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# 4D electrical resistivity tomography (ERT) for aquifer thermal energy storage monitoring

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## Abstract

In the context of aquifer thermal energy storage, we conducted a hydrogeophysical experiment emulating the functioning of a groundwater heat pump for heat storage into an aquifer. This experiment allowed the assessment of surface electrical resistivity tomography (ERT) ability to monitor the 3D development over time of the aquifer thermally affected zone. The resistivity images were converted into temperature. The images reliability was evaluated using synthetic tests and the temperature estimates were compared to direct temperature measurements. Results showed the capacity of surface ERT to characterize the thermal plume and to reveal the spatial variability of the aquifer hydraulic properties, not captured from borehole measurements. A simulation of the experiment was also performed using a groundwater flow and heat transport model calibrated with a larger set-up. Comparisons of the simulation with measurements highlighted the presence of smaller heterogeneities that strongly influenced the groundwater flow and heat transport.

32 Keywords: aquifer thermal energy storage, thermally affected zone, monitoring, electrical resistivity  
33 tomography, time-lapse, inversion

## 34 1 Introduction

35 The reduction of fossil fuel consumption is an objective for preserving non-renewable energy  
36 resources and reducing the impact of global warming. In this context, renewable and sustainable  
37 energies are promoted, e.g. Energy Efficiency Directive 2012/27/EU (European Council, 2012).  
38 Smart systems that use heat pumps to transfer heat to or from the ground can take advantage of the  
39 thermal stability of the subsurface to reduce energy consumption (Lo Russo et al., 2009; Vanhoudt et  
40 al., 2011; Sarbu and Sebarchievici, 2014). Beyond the different existing systems, ground source heat  
41 pumps present an inherent thermal resistance with boreholes, while groundwater heat pumps (GWHP)  
42 directly use groundwater that presents a relatively stable temperature over seasons. GWHP provide  
43 space heating or cooling, domestic hot water production, and are even used to store thermal energy,  
44 depending on the season and/or on the specific needs of the infrastructure (Allen and Milenic, 2003;  
45 Lo Russo and Civita, 2009). GWHP function through an open loop between two wells or groups of  
46 wells drilled in a shallow aquifer. During summer and/or any space cooling periods, water is extracted  
47 from the so-called cold well to cool down the infrastructure with the help of heat exchangers (geo-  
48 cooling). The thermal energy in excess, captured by heat exchangers, is then transferred to  
49 groundwater before its re-infiltration into the so-called warm well (Gringarten and Sauty, 1975;  
50 Ausseur et al., 1982; Voigt and Haefner, 1987). In winter, the GWHP functions in a reverse mode:  
51 water is pumped from the warm well and heat is transferred from groundwater to the building.  
52 Groundwater is then re-injected into the cold well with a lower temperature (Ausseur et al., 1982;  
53 Ampofo et al., 2006). Theoretically, the heat stored in the aquifer during space cooling periods allows  
54 an energy reduction for space heating and inversely with the cold stored during space heating periods  
55 for space cooling. Such systems are called aquifer thermal energy storage systems (Sommer et al.,  
56 2013, 2014; Bridger and Allen, 2014; Possemiers et al., 2015). GWHP working in shallow aquifers  
57 requires a relatively small ground surface area by comparison to ground source heat pumps. Thus the  
58 thermally affected zone (TAZ) is limited to a volume around boreholes (Sarbu and Sebarchievici,  
59 2014). However, the induced temperature variations in the aquifer likely modify the medium  
60 properties such as the chemical composition and water quality (Bonte et al., 2013; Jesušek et al.,  
61 2013). Those changes might in turn impact the system efficiency but also the biodiversity, the  
62 microbial activities and consequently the ecosystem functions (Griebler et al., 2016).

63

64 In addition to controlling the temperature impact often imposed by regulations (Haehnlein et al.,  
65 2010), the design of GWHP requires a good understanding of the aquifer and heat flow conditions.  
66 In particular, issues of thermal short-circuit or recycling between cold and warm wells have to be  
67 carefully considered (Banks, 2009; Galgaro and Cultrera, 2013; Milnes and Perrochet, 2013). The  
68 propagation of the heat and cold plumes in the aquifer is also highly sensitive to possible variations  
69 of hydraulic gradient that could be induced by existing water wells and/or the drilling of additional  
70 wells close to the pumping system (Lo Russo et al., 2014) and to the heterogeneity of the subsurface  
71 (Bridger and Allen, 2014; Sommer et al., 2013, 2014; Possemiers et al., 2015; Hermans et al., 2018).  
72 Methods supplying insights on the heat or cold plume's propagation in the aquifer have to be  
73 developed, notably for delimiting the TAZ, to better anticipate the possible difficulties arising from  
74 GWHP implementation. Models of groundwater flow and heat transport are often calibrated with  
75 empirical values or from local measurements in boreholes (Lo Russo and Civita, 2009; Liang et al.,  
76 2011; de Paly et al., 2012; Mattsson et al., 2008; Raymond et al., 2011), ignoring the heterogeneity  
77 of the hydrogeological medium. Monitoring the 4D evolution of the TAZ through time is then  
78 particularly relevant. The relative proximity of the TAZ to the surface enables a monitoring with non-  
79 invasive geophysical methods (Hermans et al., 2014).

80

81 Electrical resistivity tomography (ERT) is particularly sensitive to the porous medium temperature  
82 (Rein et al., 2004; Revil et al., 1998; Hayley et al. 2007). Moreover, ERT applied in time-lapse (TL)  
83 provides spatially distributed information on the changes over time of the porous medium and may  
84 target salinity, water content or temperature (for a review on TL ERT see Singha et al., 2015). Thus,  
85 TL ERT is specifically appropriate to monitor heat plume development (Hermans et al., 2014).  
86 Acquisition systems can work autonomously, allowing the repeated measurements required to  
87 achieve sufficient temporal resolution to follow the 3D TAZ development. The method is also  
88 minimally invasive and requires low implementation costs compared to a dense network of boreholes.  
89 So far, the ability of ERT to monitor heat plumes has been demonstrated in 3D in a laboratory  
90 experiment at a scale of a few tens of centimeters (Giordano et al., 2016). Field 2D set-ups confirmed  
91 its relevance for monitoring heat storage, tracing experiments and borehole heat exchanger either  
92 from surface and/or cross-boreholes measurements (Hermans et al., 2012; Hermans et al., 2015;  
93 Giordano et al., 2017; Cultrera et al., 2017). However, 2D interpretation can be limited by out-of-the-  
94 plane or shadow effects during inversion which can limit the quantitative assessment of temperature  
95 (Nimmer et al., 2008). This effect is probably partly responsible for discrepancies observed between  
96 ERT-derived temperatures and direct cross-boreholes measurements by Hermans et al. (2015). The

97 use of 3D surveys and subsequent inversion can largely improve imaging of complex 3D subsurface  
98 objects (e.g., Van Hoorde et al., 2017).

99

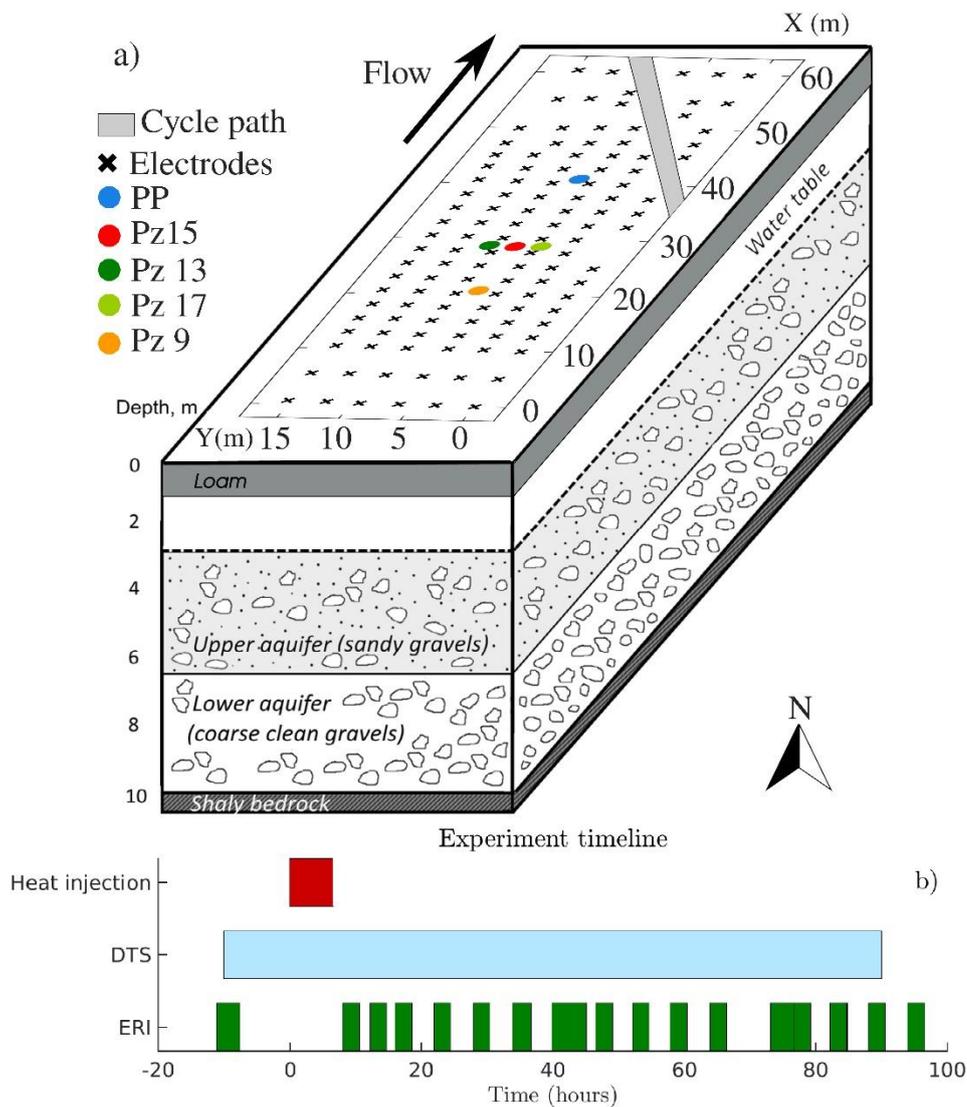
100 In this paper, we image the development of the TAZ during a heat injection and storage experiment  
101 using a 3D ERT survey. The accuracy of ERT-derived temperature estimates is explored through  
102 synthetic cases that emphasize the method's sensitivity to target depth and thickness. Results show  
103 the ability of surface TL ERT to monitor the 3D development of the TAZ in a shallow aquifer. We  
104 compare our results with direct temperature measurements which demonstrates the ability of ERT to  
105 supply complementary insights about the sub-surface spatio-temporal dynamic. The ERT-derived  
106 temperatures show a general agreement with direct observations, **although important discrepancies**  
107 **are observed in the amplitude of the measured variations.** The latter are mainly explained by the  
108 different representative volumes of the two techniques **and the limitations related to the**  
109 **regularized ERT inversion.** A groundwater flow and heat transport model calibrated during a  
110 previous experimental set-up at a larger scale, is also computed. Its comparison with direct and ERT-  
111 derived temperature measurements underline the presence of local heterogeneities at the vicinity of  
112 the injection well which should be incorporated in the flow and transport model.

## 113 2 Field experiment

### 114 2.1 Hydrogeological context

115 The study site is located on the alluvial plain of the Meuse River at Hermalle-sous-Argenteau, 13 km  
116 north-east from Liège, Belgium. From borehole logs analysis, the subsurface medium can be divided  
117 into four lithological units. The first layer is composed of loam and presents a thickness of about 1  
118 m. Below, the second unit is constituted of sandy loam, gravels and clay to a depth of 3 m. Between  
119 3 and 10 m depth, the third layer is composed of gravel and pebbles in a sandy matrix. This third layer  
120 hosts the alluvial aquifer. It can be divided in two main units: the upper aquifer, between 3 and 6 m  
121 depth, composed of sandy gravels and the lower aquifer, between 6 and 10 m depth, characterized by  
122 coarser and cleaner gravels. The water table lies approximately at a depth of 3.2 m. Below 10 m depth,  
123 the basement of the aquifer consists of low permeability carboniferous shale and sandstone (Fig. 1).

124



125 Figure 1: a) Scheme of the experimental set-up and structure of the underground medium (modified  
 126 from Klepikova et al., 2016). b) The experiment timeline shows the duration of the heat injection and  
 127 the DTS measurements as well as the moment of ERT acquisitions.

128

129 The site topography is almost flat and the natural hydraulic gradient in the aquifer is approximately  
 130 0.06% with a north-east direction (Brouyère, 2001). Previous experiments showed that the aquifer is  
 131 characterized by a high average permeability and by a horizontal and vertical heterogeneity  
 132 (Dassargues, 1997; Derouane and Dassargues, 1998; Brouyère, 2001). In particular, the upper and  
 133 the lower parts of the aquifer, ranging respectively in [3-6] m depth and [6-10] m depth present a  
 134 respective effective porosity of 4% and 8% as estimated from tracer test experiments (Brouyère,  
 135 2001). Direct measurements of Darcy fluxes were performed to estimate the groundwater flow  
 136 variations with depth. The lower part of the aquifer present Darcy fluxes in the range of  $[1-8] \times 10^{-3}$   
 137 m/s, about one order of magnitude higher than in the upper layer where values range in  $[1-10] \times 10^{-4}$   
 138 m/s (Wildemeersch et al., 2014). In a previous multiple tracer tests experiment, the monitoring of the

139 3D spreading of a heat plume resulting from the injection of heated water in the aquifer also showed  
140 lateral variations of the medium hydraulic properties (Klepikova et al., 2016), strengthened by 2D  
141 cross-borehole ERT and DTS measurements (Hermans et al., 2015). They confirmed the higher flow  
142 in the lower part of the aquifer and the lateral heterogeneity of the aquifer that presents zones of  
143 preferential flow (Hermans et al., 2015). Similarly, the thermal properties of the aquifer vary with  
144 depth as assessed during the ThermoMap project (Bertermann et al., 2013). The thermal conductivity  
145 was estimated at 1.37 and 1.86 W/mK and the volumetric heat capacity at 2.22 and 2.34 MJ/m<sup>3</sup>K in  
146 the layers located respectively at depths between 3 to 6 m and 6 to 10 m.

## 147 2.2 Heating and injection procedure

148 Water was pumped from the aquifer at the pumping well PP (Fig. 1a), its initial temperature was  
149 13.4°C and the pumping flow rate was fixed to 3 m<sup>3</sup>/h. The water was then heated using a mobile  
150 water flow heater (Swingtec AQUAMOBIL DH6 system) before being re-injected with a flow rate  
151 of 3 m<sup>3</sup>/h. The heated water was injected in the borehole Pz 15 between 4.5 and 5.5 m depth (Fig. 1).  
152 Pz 15 is located in between two boreholes (Pz 13 and Pz 17) screened on the whole aquifer depth  
153 allowing DTS measurements (Fig. 1). The water was heated to a temperature of 42°C, i.e. with an  
154 increase of 28.6°C from the initial aquifer temperature. The hot water injection lasted 6 hours, but at  
155 the end of the injection step, water was injected at a temperature of 14.5°C during 20 minutes due to  
156 a technical issue with the water heater. Afterward, a heat storage phase lasted 4 days (Fig. 1b). The  
157 injected hot water volume is about 18m<sup>3</sup> that can be used to estimate the order of magnitude of the  
158 heat plume development. The thermal plume can be assumed to develop in a 3 m height cylinder in  
159 the upper part of the aquifer that presents a porosity of 4%. Thus, the thermally affected zone should  
160 present a maximum volume of 450 m<sup>3</sup> (ignoring conduction effects), developing in a cylinder of about  
161 14 m diameter.

## 162 2.3 3D electrical resistivity data acquisition

163 Electrical resistivity measurements were performed from a grid of electrodes located at the surface to  
164 monitor the 3D heat plume evolution in the aquifer. 126 electrodes were placed along 6 profiles of  
165 21 electrodes centered on the injection well and parallel to the natural flow direction (Fig. 1a). Along  
166 each profile the electrodes spacing was 2.5 m for the 17 central electrodes. Two electrodes at either  
167 end of each profile were spaced 5 m apart. That arrangement of electrodes was selected to allow a  
168 finer resolution around the injection well together with a greater penetration depth than profiles with  
169 equidistant electrodes. Some electrodes located on the northern side of the profiles 1 to 3 could not  
170 be hammered into the ground since the field is crossed by a concrete bike path. Thus, 1 or 2 electrodes  
171 were missing on those profiles (Fig. 1a). The electrode profiles were separated by 3 m so the electrode

172 grid covers the whole space available to image the plume without degrading the resolution  
173 perpendicular to the profiles. Data were acquired with gradient and dipole-dipole protocols along  
174 each profile (2D data acquisition). Cross-line measurements were also acquired (Van Hoorde et al.,  
175 2017), but ignored during inversion due to their low quality. We used an ABEM Terrameter LS  
176 connected to a relay switch ES10-64 allowing the acquisition in one shot of the whole data set for the  
177 entire electrode grid. A first data acquisition was performed the day before the heated water injection  
178 for reconstructing the background image. During the heat storage period 16 acquisitions were  
179 performed; once every six hours (Fig. 1b). For the background acquisition, reciprocal measurements  
180 were acquired for the whole data set to estimate the measurement error (LaBrecque et al., 1996).  
181 Reciprocal measurements correspond to a swap of the electrodes used for current injection and  
182 voltage measurements during the “normal” acquisition. A complete normal data set consisted in 3045  
183 voltage measurements. For the time-lapse acquisition, the amount of reciprocal data was reduced to  
184 1119 to speed up the acquisition. The acquisition delay was set to 0.3 s and the acquisition time was  
185 0.5 s. The acquisition duration was 2.5 and 1.5 hours for the background and the time-lapse  
186 acquisitions, respectively.

## 187 2.4 Borehole measurements

188 Single-ended optical fibers were inserted in the Pz13 and Pz17 boreholes located on both side of the  
189 injection borehole (Fig. 1a). The optical fibers allow a direct monitoring of the temperature variations  
190 in the aquifer during the experiment, performed with an AP Sensing Linear Pro Series N4386. The  
191 distributed temperature sensing (DTS) measurements provide a spatial sampling of 0.2 m and a  
192 temporal resolution of 2 min. Since the aquifer vertical extension is relatively small (7 m) we choose  
193 to preserve the spatial resolution provided by the spatial sampling provided by the DTS. Instead, we  
194 preferred reducing the temporal resolution so we applied a running average through time over a  
195 20 min window. Two sections of the optical fibers were placed respectively in a chilled and a warm  
196 water bath. The cold bath was maintained near 0°C (about 13°C cooler than the aquifer) by regularly  
197 adding ice. The baths’ temperature was monitored with Pt100 sensors. The cable ends were placed at  
198 the bottom of the boreholes due to the small diameter of boreholes compared to the critical bend  
199 radius of optical fibers. Thus the differential attenuation of the light along the single-ended cables  
200 could not be estimated (Hausner et al., 2011). Nevertheless, the calibration baths allowed to correct  
201 the estimated temperatures, guaranteeing their temporal consistency throughout the experiment. DTS  
202 measurements were used for checking the relative temperature change  $\Delta T$  from the initial state and  
203 not for an absolute monitoring of the aquifer temperature. The air temperature increased by 15°C at  
204 the end of the experiment. However, this effect is attenuated with depth and, although we can observe

205 a difference at a depth of 1 m in the DTS measurements, no temperature variation below 3 m  
206 (saturated zone) is visible.

## 207 3 Groundwater flow and heat transport model

### 208 3.1 Settings of the hydrogeological model

209 We used the 3D groundwater flow and heat transport model HydroGeoSphere (HGS) (Therrien et al.,  
210 2010) developed by Klepikova et al. (2016) for the Hermalle-sous-Argenteau experimental site to  
211 simulate numerically our experiment and compare temperatures derived from ERT and measured  
212 directly with DTS to the simulation. This deterministic model has been constructed and calibrated  
213 based on historical data (Dassargues, 1997; Derouane and Dassargues, 1998; Brouyère, 2001) and a  
214 multiple tracer experiment, including heat tracer (Wildemeersch et al., 2014; Hermans et al., 2015).

215  
216 The model geometry corresponds to the one described by Klepikova et al. (2016), except that we  
217 refined the grid around the injection well to accurately model the experiment. In several aspects their  
218 experiment differs from the one presented here. They injected the heat tracer at the base of the Pz 9  
219 borehole (Fig. 1a), so in the lower unit of the aquifer that is hydraulically more conductive.  
220 Furthermore, Pz 9 is screened over the whole thickness of the aquifer (Table 1) allowing heat  
221 propagation upward along the borehole. Their experimental design was adapted to constrain the  
222 hydraulic conductivity values on the whole aquifer section. In our experiment the temperature  
223 changes occurred in the upper part of the aquifer where heat was injected. Therefore, our data were  
224 more sensitive to the spatial variability of hydraulic conductivity distribution in the upper area.  
225 Moreover, they injected the heated water at a rate of 3 m<sup>3</sup>/h during 24 hours and 20 minutes, while  
226 pumping at a constant discharge rate of 30 m<sup>3</sup>/h in the pumping well (Fig. 1) in order to speed up the  
227 heat propagation in the aquifer. Therefore their experiment supplied information on the hydraulic  
228 properties in the whole domain in between the injection borehole Pz 9 and the pumping well (Fig1).  
229 In our case, we injected and stored the heat in the Pz 15 borehole so our experiment is more sensitive  
230 to hydraulic properties distribution in a smaller region surrounding that borehole.

231

232 Table 1: Main differences between the experiment of Klepikova et al. (2016) and the one presented  
233 in this study.

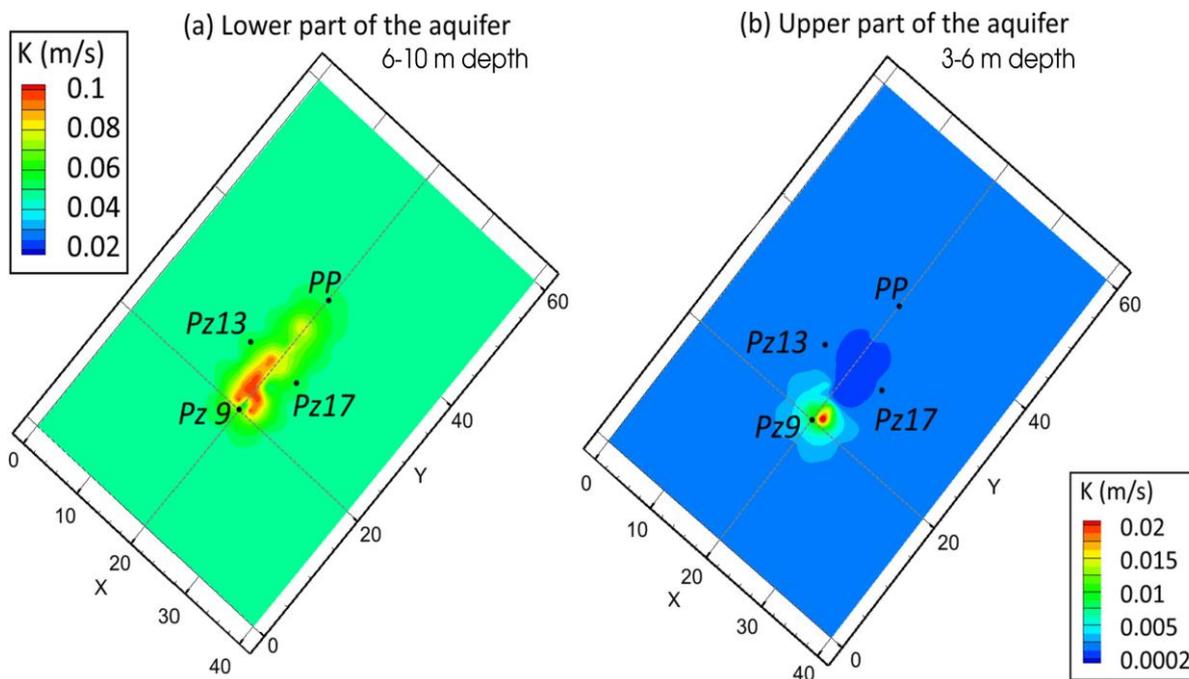
	This study	Klepikova et al., 2016
Depth of injection	5 m	9 m
Duration of injection	6 hours	24 hours and 20 minutes

Pumping time	Only during injection	All along the experiment
Pumping rate	3 m <sup>3</sup> /h	30 m <sup>3</sup> /h
Injection rate	3 m <sup>3</sup> /h	3 m <sup>3</sup> /h
Injection well	Pz 15	Pz 9
Injection well screen interval	[-5.5 ; 4.5] m	[-8.7 ; 3.2] m

234

235 The inversion process ran by Klepikova et al. (2016) sought the hydraulic conductivity distribution  
 236 of the Hermalle site aquifer, in between the Pz 9 and the pumping well PP from a monitoring of the  
 237 temperature changes in the upper and lower part of the aquifer from 11 observation boreholes. They  
 238 estimated the hydraulic property values with the pilot point method to parametrize the inversion  
 239 (Fig. 2). Here we use their resulting model to simulate the heat transport in the aquifer with the  
 240 characteristics of our experiment. The model includes the density effects due to temperature changes  
 241 (Graf and Therrien, 2005).

242

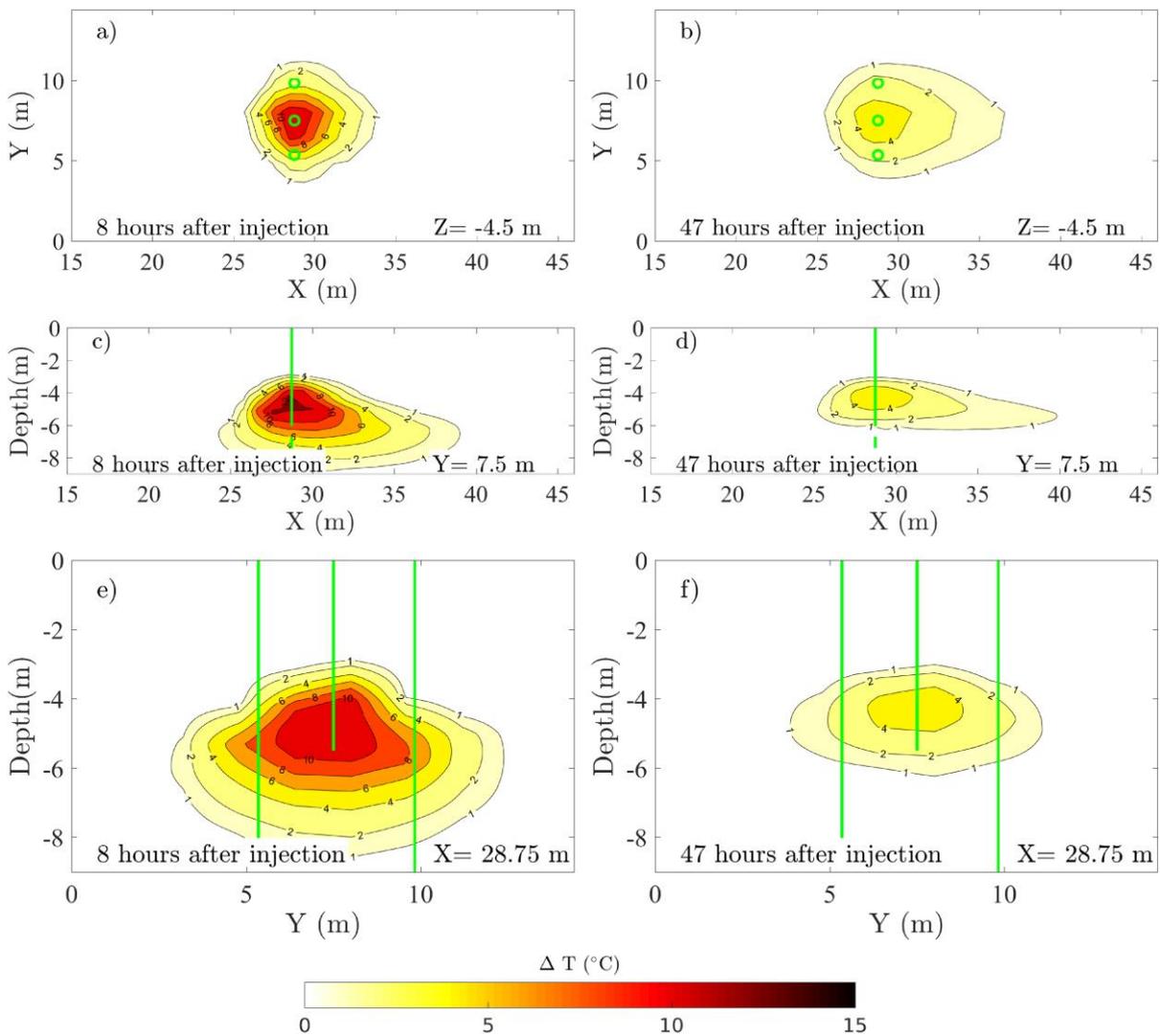


243 Figure 2: Spatial variability of the hydraulic conductivity  $K$  (m/s) models in XY planes for the lower  
 244 (a) and upper (b) parts of the aquifer obtained from the inversion of transient temperature responses  
 245 (Klepikova et al., 2016).

### 246 3.2 Temperature estimates from the HGS model

247 The temperature variations from the HGS model show a TAZ with a half-sphere geometry in the  
 248 upper part of the aquifer (Fig. 3). In the lower part of the aquifer, the TAZ presents a tail along the X  
 249 axis (towards NE which is the main flow direction; Fig. 3 c, d). The higher hydraulic conductivity

250 and the greater porosity in the lower part of the aquifer (Fig. 2 a) favors heat propagation with  
 251 groundwater flow through convection. 8 h after the beginning of the injection, the diameter of the  
 252 TAZ showing a  $\Delta T$  of  $4^{\circ}\text{C}$  at a depth of 4.5 m is about 5 m and is slightly more elongated in the X  
 253 direction. The lower diameter of the model TAZ compared to the above estimate for a cylinder of  
 254 14 m diameter can be explained by the fact that the model predicts a deeper propagation of the TAZ.  
 255 Moreover, in the lower part of the aquifer the porosity is set higher in the model than the one used for  
 256 the cylinder volume evaluation. 47 h after the injection started, the dimension of the model TAZ  
 257 showing a  $\Delta T$  of  $4^{\circ}\text{C}$  is slightly reduced to a diameter of about 3 m at a depth of 4.5 m. However at  
 258 that time,  $\Delta T$  in the middle of the TAZ is significantly lower, e.g. from  $12^{\circ}\text{C}$  after 8 h (Fig. 3 c) to  
 259  $4^{\circ}\text{C}$  for the same position after 47h (Fig. 3 d).  
 260



261 Figure 3: Temperature variation from the initial state estimated by the HydroGeoSphere model, 8 h  
 262 and 47 h after the injection started on cross-sections in a horizontal layer at -4.5 m (a, b), on cross-  
 263 section at the level of the injection borehole along the electrode profiles (c, d) and perpendicular to

264 the electrode profiles (e, f). The green circles and lines represent the injection and measurement  
 265 boreholes. The black dotted lines and crosses correspond to the electrodes.

## 266 4 ERT-derived temperature images reconstruction process

### 267 4.1 Inversion method

268 The image reconstruction process in ERT aims to determine the spatial distribution of bulk resistivity  
 269 in the medium that best reproduces the resistance measurements. An inverse problem is then solved  
 270 to iteratively minimize the objective function  $\Psi(m)$ :

$$271 \quad \Psi(m) = \Psi_d(d, m) + \lambda \Psi_m(m),$$

272 Equation 1

273 where  $\Psi_d(d, m)$  measures the data misfit,  $\Psi_m(m)$  defines model constraints for the inversion  
 274 regularization and  $\lambda$  is the damping factor that balances the weight of the regularization. We used the  
 275 EIDORS software for solving the inversion of the ERT data (Polydorides and Lionheart, 2002; Adler  
 276 et al., 2015).

277  
 278 The data misfit  $\Psi_d(d, m)$  is evaluated between observed measurements  $d$  (electrical resistances) and  
 279 calculated data  $f(m)$  computed from a model of resistivity  $m$ . Model and data are both expressed  
 280 using their base 10 logarithm.

281  
 282 For the reconstruction of the medium background, before the introduction of perturbation through  
 283 heat injection, the absolute values of the medium resistivity are sought and the data misfit is expressed  
 284 as:

$$285 \quad \Psi_d(d, m) = \sum_{i=1}^N \frac{|d_i - f_i(m)|^2}{|\epsilon_{B,i}|^2},$$

286 Equation 2

287 where  $\epsilon_B$  are weighting factors accounting for the uncertainty of the measurements.

288 For time-lapse inversions, the inversion seeks the variations of resistivity from the background model.  
 289 The measure of the data misfit is here defined using the difference between the background data  $d_{0i}$   
 290 (the subscript 0 refers to the background data) and the monitored data  $d_i$ :

$$291 \quad \Psi_d(d, m) = \sum_{i=1}^N \frac{|(d_i - d_{0i}) - (f_i(m) - f_i(m_0))|^2}{|\epsilon_{TL,i}|^2}$$

Equation 3

294  
 295 where  $m_0$  stands for the model parameter distribution of the background state and  $\epsilon_{TL}$  corresponds  
 296 to the time-lapse data weighting.

297  
 298 The insertion of prior information on the medium structure is performed by  $\Psi_m(m)$  in the inversion  
 299 to stabilize the inverse problem:

$$\Psi_m = ||W_m(m - m_0)||^2,$$

Equation 4

300  
 301  
 302 where  $W_m$  represents the regularization matrix that is here the identity matrix (Tikhonov, 1963). The  
 303 background image  $m_0$  was the result of a first inversion seeking the resistivity values of a medium  
 304 constituted by four horizontal layers as highlighted by borehole logs analysis and previous ERT  
 305 experiments (Hermans & Irving, 2017). The four layers correspond to a superficial unit above 1 m  
 306 depth, below the unsaturated zone extends to 3 m depth, then to a depth of 10 m lies the aquifer over  
 307 the bedrock. The depth of the interfaces are taken from borehole logs analysis. In the time-lapse  
 308 reconstruction scheme,  $m_0$  corresponds to the background image that refers to the initial state of the  
 309 medium.

310  
 311 The computation of  $f(m)$  is performed using a 3D finite element model built using the Netgen  
 312 software (Schöberl, 1997). The unstructured mesh is refined close to the electrodes for a better  
 313 accuracy of the resistance estimation, while elements are coarser further away from the electrodes  
 314 allowing a reduced computation time. A regular coarse mesh was designed for the inversion in order  
 315 to reduce the number of model parameters and thus stabilize the inversion procedure.

316  
 317 The inversion is performed using an iterative Gauss-Newton scheme and at each step a linear search  
 318 against data misfit seeks the optimum value of the regularization parameter  $\lambda$  that minimizes the  
 319 weighted residuals  $\chi$ :

$$\chi = \sqrt{\frac{1}{N} \Psi_d(d, m)}.$$

Equation 5

320  
 321  
 322 The inversion is stopped either when the inversion converges, that is when further iteration do not  
 323 provide a better data fit  $\chi$ , i.e. no reduction of the objective function  $\Psi(m)$  is observed, or when the  
 324 desired misfit  $\chi = |1 - \xi|$  is reached, with  $\xi = 10^{-2}$ . Such a stopping criteria avoids over-fitting data

325 and the presence of artefacts in the resulting image (Kemna, 2000). The data analysis and the method  
326 used to determine the error models weighting the data misfit are described in the appendix A.1.

## 327 4.2 ERT conversion to temperature images

328 Temperature variations in the aquifer induce a change of the medium electrical properties, i.e. the  
329 resistivity and its inverse the conductivity. The conductivity variations from the initial state and the  
330 temperature changes are related by a linear petrophysical relationship so temperature images can be  
331 derived from ERT (Hermans et al., 2014). We consider that bulk conductivity variations in the aquifer  
332 are due to changes of the fluid conductivity that depends on fluid salinity and temperature variations.  
333 We checked that chemical perturbations produced by temperature changes have a negligible effect  
334 on the fluid salinity (Hermans et al., 2015). Thus, the conductivity increase can be quantitatively  
335 interpreted in terms of temperature change using the linear relationship:

$$336 \quad \frac{\sigma_{f,T}}{\sigma_{f,25}} = m_f(T - 25) + 1$$

337 Equation 6

338 where  $\sigma_{f,T}$  stands for the fluid conductivity at a temperature  $T$  and  $m_f$  corresponds to the fractional  
339 change of the fluid conductivity per degree Celsius around the reference temperature  $T = 25^\circ\text{C}$ . From  
340 water samples taken on site the trend between the temperature and the fluid conductivity was  
341 estimated to be  $m_f = 0.0194$  and the fluid conductivity at  $T = 25^\circ\text{C}$  was evaluated at  $\sigma_{f,25} =$   
342  $0.0791 \text{ S/m}$  (Hermans et al., 2015).

343  
344 Temperature change from the initial state  $\Delta T$  images can be constructed from the observed variations  
345 of the bulk conductivity  $\sigma_b$  by converting them with (Hermans et al., 2014; 2015):

$$346 \quad \Delta T = \frac{1}{m_f} \left( \frac{\sigma_{b,TL} \sigma_{f,B}}{\sigma_{b,B} \sigma_{f,25}} - 1 \right) + 25 - T_{init}$$

347 Equation 7

348  $\sigma_{b,B}$  and  $\sigma_{b,TL}$  correspond respectively to the bulk conductivity of the background state and of a time-  
349 lapse acquisition.  $\sigma_{f,B} = 0.0614 \text{ S/m}$  represents the fluid conductivity at the initial state and was  
350 estimated from Eq. (6) with an initial temperature of  $T_{init} = 13.44^\circ\text{C}$ . That initial temperature value  
351 corresponds to the average of the temperatures measured along both boreholes using DTS before the  
352 heat injection. The estimate of the initial conductivity of the fluid was validated with a value of  $\sigma_{f,B} =$   
353  $0.0598 \text{ S/m}$  by direct measurement with a CTD probe in the injection borehole.

### 354 4.3 Sensitivity analysis of the ERT-derived temperature images

355 We ran synthetic simulations computed from models of a cylindrical plume with different thicknesses  
356 and positions in the aquifer in order to evaluate the sensitivity of the electrode array to the target depth  
357 and thickness. The temperature increase in the cylinder was fixed to 17°C and the cylinder diameter  
358 to 8 m based on the outcomes of the HGS model. The cylinder was centered on the injection borehole  
359 Pz 15 (Fig. 1). The tested cylinder heights varied between 1.5 m and 5 m. The cylinder upper limit  
360 positions checked was of -3 m and -4 m so it is always located below the unsaturated zone (Fig. 4).  
361 On all images, we observe that the inversion allows a fairly correct reconstruction of the target shape  
362 when the target is located close enough to the electrodes (approximately the distance corresponding  
363 to the electrode spacing) or sufficiently thick to be detectable (at least the electrode spacing). We note  
364 that the target upper limit is always overestimated by about 2 m and that the amplitude of  $\Delta T$   
365 decreases with depth. The region of higher  $\Delta T$  in the reconstructed target is also shifted upward the  
366 center of the actual target. Moreover, the temperature increase leaks out of the synthetic target  
367 delimitation, except in the shallow region where artifacts induce a reduction of the electrical  
368 conductivity (Fig. 4). Quantitatively we can compare the estimated temperature increase at the level  
369 of the injection borehole Pz15 to the synthetic target as summarized in Table 2. We note that when  
370 the target is relatively shallow, that is for an upper limit at -3 m, the increase of the target thickness  
371 greatly helps determining more closely the target true temperature. However if the target is only 1 m  
372 deeper, the target thickness increase does not allow accurate estimates of the target temperature.

373

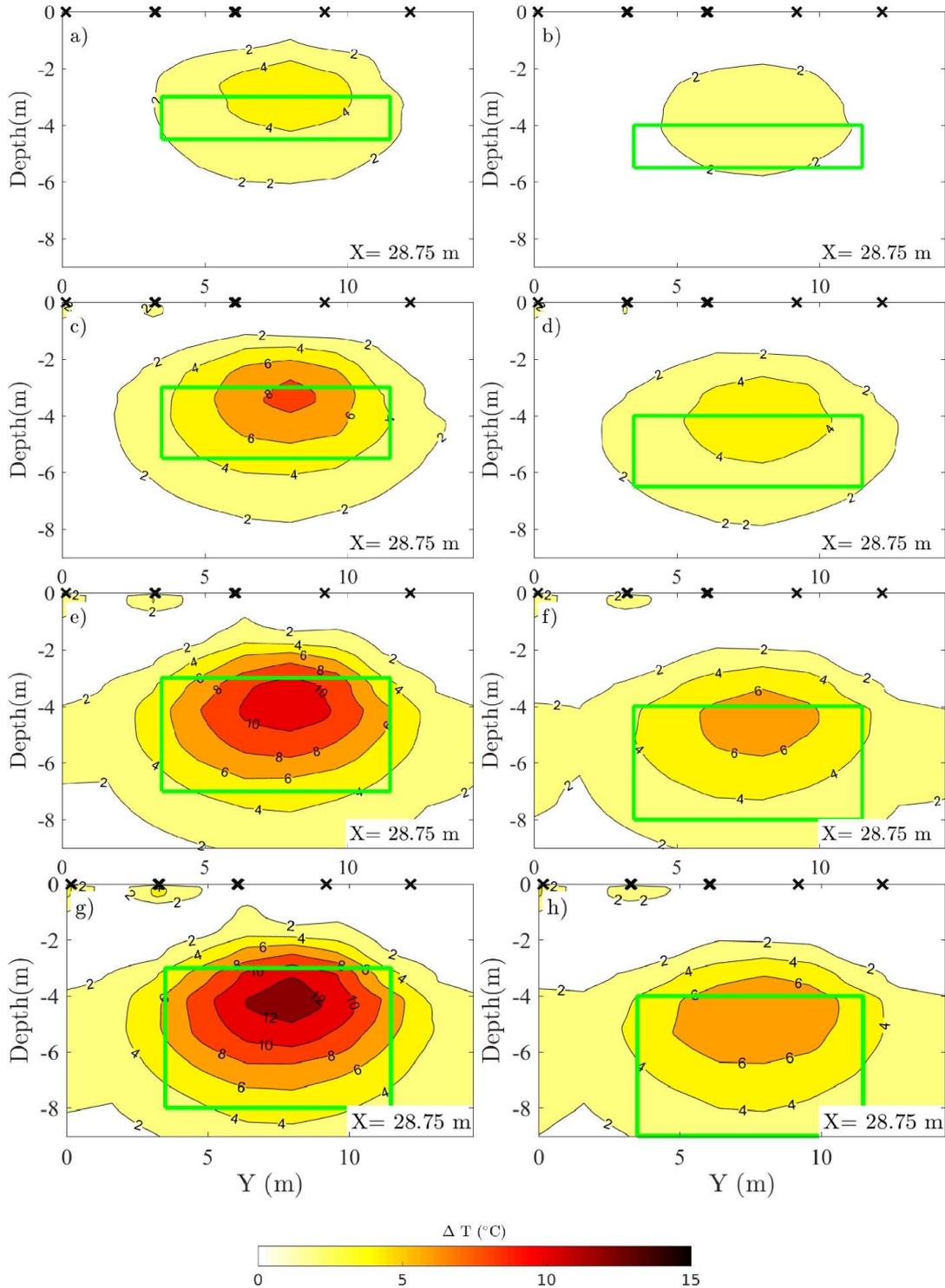
374 Table 2: Percentage of the temperature increase estimated by ERT from the synthetic target increase  
375 for different cylinder height and upper position.

Height \ Upper limit	1.5 m	2.5 m	4 m	5 m
-3 m	24 %	44 %	67 %	78 %
-4 m	17 %	30 %	41 %	46 %

376

377 The target blur and the global temperature underestimation are related to the regularization that favors  
378 a smooth change from the background model. The smoothing effect hinders a precise location of the  
379 target and thus an overestimation of the target depth. Moreover, the estimation of  $\Delta T$  at depth is  
380 degraded by the reduction of the ERT sensitivity with depth (Fig. 4). From those tests, we infer that  
381 the geometry of the plume (its size and depth) is fairly estimated, with an overestimation of the plume  
382 vertical position when the target depth and thickness correspond to the electrode inter-distances.  
383 Furthermore, images underestimate the actual temperature changes in the aquifer. Smoothing effects

384 and the underestimation of the aquifer parameters using TL-ERT are well known and results from the  
 385 regularization procedure in the inversion (Singha and Moysey, 2006).  
 386

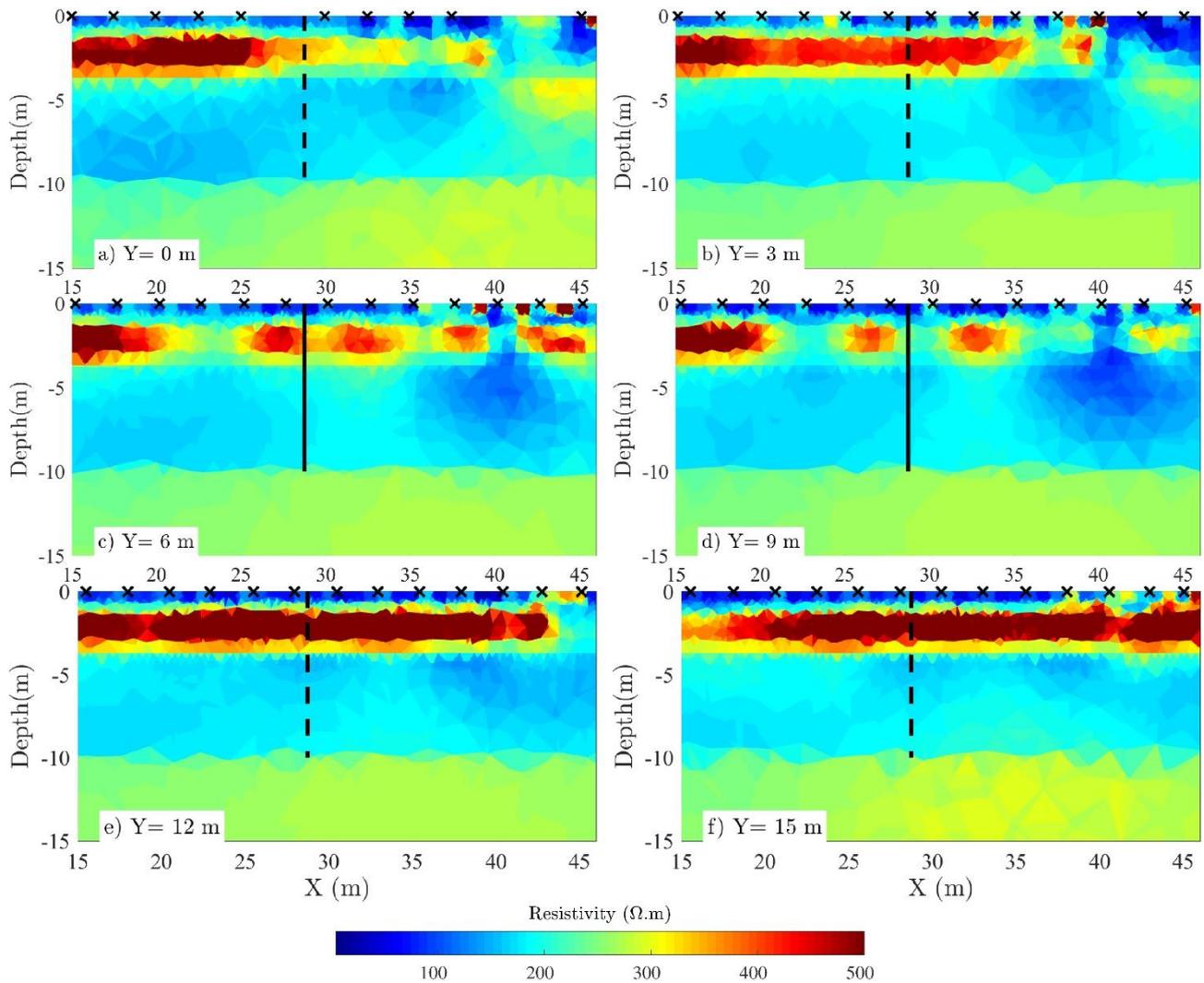


387 Figure 4: Temperature variation estimates from synthetics built as a vertical cylindrical heat plume.  
 388 The cylinder thickness is of 1.5 (a, b), 2.5 (c, d), 4 (e, f) and 5 m (g, h) and its upper limit is at -3 m  
 389 (a, c, e, g) and -4 m (b, d, f, h). The green lines indicate the shape of the cylindrical target, the black  
 390 crosses correspond to the electrodes.

## 391 5 Results from field data

### 392 5.1 Background images

393 The 3D inversion of the background ERT data set provides the distribution of the subsurface  
394 resistivity with a root mean square of the data misfit of 1.03%. The results are presented by 2D vertical  
395 cross-sections corresponding to each acquisition profile (Fig. 5). The images show the vertical  
396 variations of the electrical resistivity corresponding to the four lithological units. The superficial  
397 conductive layer with a thickness of 1 m and a resistivity of about 115  $\Omega$ .m corresponds to the loam  
398 unit. Between 1 and 3 m deep the medium presents heterogeneous variations of resistivity with zones  
399 of a few meters length showing a resistivity higher than 400  $\Omega$ .m and locally values reaching  
400 1000  $\Omega$ .m while the surrounding medium presents a resistivity of 250  $\Omega$ .m. Here the resistive regions  
401 might be interpreted by lenses of clean unsaturated gravels in a medium of sandy loam and gravels.  
402 Below, in the aquifer layer at a depth between 3 and 10 m the medium presents a median resistivity  
403 of 180  $\Omega$ .m with some more conductive areas notably below profiles located on Y= 6 and 9 m where  
404 the resistivity is around 100  $\Omega$ .m. Those conductive anomalies are likely an effect of the pumping  
405 well metallic-casing. We note that we do not distinguish any clear resistivity variations related to the  
406 different nature of the aquifer upper and lower units. Previous ERT experiments discriminated  
407 between the upper and lower regions through a higher resolution cross-borehole survey (Hermans et  
408 al., 2015), or through a specific analysis developed for discriminating the different facies of the  
409 medium (Hermans and Irving, 2017), which were not used in this study. Finally, the lower layer  
410 located below 10 m depth in the bedrock shows a median resistivity of 280  $\Omega$ .m. Those values are in  
411 accordance with previous ERT investigations on the site (Hermans and Irving, 2017).

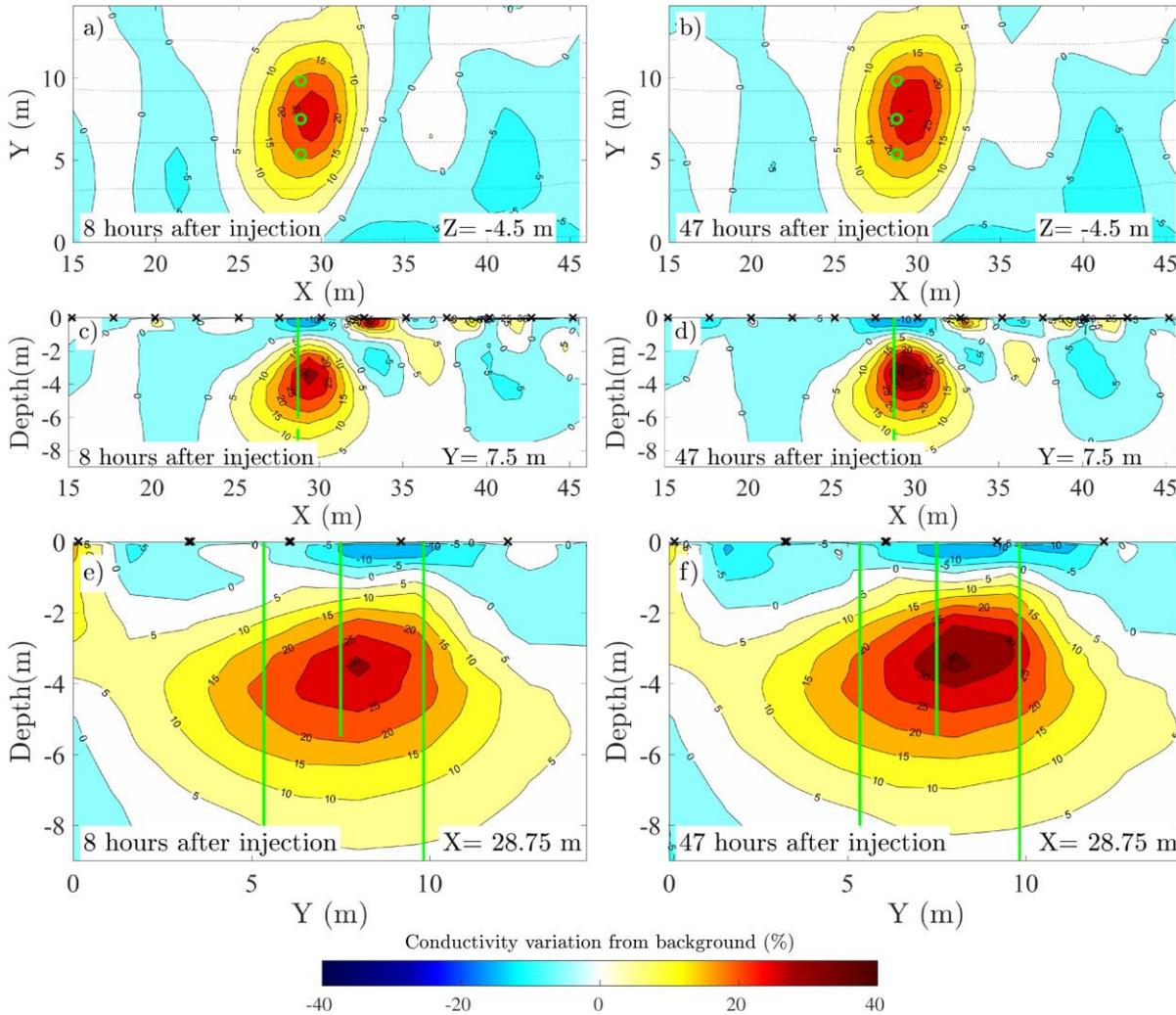


412 Figure 5: Electrical resistivity cross-sections extracted below each electrode profile from the 3D  
 413 model. The black hashed or solid lines represent the injection and measurement boreholes projection.  
 414 The black crosses correspond to the electrodes.

415  
 416 **5.2 Time-lapse conductivity variation images**

417 Since conductivity variations vary linearly with temperature we choose to represent that property  
 418 instead of resistivity variations over selected time steps (8 h and 47 h after the injection started) on  
 419 sections of the 3D model (Fig. 6). A conductive plume, with a variation between 10 and 30% from  
 420 the initial state, is observed in the upper part of the aquifer (Fig. 6c to f). The region showing a  
 421 conductivity increase higher than 10% presents an oval geometry with a horizontal extent of about  
 422 5 m along the X axis and 10 m in the Y direction (Fig. 6a and b). The plume is not centered on the  
 423 injection borehole, but is slightly shifted along the X axis, in the same direction as the natural  
 424 groundwater flow. The region with the highest conductivity change is also slightly shifted in the Y  
 425 direction. The global shape of the zone affected by conductivity changes does not significantly evolve  
 426 with time, but the conductivity further increases more than 50 hours after the end of the injection. A

427 second conductive target is observed at  $Y=0$  m, below the first profile which is lacking central  
 428 electrodes. We interpret this more conductive zone as an artifact related to a lower sensitivity in that  
 429 region due to the missing electrodes (Fig. 6). The images also present regions with a conductivity  
 430 decrease from the initial state, mostly in the shallow region but also at depth below profile 1 and at  
 431 the extremity of the central profiles. In the shallow region, they might be related to changes in the  
 432 medium saturation above the aquifer. Below, we interpret them as artifacts that are strengthened by  
 433 our less conservative choice of error model (see Appendix A.1).  
 434

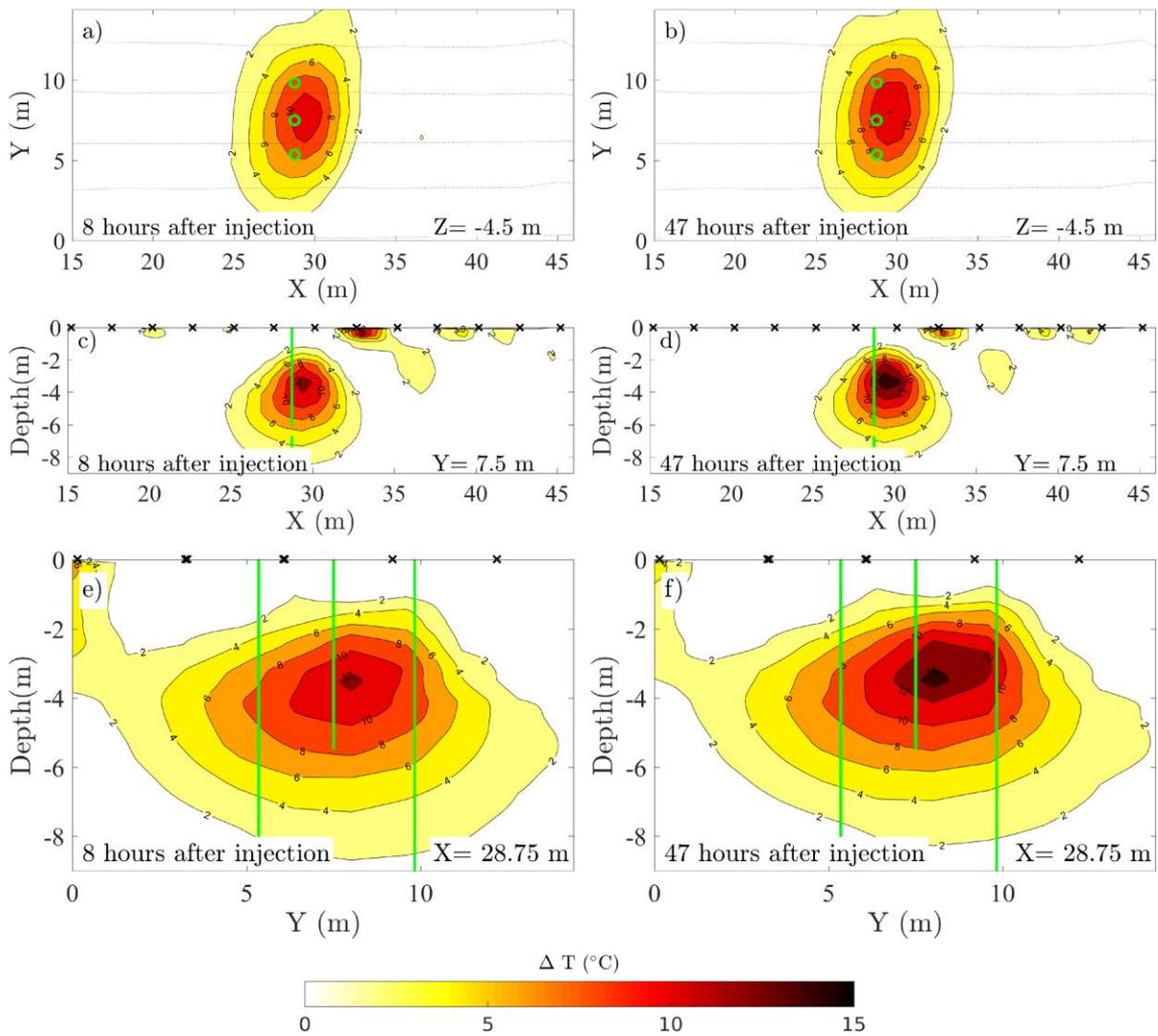


435 Figure 6: Conductivity variations 8 h and 47 h after the injection started with respect to the  
 436 background measurements in a horizontal layer at  $-4.5$  m (a, b), on cross-section at the level of the  
 437 injection borehole along the electrode profiles (c, d) and perpendicular to the electrode profiles (e, f).  
 438 The green circles and lines represent the injection and measurement boreholes. The black dotted lines  
 439 and crosses correspond to the electrodes.

### 440 5.3 ERT-derived temperature images

441 ERT-derived  $\Delta T$  images show a similar shape and evolution of the TAZ as the conductivity variation  
442 images (Fig. 7). On those images,  $\Delta T$  is fixed to  $0^{\circ}\text{C}$  in regions showing negative contrasts of  
443 conductivity since we interpret them as artifacts. The TAZ with a  $\Delta T$  increase of  $4^{\circ}\text{C}$  from the initial  
444 state presents an oval shape in the horizontal plane at  $Z=-4.5$  m with a length of 10 m in the Y  
445 direction and a width of 5 m along the X axis (Fig. 7a, b). As on the conductivity variation images,  
446 the shift along the X and Y directions of the TAZ with a highest  $\Delta T$  is observed (Fig. 7c, d, e, f).  
447 From the synthetic tests, we deduce that the TAZ oval shape is not an effect of the electrode design  
448 or inversion, but reflects the actual shape of the TAZ. This highlights the usefulness of ERT in  
449 providing information on the plume anisotropy. ERT-derived  $\Delta T$  images show a relatively shallow  
450 TAZ upper limit, in the region where the method sensitivity is still adequate for detection. However,  
451 as observed from the synthetic inversions, the TAZ upper limit deduced from ERT might be  
452 overestimated by 2 m. The TAZ should indeed be confined below the water table (3 m depth, Fig. 1).  
453 Similarly to the synthetic cases, it is difficult to properly image the shape of the TAZ in the lower  
454 part of the aquifer from ERT inversions, due to the low sensitivity (Fig. 4). Quantitatively, 8 hours  
455 after the beginning of injection, the derived  $\Delta T$  reaches locally  $12^{\circ}\text{C}$  in the upper part of the aquifer  
456 between the injection borehole and the electrode profile at  $Y=9$  m (Fig. 7). 40 hours later, that region  
457 shows a higher  $\Delta T$  value of  $14^{\circ}\text{C}$ . However, a difference of only  $2^{\circ}\text{C}$  is in the range of uncertainty  
458 observed in the synthetic tests, it is thus difficult to confirm this  $2^{\circ}\text{C}$  increase. Nevertheless, we note  
459 the persistence of a TAZ on the ERT-derived  $\Delta T$  images with a similar shape 8 h and 47 h after the  
460 injection (Fig. 7).

461

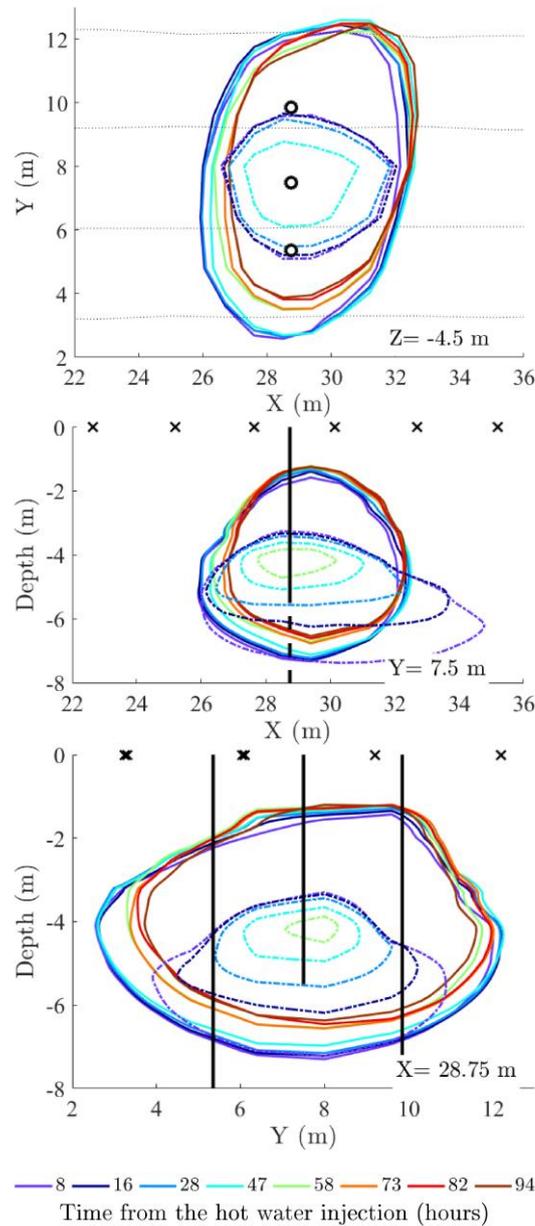


462 Figure 7: Temperature variation from the initial state estimated from ERT 8 h and 47 h after the  
 463 injection started in a horizontal layer at -4.5 m (a, b), on cross-section at the level of the injection  
 464 borehole along the electrode profiles (c, d) and perpendicular to the electrode profiles (e, f). The green  
 465 circles and lines represent the injection and measurement boreholes. The black dotted lines and  
 466 crosses correspond to the electrodes.

467

468 Predictions from the HGS model and ERT-derived  $\Delta T$  results are represented on cross-sections to  
 469 ease their comparison (Fig. 8). The edges correspond to the TAZ showing a  $\Delta T$  of  $4^{\circ}\text{C}$  from the initial  
 470 state, at different times after the injection. Estimates derived from ERT show a coarser TAZ with a  
 471 shape preserved all along the experiment, probably due to the smoothing effect of the inversion as  
 472 demonstrated by the synthetic tests. They illustrate the gradual plume shift along the  $X$  and  $Y$   
 473 directions. Predictions from the HGS model show a smaller TAZ confined at the level of the hot water  
 474 injection. The dimension of the predicted TAZ shrinks progressively and vanishes after 58 h.  
 475 Compared to ERT-derived  $\Delta T$  contours, the HGS predictions show also a stronger influence of the

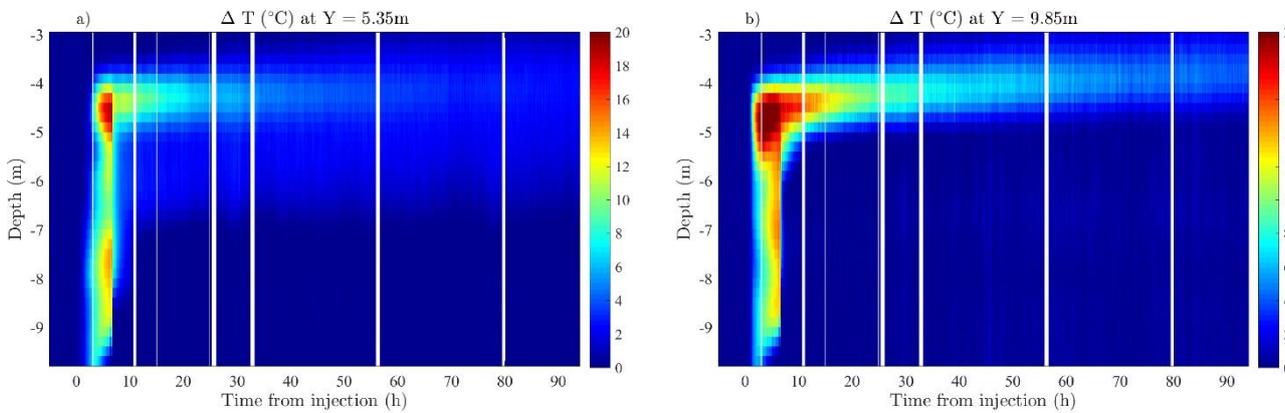
476 natural groundwater flow that transports the plume in the X direction, notably in the lower part of the  
 477 aquifer. Note that the natural flow was estimated from a regional model (Brouyère, 2001) and has a  
 478 larger influence in our experiment than in the one used for calibration given the lower pumping rate  
 479 (Table 1). Finally, the HGS predicted plumes do not show the significant anisotropy with an  
 480 elongation in the Y direction in the horizontal plane observed on the ERT results.



481 Figure 8: Evolution of the plume shape over time, with a contour line at  $\Delta T = 4^{\circ}\text{C}$ . The colored lines  
 482 represent estimates from ERT and the hashed colored lines predictions from the HGS model. Data  
 483 are presented on cross-sections in a horizontal layer at  $-4.5\text{ m}$  (a), at the level of the injection borehole  
 484 along the electrode profiles (b) and perpendicular to the electrode profiles (c). The black circles (a)  
 485 and lines (b,c) represent the injection and measurement boreholes. The black dotted lines (a) and  
 486 crosses (b,c) correspond to the electrodes.

487 **5.4 Direct temperature measurements**

488 DTS direct temperature measurements in the aquifer show that the temperature variation from the  
489 initial state  $\Delta T$  reaches 20°C in the upper part of the aquifer, while in the lower part  $\Delta T$  is about 10°C  
490 with a local increase to 15°C (Fig. 9). From the end of the injection,  $\Delta T$  decreases abruptly at depth  
491 below 5 m, while in the upper part  $\Delta T$  decreases gradually with time. We also note that measurements  
492 performed at Y=9.85 m show a larger and stronger print of the TAZ than measurements performed  
493 at Y=5.35 m. Thus DTS measurements confirm the plume shift in the Y direction observed from  
494 ERT. During the injection, the heat plume propagates at depth until the bottom of the boreholes  
495 (Fig. 9). However, the heat increase in depth might be related to local convection around boreholes  
496 and not due to heat propagation through the whole aquifer (Hermans et al., 2015; Klepikova et al.,  
497 2016).

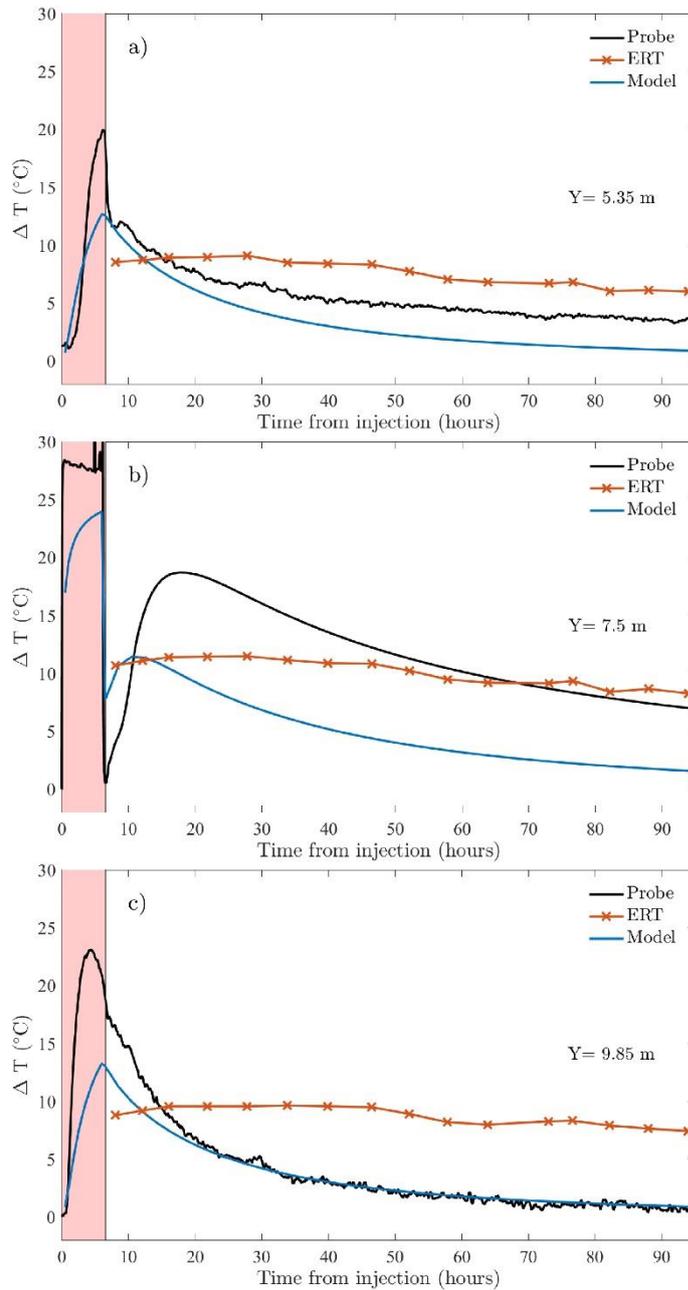


498 Figure 9: Temperature variation from the initial state estimated with the DTS all along the aquifer  
499 section at the measurement boreholes Pz17 (a) and Pz13 (b).

500 **5.5 Comparison of temperature dynamics**

501 The dynamics of  $\Delta T$  from direct measurements, ERT estimates and the HGS predictions are compared  
502 at the level of the measurement and injection boreholes (Pz 13, 15, 17 Fig. 1) at a depth of 4.5 m (Fig.  
503 10). That depth corresponds to the highest temperature measured with DTS (Fig. 9). Direct  $\Delta T$   
504 measurements are acquired using DTS data for Pz 13 and Pz 17 and a CTD probe for the injection  
505 borehole Pz 15 (Fig. 1). ERT-derived  $\Delta T$  estimates correspond to values extracted from the voxels of  
506 the 3D models obtained by the inversion of each data sets acquired after the hot water injection.  
507 Similarly, predictions correspond to values extracted from the HGS model voxels at different  
508 simulation times.

509



510 Figure 10: Temperature variations from the initial state at a depth of 4.5 m at the level of Y=5.35 m  
 511 (a), Y=7.5 m (injection borehole), b) and Y=9.85 m (c). The red zone represents the duration of the  
 512 heat injection. ERT estimates are obtained from the 3D time-lapse inversions. Model predictions were  
 513 computed with the HGS model.

514

515 The probe located in the injection well (Y=7.5 m, Fig. 10b) shows an average  $\Delta T$  value of 27°C  
 516 during the injection phase. Then  $\Delta T$  decreases sharply to 0.5°C and softly increases to a peak value  
 517 of 19°C due to the injection of cold water.  $\Delta T$  decreases then gradually to a value of 7°C at the end of  
 518 the experiment. Those direct measurements confirm the persistence of the TAZ all along the  
 519 experiment as observed with ERT (Fig. 7, 8, 10). DTS measurements in neighboring wells do not  
 520 show the same rebound pattern but instead a faster  $\Delta T$  decrease at the end of the injection phase. In

521 the measurement borehole at  $Y=5.35$  m, direct  $\Delta T$  measurements reach  $20^{\circ}\text{C}$  during the injection  
522 phase and decrease to  $3.5^{\circ}\text{C}$  at the end of the experiment (Fig. 10a). The second measurement  
523 borehole at  $Y=9.85$  m shows a different dynamic with a higher  $\Delta T$  value that reaches  $23^{\circ}\text{C}$  during  
524 the injection phase but decreases to a lower value of  $0.5^{\circ}\text{C}$  at the end of the experiment (Fig. 10c).

525

526 ERT-derived  $\Delta T$  clearly miss to capture the dynamics of the temperature decrease within the TAZ.  
527  $\Delta T$  values are underestimated a few hours after the injection phase and overestimated later (Fig. 10).  
528 In general, all  $\Delta T$  curves derived from ERT are rather stable, although showing a slight decreasing  
529 trend, compared to the direct  $\Delta T$  measurements. In particular ERT-derived  $\Delta T$  remains higher at the  
530 end of the experiment ( $\Delta T$  of  $6^{\circ}\text{C}$  and  $7.5^{\circ}\text{C}$  in borehole at  $Y=5.35$  m and  $Y=9.85$  m respectively, Fig.  
531 10). Those observations confirm the results of the synthetic tests, showing that the survey design is  
532 mostly sensitive to the global temperature change around the well.

533

534 The comparison of the HGS model prediction with the  $\Delta T$  direct measurements shows that the model  
535 globally underestimates  $\Delta T$  values compared to direct measurements (Fig. 10). The HGS model  
536 predicts lower  $\Delta T$  values at the end of the injection phase in the three boreholes of  $12.5^{\circ}\text{C}$ ,  $24^{\circ}\text{C}$  and  
537  $13.5^{\circ}\text{C}$  for a  $Y$  position of 5.35, 7.5 and 9.85 m respectively (Fig. 10a). At the end of the experiment  
538 the HGS model predicts a  $\Delta T$  value of  $1.5^{\circ}\text{C}$  in the injection borehole, confirming its inability to  
539 represent the TAZ persistence. At that time, in both measurement boreholes show similar  $\Delta T$  values  
540 of about  $1^{\circ}\text{C}$ .

## 541 6 Discussion

542 The experiment performed here demonstrates the capacity of ERT to deliver useful information about  
543 the 4D evolution of a thermal plume in a shallow aquifer. ERT-derived  $\Delta T$  images supply information  
544 on the plume anisotropy in a horizontal plane and its persistence all along the experiment (Fig. 7, 8,  
545 10). The plume anisotropy is attested with synthetic tests but cannot be deduced from direct  
546 measurements as boreholes close to the injection well are lacking along the  $X$  axis. However, direct  
547 measurements in the injection well confirm the TAZ persistence through time (Fig. 10b). So the  
548 performed experiment provides an interesting insight on the aquifer storage heat capacity not  
549 expected by the HGS model. The revealed anisotropy in the  $XY$  direction by ERT might indicate the  
550 presence of a preferential water flow path bypassing hydraulic barriers. Such a preferential flow path  
551 in the upper part of the aquifer can be related to clay lenses channeling groundwater flow in the  $Y$   
552 direction. Despite TL ERT supplying smoothed results due to regularization, the method has the  
553 capacity to provide useful insights about local hydraulic conductivity heterogeneity. The suspected

554 local heterogeneities have a strong influence on the groundwater flow and hence on heat transport.  
555 However, those small heterogeneities are not taken into account in the existing HGS model. The  
556 hydraulic parameters used for the simulation should be adjusted around the injection well to  
557 reproduce the plume evolution. Indeed, the hydraulic parameters were determined during a previous  
558 experiment representative of a larger scale. The integration of hydraulic parameter variations at that  
559 smaller scale into the model could help reproducing the horizontal anisotropy of the TAZ  
560 development revealed by ERT.

561

562 **In this specific experiment, although ERT identifies the trend, it does not accurately identify**  
563 **the temporal fluctuations of  $\Delta T$ .** Indeed, DTS measurements show a significant  $\Delta T$  decrease in both  
564 monitored boreholes, while ERT only indicates a slight decreasing trend (Fig. 10). **The agreement**  
565 **is better in the injection well.** Such examples illustrate the difficulty in comparing data  
566 representative of  $\Delta T$  changes in different volumes. On the one hand, heat loss towards the atmosphere  
567 might have favored the  $\Delta T$  decrease observed with DTS within wells. On the other hand, since ERT  
568 acquisitions are sensitive to temperature changes in a large volume, their estimates might better reflect  
569 the matrix heat storage and release. DTS and CTD measurements are bathed in the borehole water  
570 and thus not directly connected to the conditions in the aquifer. **In addition, the injection of**  
571 **unwanted cold water further complicated the distribution of temperature in the aquifer,**  
572 **creating an anomaly below the resolution of ERT.**

573

574 Quantitatively, the ability of ERT to image the dynamics compared to direct measurements is also  
575 limited since we note an underestimation of ERT-derived  $\Delta T$  values a few hours after the injection  
576 phase and an overestimation later. This is a direct consequence of the regularization, as highlighted  
577 by the synthetic inversions (Singha and Moysey, 2006). **In practice, a small plume with high**  
578 **temperature yields the same image as a large plume with a lower temperature. Combined with**  
579 **the slow dynamic characteristic of a storage experiment, it generates ERT images with limited**  
580 **amplitude variations, but with the correct trend.**

581

582 Due to regularization, ERT-derived  $\Delta T$  images also present a shift in the vertical position of the plume  
583 and a coarse shape of the TAZ, both identified with synthetic tests. The effect of blur could be  
584 attenuated by reducing the acquisition time, by acquiring a limited amount of reciprocal data (Rucker,  
585 2014). Other regularization method such as covariance constraints or minimum-gradient support  
586 might help overcome the smoothing effect and improve the delineation of the TAZ (Hermans et al.,  
587 2016b; Fiandaca et al., 2015; Nguyen et al., 2016). Furthermore, ERT lacks sensitivity in the

588 horizontal dimension due to the electrode inter-distance and the volume integrative sounding of the  
589 method. Observed changes with ERT are therefore representative of an average temperature variation  
590 in a 3D volume that prevents the reconstruction of an accurate TAZ shape and temperature estimate.  
591 The decrease of the ERT sensitivity with depth also hinders delimiting precisely the TAZ in the deeper  
592 region. This is further complicated due to the higher groundwater flux easing the plume transport as  
593 shown in the HGS model and DTS measurements (Fig. 2, 3, 9).

594

595 Discrepancy with direct measurements, providing local information of the medium properties  
596 surrounding the boreholes might be partly explained by perturbations in water circulation related to  
597 boreholes drilling and screening. Indeed, we suspect that strong temperature variations observed in  
598 the lower part of the aquifer during the hot water injection might be due to internal flow (Fig. 9). The  
599 integrative nature of ERT might also affect the comparison with local measurements. The latter,  
600 coupled to the adverse effects of inversion also hinders ERT-derived  $\Delta T$  images to reflect properly  
601 the TAZ dynamic. However, beyond the production of ERT-derived  $\Delta T$  images, discrepancies in the  
602 sensitivity of ERT and DTS could be exploited to better constrain hydraulic properties of the aquifer  
603 close to the injection borehole, considering they supply complementary insights on the medium  
604 properties. The information about the TAZ shape development from ERT can help conceptualizing  
605 the spatial variability of hydraulic properties in the aquifer, while direct measurements furnish a  
606 quantitative estimate of the temperature variations. So coupled inversion fitting simultaneously the  
607 surface resistance measurements and the direct temperature acquisitions is worth considering to  
608 evaluate accurately the TAZ location and geometry (Pollock and Cirpka, 2012). The coupled  
609 inversion would remove the regularization smoothing and would help refining the hydraulic property  
610 distribution around the injection well, which strongly affects predictions. Alternatively, the set-up of  
611 a stochastic inversion by model falsification such as a prediction focused approach would also benefit  
612 from the complementary nature of ERT and DTS data (Hermans et al., 2016a, 2018). Such a  
613 methodology presents the advantage that the developed analysis directly focuses on the prediction of  
614 the TAZ with the different measurement types, also avoiding any regularization.

## 615 7 Conclusion

616 Although a quantitative estimation of temperature was not possible in this experiment, due to the  
617 inherent limitations of inversion, our study demonstrates the pertinence of the minimally invasive 4D  
618 ERT method to provide qualitative, complementary insights on the development of a thermal plume  
619 in a shallow aquifer. The ERT-derived  $\Delta T$  images obtained here inform about the anisotropic shape  
620 of the thermal plume and its persistence all along the experiment. Hence, the method indicates the

621 presence of local heterogeneities at the vicinity of the injection well and the aquifer heat storage  
622 capacity confirmed by direct measurements. The anisotropic behavior cannot be validated using direct  
623 temperature measurements as it would have required a denser network of boreholes close to the  
624 injection well (Fig. 1). Thus, ERT measurements provide crucial information for the set-up of  
625 groundwater heat pumps (GWHP), which requires an accurate knowledge of the spatial variability of  
626 the aquifer's hydraulic and heat capacity properties. So, the insertion of a correct distribution of the  
627 hydraulic properties in the hydrogeological model has to be performed for estimating the likely shape  
628 and extension of the TAZ for different GWHP injection regimes. Surface ERT could be used as a  
629 control tool for monitoring the successful functioning of GWHP in operation.

630

631 However, the sensitive analysis we performed show that the ERT sensitivity to the plume decreases  
632 strongly with the plume depth and depends also on its dimension. Thus, the qualitative  
633 characterization of the TAZ performed through ERT inversion is strongly influenced by the TAZ  
634 depth. Indeed, the method struggles to provide the shape of the TAZ in the lower part of the aquifer.  
635 In our case, most of the TAZ is really shallow, reaching the groundwater table 3 m deep, so the  
636 method is sensitive to changes induced in the subsurface by the TAZ development. Nevertheless, for  
637 a deeper TAZ, the electrode design should be adapted to explore the medium at greater depths. This  
638 can be done by using a larger distance between electrodes but at the cost of a lower spatial resolution.

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## 646 Appendix A.1 Data analysis and error model

647 We assume that the data (resistance measurements) are uncertain due to noise and that the latter is  
648 composed of a random component and a systematic one correlated over time. The effect of systematic  
649 error, including modeling errors, can be canceled out by subtraction when working with data-  
650 differences inversion (LaBrecque and Yang, 2001). The weighting factors  $\epsilon_B$  and  $\epsilon_{TL}$  for the  
651 background and time-lapse inversions have to be correctly evaluated. As suggested by Slater et al.  
652 (2000) and Lesparre et al. (2017), we use the analysis of normal to reciprocal disparities. First,

653 resistance data were sorted to remove outliers and the data were selected so they presented a repetitive  
 654 error lower than 1% and a reciprocal error lower than 5% for each time step. The total number of  
 655 filtered data at each time step was reduced from 3045 to 1948 measurements.

656

657 For the background image, an error model  $\epsilon_B$  was fit to the measured difference  $|R_N - R_R|$  between  
 658 the normal  $R_N$  and reciprocal resistance  $R_R$  after removing outliers and binning by  $\langle R \rangle$ . The error  $\epsilon_B$   
 659 varies linearly with resistance (Fig. A.1 a, Slater et al., 2000):

660

$$\epsilon_B = a_B + b_B \langle R \rangle,$$

661

$$\text{Equation A.1}$$

662 with

663

$$\langle R_i \rangle = \frac{(R_{N,i} + R_{R,i})}{2} \text{ for all } i \text{ where } \frac{|R_{N,i} - R_{R,i}|}{\langle R_i \rangle} < 0.05 .$$

664

$$\text{Equation A.2}$$

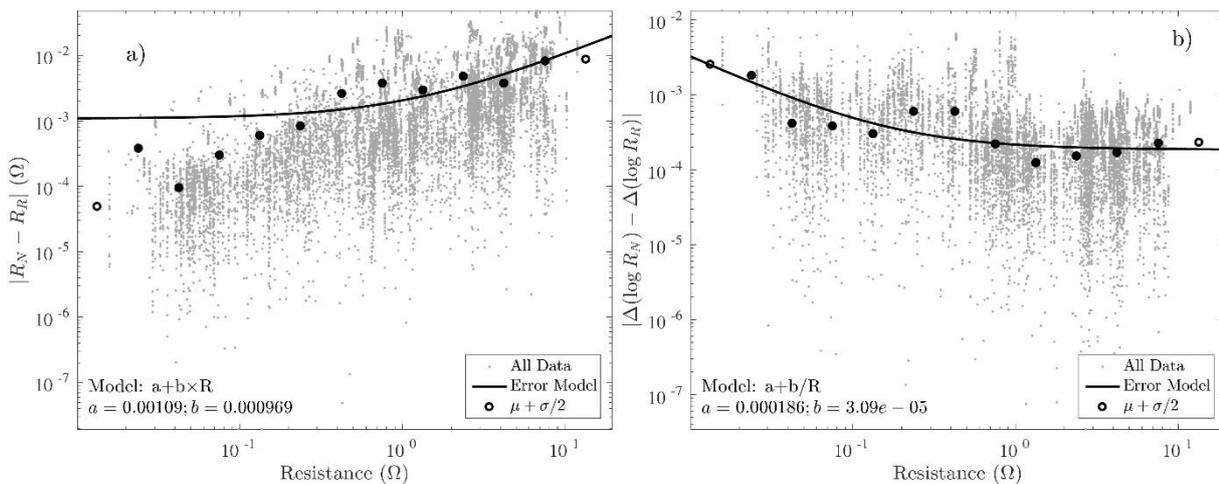
665 For the estimate of an error model of the time lapse inversions, normal and reciprocal data at a given  
 666 time were compared to the background measurements (Lesparre et al., 2017). The normal difference  
 667 between times  $t_0$  and  $t_i$  as  $\Delta \log R_N = \log R_{N,i} - \log R_{N,0}$ , and the reciprocal difference as  $\Delta \log R_R =$   
 668  $\log R_{R,i} - \log R_{R,0}$  were used in the time lapse inversions because the data fit were in the logarithmic  
 669 domain. The difference error model  $\epsilon_{TL}$  was then fit to the measured value of the normal-reciprocal  
 670 discrepancy  $|\Delta \log R_N - \Delta \log R_R|$  that varies linearly with the inverse of the resistance (Fig. A.1 b,  
 671 Lesparre et al., 2017):

672

$$\epsilon_{TL} = a_{TL} + \frac{b_{TL}}{\langle R \rangle}$$

673

$$\text{Equation A.3}$$



674 Figure A.1: Measured and model error from the normal and reciprocal acquisition for the background  
 675 (a) and the time-lapse (b) inversions.

676

677 The error model parameters  $a_B$ ,  $b_B$ ,  $a_{TL}$  and  $b_{TL}$  were estimated by fitting the measured errors as a  
678 function of  $\langle R \rangle$  (Fig. A.1). For the time-lapse error estimate,  $\langle R \rangle$  also expresses as stated in Eq. A.2.  
679 Error data were divided into classes of  $\langle R \rangle$  with four bins per decade of  $\bar{R}_i$ , logarithmically equally  
680 spaced. For each bin the average  $\mu$  and the standard deviation  $\sigma$  were estimated (Koestel et al., 2008).  
681 The choice of the error threshold on which the error model is fitted impacts the data weighting and  
682 so the residuals  $\chi$  that are used as a stopping criteria (see Eq. 5). We choose to fit both error models  
683 to  $\mu + \sigma/2$  (Fig. A.1) in order to be more sensitive to conductivity variations due to the hot water  
684 injection.

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