelling is unable to accommodate for the complexities present within an EGS. To circumvent this problem, we introduce an electrical anisotropy representation to allow better characterisation of volumes at depth. Forward modelling shows that MT measurements are sensitive to subtle variations in anisotropy. Subsequent two-dimensional anisotropic forward modelling shows that electrical anisotropy is able to reproduce the directional response associated with fractures generated by fluid injection experiments at the PGS. As such, we conclude that MT monitoring combined with anisotropic modelling is a promising alternative to the micro-seismic method when characterising fluid reservoirs within geothermal and coal seam gas reservoirs.

1. Introduction

With global energy production shifting away from fossil fuels, the development of economically viable renewable energy sources is gaining significant interest. One method which is showing promising results is geothermal, more specifically enhanced geothermal systems (EGS), with the Habanero EGS in Australia generating net power output without assistance (Geodynamics Ltd, 2013) and power production from EGS estimated to provide 100,000 MW of energy to the United States by 2050 (Tester et al., 2006). In a geothermal system, the naturally occurring temperature gradient within the Earth is utilised as a means of generating electricity through the heating of injected fluids. To generate energy efficiently, permeable pathways within hot lithologies must first be established through hydraulic fracturing. As such, the monitoring of injected fluids during fracturing and the characterisation of the resulting permeable zones are both crucial to the feasibility of an EGS.

Currently, micro-seismic tomography is the primary geophysical technique utilised when characterising fractures generated during hydraulic fracturing (House, 1987; Albaric et al., 2013). From this method we are able to estimate spatial distribution of fractures. However, as this method is not directly sensitive to the fluids present within those fractures, it is difficult to gain any information regarding the permeability or fluid motion occurring at depth. To rectify this issue, studies have begun to use electromagnetic (EM) methods which are directly sensitive to subsurface conductivity contrasts (Zlotnicki et al., 2003; Aizawa et al., 2005; Yasukawa et al., 2005; Peacock et al., 2012, 2013). One such method showing potential is the passive EM technique, known as MT with multiple authors using this method to locate and characterise potential geothermal targets (Heise et al., 2008; Newman et al., 2008; Arango et al., 2009; Spichak and Manzella, 2009; Geiermann and Schill, 2010). Furthermore, studies have also begun to identify the monitoring capabilities of MT Aizawa et al., 2011 with Peacock et al. (2012, 2013) measuring subtle conductivity contrasts within an EGS which were attributable to hydraulic fracturing. However, the spatial resolution of MT measurements decreases with depth. Therefore, EGS targets which are typically small and located at 3–5 km depth are unresolvable using conventional isotropic modelling.

One recent example of this is a study by Peacock et al. (2012, 2013) of the Paralana geothermal system (PGS), an EGS at Paralana, South Australia. During July, 2011, Petratherm Ltd. and joint venture partners Beach Energy and TruEnergy injected 3.1 million litres of fluid over 4 days into a metasedimentary package at 3680 m depth. The PGS is situated in a dilational zone along a splay off the eastward thrusting Paralana fault system (Paul et al., 1999; McLaren et al., 2002; Brugger et al., 2005) bounding the eastern margin of the Mt. Painter Domain

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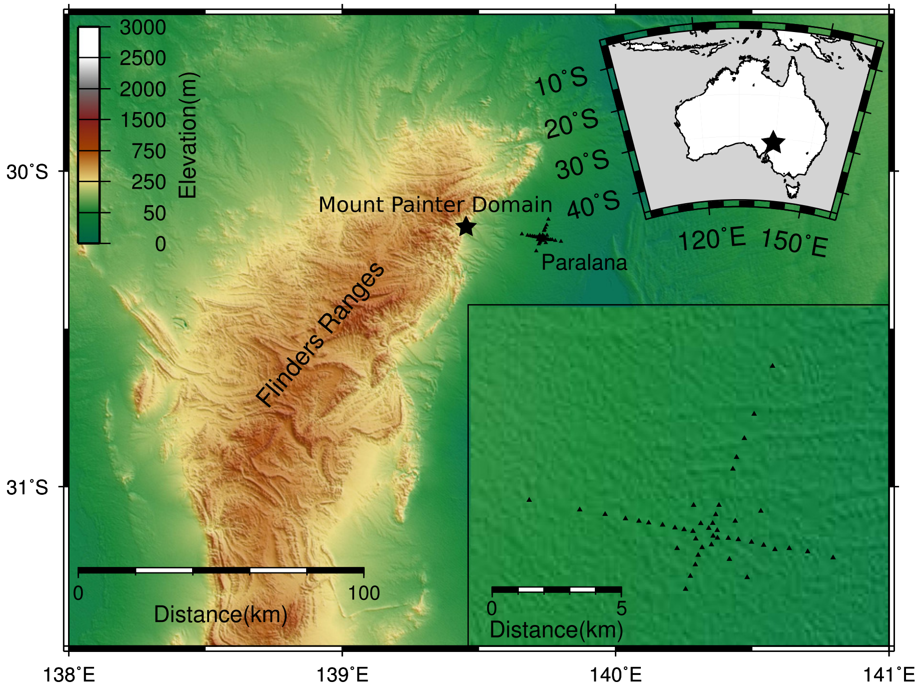

Fig. 1. Topographic location map of the Paralana geothermal system in South Australia with MT stations (from Peacock et al., 2012) displayed as black triangles and a star marking the Mount Painter Domain. The top right inset displays a map of Australia specifying the location of the Paralana geothermal system using a star. The bottom right inset displays a section of the station array with increased magnification.

(Brugger et al., 2005) in the Northern Flinders Ranges, South Australia (Fig. 1). The Mt. Painter Domain consists of granites, gneisses and meta sediments dated at approximately 1600 Ma to 1580 Ma in age (Kromkhun, 2010). These units were overlain by sediments (Paul et al., 1999; McLaren et al., 2002; Brugger et al., 2005; Wülser, 2009) with a maximum age of 800 Ma (Wülser, 2009) and a lower age limit constrained to the initiation of the Delamerian Orogen (Wülser, 2009), which occurred between 514 Ma and 492 Ma (Foden et al., 2006). Further granitic intrusions and tectonothermal events have been recorded throughout the history of the Mt. Painter Domain (Wülser, 2009) with the British Empire Granite intruding at approximately 460 to 440 Ma (Elburg et al., 2003; McLaren et al., 2006; Wülser, 2009).

To understand the pre- and post-fluid injection MT data collected at the PGS, micro-seismic (Albaric et al., 2013) and MT (Peacock et al., 2012) responses were collected before and after hydraulic fracture stimulation. From these responses a series of micro-fractures preferentially aligned towards the north-east were interpreted to have opened in response to the injected fluids from phase tensor misfit ellipses calculated from MT data measured before and after fluid injection. However, subsequent isotropic forward MT modelling did not adequately reproduce the measured responses (Peacock et al., 2013). As such, this study utilises results presented by Wannamaker (2005) to hypothesise that the preferentially orientated fracture network generated within the PGS produces an anisotropic response which can be characterised by a 2-dimensional anisotropic forward model.

A series of initial forward models are first generated to understand how each anisotropic modelling parameter influences the synthetic MT response. Preliminary results from these models along with results presented by Albaric et al. (2013) and Peacock et al. (2012, 2013) are then used to constrain an anisotropic forward model which adequately approximates MT data obtained by Peacock et al. (2012, 2013). From these results, we show that anisotropic MT modelling is able to overcome the limitations inherent of isotropic modelling and characterise the permeable pathways generated during hydraulic fracture stimulation within an EGS.

2. Hypothesis testing with forward models

To understand the fracture geometries generated at the PGS, a series of forward models are generated. Subtle variations in the modelled structures allowed us to test whether the responses associated with various anisotropic structures are capable of approaching the measured response. Each model calculated using the 2-dimensional MT direct code for conductors with arbitrary anisotropy of Pek and Verner (1997) has a common 1-dimensional, layered background defined by geological information from previous studies (Paul et al., 1999; McLaren et al., 2002; Brugger et al., 2005; Wülser, 2009; Kromkhun, 2010). This layered background consists of two sedimentary layers, the first extending to a depth of 1.06 km with a resistivity of 5 Ω m and the second from 1.06 km to 2.06 km with a resistivity of 10 Ω m following previous forward modelling by Peacock et al. (2013). These layers overlie a highly-compacted sedimentary layer from 2.06 km to 7.1 km with a resistivity of 200 Ω m, defined by 2-dimensional MT inversions of the PGS presented by Peacock et al. (2012). The remainder of the model, from 7.1 km to 417 km, is defined by a homogeneous resistive half space with a resistivity of 10,000 Ω m.

In order to represent the fluid injection into a fractured rock measured by Albaric et al. (2013) and Peacock et al. (2012), a 1 km wide block is introduced from 3.66 km to 4.46 km depth (Fig. 2). As the conductivity defining this volume is dependent on numerous variables, upper (σ_{HS}^{+}) and lower (σ_{HS}^{-}) Hashin–Shtrikman bounds for the electrical conductivity are calculated by (Hashin and Shtrikman, 1962)

$$\sigma_{HS}^{+} = \sigma_w + \frac{1-\phi}{\frac{1}{\sigma_m-\sigma_w}+\frac{\phi}{3\sigma_w}} \tag{1}$$

and

$$\sigma_{HS}^{-} = \sigma_m + \frac{\phi}{\frac{1}{\sigma_w-\sigma_m}+\frac{1-\phi}{3\sigma_m}} \tag{2}$$