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Quantitative temperature monitoring of a heat tracing experiment using cross-borehole ERT

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ABSTRACT

The growing demand for renewable energy leads to an increase in the development of geothermal energy projects and heat has become a common tracer in hydrology and hydrogeology. Designing geothermal systems requires a multidisciplinary approach including geological and hydrogeological aspects. In this context, electrical resistivity tomography (ERT) can bring relevant, qualitative and quantitative information on the temperature distribution in operating shallow geothermal systems or during heat tracing experiments. We followed a heat tracing experiment in an alluvial aquifer using cross-borehole time-lapse ERT. Heated water was injected in a well while water of the aquifer was extracted at another well. An ERT section was set up across the main flow direction. The results of ERT were transformed into temperature using calibrated petrophysical relationships. These ERT-derived temperatures were then compared to direct temperature measurements in control piezometers collected with distributed temperature sensing (DTS) and groundwater temperature loggers. Spatially, it enabled to map the horizontal and vertical extent of the heated water plume, as well as the zones where maximum temperatures occurred. Quantitatively, the temperatures and breakthrough curves estimated from ERT were in good agreement with the ones observed directly during the rise and maximum of the curve. An overestimation, likely related to 3D effects, was observed for the tail of the heat breakthrough curve. The error made on temperature can be estimated to be between 10 and 20%, which is a fair value for indirect measurements. From our data, we estimated a quantification threshold for temperature variation of 1.2 °C. These results suggest that ERT should be considered when designing heat tracing experiments or geothermal systems. It could help also to assess the geometrical complexity of the concerned reservoirs. It also appears that ERT could be a useful tool to monitor and control geothermal systems once they are in operation.

1. Introduction

Shallow alluvial aquifers constitute potential shallow geothermal energy reservoirs, relatively abundant and easily accessible. In these low temperature systems, groundwater has an average temperature ranging from 5 to 30 °C and may be used for domestic or industrial cooling and heating (Allen and Milenic, 2003; Haehnlein et al., 2010).

The two main techniques to exploit shallow geothermal energy systems are ground source heat pump (GSHP), which are closed systems with a vertical or horizontal heat exchanger, and groundwater heat pump (GWHP), which are open systems circulating groundwater between production and injection wells. Designing such systems requires a multidisciplinary approach including geological and hydrogeological aspects. The most common approach is to model the system using a coupled groundwater and heat flow simulator. However, such models require estimating parameters governing heat transport such as heat capacity, thermal conductivity and density. Due to a lack of data, authors often have to rely on standard calculation charts, values found in the literature or default values implemented in softwares (e.g. Busby et al., 2009; Lo Russo and Civita, 2009; Liang et al., 2011; de Paly et al., 2012). In situ tests, such as thermal response tests (Raymond et al., 2011; Mattsson et al., 2008), or laboratory measurements (e.g. Haffen et al., 2013) are sometimes possible but the deduced values may deliver only well-centered information or may not always be representative of in situ conditions.

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Thermal tracing experiments are performed for decades in hydrogeology (Anderson, 2005; Saar, 2011). Such experiments are used to improve the characterization of hydrogeological parameters (e.g. hydraulic conductivity or dispersivity), but the same methodologies may be used to study the thermal properties of shallow geothermal systems (e.g. Vandenbohede et al., 2009, 2011; Giambastiani et al., 2012). However, the heterogeneity of geothermal and hydrogeological systems may be too complex to be fully caught by thermal or solute tracer experiments alone (e.g. Brouyère, 2001).

In this context, electrical resistivity tomography (ERT) can bring relevant and spatially distributed information both on the heterogeneity of aquifers and on the temporal behavior of tracers. Indeed, ERT has proven its efficiency to image and/or monitor spatial phenomena (Vereecken et al., 2006) such as salt water intrusions (Nguyen et al., 2009; Hermans et al., 2012c), variations in moisture content (Binley et al., 2002), biodegradation of hydrocarbons (Atekwana et al., 2000), salt tracer experiments (Kemna et al., 2002; Robert et al., 2012) and heat injection experiments (Hermans et al., 2012b). It was also used in the characterization of geological structures, for example in the exploration of geothermal systems, where hydrothermal fluids may generate high contrasts of resistivity (Pérez Flores and Gomez Trevino, 1997; Bruno et al., 2000; Garg et al., 2007; Arango-Galván et al., 2011).

Besides the characterization of shallow geothermal systems themselves, their impact on the groundwater temperatures in the aquifer may be important since their exploitation yields cold and heat plumes (Molson et al., 1992; Palmer et al., 1992; Warner and Algan, 1984) which may influence aquifer properties and groundwater chemistry (e.g. Jesußek et al., 2013) and microbiology (Brielmann et al., 2009). Haehnlein et al. (2010) pointed out that, if laws and rules exist in some countries to limit the temperature difference caused by the use of geothermal systems, the development of anomalies is rarely monitored. With the growth of the demand for renewable energy, we can expect that regulations will become stricter and controls of installations more common. New monitoring technologies will be needed and ERT may play an important role to monitor spatially, i.e. not only in wells, the variations of temperature in the aquifer. For example, the temperature changes observed on operating GWHP systems (e.g. Vanhoudt et al., 2011) are typically in the range of temperature that could be detected by ERT.

ERT aims at imaging the electrical resistivity distribution of the subsurface. Using petrophysical relationships such as Archie's law, one may recover indirect parameters such as saturation, water electrical conductivity or total dissolved solid content. Bulk electrical resistivity also decreases with temperature (e.g. Revil et al., 1998). In most studies, temperature effects are undesirable and may create artifacts in the interpretation, a correction term is applied to remove the influence of temperature variations (Hayley et al., 2007; Sherrod et al., 2012). Few studies used time-lapse ERT to monitor directly temperature changes (Ramirez et al., 1993; LaBrecque et al., 1996b), generally in a context quite different from GWHP or GSHP systems.

Hermans et al. (2012b) monitored with time-lapse surface ERT a heat injection experiment at a relatively small scale (45 m) and at shallow depth (2–4.5 m). Their results show that ERT is a reliable tool to monitor temperature changes and may be a method of choice for the design and the monitoring of geothermal systems. However, the results need to be extended to deeper and more complex, heterogeneous reservoirs, as it will be considered in this paper. ERT-derived temperatures were very close to temperatures modeled using a calibrated coupled groundwater and heat flow and transport model bringing additional constraints on the thermal properties of the aquifer.

For deeper reservoirs, the rapid decrease in resolution and sensitivity of surface ERT becomes a major drawback (Caterina et al., 2013). It is then necessary to consider borehole ERT to improve resolution (Perri et al., 2012). For example, Prevedel et al. (2009) installed deep (600–750 m) borehole electrodes to monitor the migration of CO_2 within a storage reservoir (Bergmann et al., 2012). For cross-hole ERT, the results obtained for a specific study are more easily extendable than for surface ERT because resolution patterns are not depth dependent.

In borehole ERT, electrodes are located under the ground surface, either fixed at the outer-edge of the casing or mounted on cables with the borehole fluid ensuring the electrical contact with the surrounding rock. In the latter case, borehole fluid is generally more conductive than the rock and may influence resistance measurements (Doetsch et al., 2010). Using time-lapse ERT, the relative fluid effect will be almost similar at each time-step and should be insignificant in inversion results (Nimmer et al., 2008).

In this paper, we study the ability of ERT to monitor temperature changes in a heterogeneous aquifer and follow thermal tracing experiments. We pumped water from a gravel aquifer, heated it and reinjected it in a second well, similar to a GWHP system operation.

The paper is organized as follows: first, the field site is described; second, the methodology is presented; then, the results of the ERT monitoring are compared with direct measurements in wells; finally, conclusions are presented.

2. Field site

The study site is located in Hermalle-sous-Argenteau in Belgium near the Belgian-Dutch border (Fig. 1). It lies on the alluvial aquifer of the Meuse River. A pumping well and 8 piezometers were already present on the site since the 1980s and 11 new piezometers were drilled in June 2012 together with an injection well. They were arranged in three different panels crossing the main flow direction between the injection well and the pumping well in order to study the spatial variability during tracing experiments (Pz10–20, Fig. 2).

Borehole logs enabled to divide the deposits in four different units. The first layer consists of loam and clay with a thickness between 1 and 1.5 m. The second layer is composed of gravel in a clayey matrix. The bottom of this layer is found at depth between 2 and 3.2 m. These two first layers have little importance in this study because they are located in the unsaturated zone. The water table lies at approximately 3.2 m depth, with a very small gradient toward the northeast which is the main direction of flow (Fig. 2). The third unit is composed of gravel and pebbles in a sandy matrix. The quantity of sand decreases with depth, whereas the size of the pebbles increases with depth, a vertical variability is thus present. Lateral variability in the grain size distribution of the deposits is also expected in this heterogeneous aquifer, leading to variable hydrogeological parameters. Between 9.7 and 10.1 m, the Carboniferous bedrock composed of folded shales and sandstones is found.

In the middle panel, the outer piezometers are screened on the whole thickness of the alluvial aquifer. This is also the case for the injection and pumping wells. Except for the latter, they were equipped with a distributed temperature sensing (DTS) system to monitor the temperature during the experiment (Leaf et al., 2012 and references therein) with a spatial resolution of 0.5 m. Pz14 and Pz16 were screened at two different levels, with a 2 m long screen between 4 and 6 m depth and 1 m screen between 8.5 and 9.5 m. All other piezometers were screened at two different levels, with a 1 m screen between 4.5 and 5.5 m depth and a 2 m screen between 8 and 10 m depth. In the middle of each screened zone, a groundwater temperature logger was placed to monitor the temperature and the pressure during all the experiment.

Previous studies have shown that the gravel aquifer is very permeable. Calibrated hydraulic conductivity values were found