

CHARACTERIZATION OF SEAWATER INTRUSION USING 2D ELECTRICAL IMAGING

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ABSTRACT. We have investigated the potential of 2D electrical imaging for the characterization of seawater intrusion using field data from a site in Almeria, SE Spain. Numerical simulations have been run for several scenarios, with a hydrogeological model reflecting the local site conditions. The simulations showed that only the lower salt concentrations of the seawater-freshwater transition zone could be recovered, due to the loss of resolution with depth. We quantified this capability in terms of the cumulative sensitivity associated with the measurement setup and showed that the mismatch between the targeted and imaged parameter values occurs from a certain sensitivity threshold. Similarly, heterogeneity may only be determined accurately if located in an adequately sensitive area. At the field site, we identified seawater intrusion at the scale of a few kilometres down to a hundred metres. Borehole logs show a remarkable correlation with the image obtained from surface data but indicate that the electrically derived mass fraction of pure seawater could not be recovered due to the discrepancy between the in-situ and laboratory-derived petrophysical relationships. Surface-to-hole inversion results suggest that the laterally varying resolution pattern associated with such a setup dominates the image characteristics compared to the laterally more homogeneous resolution pattern of surface only inversion results, and hence surface-to-hole images are not easily interpretable in terms of larger-scale features. Our results indicate that electrical imaging can be used to constrain seawater intrusion models if image appraisal tools are appropriately used to quantify the spatial variation of sensitivity and resolution. The most crucial limitation is probably the apparent non stationarity of the petrophysical relationship during the imaging process.

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1. INTRODUCTION

1
2 Seawater intrusion is a natural process occurring along shores and results from the interaction between
3 sea water and coastal aquifers (figure 1a). The classic analytical description of Ghyben-Herzberg allows a
4 quick understanding of the system: due to the fact that dissolved salts give sea water a greater density than
5 fresh water, sea water flows inland in fresh water aquifers to form a wedge at equilibrium. The increasing
6 development of coastal regions (Bear et al., 1999; Cheng and Ouazar, 2004), which generates higher water
7 demands, and the sea-level rise due to climate changes (Melloul and Collin, 2006) stress such systems and
8 contribute in migrating the intrusion further inland (figure 1b). The direct consequence of seawater intrusion
9 is the degradation of water quality which becomes unsuitable for drinking or irrigation (Bear et al., 1999).
10 In the longer term, coastal ecosystems (both marine and fresh), which are strongly affected by nutrients,
11 contaminants and trace elements brought by submarine groundwater discharge (Michael et al., 2005), may
12 also be affected by seawater intrusion dynamics. To manage or study efficiently seawater intrusion, one needs
13 both real-time observation tools (Ogilvy et al., 2009), e.g. to position the seawater freshwater interface, and
14 reliable predictive models, if additional natural (sea level variations, recharge) or man-made (groundwater
15 pumping) factors are modified. The first task requires robust measuring technologies, whereas the second
16 one depends mainly on numerical modelling and its calibration. Our global objective in this context is to
17 show with other authors (Compte and Banton, 2007; Goldman and Kafri, 2006; Kirsch, 2006; Wilson et al.,
18 2006) that both tasks may benefit from non-invasive to minimally intrusive geophysical measurements, under
19 certain conditions.

20 On a worldwide scale, seawater intrusion occurrences are generally identified by observing anomalous chlo-
21 ride concentrations in coastal aquifers, using chemical or electrical well measurements (Bear et al., 1999).
22 Depending on the concentration of Cl^- ions (Kirsch, 2006), water can be classified as fresh ($< 150\text{mg/l}$,
23 $< 500\mu\text{S/cm}$), brackish ($150 - 10,000\text{mg/l}$, $500 - 25,000\mu\text{S/cm}$), or brine ($> 10,000\text{mg/l}$, $> 25,000\mu\text{S/cm}$).
24 Since the bulk electrical resistivity/conductivity is closely connected to the pore water salinity, brackish or
25 brine water is generally a good target for electrical or electromagnetic exploration. Numerous examples
26 of case studies may be found, for example in Bear et al. (1999); Goldman and Kafri (2006); Kirsch (2006).

27 Direct-current (DC) electrical methods are generally used to explore the subsurface down to approximately
28 100 m, whereas time-domain electromagnetic soundings (TDEM) may reach depths of several hundreds of
29 meters. DC resistance data are often processed using 2D inversion algorithms, yielding a cross-sectional im-
30 age, or tomogram (electrical resistance tomography or ERT), of the area in terms of bulk electrical resistivity
31 (Yang et al., 1999). Such sections may also be created using several adjacent vertical electrical soundings
32 (VES) but these reconstructed 2D sections may suffer from inversion artefacts due to the inadequate 1D
33 modelling of the subsurface, if lateral variations are present. In the examples provided by Kirsch (2006), an
34 assessment of freshwater reservoirs is demonstrated through VES models (Ketelsen and Kirsch, 2004) iden-
35 tifying a freshwater aquifer as a 130-180 Ohm-m layer, dipping inland at shallow depths. According to the
36 authors, this indicated that the seawater intrusion was shallower than expected from the Ghyben-Herzberg
37 approximation. VES limitations in the presence of significant lateral variations have led engineers and sci-
38 entists to use 2D imaging approaches (tomography), which may take into account the spatial variability in
39 the direction of the profile. In a recent contribution, Wilson et al. (2006) used VES soundings to grossly
40 delineate the spatial extent of a seawater intrusion and 2D electrical tomography to image the interface
41 in more detail. These authors compared inverted bulk resistivity values with water conductivity and Cl-
42 concentrations to retrieve the Archie's formation factor and a relationship between water quality and earth
43 resistivity. Assuming that no clays were present, they used bulk resistivity isolines to map the interface and
44 compared them visually with 1% mixing isolines from hydrogeological models. This was in turn used to
45 understand the structure and evolution of the saline interface. Other "non-electrically sensitive" methods,
46 such as seismic reflection (Foyle et al., 2002; Shtivelman and Goldman, 2000), may also be used to assist the
47 building of a hydrogeological conceptual model by providing structural information (stratigraphy, tectonics).
48 For example, combined interpretation of electrical resistivity, reflection seismic and gravity data may allow
49 one to elucidate the relationship between phreatic and confined aquifers, to obtain detailed stratigraphic
50 information of the main aquifer, and to provide a quantitative geological section that could be used for
51 aquifer modelling (Balía et al., 2003).

52

53 Hydrogeological modelling of density-dependent flow and transport is a well-established methodology
54 (Bear et al., 1999). Reliable numerical models are now able to represent a wide variety of physicochemical
55 phenomena occurring in seawater intrusion such as chemical precipitation, heat flux exchanges or nutrient
56 transportation. Paradoxically, characterizing coastal aquifers remains a major challenge and the more so-
57 phisticated models become, the more parameters are generally needed. According to Sanz and Voss (2006),
58 the most important hydrogeological parameters to calibrate seawater intrusion models are space-dependent
59 hydraulic conductivity and dispersivity. However, measuring these parameters *in situ* at the scale of sea-
60 water intrusion (which can reach several kilometres) is still a difficult task (Dagan, 2006). To support and
61 complement hydrogeological data, geophysical techniques are increasingly used in hydrogeological model
62 parameterization (Binley et al., 2002; Deiana et al., 2008; Looms et al., 2008; Rubin and Hubbard, 2005;
63 Tronicke and Holliger, 2005; Vereecken et al., 2006) since they may provide spatially distributed parameters
64 on a finer grid than conventional wells. Scarcity of hydrogeological data at the seawater intrusion scale may
65 account for the limited number of hydrogeophysical parameter estimation studies in seawater intrusion prob-
66 lems (Antonsson et al., 2006; Compte and Banton, 2007; Koukadaki et al., 2007; Lebbe, 1999; Wilson et al.,
67 2006). As an example, Koukadaki et al. (2007) modelled a simplified saltwater-freshwater interface using
68 the Ghyben-Herzberg relationship where the calibration is performed on hydraulic heads. For this purpose,
69 they used ERT-derived hydraulic conductivities computed using Archie's law but failed to address inher-
70 ent geophysical inversion and petrophysical limitations. Another recent study is provided in Compte and
71 Banton (2007), where the authors propose a so-called cross-validation methodology. They visually compare
72 on one hand, ERT with chloride concentrations obtained from hydrogeological modelling, and on the other
73 hand apparent resistivity data sets with synthetic ones generated by linking simulated salt concentrations
74 with bulk resistivity via a site-specific Archie's law. According to the authors, this approach leads to an
75 increased confidence of hydrogeological models. However, they do not address inherent geophysical inversion
76 and petrophysical limitations either, which are critical factors in the success of hydrogeophysical parameter
77 estimation (Day-Lewis et al., 2005; Kemna et al., 2006; Linde et al., 2006).

78

79 As seen above, objectives of seawater intrusion geophysical studies may either be detection, imaging, or
80 model building. For such studies, success relies on the interpretation of spatially and temporally dependent
81 values (in this case bulk electrical resistivity) to some diagnostic values (generally salt concentration per-
82 centage or water electrical conductivity). It implies that the computed geophysical values are representative
83 of the true ones, and that a petrophysical law applies adequately for all points. In other words, this assumes
84 that the geophysical image is equally well resolved, and that the petrophysical relationship is stationary in
85 space. To address resolution issues, geophysicists may use image appraisal tools as stressed by Alumbaugh
86 and Newman (2000); Day-Lewis et al. (2005); Friedel (2003). For example, Alumbaugh and Newman (2000)
87 have analyzed the use of various linearized functions for the *a-posteriori* appraisal of 2D and 3D nonlinear
88 electromagnetic inversion (applicable to 2D ERT). According to the authors, the linear results spatially
89 appear quite reasonable and the linearized approximations are yielding valuable information about the accu-
90 racy and resolution provided by the image and the non-uniqueness of the inversion problem. This should be
91 taken into account, in particular when attempting hydrogeophysical parameter estimation, to filter part of
92 the geophysical image (for example by computing empirically a threshold (Nguyen et al., 2007)) or to weight
93 their reliability using the diagonal terms of the model covariance matrix. Petrophysical uncertainties on the
94 other hand are more difficult to compute because of the necessity to estimate a site-specific relationship and
95 the limited knowledge of the validity of this relationship away from the calibration points. However, the
96 assumption of a stationary petrophysical law is often the only possibility and several successful case studies
97 are given in the literature (Hubbard et al., 2001; Kemna et al., 2002; Purvance and R. Andricevic, 2000).

98

99 The objective of this work is to assess the use of smoothness-constrained 2D electrical tomographic images
100 for marine seawater intrusion detection, and to constrain hydrogeological parameters. We will use different
101 hydrogeological simulations, considering first a homogeneous model with several transverse dispersivity val-
102 ues, then a heterogeneous, layered model. These models are scaled to reflect the overall geometrical setup
103 of a field site, located in Almeria, Spain. We will further demonstrate ERT potential under field conditions

104 using surface, in-hole, and surface-to-hole resistance data, for which ground-truth calibration data is avail-
105 able in deep boreholes.

106

107 Following this introduction, we briefly outline the electrical imaging approach used in this study, and
108 recall associated tools for image appraisal. We will then present the seawater intrusion simulations results
109 and their conversion to geophysical earth resistivity models. The results of these transformations are then
110 discussed and analyzed before presenting the field data and their interpretation. Finally, we present our
111 conclusions and perspectives following this work.

112

2. ELECTRICAL IMAGING

113 Significant advances in the resistivity method in the 1990's (see Binley and Kemna (2005) for an overview)
114 resulted in an increasing number of engineering and environmental applications. A drawback from this
115 increasing interest is that some users non-familiar with inversion theory may use commercially available
116 softwares as a black box and fall in interpretation pitfalls. Below, we briefly outline the imaging procedure
117 applied in this study and recall approaches for image appraisal.

118 **2.1. Forward model and inversion.** We here assume that the bulk electrical conductivity distribution
119 σ_b [$\frac{S}{m}$] is constant in the direction perpendicular to the image plane, in agreement with typical seawater
120 intrusion geometries along linear shores. The corresponding 2.5D electrical forward problem is then solved,
121 for the electric potential as a response to a point-source current excitation, for given boundary conditions by
122 means of the finite-element (FE) method. See Kemna (2000) for details of the forward model implementation
123 employed in this study.

124 We used a standard smoothness-constrained inversion algorithm (e.g. LaBrecque et al. (1996); Loke
125 and Barker (1996)) to invert a set of log-transformed resistance data, $d_i = \ln(r_i)$, r_i [$\frac{V}{A}$], into a 2D log-
126 transformed bulk electrical conductivity distribution, parameterized into pixels $m_j = \ln(\sigma_{b_j})$. Here, the
127 objective function

$$(2.1) \quad \Psi(m) = \Psi_d + \lambda \Psi_m$$

128 is minimized, where λ is the regularization parameter, Ψ_d is the measure of error-weighted data misfit,

$$(2.2) \quad \Psi_d = \|W_d(d - f(m))\|^2$$

129 and Ψ_m is the measure of first-order model roughness,

$$(2.3) \quad \Psi_m = \|W_m m\|^2$$

130 with W_d denoting data weighting matrix, $f()$ forward operator, and W_m model roughness matrix. We
 131 assume uncorrelated data errors so that W_d is a diagonal matrix; its entries w_i represent the inverse relative
 132 resistance errors. Due to the nonlinearity of the problem, the minimisation of $\Psi(m)$ results in an iterative
 133 Gauss-Newton scheme where, at each iteration q , the Jacobian matrix J_q (sensitivity) is computed. We refer
 134 to Kemna (2000) for more details on the implementation of the iterative scheme (choice of λ , step-length
 135 damping etc.). We start the inversion with a homogeneous model, determined from the mean measured
 136 response. The iteration process is stopped when the RMS (root-mean-square) value of error-weighted data
 137 misfit ϵ^{RMS} reaches the value 1 for a maximum possible value of λ . The latter aspect is important to actually
 138 fulfil the optimization criterion (smoothest model, subject to fitting the data). Moreover it is important,
 139 in order to avoid imaging artefacts, to not minimize the data misfit below the desired target value. For
 140 the field data sets, we estimated the average resistance error from deviations between normal and reciprocal
 141 measurements following the approach of Koestel et al. (2008); Slater et al. (2000)).

142 **2.2. Image appraisal.** When interpreting resistivity imaging results, one must be aware of the spatial
 143 variation of the resolution throughout the image plane. Several approaches may be used for the computation
 144 of an appraisal quantity (see Binley and Kemna (2005)) but they are rarely used in practice (Compte and
 145 Banton, 2007; Wilson et al., 2006).

146 Classic inversion theory (Parker, 1994) defines the explicit computation of the resolution matrix for a
 147 non-linear problem:

$$(2.4) \quad R = (J^T W_d^T W_d J + \lambda W_m^T W_m)^{-1} J^T W_d^T W_d J$$

148 The matrix R is often represented using its diagonal or by defining a resolution radius using off-diagonal
 149 terms (Alumbaugh and Newman, 2000; Day-Lewis et al., 2005; Ramirez et al., 1995). Ideally, R should be
 150 close to the identity matrix to achieve perfect resolution. However, as stressed by Meju (1994), a perfect
 151 resolution does not imply that the obtained model is accurate or reliable. One of the major drawbacks of
 152 this approach is the relatively high computation cost.

153

154 We use an alternative, faster approach in this study based on the data error-weighted cumulative sensitivity
 155 (Kemna, 2000):

$$(2.5) \quad s = \text{diag}(J^T W_d^T W_d J)$$

156 This cumulative sensitivity distribution shows how the data set is actually influenced by the different model
 157 cells, or in other words, how specific areas of the imaging region are "covered" by the data, by analogy to
 158 ray-based tomography. It must be emphasized, however, that a high sensitivity region may not necessarily
 159 be well resolved, but rather represents a favoring factor (Kemna, 2000).

160

3. SALINE INTRUSION MODELS

161 **3.1. Density-dependent flow and transport.** The simulation of seawater intrusion into the coastal
 162 aquifer was performed using the finite-difference model HST3D (Heat and Solute Transport in 3 Dimen-
 163 sions) by Kipp (1987). The three governing equations of heat, solute and fluid mass are coupled through
 164 the interstitial pore velocity, the dependence of the fluid density on pressure, temperature, and solute-mass
 165 fraction, and the dependence of the fluid viscosity on temperature and solute-mass fraction. Heat is not
 166 considered in this study. The flow equation used in the model is based on the fluid mass-balance equation
 167 and can be written as

$$(3.1) \quad \frac{\partial \epsilon \rho}{\partial t} = \nabla \cdot \rho \frac{k}{\mu} (\nabla p + \rho g) + q \rho^*$$

168 where p is the fluid pressure [\mathbf{Pa}], t is the time [t], ϵ is the effective porosity [*dimensionless*], ρ is the fluid
 169 density [M/L^3], ρ^* is the density of a fluid source [M/L^3], k is the porous-medium intrinsic permeability
 170 tensor [L^2], μ is the fluid viscosity [M/Lt], q is the fluid source flow-rate intensity [t^{-1}] and z is elevation
 171 above a reference point [L].

172

173 After solving the flow equation, the interstitial pore velocity is obtained from Darcy's law as

$$(3.2) \quad v = -\frac{k}{\epsilon\mu}(\nabla p - \rho g \nabla z)$$

174 where v is the interstitial-velocity vector [Lt^{-1}]. In a similar manner, the solute-balance equation used in
 175 the model is:

$$(3.3) \quad \frac{\partial(\epsilon\rho\omega)}{\partial t} = \nabla \cdot \epsilon\rho D_s \nabla \omega + \nabla \cdot \epsilon\rho D_m I_w \nabla \omega - \nabla \cdot \epsilon\rho v \omega + q\rho^* \omega^*$$

176 where, ω is the mass fraction of solute in the fluid phase [*dimensionless*], ω^* is the mass fraction of solute
 177 in the fluid source [*dimensionless*], D_s is the mechanical dispersion tensor [L^2t^{-1}] and D_m is the effective
 178 molecular diffusion of the solute [L^2t^{-1}]. In HST3D the fluid density is a function of pressure, solute
 179 concentration, and temperature (excluded here). It is incorporated into the code as

$$(3.4) \quad \rho(p, T, \omega) = \rho_0 + \rho_0\beta_p(p - p_0) + \rho_0\beta_\omega(\omega - \omega_0)$$

180 where ρ_0 is the fluid density at a reference pressure, p_0 [M/L^2t], temperature, T_0 [T], and mass fraction, ω_0
 181 [*dimensionless*], β_p is the fluid compressibility [L^2t/M] and β_ω is the slope of the fluid density as a function
 182 of mass fraction divided by the reference fluid density [*dimensionless*].

183 **3.2. Discretization and boundary conditions.** A centred-in-space and fully implicit-in-time finite-difference
 184 scheme was chosen as a compromise between stability and accuracy. This scheme eliminates numerical dis-
 185 persion caused by spatial differentiation, but not by temporal differentiation. A fully-implicit scheme adds
 186 numerical dispersion approximately proportional to the time step length. An inappropriate choice of time

187 step could therefore add more numerical dispersion than physical dispersion, but no significant numerical
188 dispersion was observed with a time step size of less than 20 days. When choosing the grid discretization, two
189 contradictory criteria are required to be fulfilled, stability and manageable simulation times. To minimize
190 the risk of instability when using central-in-space approximation the grid-Peclet criterion should be fulfilled
191 (Daus et al., 1985):

$$(3.5) \quad p_e = \frac{v\Delta x}{D} \leq 2$$

192 where Δx nodal spacing [L], and D is dispersion [L^2/T]. The criterion determines the relative size of
193 advection and dispersion on model element basis and basically states that smaller hydrodynamic dispersion
194 requires smaller discretization (assuming the velocity does not change). However, the criterion can only be
195 regarded as a guideline since acceptable solutions have been obtained with Peclet numbers as high as 10
196 (Huyakorn and Pinder, 1983; Sudicky, 1989). Therefore a grid spacing determined solely on the basis of the
197 Peclet number may be conservative, (i.e. the grid may contain too many nodes) and result in impractically
198 long simulation times. A horizontal grid spacing of 10 meters and vertical spacing of 1 metre proved adequate
199 for the current intrusion model setup.

200 A 2D vertical cross-sectional model was established to simulate saltwater intrusion of dimension similar
201 to that encountered in our field experiment. As illustrated in figure 2, the model domain is rectangular
202 with a length of 3000 m and a depth of 90 m. The flow system is defined by two fixed pressure (Dirichlet)
203 boundaries. A hydrostatic pressure distribution equivalent to 0 m and 2.5 m was applied at the sea- and
204 upstream boundary, respectively. The bottom boundary is no flow and the top is a free-surface (water table)
205 boundary condition. For the solute transport, associated concentrations as scaled mass fractions are used for
206 both the sea- and upstream boundary, with values of 1 and 0, respectively. The system parameters reflect
207 site conditions at Almeria and are outlined in table 1.

208

209 In order to obtain steady state conditions the simulation ran for 50000 days. With an initial condition of
210 freshwater in the whole domain, the saltwater starts intruding from the sea boundary in the form of a wedge

211 in the lower part of the aquifer. When saltwater meets the freshwater from the upstream boundary, the two
 212 fluids mix and a transition (mixture) zone forms between them. The shape of the intruding wedge and the
 213 thickness of the transition zone is governed by the hydraulic conductivity and the dispersivity values and is
 214 influenced by the boundary conditions, in particular on the freshwater side. The mixture of the two fluids,
 215 which is lighter than the saltwater, rises and is transported towards and out of the sea boundary by the
 216 freshwater gradient. The result is an establishment of a saltwater convection cell within the aquifer, where
 217 a dynamic equilibrium is ultimately obtained when the amount of salt entering through the lower portion of
 218 the sea boundary equals the salt leaving through the upper part of the sea boundary.

219 **3.3. ERT simulation of saline intrusion synthetic model.** In order to produce electrical tomography
 220 images, we converted the water salinity distributions, expressed as scaled mass fractions, obtained from
 221 the density dependent flow and transport simulations to bulk electrical resistivity. To achieve this, we first
 222 converted the mass fraction distribution ω to water electrical conductivities σ_w by means of the following
 223 relationship involving electrical conductivities σ_{wf} and σ_{ws} of fresh and sea water, respectively:

$$(3.6) \quad \sigma_w = [(1 - \omega) \cdot \sigma_{wf} + \omega \cdot \sigma_{ws}]$$

224 In this study, we used $\sigma_{wf} = 2,000\mu S/cm$ and $\sigma_{ws} = 50,000\mu S/cm$, which correspond to measured conduc-
 225 tivities on well samples. Then, we used Archie's law and a one layer (aquifer) over a half-space (aquitard)
 226 geological model (matching the hydrogeological one, figure 2) in order to delimit the areas where two differ-
 227 ent petrophysical models apply (see text below and figure 3). Archie's law (1942) relates empirically bulk
 228 electrical conductivity σ_b and the conductivity of the pore solution σ_w :

$$(3.7) \quad \sigma_b = \frac{\Phi^m}{a} \sigma_w S_w^n$$

229 In Archie's law, the observed proportionality between the two electrical conductivities is found to be de-
 230 pendent on the porosity Φ and the water saturation S_w (1 if fully saturated). In equation (3.7), m and n
 231 are usually referred to as the cementation exponent and the saturation exponent, respectively, and a is a
 232 proportionality constant. The quantity $F = \frac{a}{\Phi^m}$ represents the so-called formation factor. These parameters

233 are mainly controlled by the pore space geometry (constrictivity, tortuosity).

234

235 Archie's law validity depends on the contribution of electrical surface conduction to the bulk electrical
236 conductivity. When surface conduction may not be neglected (e.g. in the presence of clays), one must modify
237 Archie's law in order to account for this effect (e.g., Waxman and Smits (1968)):

$$(3.8) \quad \sigma_b = \frac{\sigma_w}{F} + \sigma_{surf}(\sigma_w)$$

238 In equation (3.8), the surface conduction term $\sigma_{surf}(\sigma_w)$ may also depend on water electrical conductivity
239 (e.g., Revil and P.W.J. (1998)). Since, in practice, this dependence is difficult to separate from the first term
240 on the right-hand side of equation (3.8), one may also write equation (3.8) as

$$(3.9) \quad \sigma_b = \frac{\sigma_w}{F'} + b$$

241 where F' corresponds to an effective formation factor, which includes both the dependence of pore-water
242 related bulk electrical conductivity and of surface electrical conduction on water conductivity, and b corre-
243 sponds to a surface conduction contribution independent of the water conductivity.

244

245 For our study, we chose to use site-specific petrophysical relationships, in order to relate simulations with
246 field results. Figure 3 shows a plot of the bulk electrical conductivity versus water salinity for different sam-
247 ples from the aquifer and of impermeable Pliocene bedrock of the Andarax river site. These were measured
248 in the laboratory at various salinities on disturbed samples and may therefore be only partly representative
249 of field conditions. The formation factors of the aquifer and aquitard materials were determined by fitting
250 equation (3.9) to the data points shown in figure 3. One can see that in the relevant salinity range, the bulk
251 electrical conductivity variations of the sediments are controlled by water salinity changes, which implies
252 that the surface conduction term b may be neglected. Therefore, one can directly relate bulk electrical
253 conductivity variations to salinity concentration variations. One can also see that the aquitard materials
254 exhibit higher bulk electrical conductivity values than the aquifer materials for the same water electrical

255 conductivity. However, since the two petrophysical relationships overlap over a wide range of electrical
256 conductivities, it is generally not possible to discriminate between the two lithologies from bulk electrical
257 conductivity. However, if time-lapse data are available, one could distinguish the two formations by (1) ob-
258 serving that bulk electrical conductivity related to aquifer materials exhibits larger variations than aquitard
259 materials, due to the latter's low hydraulic conductivity, and (2) by considering independent water salinity
260 measurements.

261

262 Once the water salinity distribution were modelled, we converted the results to an electrical resistivity
263 distribution (figure 4) on the ERT forward grid (cell size in the domain of interest: 10 m horizontal and
264 1.25 m vertical) using equations (3.6) and (3.9) (neglecting the surface conduction term b) and a standard
265 interpolation function. From there, we used the electrical finite-element forward code to simulate any type of
266 desired electrical measurement. The electrode layout (figure 4) was chosen in order to fit a future permanent
267 installation at the field site. It consists of a 1 km surface line and two approximately 90 m deep boreholes.
268 The surface line is equipped with electrodes every 10 m whereas the borehole electrodes have a 5 m vertical
269 spacing, allowing better resolution at depth. The chosen measurement protocol is a standard "skip-1" dipole-
270 dipole. Simulated resistances exhibiting very low values ($< 5e - 4\Omega$), corresponding to very large geometric
271 factors, were disregarded in order to simulate a minimum voltage that can be reliably determined by typical
272 ERT instruments, e.g., 0.5 mV for a current of 1 A (Vanderborght et al., 2005). For this study, 7% Gaussian
273 noise was added to the resistance data. This level is slightly higher than the observed noise level in the
274 field data (5%) as to simulate a realistic worse case scenario. The inversion grid was defined using coarser
275 vertical meshing. This is often the case when coupling geophysical and hydrogeological models since the
276 latter generally demand a finer discretization.

277 **3.4. Results.** In a first step, we will study the imaging ability of the electrical setup defined above, to re-
278 construct the spatial electrical conductivity variations associated with simulated salinity distributions. The
279 different mass fraction distributions correspond to different steady state scenarios for a given set of boundary

280 and initial hydrological conditions and predefined hydrogeological parameters (see section 3.1).

281

282 Let us first examine the overall procedure illustrated in figure 4. At first glance, the resistivity distri-
283 bution seems pretty well resolved (compare figures 4 c and d), except for the region deeper than 100 m,
284 where no intrusion occurs. To detect or to position the interface, which is a primary objective of ERT in this
285 context, one must relate bulk electrical resistivity values to either water electrical resistivity (or its inverse,
286 conductivity), or mass fraction, diagnostic of seawater contamination. We again performed this step on the
287 basis of equation (3.9) using the same parameters as for the "forward step" and obtained the result shown
288 in figure 4b. Comparing now figures 4 a and b, the ERT-derived mass fraction image appears less correlated
289 to the true one than the resistivity images are. If we examine carefully the mass fraction distribution up
290 to 0.2, we can see that there is a good agreement between the recovered image (figure 4 b) and the original
291 one (figure 4 a), despite the introduction of a 7% Gaussian noise in the electrical data. Below this level, a
292 distortion of the interface clearly appears, in particular around 700m in the x direction, where sensitivity
293 (resolution) reaches a critical point as explained hereafter.

294

295 In figure 5, one can see that the cumulative sensitivity distribution of the resistivity model is asymmetric,
296 and reflects the overall ERT configuration. Let us now examine more carefully the relation between the
297 sensitivity distribution along two vertical lines, one located adjacent to a borehole and the other located be-
298 tween the two boreholes , and the correctness of the ERT-derived mass fractions (figure 5). At the bottom of
299 figure 5, two graphs are shown in which the recovered mass fractions is plotted against the true mass fraction
300 for the two vertical boxes shown in the sensitivity image (note that here, sensitivity, electrical conductivity,
301 and depth are all positively correlated due to the general shape of seawater intrusion). For the line where
302 sensitivity is constantly high, the correlation is excellent. For the second graph, this agreement gradually
303 breaks down (reflecting the gradual variation in the sensitivity distribution) for mass fractions above 0.4.
304 This deviation, sometimes miscalled "petrophysical breakdown", is a matter of resolution as suggested by
305 earlier authors (Day-Lewis et al., 2005). The discrepancy between the recovered and true mass fractions
306 originates from the incorrect inverted resistivity values; the petrophysical relationship, on the other hand

307 still holds (if still valid - like in this example). We must stress that the problem of determining this fit is
308 of crucial importance for using ERT to characterize seawater intrusion in terms of presence, position, and
309 shape, but is not often taken into account (Compte and Banton, 2007; Wilson et al., 2006). It can only be
310 addressed through numerical modelling or via calibration if ground-truth data is available.

311

312 As a second task, we will investigate the ability of ERT to differentiate between different seawater intrusion
313 shapes corresponding to hydrological models run using different transverse dispersivity values α_T . From a
314 geometrical point of view, the seawater-freshwater transition zone ranges from diffuse to sharp, corresponding
315 to α_T values ranging from 2.5 m to 0.5 m, respectively. This range of values is in agreement with values
316 adopted in the literature (Oude Essink, 2003; Panday et al., 1993; Reily and Goodman, 1987; Voss and Souza,
317 1987). Transverse dispersion is important, since it generates a transition zone, responsible for seawater
318 circulation and for salinity of pumped water (Dagan, 2006). The results of these simulations and electrical
319 imaging are summarized in figure 6 where two sets of mass fraction isolines (true and ERT-recovered) are
320 shown. In general, the ERT-recovered mass fraction isolines correspond relatively well with the true ones
321 for the different scenarios, except for the area around 700m mark in the x direction, and for the sides where
322 no electrodes are located. In addition, figure 6 indicates that the chosen electrode configuration only allows
323 mass fractions up to 0.2 to be recovered correctly in all different scenarios involved. If the transition zone is
324 entirely located in well resolved areas, as assessed e.g. by cumulative sensitivity distribution, it is possible
325 to correctly determine the width of the transition. In our case, ERT yields less reliable results for diffuse
326 interface simulations since the end of the transition is then located in deeper areas. Paradoxically, although
327 we are using a smoothness constraint to regularize the ERT inverse problem, in our simulations the recovery
328 of the transition zone is better the sharper it is.

329 A critical task for ERT, for which it is often "promoted", is the recovery of heterogeneity. We illustrate
330 this point by attempting to detect the presence of a low permeability layer or, more accurately, its lateral
331 continuity (since its presence would be already given by borehole information) for two scenarios where the
332 electrical conductivity σ_{wf} varies (equation (3.6)). The formation factor (equation (3.9)) is taken as uniform
333 for the recovery over the entire image plane (but depends on the lithology for the simulation), assuming that

334 we have limited knowledge of the spatial stationarity of the petrophysical relationship. This could illustrate
335 two regional settings, one for which the freshwater is relatively free of chlorine ions and one where the "fresh"
336 water is more brackish (as for the region of Almeria). Figure 7 shows an exemplary mass fraction isoline
337 (0.01) for these scenarios in conjunction with the corresponding cumulative sensitivity distribution, clipped
338 for values smaller than 10^{-5} of the maximum value. This threshold was determined empirically using the
339 breakdown limit observed in figure 5 (bottom right). The difference between the two sensitivity distributions
340 in figure 7 arises from the two different fresh water values used in the petrophysical relationships. From the
341 top figure, we see that the true isoline is located far in the clipped (low) sensitivity region and that the
342 recovered mass fraction isoline is far from the true one. For the other scenario (bottom figure), the ERT
343 recovered isoline is closer to the true one and could indicate the continuity of the low permeability layer.
344 The latter isoline is, in comparison with the top figure, located nearer to the edge of the clipped sensitivity
345 region, indicating a better resolution. This clearly further supports the necessity to use an image appraisal
346 tool in order to validate the derived information (heterogeneity) from ERT images.

347

4. THE SITE OF ALMERIA

348 **4.1. Presentation of the site.** Situated in the Southeast of Spain (figure 8), the area coincides with a
349 valley in the lower part of the river Andarax, which in this stretch supports a water flow only during flood
350 conditions. Precipitation in the area is low, with mean values in the order of 300 mm/year, concentrated
351 over short periods (Pulido-Bosch et al., 2004). The Mediterranean coastal region of southern Spain has un-
352 dergone a startling economic development over the last 20 years. This is due not only to the long-established
353 tourist industry but also to changes in the pattern of agriculture in which traditional farming has given way
354 to greenhouse cultivation. Because of the high profitability of this method of agriculture, the surface area
355 under cultivation has increased notably and so has the water demand. The main source of water in the
356 area is from groundwater and in many cases the withdrawal has led to aquifer overexploitation and seawater
357 intrusion (Benavente, 1985). The economy of the region depends to a large extent on the availability of
358 water of adequate quality for crop irrigation, so that marine intrusion and salinization could endanger the

359 economic future of the coastal areas (Pulido-Leboeuf, 2004; Van Cauwenbergh et al., 2006).

360

361 The Lower Andarax valley is enclosed by the Sierra Alhamilla, with its mainly metapelitic outcrops and
362 by the Sierra de Gador, which constitutes a limestone/dolomite massif. The area is infilled by Miocene
363 and Pliocene detritic deposits of quite diverse lithology (marls, sandy silts, sands and conglomerates) and
364 numerous gypsiferous evaporite intercalations. The micashists and quartzites are practically impervious and
365 the carbonate formation is highly porous, with permeability due to fissuring and/or karstification. The post-
366 orogenic materials show a highly variable permeability. The Miocene and Pliocene marly formations have
367 very low permeability, whereas the Pliocene deltaic deposits, the Quaternary and Plioquaternary formations
368 are water-bearing. The hydrogeological units are defined in terms of three units, the detrital, the carbonate
369 and the deep aquifers (Pulido-Bosch et al., 1992). The Detrital Aquifer runs the entire length of the valley.
370 It includes Quaternary materials, both alluvial and deltaic, together with Pliocene, deltaic, sandy-silt con-
371 glomerates. The thickness of the aquifer varies from 200 m in the Pliocene materials in the area of Santa Fe,
372 to 40-80 m in the Quaternary deltaic materials. Consistent with the piezometric evolution, the aquifer shows
373 sharp fluctuations in level, with a clear seasonal recovery. In low-lying areas, the area registering piezometric
374 head drop has increased over the last few years and even hydraulic head cones below the sea level have
375 appeared in the areas of greatest exploitation. The Carbonate Aquifer of the Sierra de Gador consists of
376 Alpujarride limestones and dolomites, with local Miocene calcarenites that extend along the entire edge of
377 the sierra. The Deep Aquifer, situated in the centre of the valley is confined and is compartmentalized into
378 blocks. It is composed of limestone dolomite strata with some layers of quartzite that form the basement of
379 the depression.

380 **4.2. Electrical measurement setup.** The electrical profile was deployed in the bed of the River Andarax
381 (figure 8) in the lower part of the delta, where it is invariably dry for most of the year except for a few
382 flash floods in the autumn-winter months. The measurement campaign was performed for one part using
383 a temporary electrode surface layout with 20 m spacing of 2.3 km long with a Supersting system and for
384 another part, using a research prototype system (Ogilvy et al., 2009) and a permanent electrode array. The

385 latter was buried in a trench about 1m deep and 1.6 km in length, extending upstream from a point about
386 300m from the shoreline. The subsurface array has electrode take-outs at 10 m intervals. The measurement
387 system used was developed under the EU ALERT project (Ogilvy et al., 2007) and has 288 electrodes, all of
388 which are addressable so that any electrode array may be selected. For control purposes, 3 deep boreholes
389 (B320, B820, and B1320, shown in figure 8) were drilled through the Quaternary aquifer sediments into the
390 relatively impermeable Pliocene marls at a depth of 100 m. These boreholes were spaced about 500m apart
391 on the electrical line to provide ground-truth and to assist image calibration. Electrode arrays with take-outs
392 at 1 m spacing were attached to the outer PVC casing of two of the boreholes to facilitate surface-to-hole
393 measurements (figure 8) for the permanent array.

394 Data have been collected using both surface (dipole-dipole and Wenner) and borehole electrode configura-
395 tions. Data quality has been checked using standard protocols (stack and reciprocals) and proved to be
396 satisfactory (figure 9) for both surface and surface-to-borehole configurations. Water electrical conductivity
397 measurements were also conducted in the deep borehole along with bulk electrical resistivity measurements
398 (in-hole measurements using borehole electrodes).

399 **4.3. Results.** We will now examine the potential of electrical imaging in seawater intrusion studies under
400 field conditions. Figure 10 shows the result of the inversion of a dipole-dipole surface dataset (with a max-
401 imum inter-dipole spacing-number of 6) gathered along the 2km line. Based on an error analysis of the
402 datasets (figure 9), a resistance noise level of 6% was assumed for the inversion. However, tests have shown
403 that decreasing or increasing the noise level by a few per cent does not affect the inversion result in qualita-
404 tive terms. From figure 10a, we clearly see a low resistivity body forming a wedge in the subsurface from the
405 shoreline to 1.5 to 2 km inland. The shape of this feature clearly reflects a seawater intrusion, similar to the
406 Ghyben-Herzberg approximation. In order to convert the electrical image into mass fraction, we applied the
407 petrophysical relationships given in equations (3.6) and (3.9), using the experimentally determined parame-
408 ters σ_{ws} , σ_{wf} and F . The result of the conversion is shown in figure 10 b. Assuming that seawater intrusion
409 is defined by a mass fraction of chlorine higher than 0.1, we can see that we can identify an intrusion toe
410 at approximately 1.5 km inland. Figure 10b also reveals that a mass fraction of 1, corresponding to pure

411 seawater, is not retrieved. This must be attributed to a lack of resolution and/or inaccurate petrophysical
412 parameters. We also note the good correlation in figure 10b between the image mass fraction lines and the
413 water electrical conductivity logs in the boreholes.

414

415 To assess further the quality of the ERT result, we compared it with geological, water electrical conduc-
416 tivity, and bulk electrical resistivity logs in two boreholes at 320 and 820 m along the ERT section (figure
417 11. We plotted the vertical profiles of the resistivity distributions inverted from the surface data and from
418 the in-hole data at the positions of the boreholes. These ERT "logs" show a good correlation. Moreover,
419 they correlate well with the respective water electrical conductivity log, indicating that ERT images are
420 mainly sensitive to the amount of dissolved salt rather than to lithology. This suggests that the width of
421 the transition zone may be estimated at this site to a reasonable degree using surface ERT. However, figure
422 11 also shows that the surface ERT only resolves the larger-scale variation but fails to capture the meter-
423 scale variations (most likely related to lithological variations). Consequently, surface ERT may only be able
424 to constrain upscaled hydrological parameters or features. By comparing the formation factors computed
425 in-situ (using the water and bulk electrical resistivities from boreholes measurements; figure 11) and from
426 samples in the laboratory (figure 3), we can see that there is a relatively large discrepancy between the two,
427 which may explain the fact that the mass fraction does not reach a maximum value of 1 where seawater is
428 located (see above).

429

430 The permanent electrode array allowed us to test surface-to-borehole ERT; however, cross-hole ERT
431 surveys were not possible, due to the large distance between the boreholes relative to their depth. Figure 12
432 shows the image inverted from the surface-to-borehole (dipole-dipole) data (assumed resistance noise level of
433 4%; see figure 9). There is a general agreement between the surface and surface-to-borehole images; however,
434 the interpretation in terms of the shape of the intrusion front is not obvious for the surface-to-borehole
435 image. This may be attributed to the laterally much more heterogeneous sensitivity (resolution) pattern of
436 a surface-to-borehole survey compared to a surface survey (Tsourlos et al., 2004), making interpretations of

437 surface-to-borehole images more difficult. In addition, the lateral coverage of the surface-to-borehole ERT
438 survey is relatively weak given the overall extent of the seawater intrusion.

439

5. CONCLUSIONS AND PERSPECTIVES

440 In this work, we evaluated the potential of smoothness-constrained 2D electrical imaging to characterize
441 seawater intrusion. We used both synthetic data from hydrological simulations of seawater intrusion, field
442 data, and deep boreholes logs (geology, bulk resistivity, water electrical conductivity). The setup that we
443 adopted in the numerical study is representative of the Almeria field site; however our results and conclusions
444 may be extrapolated to coastal aquifer studies in general, if clays play a negligible part in the conduction of
445 electric current due to high salinity content. The latter should often be the case in arid to semi-arid coastal
446 aquifers where groundwater resources are vital to the population.

447 Our simulation results indicate that the subsurface resistivity distribution associated with saline intrusion
448 can be reasonably well recovered by ERT. However, mass fraction could only be accurately recovered up
449 to values of 0.2. The higher mass fractions at larger depths are only poorly resolved. The low cumulative
450 sensitivity values in these regions clearly indicate that the loss of resolution with depth (away from the
451 electrodes) accounts for the mismatch between recovered and true mass fractions. Nevertheless, if the
452 transition zone is entirely located in fairly sensitive (resolved) areas, it is possible to capture its extent.
453 Paradoxically, although we used a smoothness-constrained inversion scheme we were better able to recover
454 sharper (but still smooth) transition zones since these were located closer to the electrodes.

455 A clear recommendation that emerges from this study is that if one uses ERT images to constrain a hydro-
456 geological model (to calibrate dispersivity, to position the interface, to detect heterogeneity, or to determine
457 the width of the transition zone), one must not make use of the entire image plane but only consider, on the
458 basis of image appraisal tools, reliable regions. We here employed an empirical approach to determine at
459 which sensitivity value the correlation between the targeted mass fraction and the ERT-recovered one breaks
460 down, based on synthetic modelling. We then used this value as a threshold to restrain the interpretation of
461 the ERT image at a certain location. This is one approach and other alternatives exist, such as the use of

462 the resolution matrix or the comparison with ground truth data (if available).

463

464 We tested ERT at the site of Almeria and assessed its potential using ground truth information. ERT
465 detected the presence of seawater intrusion and delineated its shape, forming a wedge whose toe reaches
466 approximately 1.5 km inland. This is confirmed by borehole data. Interestingly, we were not able to recover
467 a mass fraction of 1, although boreholes indicated pure seawater. Comparison between in-situ and labora-
468 tory formation factors indicates a discrepancy which could partly explain this. We believe that this is not
469 an issue of limited depth resolution of surface ERT, since the extracted surface ERT and in-hole ERT (not
470 suffering from a loss of resolution with depth) bulk resistivity logs show an excellent agreement in terms of
471 the overall behaviour as dominated by chloride concentration. Moreover, the ERT "logs" also correlate well
472 with the water electrical conductivity logs, indicating that ERT images are mainly sensitive to the amount
473 of dissolved salt (large-scale variations) rather than to lithology (small-scale variations). The log compari-
474 son further showed that we cannot expect to resolve meter-scale variations at depth and that surface ERT
475 may only be able to constrain upscaled hydrological parameters or features. The upscaling factor may be
476 determined using image appraisal tools.

477

478 The surface-to-borehole imaging result suggests that the laterally heterogeneous resolution pattern of
479 this setup may dominate the image characteristics compared to the laterally more homogeneous resolution
480 pattern of surface ERT, complicating - in addition to only little lateral coverage - reliable interpretations in
481 terms of the shape of the intrusion front.

482

483 We are continuing to study the use of ERT images as data for constraining the hydrological inversion
484 of seawater intrusion model parameters. This will in particular require the inclusion of the petrophysical
485 parameter relationships in the inversion process.

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638	1	Parameters for the density-dependent flow and transport model.	32
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- 640 1 Ghyben-Herzberg analytical seawater intrusion model (Bear et al., 1999; Guo and Jiao, 2007) with
641 (a) higher and (b) lower inflow q . h is the water head above sea level. Note that in real situation,
642 the interface is diffuse and not sharp as represented here for simplicity. 33
- 643 2 Domain geometry and boundary conditions (BC) for the finite-difference seawater intrusion model
644 (not to scale). Pressure-depth curves are shown on both sea and inland sides. 34
- 645 3 Petrophysical relationships between bulk electrical conductivity (σ_b) and water conductivity (σ_w)
646 for different samples of the Almeria test site. F' denotes the effective formation factor computed
647 from fitting the data using equation (3.9) (neglecting the surface conduction term b). 35
- 648 4 Conversion from simulated mass fraction distribution (a) to electrical resistivity model (c) using
649 the petrophysical relationships given in equations (3.6) and (3.9). ERT imaging results (d) and
650 ERT-derived mass fraction (b) (using equations (3.6) and (3.9)). 36
- 651 5 Illustration of the behavior of ERT-derived mass fraction versus true mass fraction (bottom) with
652 increasing depth. The two bottom plots were extracted from boxes A and B in the upper image,
653 showing the spatial variation of the cumulative sensitivity as defined by equation 2.5. 37
- 654 6 Comparison between true and ERT recovered mass fraction isolines for different hydrogeological
655 simulations where the transverse dispersivity varies. 38
- 656 7 Influence of the cumulative sensitivity distribution on the true (dashed red line) and recovered
657 (solid red line) mass fraction 0.01 isoline for a layered model (the layer is indicated by the solid
658 black lines). The sensitivity is clipped for values smaller than $10e-5$ of the maximum value. The
659 difference in sensitivity arises from different electrical contrasts corresponding to brackish (top) and
660 fresh (bottom) upstream water. 39
- 661 8 Spain is shown on the upper left map, the Province of Andalusia (where the field site is located)
662 is highlighted. The image map shows the location of the Andarax river bed (blue line), where the
663 ERT profile is located and positions of the deep boreholes B320, B520 and B1320. The borehole

664 numbers correspond to the horizontal distance along the ERT profile. Two pictures (right) show
 665 the installation of the electrodes on the PVC casing of the deep boreholes. 40

666 9 Histograms of relative resistance errors for the manual surface dataset (inferred from stacking (dark
 667 blue) and normal-reciprocal analysis (red)) and the permanent surface-to-hole dataset (inferred from
 668 stacking (light blue)). The inner box shows that the relative histograms for the surface datasets are
 669 similar. 41

670 10 Field ERT results (a) and derived mass fraction distribution (b). The black lines graphs in figure b
 671 show the normalized water electrical conductivity logs at scale for boreholes B320, B820, and B1320
 672 located on figure 8. The dashed line indicates the separation between two domains of different
 673 Archie's formation factors (see figure 3). 42

674 11 Wells B320 (a) and B820 (b) information (see figure 8). For each subplots, from left to right: in situ
 675 formation factors computed from in-hole bulk resistivity and water electrical resistivity (dot-dash
 676 line correspond the laboratory formation factor; see figure 3), water electrical resistivity, in-hole
 677 ERT (green line) and surface ERT (red dots) bulk electrical resistivity logs, and geology. 43

678 12 Surface imaging result (b) and surface-to-borehole imaging result (a) at the site of Almeria. The
 679 red rectangle in (a) shows the position of the surface-to-borehole image. 44

TABLE 1. Parameters for the density-dependent flow and transport model.

Parameter	Value
Horizontal hydraulic conductivity K_h	100 m/d
Vertical hydraulic conductivity K_v	1 m/d
Permeability k	1.2e-10 m^2
Effective porosity ϵ	0.30
Horizontal dispersivity α_L	10 m
Transversal dispersivity α_T	0.5 m
Freshwater density ρ_f	1000 kg/m^3
Seawater density ρ_s	1025 kg/m^3
Seawater mass fraction ω	0.0357
Diffusion coefficient D_m	0 m^2/s
Viscosity μ	1.0e-3 kg/ms
Fluid compressibility	0 kg/ms^2
Matrix compressibility	0 kg/ms^2

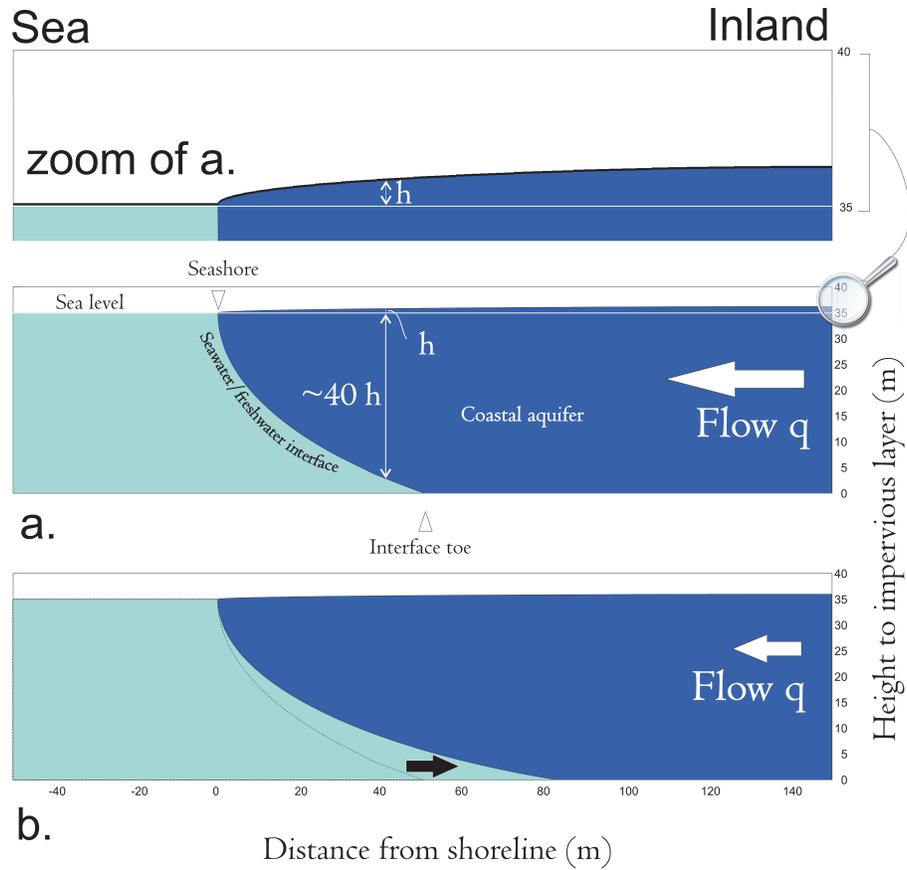


FIGURE 1. Ghyben-Herzberg analytical seawater intrusion model (Bear et al., 1999; Guo and Jiao, 2007) with (a) higher and (b) lower inflow q . h is the water head above sea level. Note that in real situation, the interface is diffuse and not sharp as represented here for simplicity.

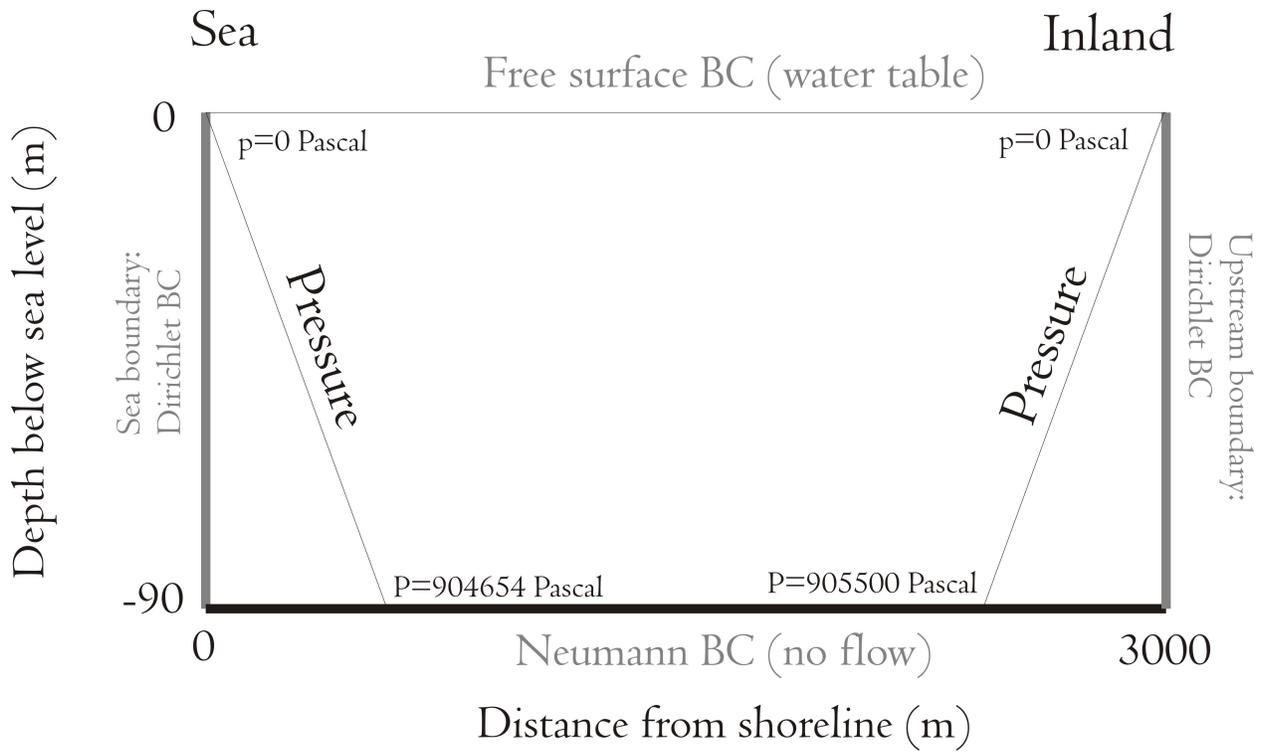


FIGURE 2. Domain geometry and boundary conditions (BC) for the finite-difference seawater intrusion model (not to scale). Pressure-depth curves are shown on both sea and inland sides.

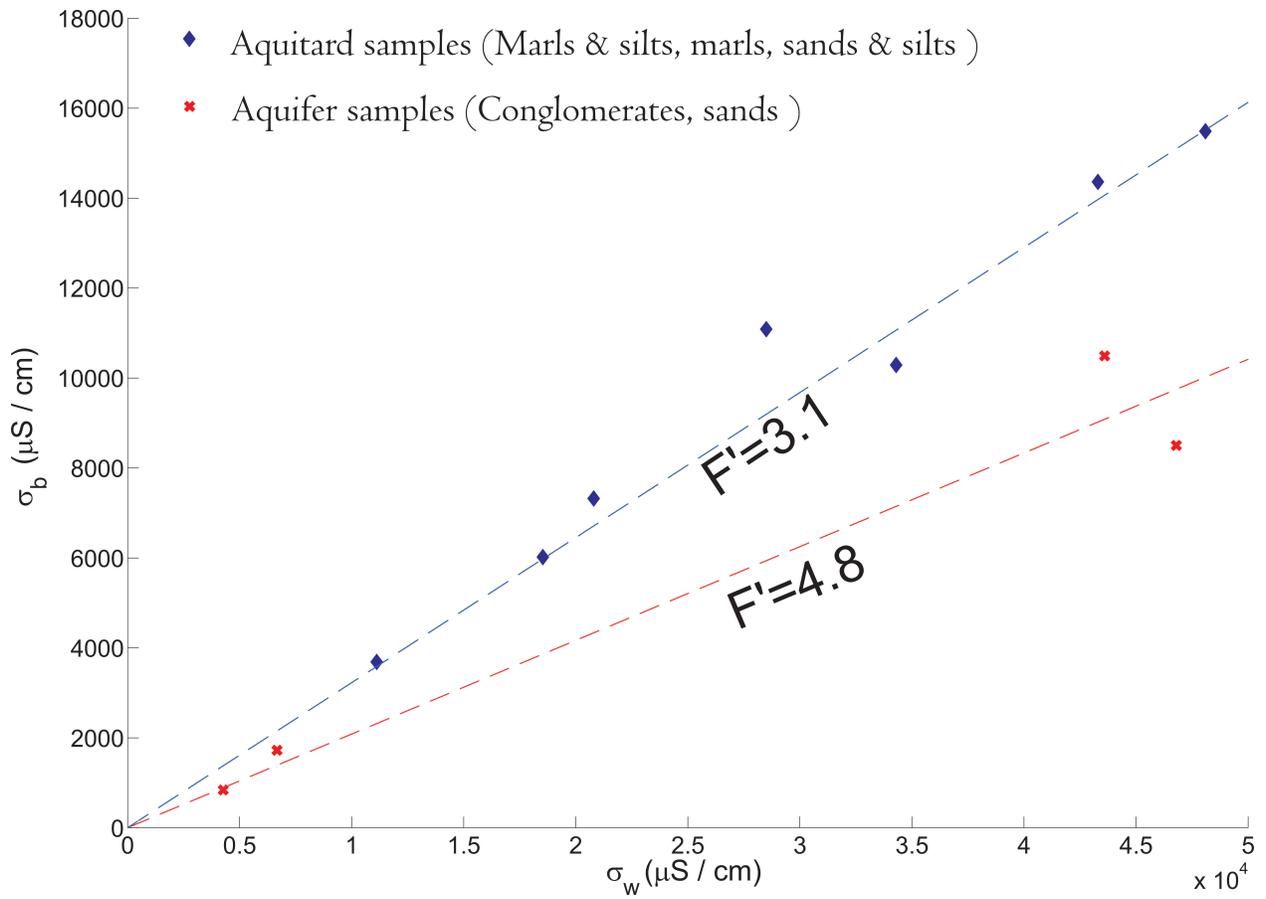


FIGURE 3. Petrophysical relationships between bulk electrical conductivity (σ_b) and water conductivity (σ_w) for different samples of the Almeria test site. F' denotes the effective formation factor computed from fitting the data using equation (3.9) (neglecting the surface conduction term b).

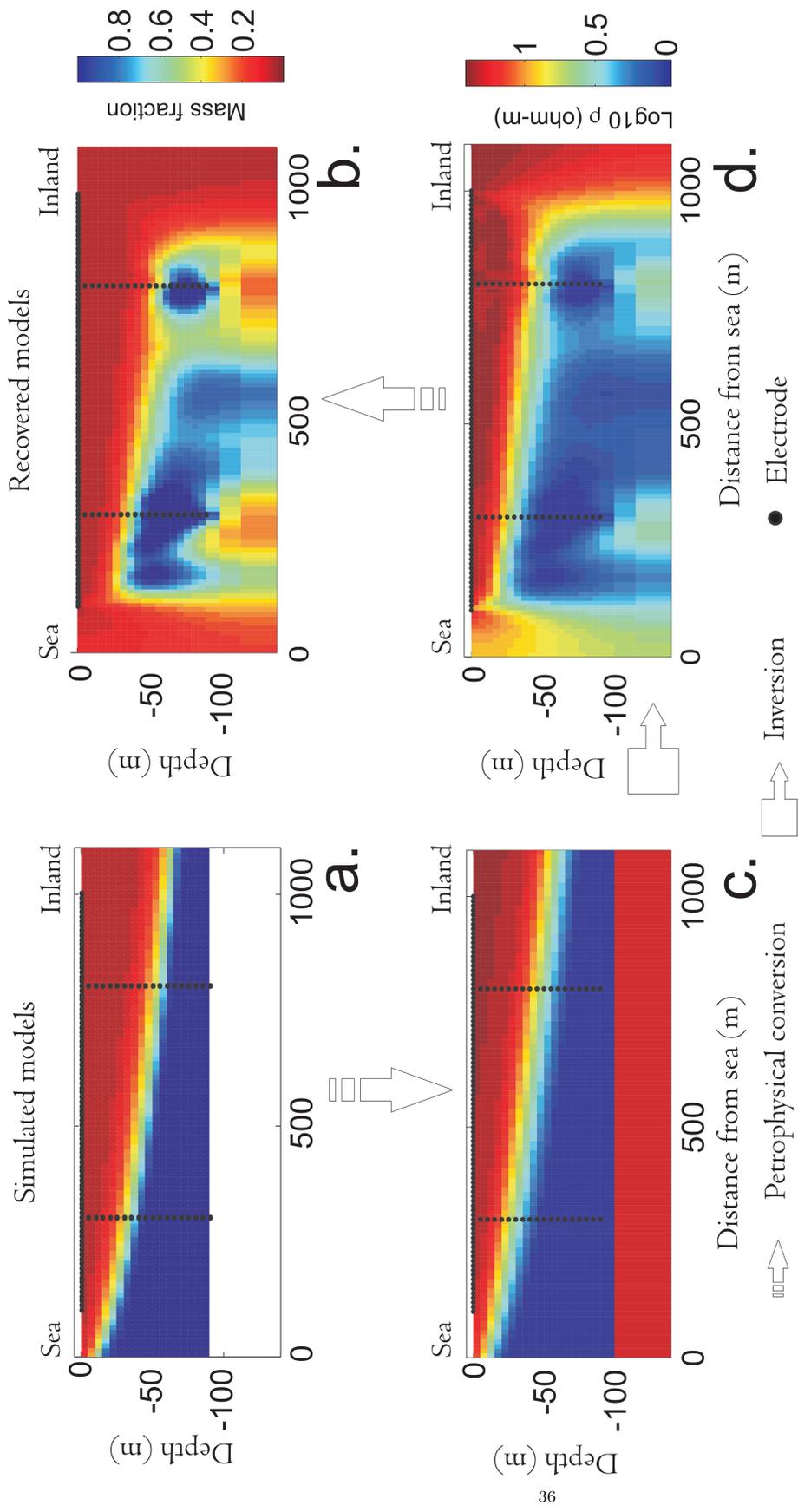


FIGURE 4. Conversion from simulated mass fraction distribution (a) to electrical resistivity model (c) using the petrophysical relationships given in equations (3.6) and (3.9). ERT imaging results (d) and ERT-derived mass fraction (b) using equations (3.6) and (3.9).

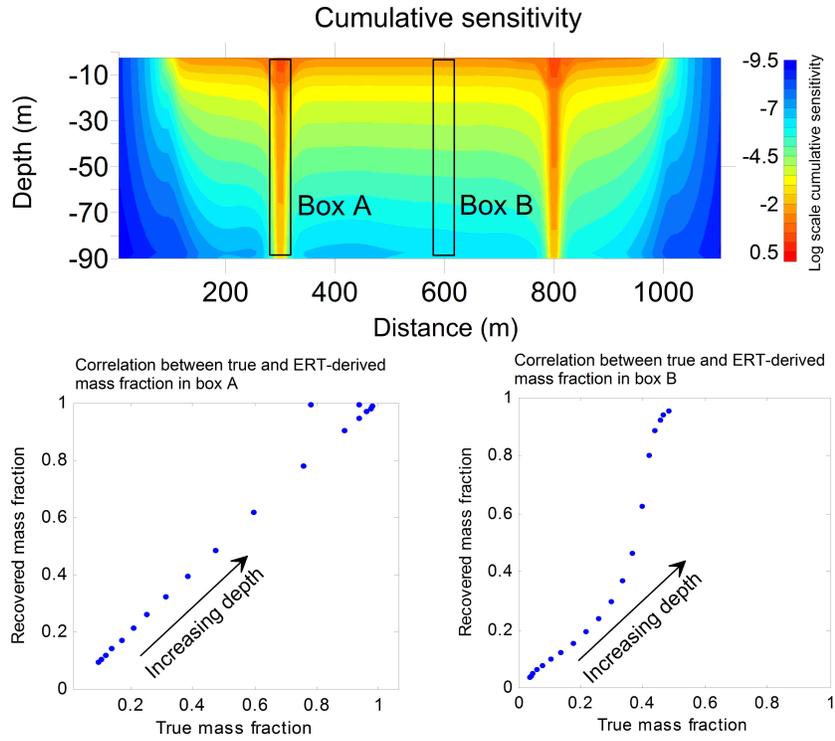


FIGURE 5. Illustration of the behavior of ERT-derived mass fraction versus true mass fraction (bottom) with increasing depth. The two bottom plots were extracted from boxes A and B in the upper image, showing the spatial variation of the cumulative sensitivity as defined by equation 2.5.

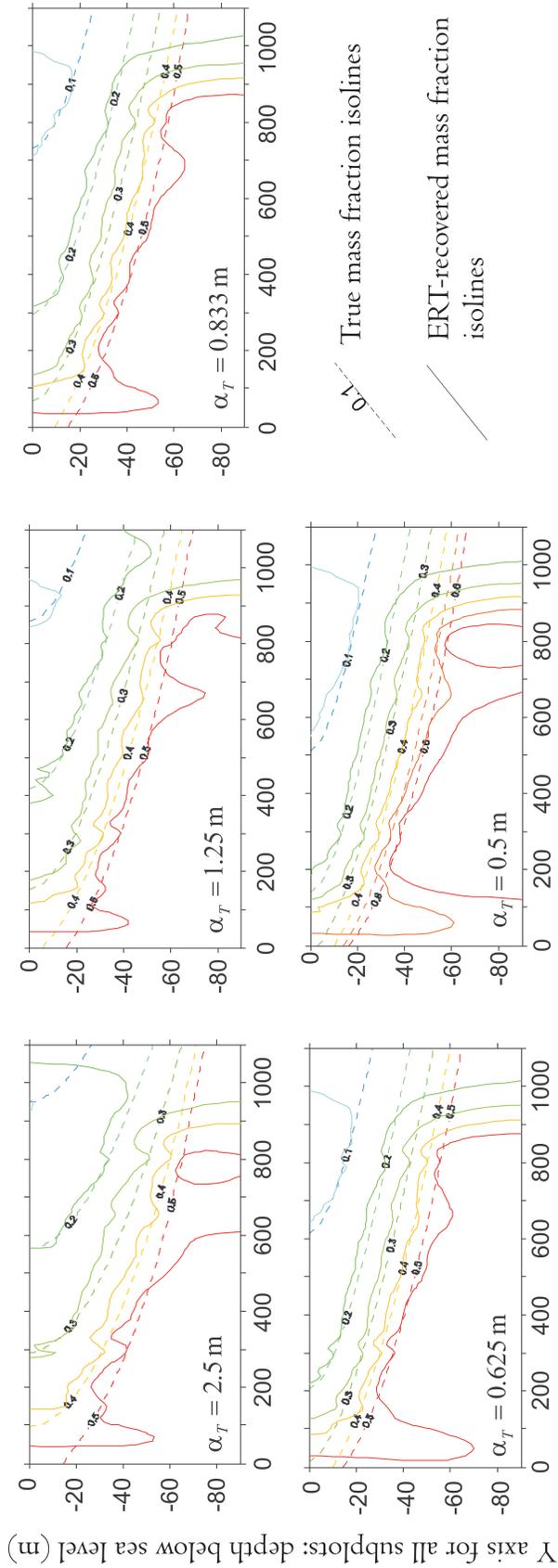


FIGURE 6. Comparison between true and ERT recovered mass fraction isolines for different hydrogeological simulations where the transverse dispersivity varies.

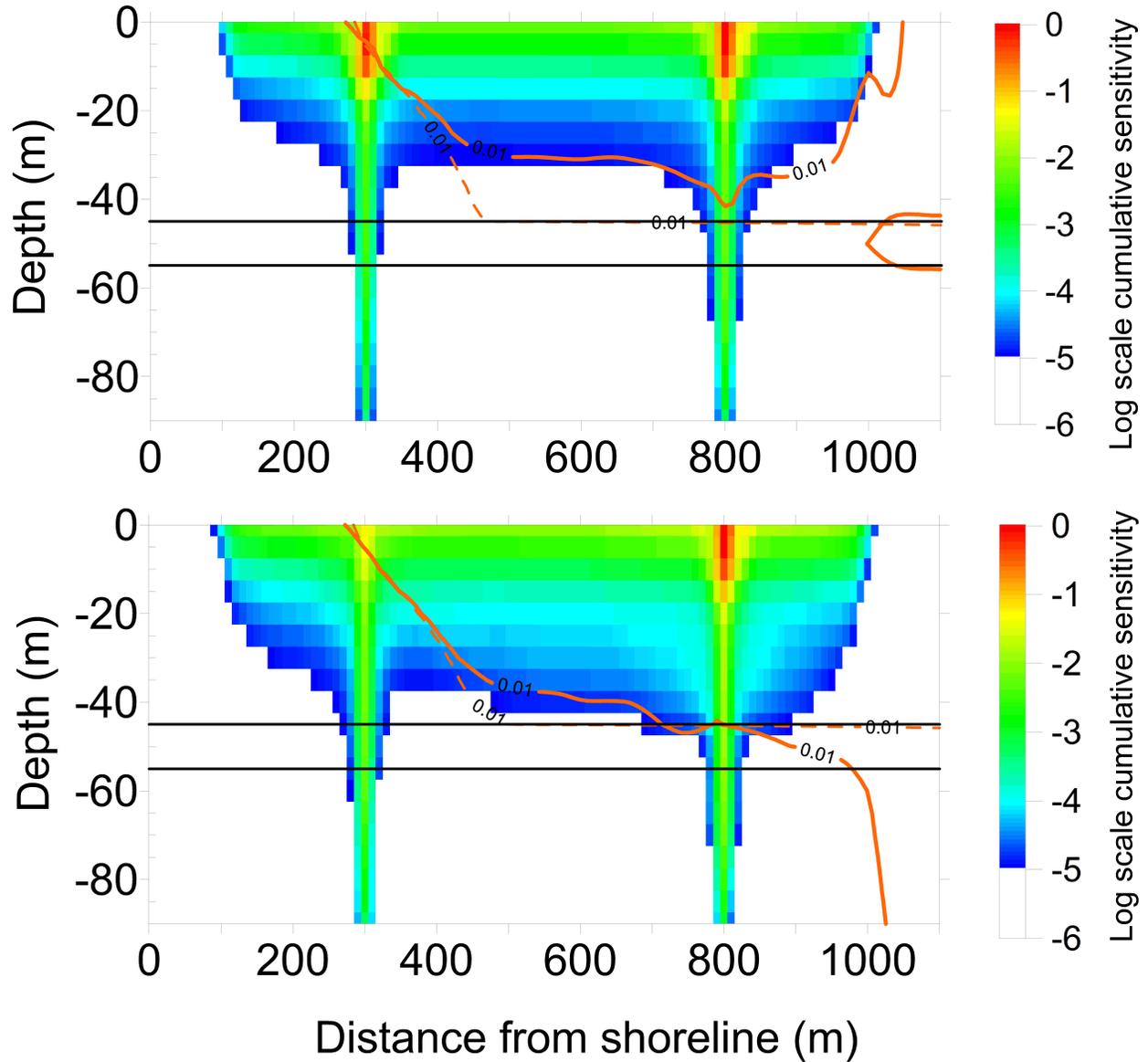


FIGURE 7. Influence of the cumulative sensitivity distribution on the true (dashed red line) and recovered (solid red line) mass fraction 0.01 isoline for a layered model (the layer is indicated by the solid black lines). The sensitivity is clipped for values smaller than $10e-5$ of the maximum value. The difference in sensitivity arises from different electrical contrasts corresponding to brackish (top) and fresh (bottom) upstream water.

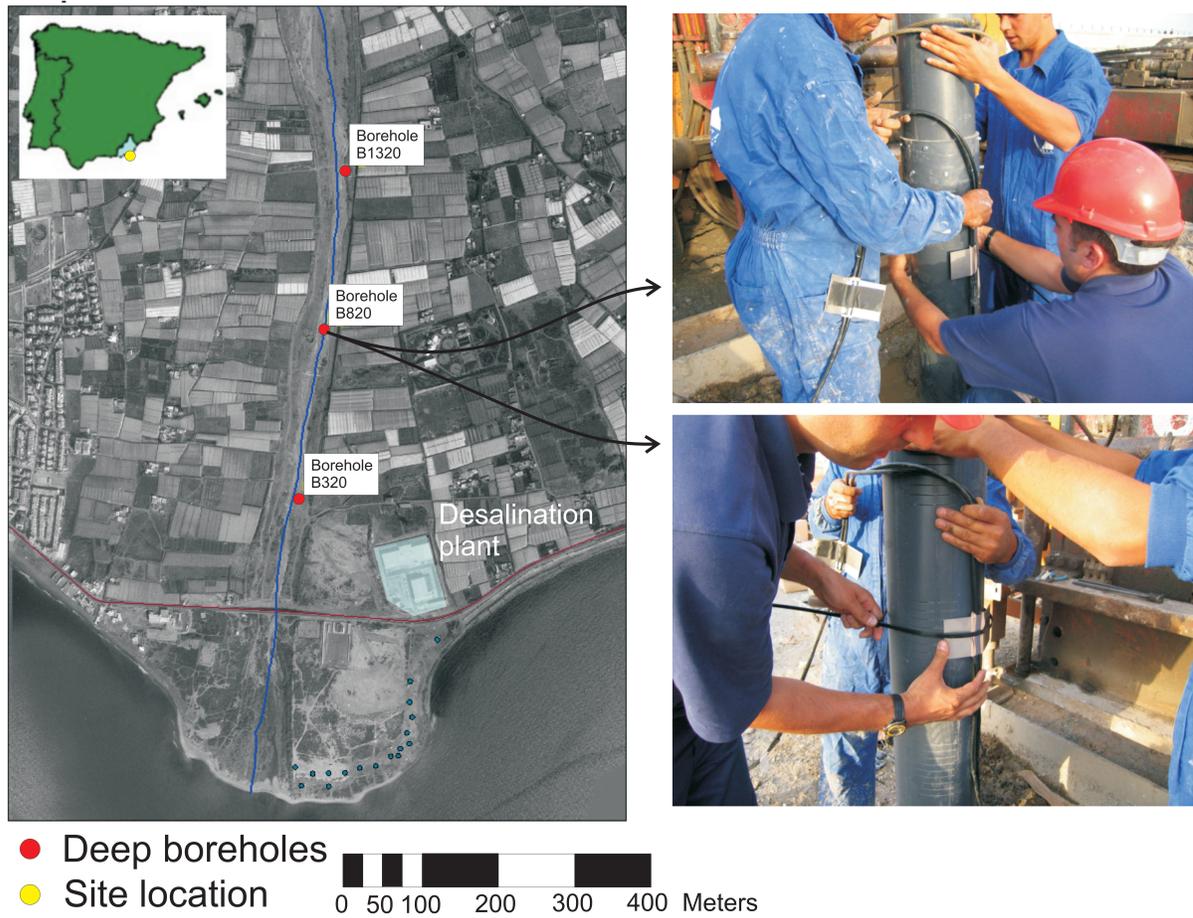


FIGURE 8. Spain is shown on the upper left map, the Province of Andalusia (where the field site is located) is highlighted. The image map shows the location of the Andarax river bed (blue line), where the ERT profile is located and positions of the deep boreholes B320, B520 and B1320. The borehole numbers correspond to the horizontal distance along the ERT profile. Two pictures (right) show the installation of the electrodes on the PVC casing of the deep boreholes.

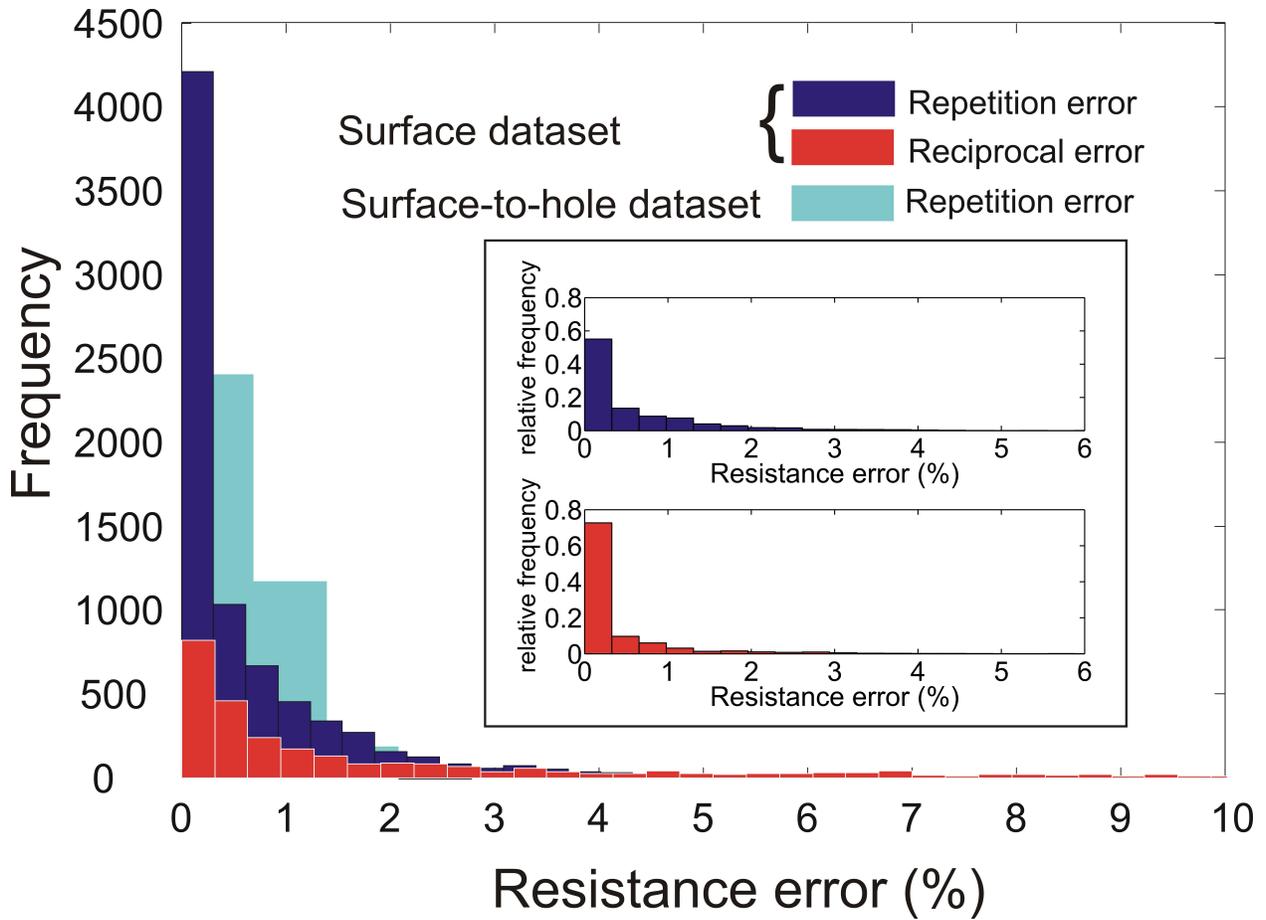


FIGURE 9. Histograms of relative resistance errors for the manual surface dataset (inferred from stacking (dark blue) and normal-reciprocal analysis (red)) and the permanent surface-to-hole dataset (inferred from stacking (light blue)). The inner box shows that the relative histograms for the surface datasets are similar.

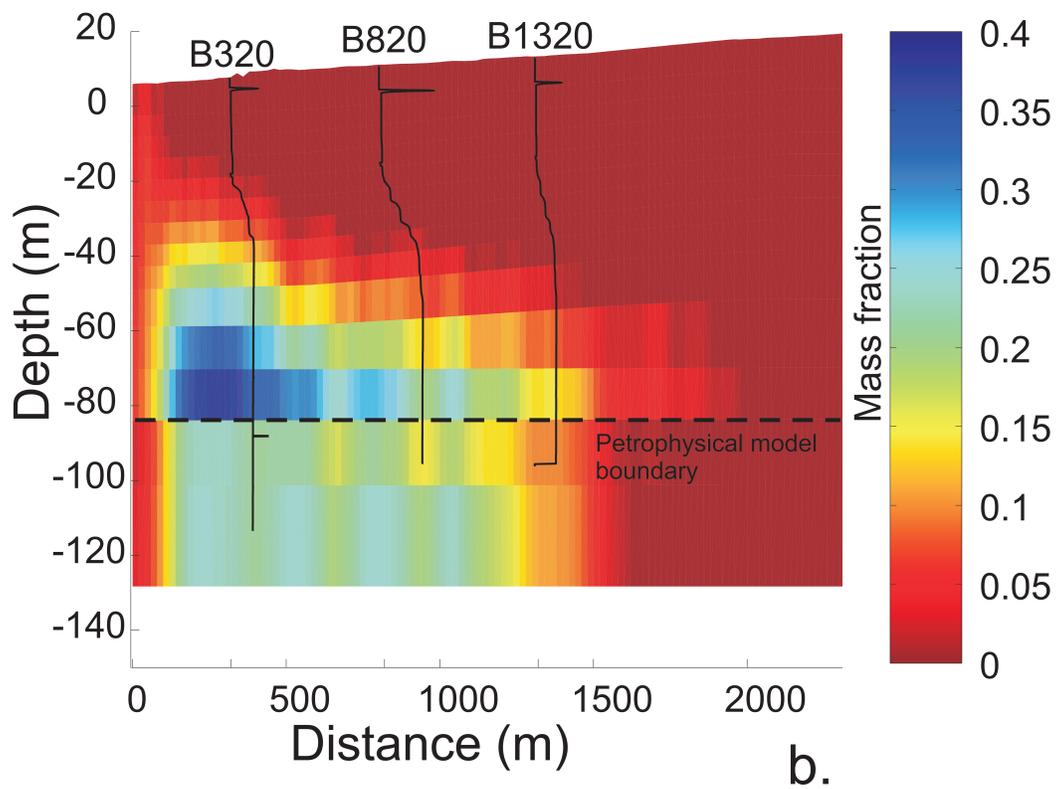
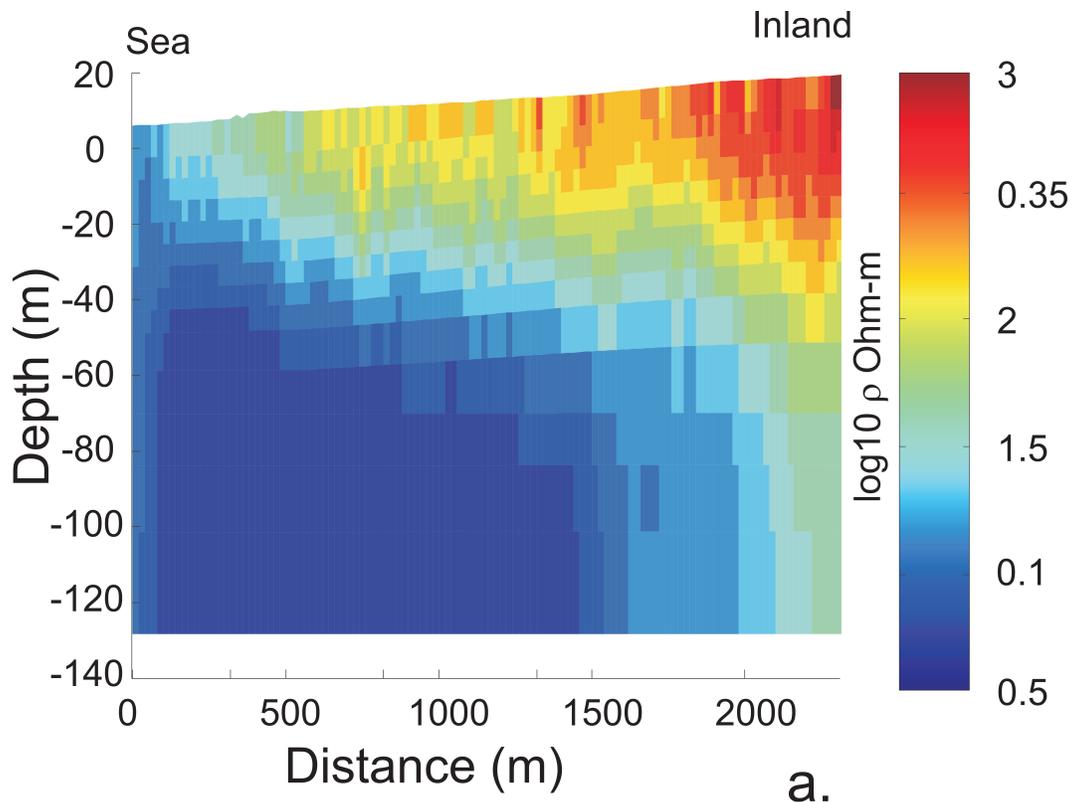


FIGURE 10. Field ERT results (a) and derived mass fraction distribution (b). The black lines graphs in figure b show the normalized water electrical conductivity logs at scale for boreholes B320, B820, and B1320 located on figure 8. The dashed line indicates the separation between two domains of different Archie's formation factors (see figure 3).

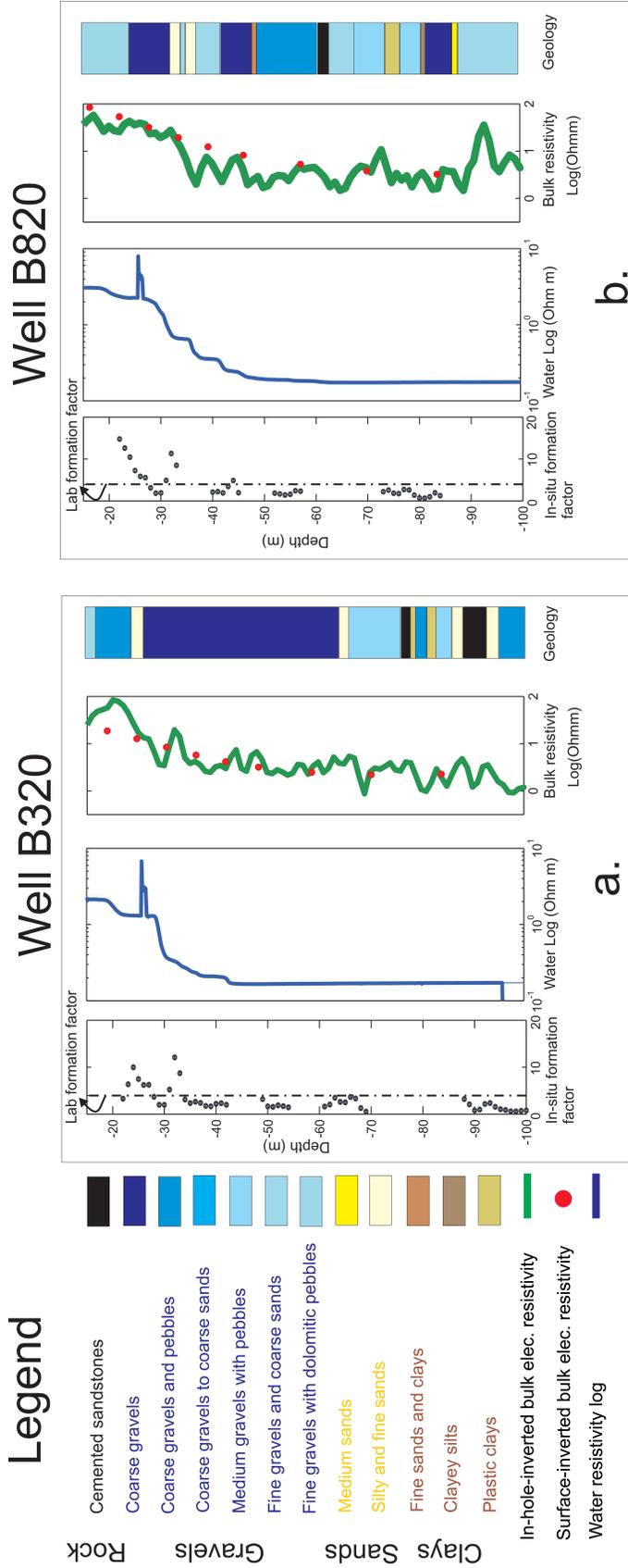


FIGURE 11. Wells B320 (a) and B820 (b) information (see figure 8). For each subplots, from left to right: in situ formation factors computed from in-hole bulk resistivity and water electrical resistivity (dot-dash line correspond the laboratory formation factor; see figure 3), water electrical resistivity, in-hole ERT (green line) and surface ERT (red dots) bulk electrical resistivity logs, and geology.

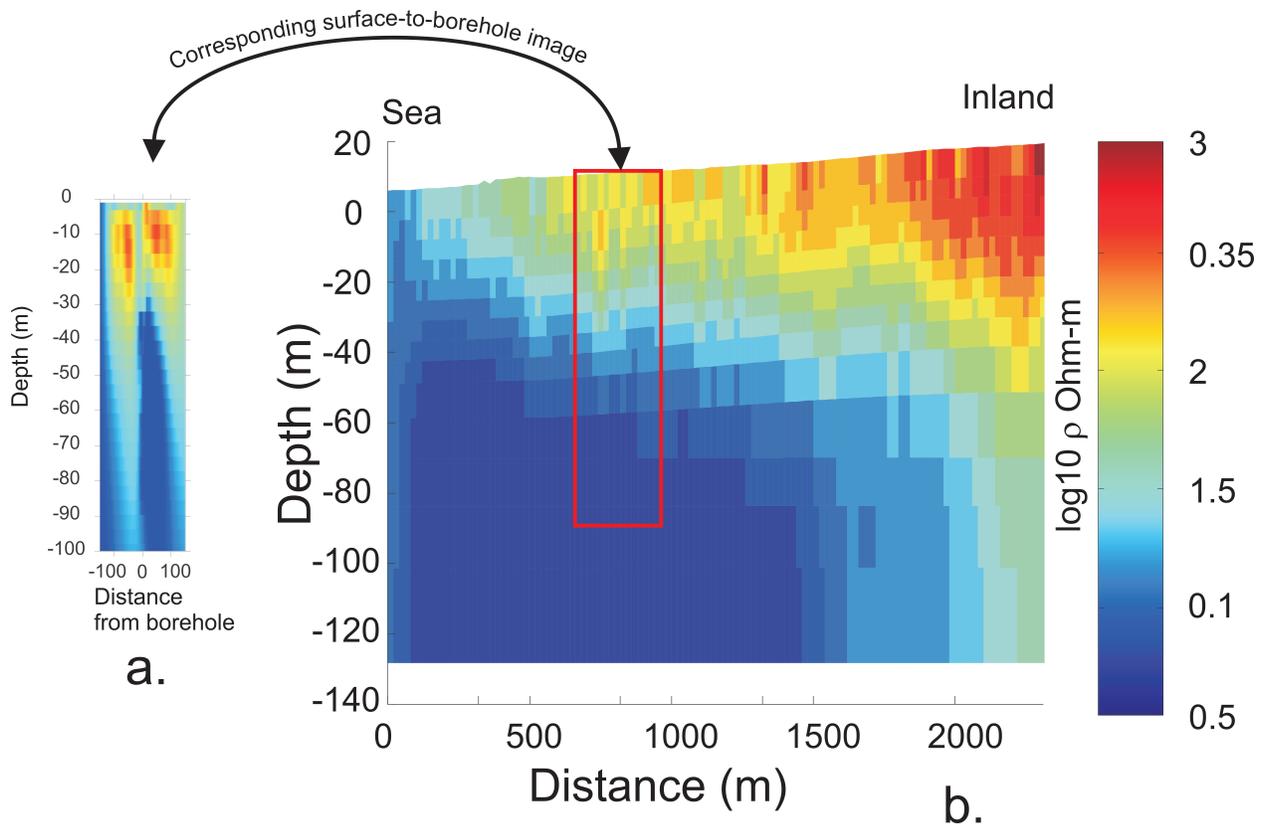


FIGURE 12. Surface imaging result (b) and surface-to-borehole imaging result (a) at the site of Almeria. The red rectangle in (a) shows the position of the surface-to-borehole image.