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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

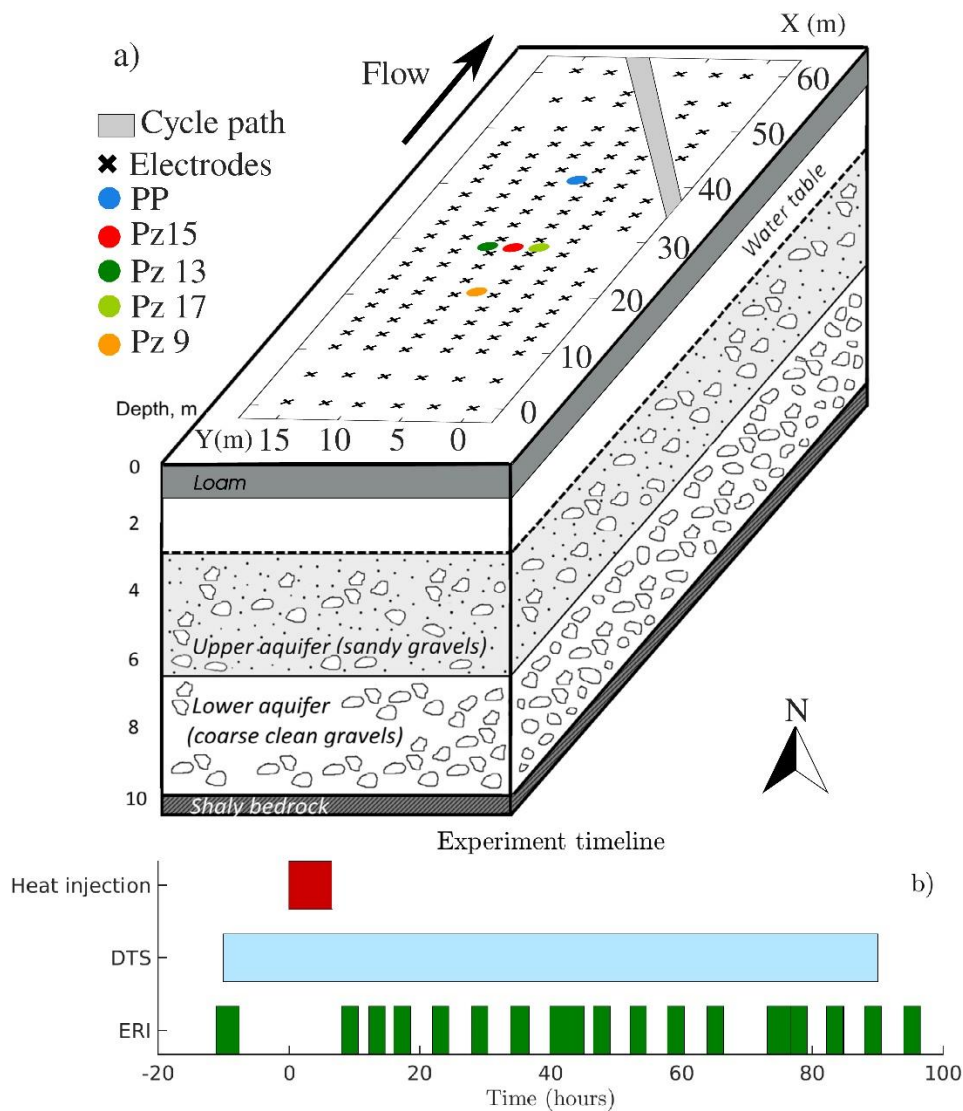
100 In this paper, we image the development of the TAZ during a heat injection and storage experiment
101 using a 3D ERT survey. The accuracy of ERT-derived temperature estimates is explored through
102 synthetic cases that emphasize the method's sensitivity to target depth and thickness. Results show
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/m³K 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 m³/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 m³/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 18m³ 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 m³ (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 ΔT 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 m³/h during 24 hours and 20 minutes, while
226 pumping at a constant discharge rate of 30 m³/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.

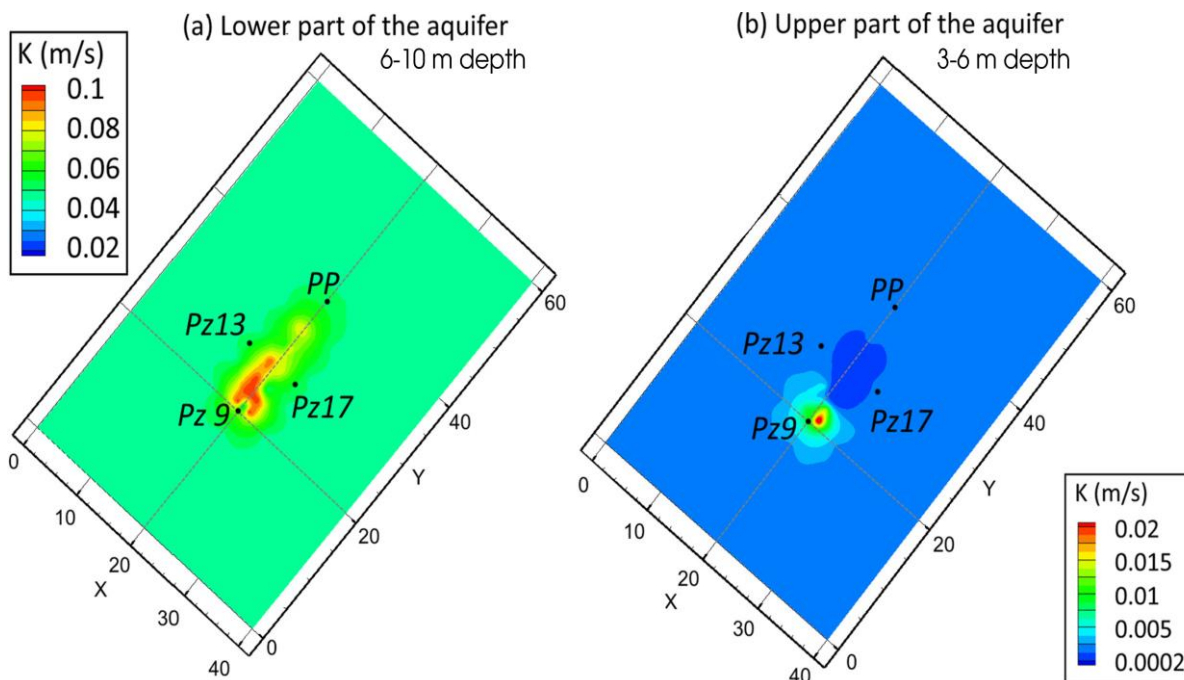
	This study	Klepikova et al., 2016
Depth of injection	5 m	9 m
Duration of injection	6 hours	24 hours and 20 minutes

Pumping time	Only during injection	All along the experiment
Pumping rate	3 m ³ /h	30 m ³ /h
Injection rate	3 m ³ /h	3 m ³ /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

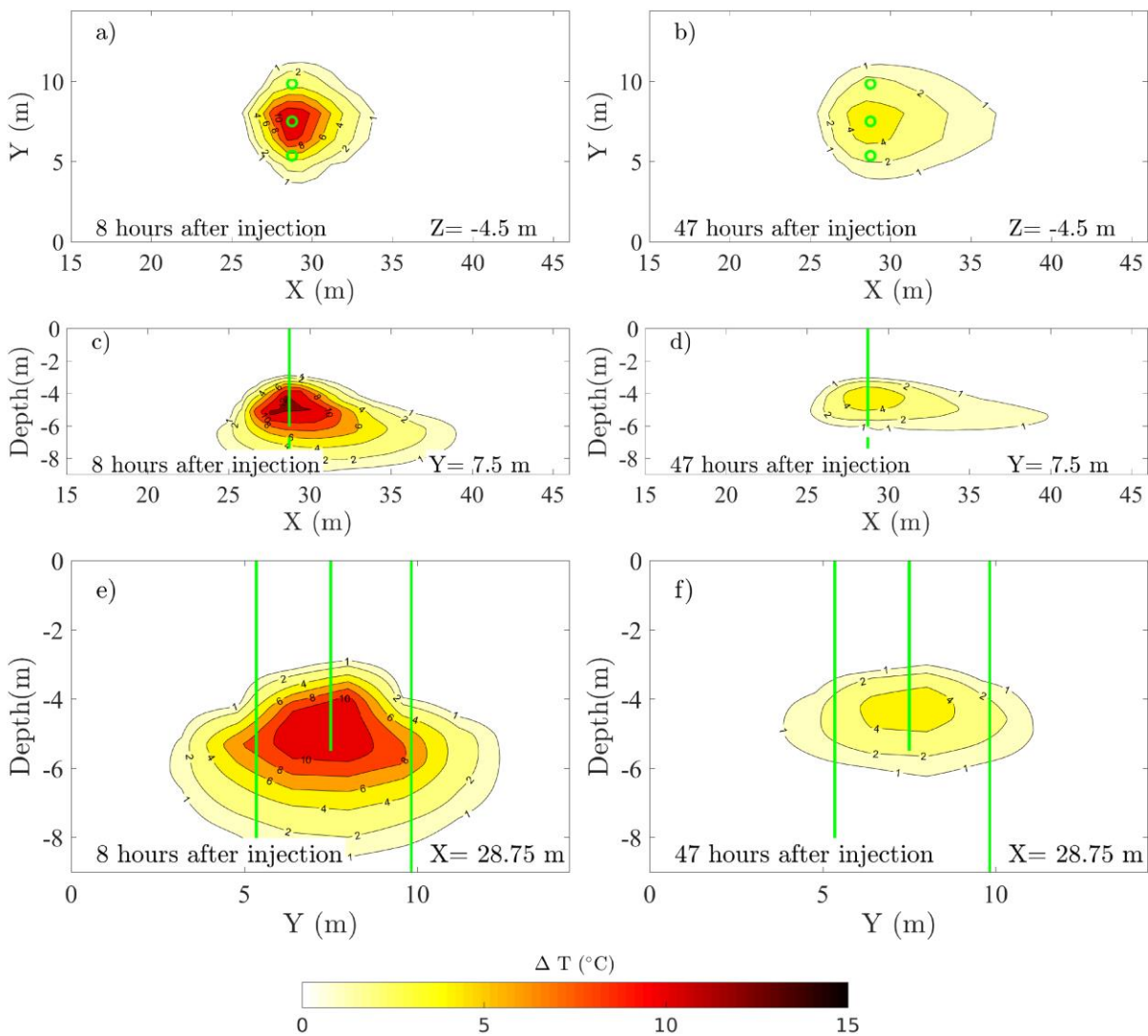


243 Figure 2: Spatial variability of the hydraulic conductivity K (m/s) models in XY planes for the lower
 244 (a) and upper (b) parts of the aquifer obtained from the inversion of transient temperature responses
 245 (Klepikova et al., 2016).

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 ΔT of 4°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 ΔT of 4°C is slightly reduced to a diameter of about 3 m at a depth of 4.5 m. However at
 258 that time, ΔT in the middle of the TAZ is significantly lower, e.g. from 12°C after 8 h (Fig. 3 c) to
 259 4°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 λ 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 ϵ_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 ϵ_{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 λ that minimizes the
 319 weighted residuals χ :

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

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 χ , 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 S/m (Hermans et al., 2015).

343
344 Temperature change from the initial state ΔT images can be constructed from the observed variations
345 of the bulk conductivity σ_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 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 ΔT
365 decreases with depth. The region of higher ΔT 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

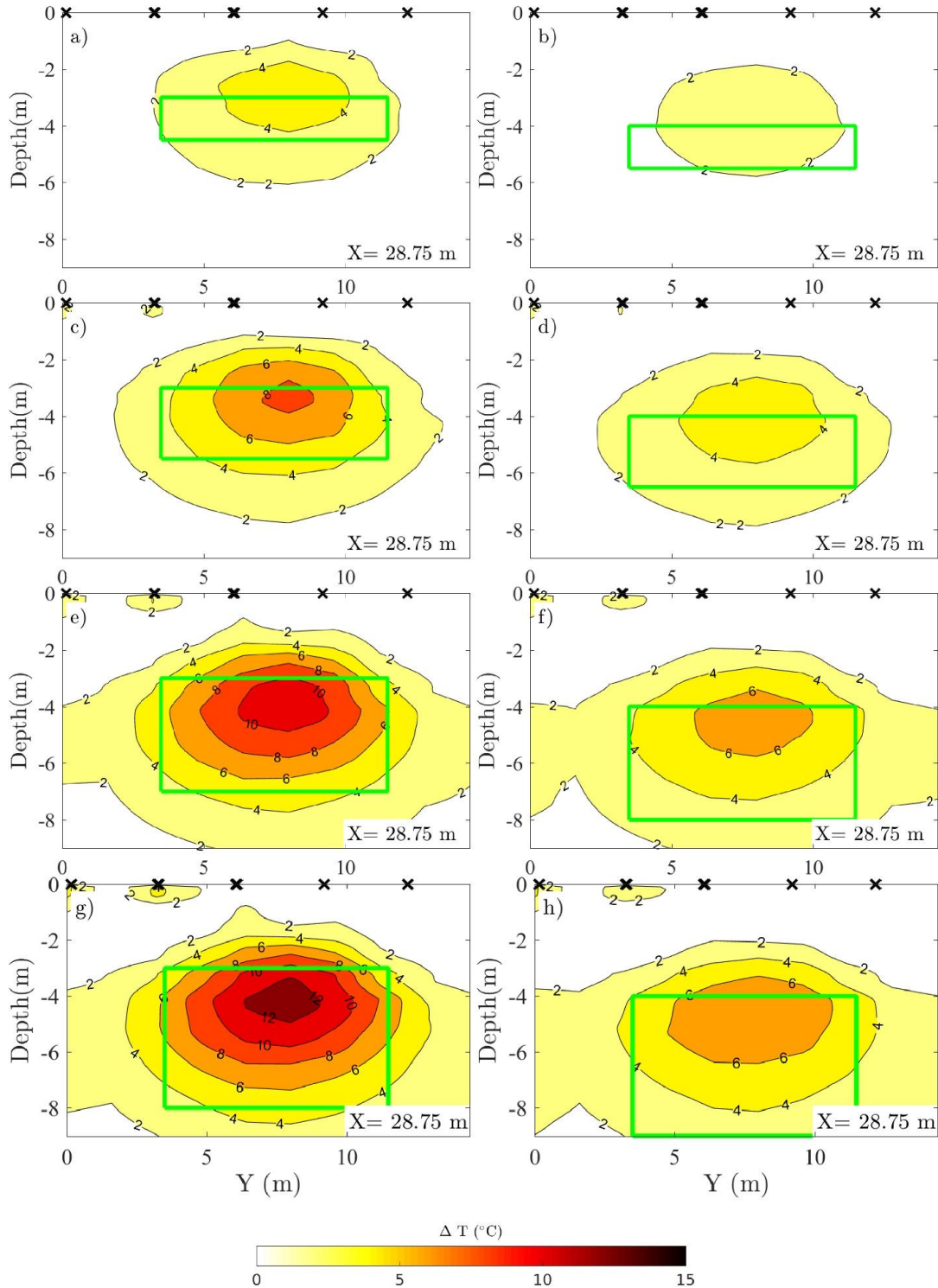
374 Table 2: Percentage of the temperature increase estimated by ERT from the synthetic target increase
375 for different cylinder height and upper position.

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 ΔT 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

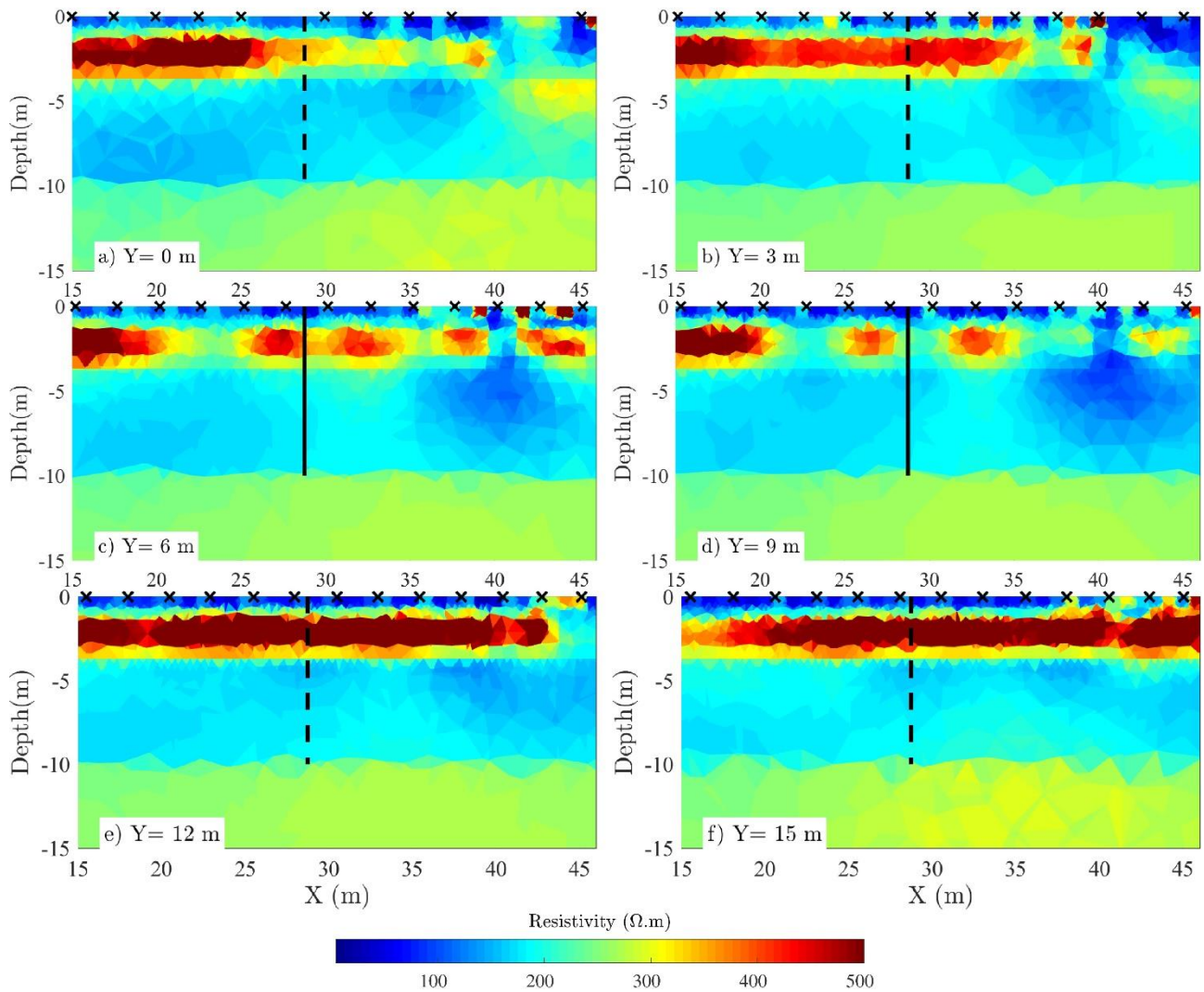


387 Figure 4: Temperature variation estimates from synthetics built as a vertical cylindrical heat plume.
 388 The cylinder thickness is of 1.5 (a, b), 2.5 (c, d), 4 (e, f) and 5 m (g, h) and its upper limit is at -3 m
 389 (a, c, e, g) and -4 m (b, d, f, h). The green lines indicate the shape of the 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 Ω .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 Ω .m and locally values reaching
400 1000 Ω .m while the surrounding medium presents a resistivity of 250 Ω .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 Ω .m with some more conductive areas notably below profiles located on Y= 6 and 9 m where
404 the resistivity is around 100 Ω .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 Ω .m. Those values are in
411 accordance with previous ERT investigations on the site (Hermans and Irving, 2017).



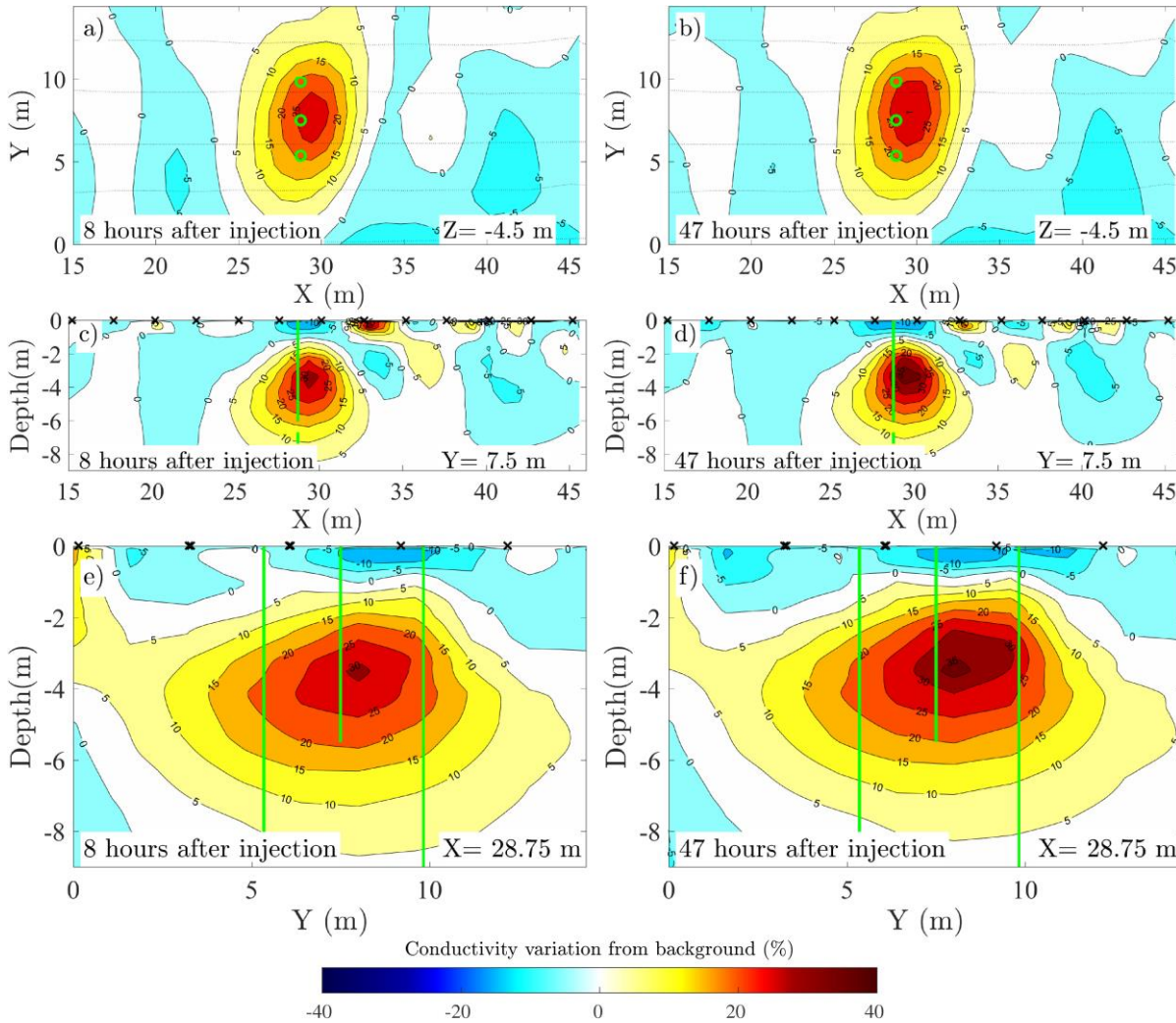
412 Figure 5: Electrical resistivity cross-sections extracted below each electrode profile from the 3D
 413 model. The black hashed or solid lines represent the injection and measurement boreholes projection.
 414 The black crosses correspond to the electrodes.

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

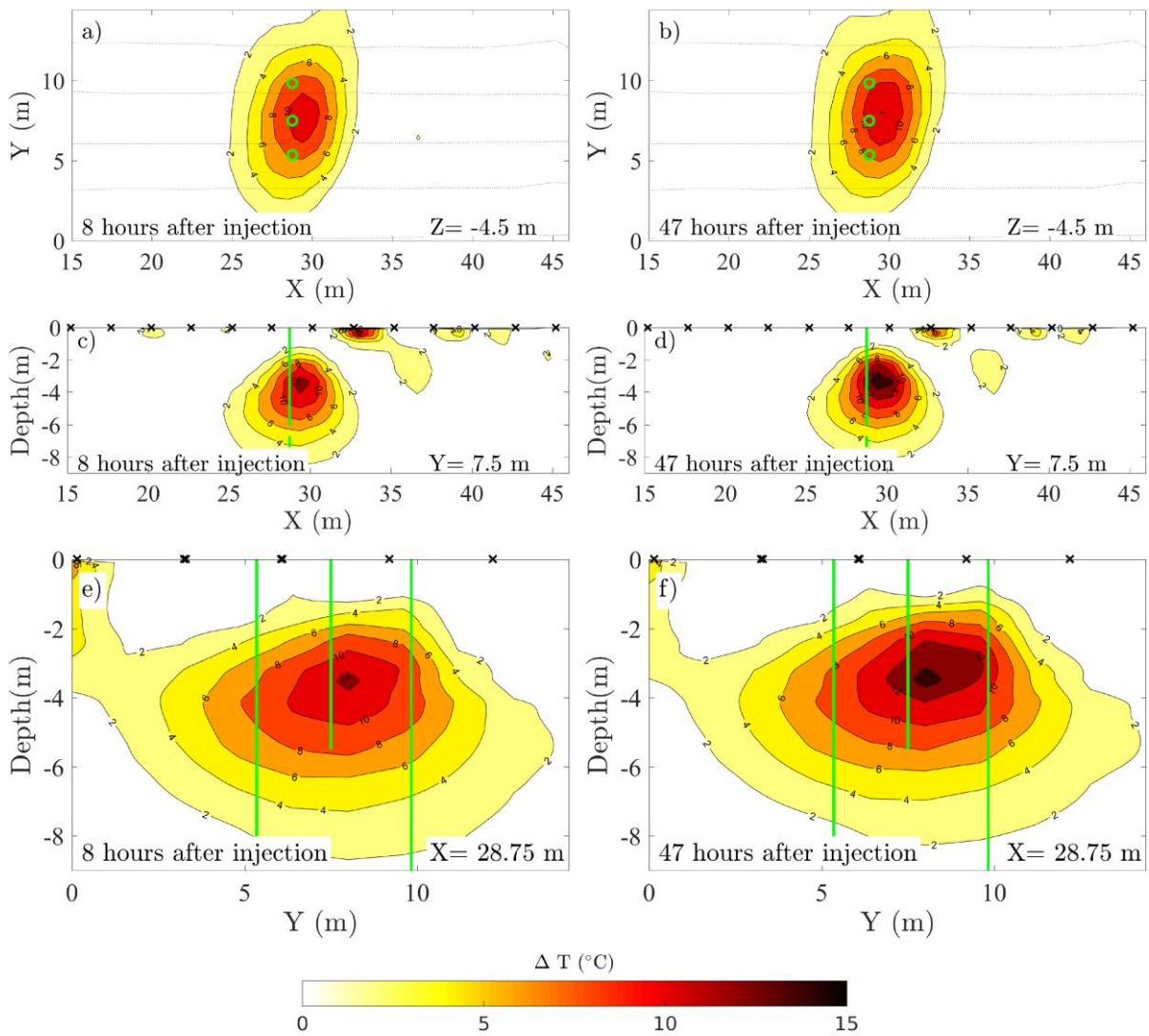


435 Figure 6: Conductivity variations 8 h and 47 h after the injection started with respect to the
 436 background measurements in a horizontal layer at -4.5 m (a, b), on cross-section at the level of the
 437 injection borehole along the electrode profiles (c, d) and perpendicular to the electrode profiles (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 ΔT images show a similar shape and evolution of the TAZ as the conductivity variation
442 images (Fig. 7). On those images, ΔT is fixed to 0°C in regions showing negative contrasts of
443 conductivity since we interpret them as artifacts. The TAZ with a ΔT increase of 4°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 ΔT 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 ΔT 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 ΔT reaches locally 12°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 ΔT value of 14°C . However, a difference of only 2°C is in the range of uncertainty
458 observed in the synthetic tests, it is thus difficult to confirm this 2°C increase. Nevertheless, we note
459 the persistence of a TAZ on the ERT-derived ΔT images with a similar shape 8 h and 47 h after the
460 injection (Fig. 7).

461

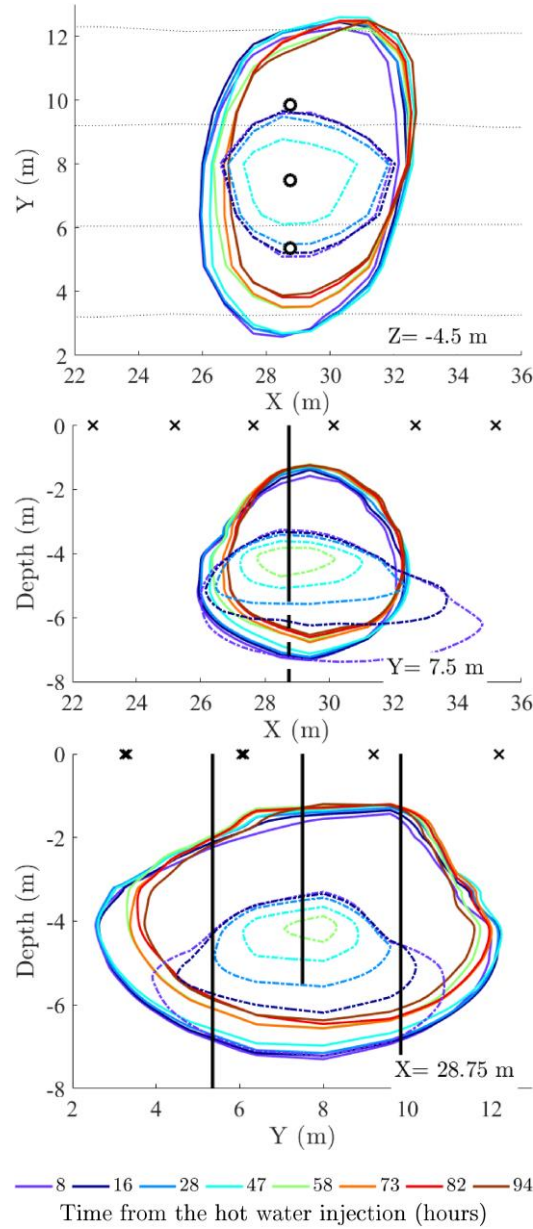


462 Figure 7: Temperature variation from the initial state estimated from ERT 8 h and 47 h after the
 463 injection started in a horizontal layer at -4.5 m (a, b), on cross-section at the level of the injection
 464 borehole along the electrode profiles (c, d) and perpendicular to the electrode profiles (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 ΔT results are represented on cross-sections to
 469 ease their comparison (Fig. 8). The edges correspond to the TAZ showing a ΔT of 4°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 ΔT contours, the HGS predictions show also a stronger influence of the

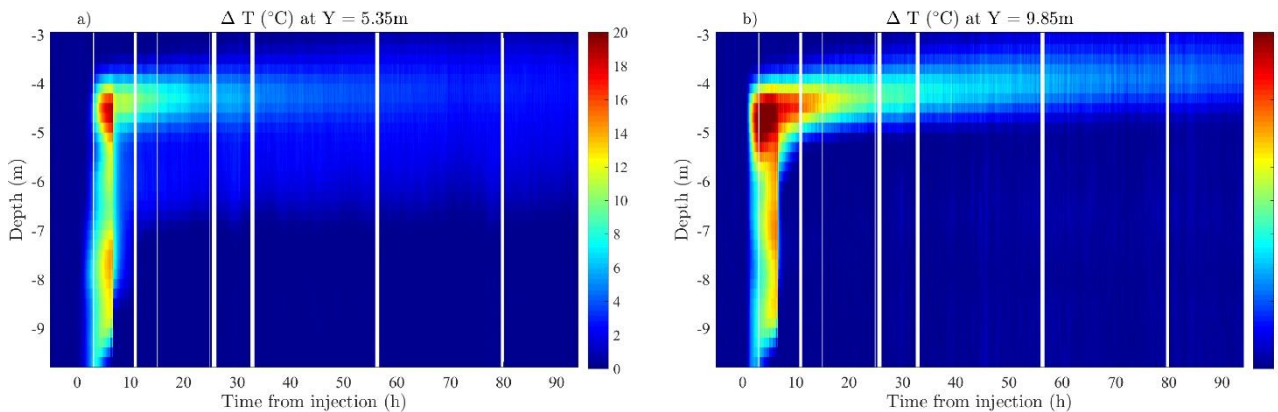
476 natural groundwater flow that transports the plume in the X direction, notably in the lower part of the
 477 aquifer. Note that the natural flow was estimated from a regional model (Brouyère, 2001) and has a
 478 larger influence in our experiment than in the one used for calibration given the lower 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 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 ΔT reaches 20°C in the upper part of the aquifer, while in the lower part ΔT is about 10°C
490 with a local increase to 15°C (Fig. 9). From the end of the injection, ΔT decreases abruptly at depth
491 below 5 m, while in the upper part ΔT 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).

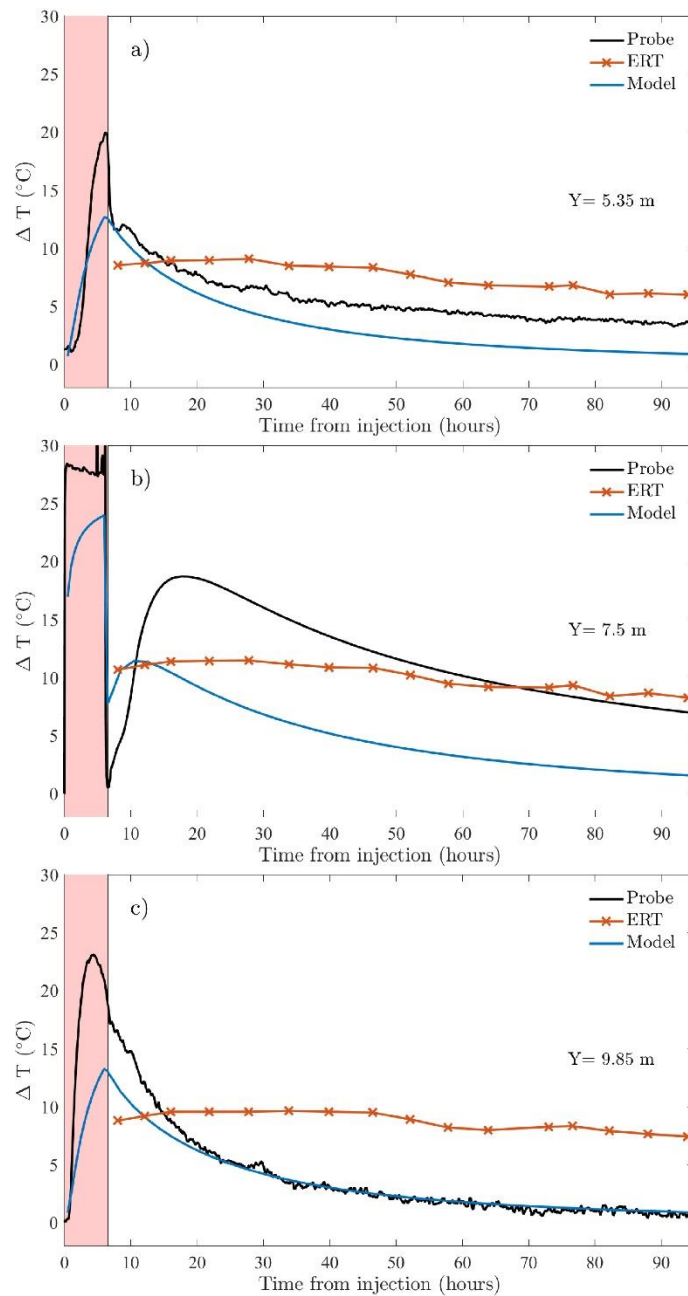


498 Figure 9: Temperature variation from the initial state estimated with the DTS all along the aquifer
499 section at the measurement boreholes Pz17 (a) and Pz13 (b).

500 5.5 Comparison of temperature dynamics

501 The dynamics of ΔT 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 ΔT
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 ΔT 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 ΔT value of 27°C
 516 during the injection phase. Then ΔT decreases sharply to 0.5°C and softly increases to a peak value
 517 of 19°C due to the injection of cold water. ΔT 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 ΔT decrease at the end of the injection phase. In

521 the measurement borehole at Y=5.35 m, direct ΔT measurements reach 20°C during the injection
522 phase and decrease to 3.5°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 ΔT value that reaches 23°C during
524 the injection phase but decreases to a lower value of 0.5°C at the end of the experiment (Fig. 10c).

525

526 ERT-derived ΔT clearly miss to capture the dynamics of the temperature decrease within the TAZ.
527 ΔT values are underestimated a few hours after the injection phase and overestimated later (Fig. 10).
528 In general, all ΔT curves derived from ERT are rather stable, although showing a slight decreasing
529 trend, compared to the direct ΔT measurements. In particular ERT-derived ΔT remains higher at the
530 end of the experiment (ΔT of 6°C and 7.5°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 ΔT direct measurements shows that the model
535 globally underestimates ΔT values compared to direct measurements (Fig. 10). The HGS model
536 predicts lower ΔT values at the end of the injection phase in the three boreholes of 12.5°C, 24°C and
537 13.5°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 ΔT value of 1.5°C in the injection borehole, confirming its inability to
539 represent the TAZ persistence. At that time, in both measurement boreholes show similar ΔT values
540 of about 1°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 ΔT 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 ΔT .** Indeed, DTS measurements show a significant ΔT 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 ΔT changes in different volumes. On the one hand, heat loss towards the atmosphere
567 might have favored the ΔT 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 ΔT 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 ΔT 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 ΔT images to reflect properly
601 the TAZ dynamic. However, beyond the production of ERT-derived ΔT 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 ΔT 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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644 We gratefully acknowledge Thomas Kremer and Solomon Ehosioke who participated to the field
645 works and data acquisition supervision.

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 ϵ_B and ϵ_{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 ϵ_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 ϵ_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

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

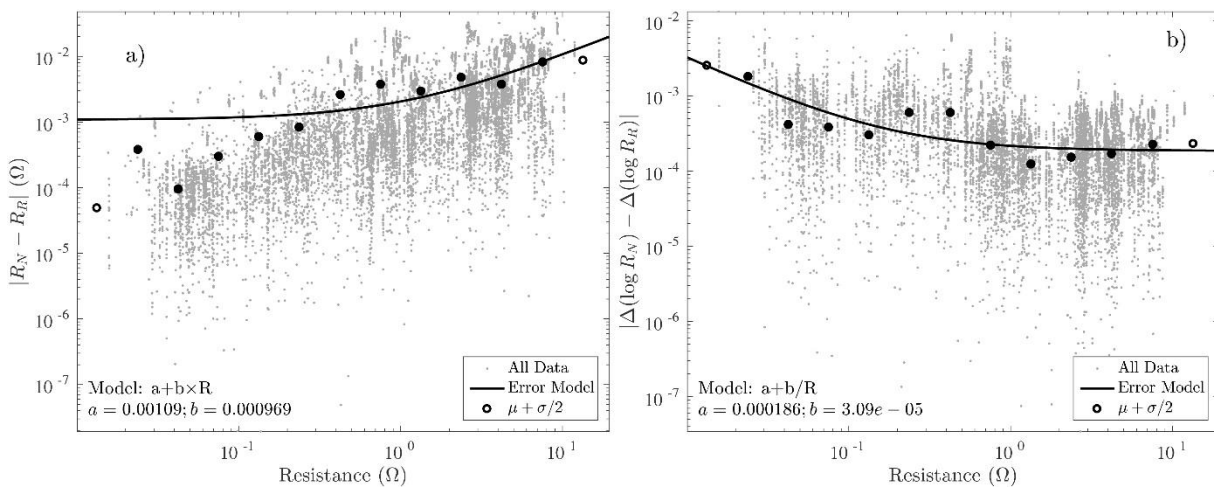
665 For the estimate of an error model of the time lapse inversions, normal and reciprocal data at a given
 666 time were compared to the background measurements (Lesparre et al., 2017). The normal difference
 667 between times t_0 and t_i as $\Delta \log R_N = \log R_{N,i} - \log R_{N,0}$, and the reciprocal 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 ϵ_{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 μ and the standard deviation σ 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 χ 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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