

1 Borehole seismoelectric logging using a shear-wave source: Possible application to  
2 CO<sub>2</sub> disposal?

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11 **Abstract**

12 The behaviour of CO<sub>2</sub> deposition sites -and their surroundings- during and after carbon dioxide injection has  
13 been matter of study for several years, and several geophysical prospection techniques like surface and crosshole  
14 seismics, geoelectrics, controlled source electromagnetics among others, have been applied to characterize  
15 the behaviour of the gas in the reservoirs. Until now, Seismoelectromagnetic wave conversions occurring in  
16 poroelastic media via electrokinetic coupling have not been tested for this purpose. In this work, by means  
17 of numerical experiments using Pride's equations -extended to deal with partial saturations- we show that  
18 the seismoelectric and seismomagnetic interface responses (IR) generated at boundaries of a layer containing  
19 carbon dioxide are sensitive to its CO<sub>2</sub> content. Further, modeling shear wave sources in surface to borehole  
20 seismoelectric layouts and employing two different models for the saturation dependence of the electrokinetic  
21 coefficient, we observe that the IR are sensitive to CO<sub>2</sub> saturations ranging between 10% and 90%, and that the  
22 CO<sub>2</sub> saturation at which the IR maxima are reached depends on the aforementioned models. Moreover, the IR  
23 are still sensitive to different CO<sub>2</sub> saturations for a sealed CO<sub>2</sub> reservoir covered by a clay layer. These results,  
24 which should be complemented by the analysis of the IR absolute amplitude, could lead, once confirmed on the  
25 field, to a new monitoring tool complementing existing ones.

26 **Keywords:** Permeability and porosity, Numerical approximations and analysis, Wave propagation, Acoustic

28 **1. Introduction**

29 Injection of large amounts of man-produced CO<sub>2</sub> in depleted oil wells below the sea floor and in other  
30 appropriate geological formations has been used, for several years, as a means of reducing the carbon dioxide  
31 emissions into the atmosphere. For example, CO<sub>2</sub> is being injected in the Sleipner field in the North Sea since  
32 1996 at a rate of 0.85 Mt per year (Ellis, 2010), and also beneath the Sahara desert, at In Salah in Algeria  
33 (Ringrose et al., 2009). The former has been a subject of extensive theoretical and experimental studies, includ-  
34 ing laboratory rock sample analysis, seismic monitoring, etc. We mention, from the large literature concerning  
35 this deposition site, the studies of Chadwick et al. (2009, 2010) where time-lapse seismic is employed to char-  
36 acterize CO<sub>2</sub> plume development, and the studies of Gomez and Ravazzoli (2011), where CO<sub>2</sub> content related  
37 to seismic attributes were investigated. Moreover, a test site in Ketzin, Germany, is being run and extensively  
38 studied in order to monitor the CO<sub>2</sub> behaviour during injection and afterwards, see Martens et al. (2012, 2013)  
39 and references therein. Scientists from different areas have been studying this topic, and a still open problem  
40 is to predict the behaviour of the gas once set into the reservoir. Will it remain stable? Will it migrate, and  
41 make its way back to the surface? How the stored CO<sub>2</sub> can be efficiently monitored in order to avoid pollution  
42 of overlying aquifers by leaked gas, among other issues (Thibeau and Mucha, 2011) is still a topic of intense  
43 research.

44  
45 Among other works implemented at Ketzin, Wiese et al. (2010) studied the hydraulic properties of the stor-  
46 age reservoir, Kazemeini et al. (2010) carried out some rock physics and seismic modeling studies of surface  
47 seismic CO<sub>2</sub> monitoring, and cross-well seismic tomography has been also performed (Zhang et al., 2012);  
48 more recently Fischer et al. (2013) made laboratory studies of geochemical changes induced in Ketzin rock  
49 matrix samples by the presence of the stored carbon dioxide, and Wiese et al. (2013) studied -at the same site-  
50 not only the geochemical but also the hydraulic changes induced in the overburden by deposited CO<sub>2</sub>. We  
51 can also mention that both seismic and electric methods are potentially appropriate to study the CO<sub>2</sub> reservoir

52 (Fabriol et al., 2011; Girard et al., 2011; Carcione et al., 2012). Martens et al. (2012) describe not only the  
53 results of different campaigns including seismic, surface and borehole monitoring, but also some seismic simu-  
54 lation runs in order to check previous models; on the other hand synthetic and field geoelectrical methods were  
55 applied to study possible gas migration (Kiessling et al., 2010). Moreover Ishido et al. (2013) have numerically  
56 investigated the application of self potential methods to monitor the migration of CO<sub>2</sub> sequestrated into saline  
57 aquifers, concluding that the used methods are effective for sensing the approach of CO<sub>2</sub> to the well casings  
58 deep within the subsurface. We finally point out that in recent studies it was shown that seismics was useful to  
59 detect CO<sub>2</sub> saturation below 15% and that electrical resistivity was useful to detect CO<sub>2</sub> saturation above 15%  
60 (Kim et al., 2013).

61  
62 Seismoelectric signals are electrokinetically generated by the propagation of seismic waves within a porous  
63 material. They can be recorded using a seismic source and electric receivers. The seismoelectric strategy aims  
64 to combine the resolution of the seismics to the sensitivity of the electric methods to fluid content. A specific  
65 seismoelectric signal, denoted the interfacial response, is expected to be induced at contrasts between rock  
66 properties (Garambois and Dietrich, 2002), including different fluids and different fluid-contents. This signal is  
67 usually weak compared to the so-called coseismic signal, which is the seismo-electric signal travelling within  
68 the seismic wave directly induced by the source. Several authors have investigated the benefits of surface-to-  
69 borehole seismoelectric layouts to accomplish efficient measurements of the interfacial response, as opposed to  
70 layouts for which both the seismic source and the receiving electrodes are laid at the surface.

71 The aim of this work is to provide numerical evidence that borehole seismoelectrics can discern carbon  
72 dioxide concentrations in a broader range than seismics allow, detecting at the same time salinity contrasts, task  
73 up to now fulfilled by geoelectrics. The pure SH seismic source considered in the present study could achieve a  
74 better resolution than the one obtained through the usual P-driven experiments because of shorter wavelengths.  
75 We start our work by reviewing the most important theoretical concepts of seismoelectrics, and by proposing  
76 a possible appropriate field experimental setup. We follow by analyzing shear-wave driven interface responses  
77 generated between two consecutive units saturated with water, using a one dimensional finite element method  
78 to approximate the solution to Pride's equations. We study the sensitivity of these responses to contrasts in

79 relevant parameters, such as porosity, salinity and viscosity; and continue by investigating the coseismic waves  
80 and interface response amplitudes of tabular media when one layer is partially saturated with carbon dioxide,  
81 employing in this analysis different models to take into account this situation in the electrokinetic coupling.  
82 Finally, we consider a layered model including a seal layer, in order to simulate a realistic CO<sub>2</sub> deposition site.

## 83 **2. Theoretical background**

84 The seismoelectric method relies on electrokinetically induced seismic-to-electric energy conversions oc-  
85 ccurring in fluid-containing porous media. The reader can find a tutorial on electrokinetics in Jouniaux and  
86 Ishido (2012).

### 87 *2.1. Theoretical aspects*

88 When a compressional wave travels through a porous medium, it creates a fluid-pressure gradient and an  
89 acceleration of the solid matrix, inducing a relative motion between the immobile ions adsorbed at the grain  
90 surface and the counter-ions in the diffuse layer. This charge separation at the scale of the seismic wavelet  
91 creates an electrical potential difference known as the streaming potential. The electric field arising from  
92 this potential is known as the coseismic wave, as it travels within the passing compressional seismic waves.  
93 Therefore coseismic electric fields do not extend outside the seismic waves creating them, and may only help  
94 characterize the medium near the receivers. For borehole seismoelectric measurements they give information  
95 about the medium in the vicinity of the well (Mikhailov et al., 2000).

96 Another type of seismoelectric conversions arises when a seismic wave crosses a contrast between mechanical  
97 or electrical properties (Haartsen and Pride, 1997; Chen and Mu, 2005; Block and Harris, 2006). In this situation  
98 a transient localized charge separation across the interface is created, which acts as a secondary source that can  
99 be approximated as an electrical dipole oscillating at the center of the first Fresnel zone (Thompson and Gist,  
100 1993; Garambois and Dietrich, 2002). The resulting electromagnetic (EM) wave is known as the interface  
101 response (IR), and diffuses independently from the seismic wavefield: the velocity at which it travels is several  
102 orders of magnitude greater than seismic velocities. This IR may provide information about the contrasts in the  
103 medium's properties at depth.

104 The equations governing the coupled seismic and electromagnetic wave propagation in fluid-filled porous media  
 105 were derived by Pride (1994) by combining Maxwell's equations with Biot's equations for poroelasticity (Biot,  
 106 1956a,b). Two coupled transport equations were derived (Eq.251 and 252 in Pride (1994)):

$$107 \quad \mathbf{J} = \sigma(\omega)\mathbf{E} + L(\omega) (-\nabla p + \omega^2 \rho_w \mathbf{u}_s) \quad (1)$$

$$108 \quad -i\omega \mathbf{u}_f = L(\omega)\mathbf{E} + \frac{k(\omega)}{\eta_w} (-\nabla p + \omega^2 \rho_w \mathbf{u}_s) \quad (2)$$

109  
 110 The macroscopic electrical current density  $\mathbf{J}$  [A/m<sup>2</sup>] is given in Eq.1 as the sum of the average conduction and  
 111 streaming current densities, respectively the first and second term of its right-hand side. Both the above equa-  
 112 tions assume a  $e^{-i\omega t}$  time dependence of the propagating wave, where  $\omega$  [rad/s] denotes the angular frequency.  
 113 The parameter  $\mathbf{E}$  [V/m] denotes the electric field and  $\sigma(\omega)$  [S/m] is the frequency-dependent conductivity of the  
 114 material. Streaming currents may be induced by both the pressure gradient  $-\nabla p$ , where  $p$  [Pa] is the pore-fluid  
 115 pressure, and the acceleration of the solid frame  $\omega^2 \rho_w \mathbf{u}_s$ , where  $\rho_w$  [kg/m<sup>3</sup>] is the density of the fluid (water)  
 116 and  $\mathbf{u}_s$  [m] denotes the solid displacement. The fluid velocity  $-i\omega \mathbf{u}_f$  [m/s] is written in Eq.2 as the sum of  
 117 electrically and mechanically induced contributions. The frequency-dependent permeability is written as  $k(\omega)$   
 118 [m<sup>2</sup>] and the dynamic viscosity of the fluid is expressed as  $\eta_w$  [Pa.s]. The complex and frequency-dependent  
 119 coupling  $L(\omega)$  links Eq.1 and Eq.2:

$$120 \quad L(\omega) = L_0 \left[ 1 - i \frac{\omega}{\omega_t} \frac{b}{4} \left( 1 - 2 \frac{d}{\Lambda} \right)^2 \left( 1 - i^{3/2} d \sqrt{\frac{\omega \rho_w}{\eta_w}} \right)^2 \right]^{-\frac{1}{2}} \quad (3)$$

121 In Eq.3,  $\Lambda$  [m] is a pore geometrical parameter, defined in Johnson et al. (1987), whereas  $b$  is a dimensionless  
 122 parameter defined in terms of the latter, the porosity  $\phi$ , the absolute permeability  $k_0$  and the tortuosity  $\alpha_\infty$  as  
 123  $b = \frac{\phi}{\alpha_\infty k_0} \Lambda^2$  and consisting only of the pore-space geometry terms. This parameter  $b$  was originally denoted  $m$  in  
 124 Pride (1994). When  $k_0$ ,  $\phi$ ,  $\alpha_\infty$  and  $\Lambda$  are independently measured,  $b$  is comprised between 4 and 8 for a variety  
 125 of porous media ranging from grain packing to capillary networks consisting of tubes of variable radii (Johnson  
 126 et al., 1987). The parameter  $d$  [m] denotes the Debye length, while  $\omega_t$  [rad/s] is the permeability-dependent  
 127 transition angular frequency between the low-frequency viscous flow and high-frequency inertial flow. Finally,  
 128  $L_0$  denotes the electrokinetic coupling which expression we give below. The coupling  $L(\omega)$  was studied by

129 Reppert et al. (2001), Schoemaker et al. (2007), Jouniaux and Bordes (2012) and Glover et al. (2012). When  
 130 this coefficient is set to zero, the two subsets of equations describing the behavior of EM and seismic waves  
 131 are decoupled. Different modellings have been developed to resolve the seismoelectric conversions, see for  
 132 example (Guan et al., 2013; Schakel et al., 2012, 2011; Gao and Hu, 2010; Guan and Hu, 2008).

## 133 2.2. Transfer functions

134 The displacement and EM fields in an isotropic and homogeneous wholespace were derived by Pride and  
 135 Haartsen (1996) using a plane-wave solution of the governing equations. Later, Garambois and Dietrich (2001)  
 136 making use of these results, derived the electric and magnetic fields  $\mathbf{E}$  and  $\mathbf{H}$  as a function of the seismic  
 137 displacement  $\mathbf{u}$ . They demonstrated that low-frequency approximations of these relationships lead to a seismo-  
 138 electric field  $\mathbf{E}$  proportional to the grain acceleration  $\ddot{\mathbf{u}}$  associated to longitudinal fast  $P$ -waves as:

$$139 \quad \mathbf{E} \simeq \frac{\epsilon_0 \kappa_w \zeta}{\eta_w \sigma_w} \rho_w \left( 1 - \frac{\rho}{\rho_w} \frac{C}{H} \right) \ddot{\mathbf{u}} = C_K \rho_w \left( 1 - \frac{\rho}{\rho_w} \frac{C}{H} \right) \ddot{\mathbf{u}}; \quad (4)$$

140 moreover, Garambois and Dietrich (2001) verified the consistency of this relation on real field P-wave volume  
 141 waves. The definitions of the  $C$  and  $H$  moduli are those of Biot (1962);  $\epsilon_0$  is the vacuum permittivity,  $\kappa_w$  and  
 142  $\sigma_w$  are the dimensionless dielectric constant and the electrical conductivity of the saturating fluid respectively.  
 143 The zeta potential  $\zeta$  [V] is the electric potential on the slipping plane within the electric double layer.  
 144 Therefore the coseismic electric field is also proportional to the electrokinetic coefficient  $C_K$  largely studied in  
 145 laboratory and modeled (Vinogradov and Jackson, 2011; Vinogradov et al., 2010; Aizawa et al., 2008; Guichet  
 146 et al., 2006; Mainault et al., 2006; Jouniaux et al., 1999; Pozzi and Jouniaux, 1994; Jouniaux et al., 1994; Ishido  
 147 and Mizutani, 1981).

148 The magnetic field  $\mathbf{H}$  has been shown to be proportional to the velocity  $\dot{\mathbf{u}}$  associated to transverse  $SH$ - and  
 149  $SV$  -waves as:

$$150 \quad |\mathbf{H}| \simeq \frac{\phi}{\alpha_\infty} \frac{\epsilon_0 \kappa_w |\zeta|}{\eta_w} \rho_w \sqrt{\frac{G}{\rho}} |\dot{\mathbf{u}}| \quad (5)$$

151 where  $G$  is the shear modulus of the framework. The tortuosity  $\alpha_\infty$  is usually taken equal to the product of the  
 152 porosity by the formation factor  $F$ . The magnetic field can also be expressed as a function of the electrokinetic

153 coefficient  $C_K$  as:

$$154 \quad |\mathbf{H}| \simeq \frac{\varepsilon_0 \kappa_w |\zeta|}{\eta_w \sigma_w} \frac{\sigma_w}{F} \rho_w \sqrt{\frac{G}{\rho}} |\dot{\mathbf{u}}| = C_K \frac{\sigma_w}{F} \rho_w \sqrt{\frac{G}{\rho}} |\dot{\mathbf{u}}| \quad (6)$$

155 Therefore the coseismic magnetic field is also proportional to the electrokinetic coefficient, considering that the  
156 water density and conductivity are constant, as are the formation factor and the  $G$ ,  $C$ ,  $H$  moduli.

### 157 **3. Appropriate field experimental setup**

158 Although performing a field experiment is beyond the scope of this paper, we would like to emphasize  
159 what would be the most appropriate geometry to be developed to detect seismo-electromagnetic conversions  
160 for CO<sub>2</sub> disposal monitoring. The interfacial response can provide information about the formations at depth  
161 while the co-seismic signal provides only information of the soil in the vicinity of the electrodes. The challenge  
162 is therefore to isolate the interfacial response, which is often of the order of 1-100  $\mu\text{V/m}$  (Mikhailov et al.,  
163 2000; Chen and Mu, 2005; ?).

#### 164 *3.1. Signal processing*

165 The first step in processing the seismoelectric data is to remove the noise from power lines, which can  
166 be of the order of 1 mV/m. The estimate of the harmonic noise can be performed on the data recorded just  
167 before the shot, using a pre-trigger recording. The filtering of this noise can be performed by applying a single  
168 frequency adaptative noise cancellation filter. Butler et al. (1996) proposed to apply the techniques of block  
169 and sinusoidal subtraction. Presently the most efficient method which is used for most of the observations is  
170 to routinely reduce the harmonic noise using the algorithm of Butler et al. (1996); Butler and Russell (2003);  
171 Butler et al. (2007) applied to individual shots before the stacking. Wiener and bandpass filters can be used  
172 to reduce high-frequency noise. Supplementary techniques as delay-line filtering in case of severe noise, and  
173 low-pass filtering in case of strong high-frequency noise contamination can be used.

#### 174 *3.2. Benefit of surface-to-borehole measurements*

175 The main issue for detecting the weak IR signal is often the high amplitude of the co-seismic signal. We  
176 propose to perform the electromagnetic measurements in borehole so that this IR signal can be recorded before

177 the arriving of the co-seismic signal. Indeed the interfacial response can be observed free from the coseismic  
178 signal when the electrodes are located below the interface of interest (Dupuis et al., 2007), by measuring the  
179 electric field within a borehole. In (Haines, 2004; Haines et al., 2007) field experiments were conducted in  
180 which the source and the receivers were laid on opposite sides of one or two man-made sand-filled trenches dug  
181 in a clayey background. This fan-shape layout enabled the measurement of the Interface Response as it reached  
182 the electric receivers before the typically stronger Coseismic wavefield. Therefore, the authors suggested that  
183 by setting the source at the surface and the electrodes in a borehole *below* the interfaces of interest, one may  
184 separate the different types of seismoelectric waves without resorting to numerical wave separation techniques.  
185 This layout-related separation may better preserve the amplitudes and waveforms of seismoelectric signals than  
186 numerical data processing such as f-k or  $\tau - p$  filtering, often distorting seismoelectric signals (Warden et al.,  
187 2012). Moreover, with this layout the influence of surface waves is strongly decreased; however Stoneley waves  
188 may appear; which should be dealt with.

189 Another benefit of surface-to-borehole geometries is related to the amplitudes of the Interface Response.  
190 When working with surface geometries, deep interfaces become rather harder to track as the amplitudes of the  
191 associated interfacial signals may have decayed below the noise level by the time they reach the surface. Setting  
192 the receivers close to the target interfaces therefore allows to pick up stronger signals (Haines and Pride, 2006).  
193 Furthermore, when working with uncased wells (Zhu et al., 1999), deploying the receivers at depth also allows  
194 to take advantage of the Coseismic signal, which provides information about the subsurface in the vicinity of  
195 the receivers (Garambois and Dietrich, 2001; Haines et al., 2007; Bordes et al., 2008). For instance, by hitting  
196 the upper casing of a borehole with a sledgehammer, Mikhailov et al. (2000) triggered Stoneley waves, which  
197 in turn induced a pore fluid flow in the permeables zones intersecting the borehole. The authors were able to  
198 measure the small (tens of microvolts) electrical signals associated with this flow.

### 199 3.3. *Seismic source*

200 We propose to use a pure SH seismic source that could achieve a better resolution than the one obtained  
201 through the usual P-driven experiments because of shorter wavelengths. As stated in Haines and Pride (2006),  
202 there is no coseismic electric field for S-waves, but the coseismic magnetic field is present; therefore the electric



203 IR may be easier to be detected than the magnetic IR.

204 In the next sections, we use a 120 Hz peak frequency for the source to keep the response in time fairly sharp,  
205 so that the different events could be easily resolved. But, as for the whole range of seismic frequencies both  
206 the dynamic permeability and the electrokinetic coupling coefficient are fairly independent of the frequency,  
207 qualitatively the same responses would be obtained for a source with peak frequency of 40 Hz or 60 Hz. A  
208 pure shear wave source is difficult to achieve; however surface vibrators for SH-source do exist, which can emit  
209 SH-waves as well as SV-waves, being designed for a peak force of approximately 30 kN (equivalent to the free  
210 fall of a 3 t mass from a height of 1 m) and a frequency range of 16 Hz to 300 Hz.

211 Furthermore, there is presently interest in the seismics community in S-wave exploration because of its appli-  
212 cation in unconventional reservoirs; an application in heavy oil production management through S-wave data  
213 monitoring of stress effects in the reservoir has been reported (Bale et al., 2013), as well as monitoring of the  
214 seal of a CO<sub>2</sub> deposition site (Davis et al., 2013). This situation could contribute to facilitate further progress  
215 in field measurements in seismoelectrics using shear wave sources.

#### 216 **4. Modeling seismoelectric and seismomagnetic signals measured at depth using a shear-wave source**

217 In this section we use a numerical simulator, which features infinite shear sources generating 1D wave fields  
218 in likewise layered media for the modeling of the seismoelectric conversions; see the appendix for details in the  
219 1D SHTE formulation. We model the seismoelectric and seismomagnetic conversions induced by a shear-wave  
220 source within a tabular model consisting of a sand layer over a sandstone layer. We then describe the results  
221 of the horizontal displacement, the horizontal electric field, and the horizontal magnetic field as a function of  
222 depth for full water saturation conditions.

##### 223 *4.1. Model description*

224 We consider a simple tabular model consisting of a sand layer, 30.5 m thick, set on top of a sandstone  
225 half-space (Fig. 1).

226 We model a seismic transverse source of peak frequency  $f_{peak}=120$  Hz at a depth of  $z_s=1$  m, the source wavelet  
227 being a Ricker wavelet, or “Mexican hat” wavelet, which generates seismoelectric signals recorded by a vertical

228 array located right under the source; this array consists of 51 seismic and electromagnetic transverse receivers,  
 229 evenly spaced between a depth of 1 and 51 m so that there are receivers on either side of the interface.

230 The considered source is just a shearing force per unit volume applied along  $y$  on a whole horizontal plane  
 231 located at the source depth with the already described signature, its implementation can be seen in the appendix.

232 Both sand layer and sandstone layer -called Sand and Sandstone I in Table 1, where their properties are detailed-  
 233 are fully saturated with a moderately briny water ( $C_0=10^{-3}$  mol/L). As there is no salinity contrast between the  
 234 two layers and since we compute the  $\zeta$  potential as  $\zeta = 0.008 + 0.026 \log_{10}(C_0)$  (Pride and Morgan, 1991),  $\zeta=-$   
 235 70 mV throughout the entire model. The frame bulk modulus  $K_{fr}$  [Pa] is deduced from the solid bulk modulus  
 236  $K_s$  [Pa], following Pride (2005) and assuming a consolidation parameter of 20 for sand and 5 for sandstone as

$$237 \quad K_{fr} = K_s \frac{1 - \phi}{1 + c_s \phi}. \quad (7)$$

238 The frame shear modulus  $G_{fr}$  [Pa] is linked to the solid shear modulus modulus  $G_s$  in a similar fashion:

$$239 \quad G_{fr} = G_s \frac{1 - \phi}{1 + 1.5c_s \phi}. \quad (8)$$

240 As we deal in next sections with porous media saturated with mixtures of water and  $CO_2$ , it is necessary to  
 241 introduce appropriate effective properties in order to use them within Pride's formulation for electroseismics.  
 242 Therefore, for the effective fluid mass density we use

$$243 \quad \rho_f = \rho_w S_w + \rho_{CO_2} (1 - S_w), \quad (9)$$

244 where  $S_w + S_{CO_2} = 1$  is assumed and subscripts  $CO_2$  and  $w$  refer to carbon dioxide and water respectively,  
 245  $S_w$  denotes water saturation. For the effective bulk modulus of such fluid mixture we use Brie et al. (1995)  
 246 approach

$$247 \quad K_f = (K_w - K_{CO_2}) S_w^5 + K_{CO_2}; \quad (10)$$

248 the power five in this expression is chosen following Carcione et al. (2006). Here the  $CO_2$  is supercritical as  
 249 explained in the section 6 and there is no gaseous phase in our model. The effective viscosity is computed  
 250 in terms of the mixture components viscosities  $\eta_l$ ,  $l = w, CO_2$  and water saturation  $S_w$  using Teja and Rice

251 (1981)

252

$$\eta_f = \eta_{CO_2} \left( \frac{\eta_w}{\eta_{CO_2}} \right)^{S_w}. \quad (11)$$

253 In order to characterize the electric conductivity of the effective fluid saturated solid matrix we use the expres-  
254 sion recently proposed by Warden et al. (2013), extending Pride's original formula (Pride, 1994, Eq.(242)) to  
255 the realm of partially saturated media:

256

$$\sigma(S_w, \omega) = \frac{S_w^n}{F} \sigma_w + \frac{2}{F} \frac{C_{em} + C_{os}(\omega)}{\Lambda} \quad (12)$$

257 The first term in this equation -where  $F = \phi^{-m}$  stands for the formation factor,  $m$  being the cementation  
258 coefficient- is Archie's law for a partially saturated medium, while the second term accounts for the surface  
259 conductivity. In the latter, -as Pride stated in his liminar work- the factor  $C_{em}$  [S] is the excess conductance as-  
260 sociated with the electromigration of double layer ions;  $C_{os}(\omega)$  [S] is the frequency-dependent electro-osmotic  
261 conductance due to electrically induced streaming of the excess double-layer ions and  $\Lambda$  [m] is the above pre-  
262 sented pore-geometry dependent factor. We remark here that, as in Brovelli et al. (2005) and Warden et al.  
263 (2013), the surface conductivity is assumed to be independent of water saturation  $S_w$ , because under realistic  
264 saturation ranges (residual water saturation  $S_{wr} \geq 10\%$ ) the thickness of the wetting phase layer on the pore  
265 surface is always larger than the Debye length  $d$ . This also means that all fluid related properties involved in  
266 the calculation of the surface conductivity and of the electrokinetic coupling -see below- are just those of water.  
267 Again, following Warden et al. (2013), we propose for the effective fluid saturated media the following elec-  
268 trokinetic coupling:

269

$$L_0(S_w) = -\frac{\phi}{\alpha_\infty} \frac{\epsilon_0 \kappa_w \zeta}{\eta_w} \left( 1 - 2 \frac{d}{\Lambda} \right) S_w^n S(S_w), \quad (13)$$

270 In this equation  $n$  is Archie's saturation exponent (taken to be equal to the cementation exponent) and  $S(S_w)$   
271 is a function relating the streaming potential coefficient obtained under partial saturation conditions to the one  
272 corresponding to full saturation conditions. Several authors investigated this relation from both theoretical and  
273 experimental viewpoints; Perrier and Morat (2000), Guichet et al. (2003), Revil et al. (2007), and Strahser  
274 et al. (2011) predict a monotonic behaviour with saturation, Jackson (2010) suggested that the coupling coeffi-  
275 cient could be either monotonic or non-monotonic depending on the properties of the saturating phases, while

276 Allègre et al. (2010) and Allègre et al. (2012) observed and modeled a non-monotonic behaviour while study-  
 277 ing laboratory drainage experiments; we select for the present work two expressions for  $S(S_w)$  displaying a  
 278 qualitatively distinct behaviour, meaning monotonic and non-monotonic:

$$279 \quad S(S_w) = \begin{cases} \frac{1}{S_w^n} \left( \frac{S_w - S_{wr}}{1 - S_{wr}} \right)^2 & S_{wr} = 0.10 \quad (\text{Perrier and Morat, 2000}) \\ \left( \frac{S_w - S_{wr}}{1 - S_{wr}} \right) (1 + 32(1 - (\frac{S_w - S_{wr}}{1 - S_{wr}}))^{0.4}) & S_{wr} = 0.305 \quad (\text{Allègre et al., 2010}). \end{cases} \quad (14)$$

#### 280 4.2. Seismic and seismo-electromagnetic results

281 In this section we present our first results; here it should be noticed that they correspond to media saturated  
 282 with water, i.e.  $S_{CO_2} = 0$ , and that free surface reflections are neglected.

283 On the synthetic recording displaying the horizontal solid displacement (Figure 2 (a)), one can notice the  
 284 downgoing direct S-wave  $u_{y,i}$ , traveling at  $v_S = 1104$  m/s. When this direct wave hits the interface located at  
 285 30.5 m depth at about 0.027 s, part of the total incident energy reflects back to the surface as an upgoing S wave  
 286  $u_{y,r}$  with the same velocity as the incident wave. The transmitted downgoing S wave  $u_{y,t}$  travels at a higher  
 287 velocity of  $v_S = 2485$  m/s.

288 On the synthetic recording displaying the horizontal electric field (Figure 2 (b)), one can distinguish three  
 289 events. An event with zero moveout -labeled as  $E_D$  in this figure- appears at the time at which the source is  
 290 triggered (0.01 s). This flat arrival may be related to the *direct field* predicted by Pride and Haartsen (1996) and  
 291 measured by Haines (2004).

292 An event we associate with a first Interface Response  $-E_{IR}$  in the figure- arises at about 0.027 s, that is, at  
 293 about the time needed for the S-wave to reach the interface. A second Interface Response  $E_{IR2}$  occurring at  
 294 the surface when the S-wave reflected at the 30.5 m deep interface  $u_{y,r}$  reaches the surface, is seen at two-way  
 295 traveltime. Its origin could be partially due to an numerical artifact caused by the boundary conditions for Biot  
 296 equations at the Earth surface; further modelling with independent codes and field experiments will help to  
 297 clarify this question.

298 It is also interesting to notice that, as stated in Haines and Pride (2006), there is no coseismic electric field for  
 299 S-waves, but the coseismic magnetic field is present, as can be seen in Figure 2 (c). The magnetic field existing  
 300 within the seismic shear wave displays the same behaviour as the latter: the incident coseismic magnetic field

301  $H_{C,i}$  arrives at the 30.5 m depth interface at about 0.027 s, and is partially transmitted -see  $H_{C,t}$  in the figure-,  
 302 and partially reflected towards the surface as  $H_{C,r}$ . It is also possible to see the flat event associated to the direct  
 303 field  $H_D$ , at about 0.01 s, and a first Interface Response  $H_{IR}$  generated simultaneously with the arrival of the  
 304 seismic wave to the interface. The amplitude of the Interface Response is negligible in the upper layer. Finally,  
 305 the second Interface Response occurring at the Earth surface, labeled  $H_{IR2}$  in the figure, is also present for the  
 306 magnetic field.

## 307 **5. Sensitivity of the Interface Response to contrasts in fluid and rock properties**

308 In this section we describe the amplitude of the interfacial response induced by a S-wave source when some  
 309 physical properties of the sandstone half-space are changed whereas an upper sandstone layer is kept with  
 310 constant parameters. The properties of the upper sandstone layer (Sandstone II) are given in the third column of  
 311 Table 1.

### 312 *5.1. Porosity contrast*

313 In this paragraph we study the influence of a porosity contrast on the amplitude of the interfacial response.  
 314 The porosity in the lower half-space is allowed to change between 2 and 24%. The empirical relation of Bourbié  
 315 et al. (1987) linking porosity and permeability in Fontainebleau sandstones is used to account for the influence  
 316 of porosity changes on permeability,

$$317 \quad k_0 = 1.66 \times 10^{-4} \phi^8 \text{ for } \phi < 6\% \quad (15)$$

$$318 \quad k_0 = 2.5 \times 10^{-10} \phi^3 \text{ for } \phi > 6\% \quad (16)$$

320 The permeability values associated with the porosity values of the sandstone half-space are given in Table 2. For  
 321 each set porosity and permeability values, we modeled the electric field along y, and measured the maximum  
 322 S-EM IR value on the synthetic electrograms. Apart from porosity and permeability varying in the lower half-  
 323 space, all other parameters are fixed.

324 In order to eliminate the influence of the source amplitude, the results obtained were normalized as follows: for

325 each fixed parameter (here porosity) value, the portion of the signal corresponding to the interfacial response  
326 is isolated, and the maximum value of the maximum amplitude of all recorded IR's is selected. By varying  
327 the analysed parameter a set of these maxima is obtained. Finally, this set is normalised by dividing all values  
328 by the maximum value in it. The obtained results are displayed in Figure 3. When the porosity value of the  
329 sandstone layer is 12%, meaning the same value as the upper sand layer, the interfacial response is zero because  
330 there is no contrast of any physical properties between the layers. When the porosity of the sand layer is either  
331 decreasing down to 2% or increasing up to 24% the maximum amplitude is increasing, because the contrast in  
332 porosity between the two layers is increasing.

### 333 5.2. $\zeta$ potential contrast

334 Following the same procedure as in the previous paragraph, we investigate the behaviour of the S-EM IR  
335 response when the  $\zeta$  potential in the top layer remains fixed at -0.035V, whilst its values are allowed to vary  
336 from -0.05V to -0.07V in the half-space.

337 Results are shown in Figure 4: when both layers have the same zeta potential, there is no contrast in physical  
338 properties to induce an interfacial response. When the contrast in the zeta potential is increased, either by  
339 decreasing or increasing the zeta potential of the lower layer, then the interfacial response is increased.

### 340 5.3. Viscosity contrast

341 We finish this section by studying the dependence fo the S-EM IR response when the viscosity  $\eta = 10^{-3}$   
342 Pa.s in the upper-layer, whilst its values are allowed to vary from to  $10^{-4}$  Pa.s to  $10^{-1}$  Pa.s in the half-space.  
343 Results are displayed in Figure 5. As for the study of the other properties, when there is no contrast in physical  
344 properties there is no interfacial response. When the constrast in viscosity is increasing between the layers, the  
345 amplitude of the interface response increases. However, when the viscosity of the half space is larger than that  
346 of the upper layer, the amplitude of the response grows slowly, contrary to what happens when the viscosity of  
347 the half space is smaller than that of the top layer.

348 This sensitivity study shows that the amplitude of the interfacial response increases with an increasing contrast  
349 in porosity and zeta potential. This amplitude is also increased by an increasing contrast in viscosity when  
350 viscosity is decreased, which is the case when dealing with CO<sub>2</sub> at supercritical conditions.

## 351 **6. Effect of a contrast between water-saturated sand and sandstone with various CO<sub>2</sub> saturations**

352 In this section we model a contrast between an upper water-saturated layer and a lower semispace with  
353 various concentrations of CO<sub>2</sub> at supercritical conditions. We describe the results of the modelling coseismic  
354 magnetic field and the electric and magnetic interfacial responses induced by a shear-wave source.

### 355 *6.1. Model including a layer with various CO<sub>2</sub> concentrations*

356 In this section we consider a simple model which consists of a 100 m thick layer on top of a half-space.  
357 Whilst the top layer remains fully saturated with water with a salinity coefficient  $C_0 = 10^{-3}$  mol/l, the CO<sub>2</sub>  
358 saturation is allowed to vary in the half-space; the salinity is the same in both layers. The effective properties  
359 necessary for Pride's equations to remain valid when dealing with more than a single fluid phase are calculated  
360 following the formulae described in Section 4.1. As it is known from the CO<sub>2</sub>-sequestration literature, see -  
361 among others- (Kiessling et al., 2010; Cairns et al., 2012), this gas is usually pumped in a supercritical state into  
362 the subsurface. Therefore, we assume here that CO<sub>2</sub> is in the mentioned supercritical state and throughout the  
363 next sections we consider the following physical properties values for the carbon dioxide  $\rho_{CO_2} = 505$  kg/m<sup>3</sup>,  
364  $\eta_{CO_2} = 1.5 \times 10^{-4}$  Pa.s,  $K_{CO_2} = 25$  MPa, (Carcione et al., 2006) which as just mentioned correspond to it  
365 being in supercritical state meaning at pressure 10 MPa and temperature 37 °C.

366 Although a hundred meters depth are not enough for this assumption to be valid (Kazemeini et al., 2010), we  
367 retain the mentioned depth value to keep a reasonable computational cost, because of the size of the model. Note  
368 that the following analysis would remain exactly the same if we increased the depth of the bottom of the top  
369 layer as much as necessary to reach the pressure and temperature conditions for the CO<sub>2</sub> to be in supercritical  
370 state.

371 It is known that when pumped into a reservoir a small portion of carbon dioxide dissolves in water (Carcione  
372 et al., 2006; Wang et al., 2013), forming weak carbonic acid which reacts with the present dissolved salt ions  
373 (Darwish and Hilal, 2010). This process alters the  $\zeta$  potential (Moore et al., 2004), which in turn changes the  
374 electrokinetic coupling  $L_0$ ; in the present work the zeta potential itself is not varying, but the effect of water-  
375 saturation is taken into account, as described above, by making  $L_0$  saturation dependent. We assumed that the  
376 electrokinetic coupling is changed when the amount of CO<sub>2</sub> is increased and water expelled, as it changes when

377 water-saturation is decreased, replaced by air (Eq.13).

378 We remark here that, as it can be inferred from Equation (12), we consider the electrical conductivity of carbon  
379 dioxide  $\sigma_{CO_2}$  negligible compared to that of salty water. Although the  $CO_2$  is not gaseous in our model, the  
380 supercritical  $CO_2$  will increase the electrical resistivity of the layer, compared to water, as will do the air, with a  
381 lower increase induced by supercritical  $CO_2$  than by gaseous  $CO_2$  (Borner et al., 2013). Further studies should  
382 investigate, if possible, the electrokinetic coefficient in presence of supercritical  $CO_2$ .

### 383 6.2. Results of the modelling: coseismic magnetic field

384 We first model the coseismic signal linked to the seismic propagation induced by the S-wave source (Fig.6a).  
385 The maximum of each trace, as a function of depth, is pointed, and the maximum of these maxima is deduced,  
386 for each  $CO_2$  saturation. Then this maximum is normalized by the value of the magnetic field for water-  
387 saturated conditions, each curve being normalized by its own maximum value. When no  $CO_2$ -dependence on  
388  $L_0$  is assumed, meaning  $S_w^n S(S_w) = 1$  in Eq.13, the coseismic magnetic field linearly decreases with increas-  
389 ing  $CO_2$  content. Even when  $L_0$  remains constant, as the effective fluid density and conductivity diminish with  
390 increasing  $CO_2$  saturation, so does the coseismic magnetic response, as can be seen from Fig.6a, reflecting the  
391 behaviour predicted in Eq. 6.

392 When assuming a monotonous decrease of the electrokinetic coefficient with decreasing water-saturation as  
393 proposed by Perrier and Morat (2000), it is expected to observe a monotonous decrease of the coseismic mag-  
394 netic field  $\mathbf{H}$  as foreseen in Eq.6 and shown in fig.6a. Then, when another behaviour of the electrokinetic  
395 coefficient as a function of the water-saturation is assumed, as the one proposed by Allègre et al. (2010), we  
396 can observe first an increase in the coseismic magnetic field when the  $CO_2$  saturation is increased and then  
397 a decrease with further increasing  $CO_2$  saturation, as expected through the Eq.6 where the magnetic field is  
398 proportional to the electrokinetic coefficient.

399

### 400 6.3. Results of the modelling: electric and magnetic interfacial responses

401 The interfacial response of the electric field and the magnetic field are shown in figures 6b and 6c. The  
402 electric interfacial response is increasing with increasing  $CO_2$  saturation, for both cases of  $S_w^n S(S_w) = 1$  and



403 for the model from Perrier and Morat (2000). Using the model of Allègre et al. (2010) the electric interfacial  
 404 response first increases to reach a maximum for a CO<sub>2</sub> saturation of about 12% and then decreases for a CO<sub>2</sub>  
 405 saturation in the range 12-55% before increasing for a CO<sub>2</sub> saturation in the range 55-70%. The magnetic in-  
 406 terfacial response increases when the CO<sub>2</sub> saturation increases up to about 70% for both cases of  $S_w^n S(S_w) = 1$   
 407 and for the model from Perrier and Morat (2000), before decreasing with further increase of CO<sub>2</sub> saturation.  
 408 Using the model of Allègre et al. (2010) the magnetic interfacial response first increases to reach a maximum  
 409 for a CO<sub>2</sub> saturation of about 10% and then decreases for a CO<sub>2</sub> saturation in the range 10-55% before increas-  
 410 ing for a CO<sub>2</sub> saturation in the range 55-70%. Therefore the relative maximum interfacial response, for both  
 411 the electric field and the magnetic field, is different according to the different models of the CO<sub>2</sub>-dependence  
 412 of the electrokinetic coupling, and occurs either around 10% or 70-90% of CO<sub>2</sub> saturation.

413 We can compare the behaviour of the interfacial response of the electrical field observed here with the one  
 414 observed for the study of a contrast in water-saturation using a P-wave source rather than a S-wave source  
 415 (Warden et al., 2013). The electric interfacial response showed also an increase and then a decrease with in-  
 416 creasing air-content using the model of Allègre et al. (2010), but showed a maximum at about 30% (Fig.9a in  
 417 Warden et al. (2013)) rather than 12% (see Fig6b). The electrical interfacial response using the model of Perrier  
 418 and Morat (2000) also showed a continuous increase with increasing air-content, although the curvatures are  
 419 different.

420 These results do not show the relative amplitudes according to the different models. Consequently we nor-  
 421 malized the magnetic coseismic field, the electric interfacial response and the magnetic interfacial response ob-  
 422 tained for both models by the field values obtained using the electrokinetic coupling  $L_0$  assuming  $S_w^n S(S_w) = 1$   
 423 in Eq.13. These results show that the amplitude of the coseismic magnetic field using the model from Allègre  
 424 et al. (2010) can be a factor 10 larger than the results using the model from Perrier and Morat (2000) for CO<sub>2</sub>  
 425 saturation around 10% (fig.7). The electric interfacial response using the model from Allègre et al. (2010) can  
 426 be about 50 times larger than the results using the model from Perrier and Morat (2000) for CO<sub>2</sub> saturation  
 427 around 5% (fig.8). The magnetic interfacial response using the model from Allègre et al. (2010) can be also  
 428 about 50 times larger than the results using the model from Perrier and Morat (2000) for CO<sub>2</sub> saturation around  
 429 5% (Fig.9). Comparing figures 7 and 9 we can deduce that the ratio between the IR amplitude and coseismic

430 amplitude of the magnetic field is about 45 and 6 using the model from Allègre et al. (2010) and Perrier and  
431 Morat (2000) respectively.

432 We can notice that in both considered cases for the CO<sub>2</sub>-dependence of the electrokinetic coupling model, the  
433 magnetic IR is larger than the coseismic magnetic field, which is not usually the case for the electric field using  
434 a P-wave source. Therefore measuring the magnetic interfacial response induced by a S-wave source, could be  
435 an efficient method to detect the interface between a water-saturated layer and a partially CO<sub>2</sub> saturated layer.  
436 Moreover, as the electric coseismic signal induced by an S-wave source is absent, the electric IR is easier to  
437 be detected. Therefore an efficient method to detect different CO<sub>2</sub> saturations would be to measure the electric  
438 interfacial response using a S-wave source. The amplitude of the response would be up to 300 times higher  
439 than the amplitude of the signal induced by a water-saturated medium, depending on the model used for the  
440 CO<sub>2</sub>-dependence of the electrokinetic, in the saturation range 5-15%, and 10 to 100 times higher in the 15-40%  
441 CO<sub>2</sub> saturation range.

## 442 **7. Seismo-electromagnetic conversions induced in a CO<sub>2</sub> reservoir with a seal layer**

443 Let us consider now a new model, shown in Figure 10, in which we intersperse a 10 m deep seal layer of  
444 very low permeability among a 100 m deep layer whose top boundary is the Air-Soil interface, and a semispace  
445 in which CO<sub>2</sub> saturation can be changed. Indeed clay layers can be present as thin intra-reservoir shales. They  
446 act as main barriers to the upward migration of CO<sub>2</sub> beneath which the the CO<sub>2</sub> accumulates at high saturations  
447 (Arts et al., 2004).

448 The three layers parameters are displayed in Table 3; we remark that the permeability of the seal layer is four  
449 orders of magnitude smaller than the one of the top layer and the semi-space. In order to better approximate a  
450 possible carbon dioxide deposition site, we strongly increase the NaCl concentration in the water saturating the  
451 semi-space, therefore enhancing its electrical conductivity, which yields a bulk conductivity of  $\sigma = 0.12$  [S/m]  
452 at full water saturation. The semi-space is then the most electrically conductive, the seal layer has a smaller  
453 bulk conductivity including a surface conductivity, and the top layer has the lowest bulk conductivity. The zeta  
454 potential, which depends on the fluid conductivity is very small within the briny semi-space. It is about  $-3$  mV

455 within the seal and  $-70$  mV within the top layer. This configuration leads to an electrokinetic coupling in the  
456 seal layer about 16 times smaller than the value of the one corresponding to the top layer, and of about twice  
457 that of the semi-space. The electrokinetic coupling does not reflect here only the zeta potential (itself linked to  
458 the fluid conductivity), but also the permeability (see eq.3) which is very low in the half-space. Therefore the  
459 contrast in the electrokinetic coupling is higher between the semi-space and the seal layer than the one between  
460 the top layer and the seal layer.

461 According to the analysis performed in previous sections, an interface response is expected to arise at both  
462 seal interfaces, with potentially a larger signal between the semi-space and the seal layer because of a larger  
463 contrast in the electrokinetic couplings. However, the method is not expected to resolve them, because the two  
464 IR's are separated about 7 milliseconds ( $v_S=1700$  m/s), the width of the central peak of the source being about  
465 4.5 milliseconds. Recall, however, that we are not mainly interested in determining the width of the seal, but in  
466 what lies beneath it. Notice that previous numerical studies (Pride and Garambois, 2005) have shown that for  
467 thin enough layers, the Interface Response can attain very large values.

468 In Figure 11 we display a borehole gather for (a) the seismic waves, (b) the electric field and (c) the magnetic  
469 field, considering a 65% carbon dioxide saturation in the semispace. It can be seen in the leftmost picture  
470 that contrary to the seismic response of our first example, shown in Figure 2(a), the amplitude of the reflected  
471 seismic wave is much smaller than the incident wave, due to the similar mechanical properties of the seal layer  
472 and the semi-space, recall that they have just different permeabilities and different fluids saturating them. How-  
473 ever, both electric interface response  $E_{IR}$  and magnetic interface response  $H_{IR}$  are clearly observable, raised  
474 simultaneously with the arrival of the incident seismic wave to the interface between the top and seal layers,  
475 at about 0.06 s. The electric IR can be detected within the whole depth range, whereas the magnetic IR can  
476 be detected only at depths below the seal layer. Also discernible is the coseismic magnetic field (see fig.11c),  
477 traveling within the incident, reflected and transmitted seismic shear waves.

478 With the goal of detecting possible changes in the  $\text{CO}_2$  saturation within the semi-space, the magnetic and  
479 electric interfacial responses were calculated for  $\text{CO}_2$  saturations of 5%, 25% and 65%. In Figure 12 we show  
480 time windows highlighting the difference between the magnetic field interface responses of two different  $\text{CO}_2$   
481 saturations; in (a) we take the difference between 65% and 5% saturations and in (b) we take the difference

482 between 65% and 25% saturations. The relative differences normalised by the value of the IR at 65% of CO<sub>2</sub>  
483 saturation (the maximum of the maxima of all traces is taken into account, as previously explained) show a  
484 variation of 30% when saturation varies from 65% to 5%, and a variation of 22% when saturation varies from  
485 65% to 25%. These results show us that even when the interface response of the bottom boundary of the seal is  
486 "entangled" with the one produced at its top boundary, the recorded magnetic IR in the well receivers -located  
487 below the lowest interface- are very sensitive to carbon dioxide saturation changes.

488 In Figure 13 we show the time windows highlighting the differences between the electric field interface re-  
489 sponses of the same two different CO<sub>2</sub> saturations. The relative differences normalised by the value of the IR  
490 at 65% of CO<sub>2</sub> saturation show a variation of 62% when saturation varies from 65% to 5%, and a variation of  
491 52% when saturation varies from 65% to 25%. Therefore the sensitivity of the electric IR to the CO<sub>2</sub> saturation  
492 variations is larger than the sensitivity of the magnetic IR.

493

494 The detection of the variation of CO<sub>2</sub> concentration could be therefore performed by measuring the electric  
495 and magnetic field in boreholes. The magnetic field measured below the interface could detect the magnetic  
496 IR induced by the contrast of the seal layer and the semi-space; the electric field could be measured below and  
497 above the interface. Moreover, as the coseismic part of the electric field is absent when using a S-wave source,  
498 the electric IR may be easier to be detected than the magnetic IR. However if the electric ambient noise is high,  
499 the electric IR may be still difficult to measure, even if a large variation is expected as a function of the CO<sub>2</sub>  
500 concentration. The measurement of the magnetic field below the interface may still help to detect the CO<sub>2</sub>  
501 saturation variations, because the magnetic IR is larger than the coseismic magnetic signal, and because of the  
502 sensitivity of the magnetic IR which is still noticeable although lower than the one of the electric IR.

## 503 **8. Conclusions**

504 -In this paper we numerically analyzed shear wave driven seismoelectromagnetic conversions in a surface-  
505 to-borehole layout, using a one dimensional finite elements code. Sensitivity analysis of the S-EM IR for  
506 porosity, permeability, zeta potential and viscosity were performed for a simple tabular medium, and normalized

507 responses were used in these analysis, in order to make them independent of the physical source used by the  
508 employed method.

509 -It was observed that -as expected- no contrast in the properties leads to the absence of interface response; and  
510 that the response increases when the contrast in porosity and zeta potential is increased, while for relatively  
511 large values of the viscosity the response is asymptotically constant.

512 -We studied also the behaviour of the electromagnetic responses for a model considering realistic partial CO<sub>2</sub>  
513 saturations. We used the extended Pride's formulation for the electrokinetic coupling for the case of partially  
514 saturated media recently presented in Warden et al. (2013) to take into account the presence of carbon dioxide  
515 in our model. Moreover, we studied the magnetic coseismic response and electric and magnetic interface  
516 responses using Perrier-Morat and Allègre formulas in the partial CO<sub>2</sub> saturation version of the electrokinetic  
517 coupling. We observed that the relative maximum in the interface response for both the electric and magnetic  
518 fields is different according to the different models of the CO<sub>2</sub>-dependence for the electrokinetic coupling, and  
519 occurs at either around 10% or 70-90% of carbon dioxide saturation.

520 -These results are obtained assuming that the injection of supercritical CO<sub>2</sub> in water-saturated sandstone  
521 decreases the electrical conductivity, as shown by Borner et al. (2013) during short time experiments. However  
522 it has to be noted that the injection of supercritical CO<sub>2</sub> in brine solutions increases its electrical conductivity,  
523 because of the dissolution of CO<sub>2</sub> in water, with a larger effect on fresh water solution than on saline solutions  
524 (Borner et al., 2013). This effect was assumed not to take place in the short-time experiments. Therefore  
525 considering long-time period of CO<sub>2</sub> storage and dissolution process, further studies may consider different  
526 hypotheses on the effect of supercritical CO<sub>2</sub> on the electrical conductivity.

527 -When studying the magnetic and electric interface responses of a model for a sealed CO<sub>2</sub> reservoir with  
528 different saturations, we observed that they are sensitive to CO<sub>2</sub> saturation variations, even when they are  
529 superimposed with the interface response of the boundary between the overburden and the seal; the sensitivity  
530 of the electric IR is higher than the one of the magnetic IR. The electric IR variations are observable with  
531 recorders located in the reservoir above or below the seal layer, whereas the magnetic IR is only observed  
532 below the seal layer. Both magnetic and electric field measured in borehole could detect the variations of CO<sub>2</sub>  
533 saturation below the seal.

534 -Noting that up to now seismomagnetic signals were recorded in experiments performed in the laboratory by  
535 Bordes et al. (2008) or under special conditions (Gaffet et al., 2003), and further developments are needed to  
536 measure seismomagnetic signals in the field, we consider that an efficient method to detect a CO<sub>2</sub> saturation  
537 in the range 5-40% would be to measure the electric interfacial response using a S-wave source, expecting a  
538 signal 5 to 300 higher than the signal induced by a water-saturated medium.

539 -We expect the novel results presented in this work will be followed by other necessary developments, such as  
540 the analysis of the absolute amplitudes of the IR responses, to show if they are large enough to be detected.  
541 Moreover more realistic geological environments should be modeled, eventually leading to a new monitoring  
542 tool which complements the existing ones.

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546 **9. Bibliography**

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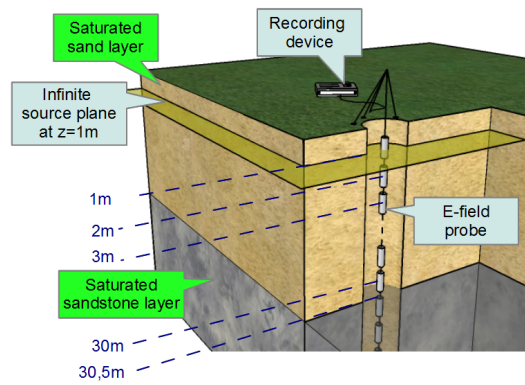


Figure 1: *Tabular model and seismoelectric vertical profiling layout. The subsurface consists of a fully saturated 30.5 m thick sandstone layer on top of a saturated sand half-space. The source is oriented along y, 51 dipole receivers -set 1 m apart- are deployed inside a vertical uncased borehole.*



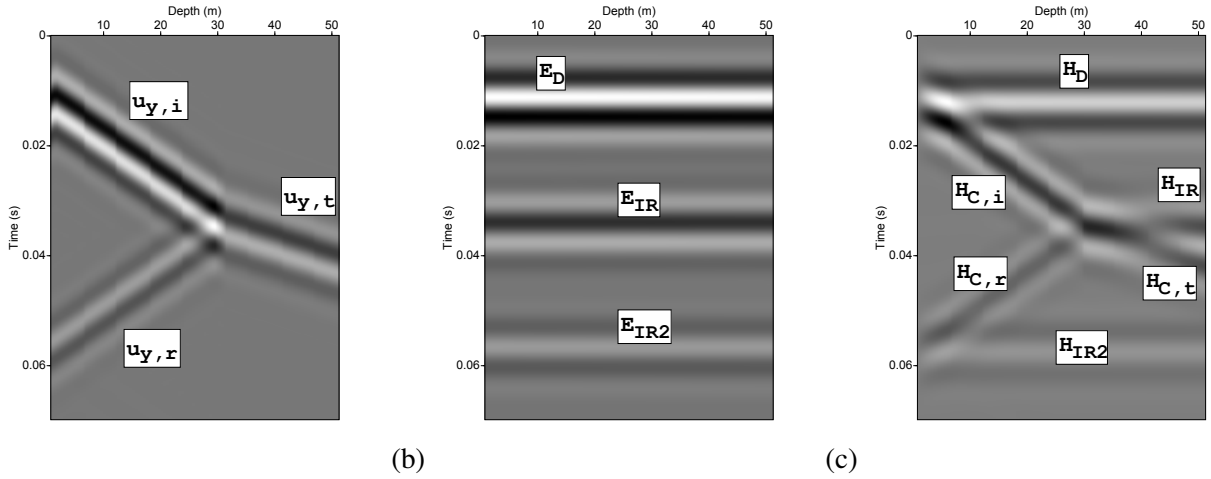


Figure 2: Results obtained with the FE-algorithm using an infinite source along  $y$ , as described in Section 4.1. (a) Horizontal solid displacement  $u_{y,j}$ ,  $j = i, r, t$ ,  $i$ : incident,  $r$ : reflected,  $t$ : transmitted, (b) Horizontal electric field  $E_y$ , subscript  $D$  stands for direct, i.e. the field originated as a conversion at the source; subscript  $IR$  is associated to the interface response generated at 30.5 m depth, and the signal with subscript  $IR2$  is assumed to be an interface response generated at the surface when  $u_{y,r}$  reaches it at about 0.057 s, (c) horizontal magnetic Field  $H_x$ ; the subscript  $D$  stands for direct,  $C$  for coseismic ( $i$ : incident,  $r$ : reflected and  $t$ : transmitted), subscript  $IR$  is associated to the magnetic interface response generated at 30.5 m depth,  $IR2$  is the interface response generated at the surface, when  $u_{y,r}$  reaches it.

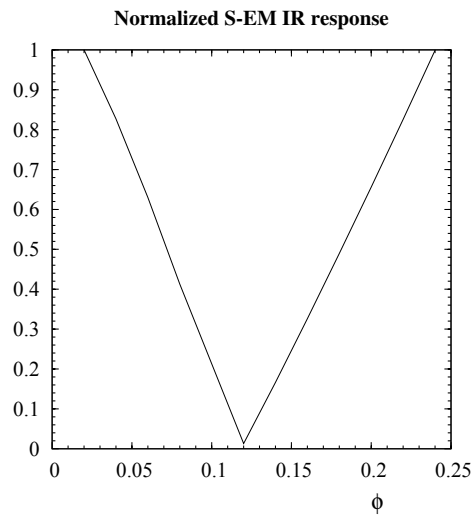


Figure 3: *Maximum S-EM IR amplitude versus porosity in the half-space of model described in Section 5.1.*

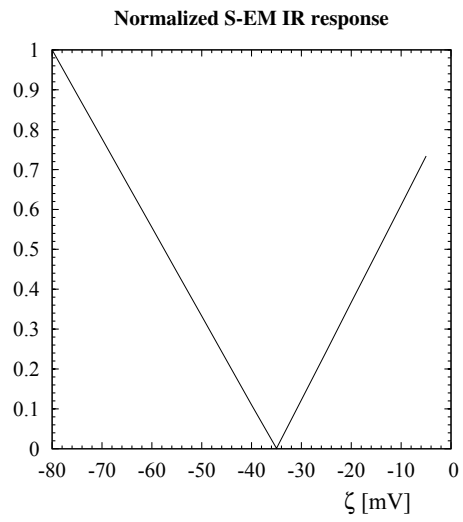


Figure 4: *Maximum S-EM IR amplitude versus  $\zeta$  potential in the half-space of model described in Section 5.2.*

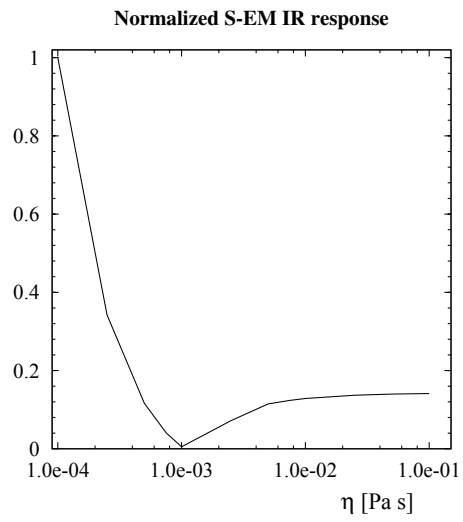


Figure 5: Maximum S-EM IR amplitude versus viscosity  $\eta$  in the half-space of model described in Section 5.3.

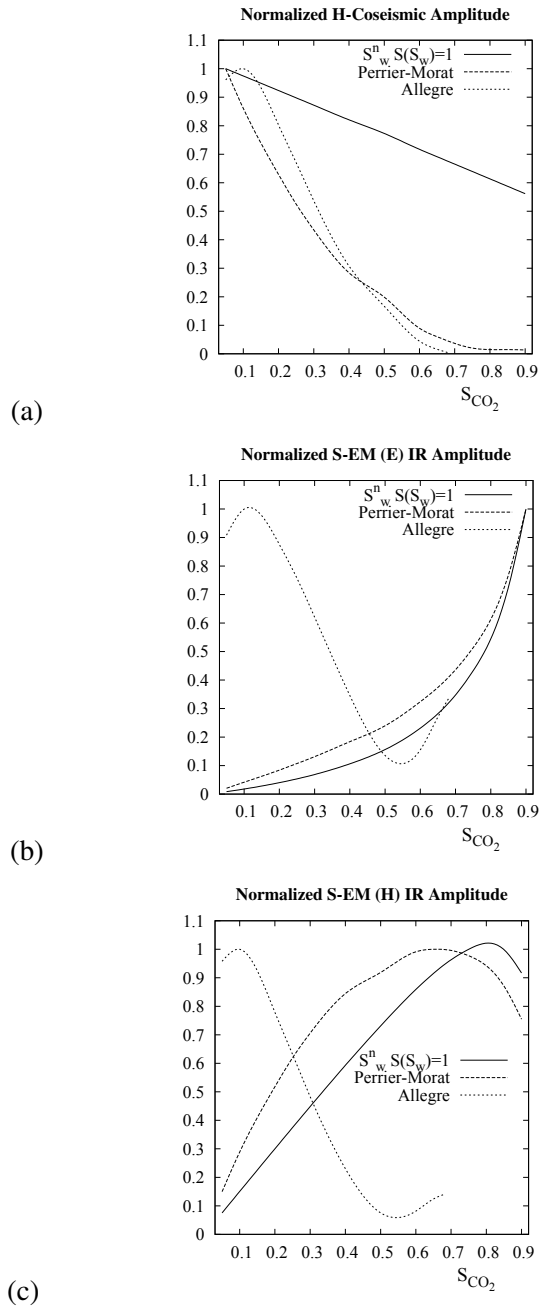


Figure 6:  $CO_2$  saturation dependence of the (a) Coseismic Magnetic Response, (b) S-EM Interface Response for the electric field and (c) S-EM Interface Response for the magnetic field for the different electrokinetic coupling models assumed in this work.

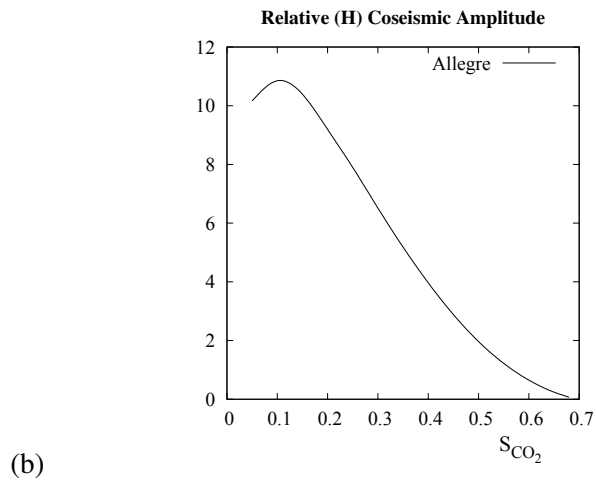
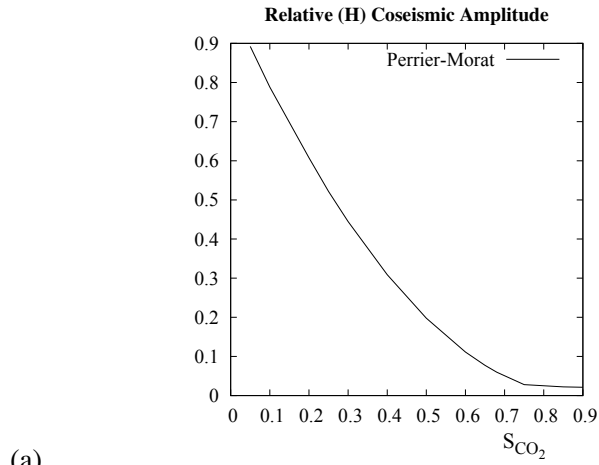
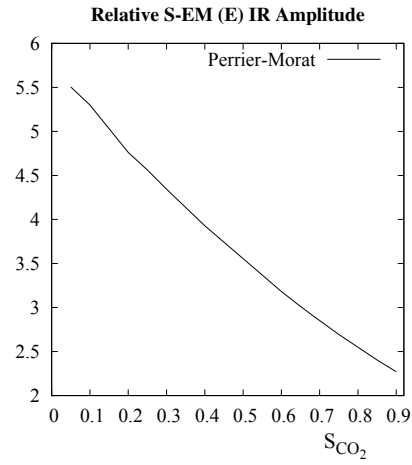
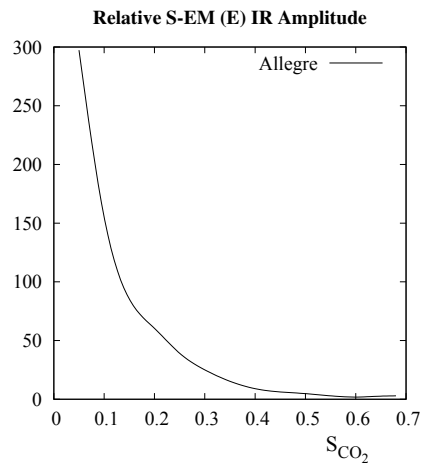


Figure 7: Relative amplitude of the coseismic magnetic field for (a) Perrier-Morat model and (b) Allègre model; both cases are compared to the amplitude of the H-coseismic response when using the electrokinetic coupling  $L_0$  assuming  $S_w^n S(w) = 1$  in Eq.13.



(a)



(b)

Figure 8: *Relative amplitude of the Interface Response electric field for (a) Perrier-Morat model and (b) Allègre model; both cases are compared to the amplitude of the S-EM (E) response when using the electrokinetic coupling  $L_0$  assuming  $S_w^n S(w) = 1$  in Eq.13.*

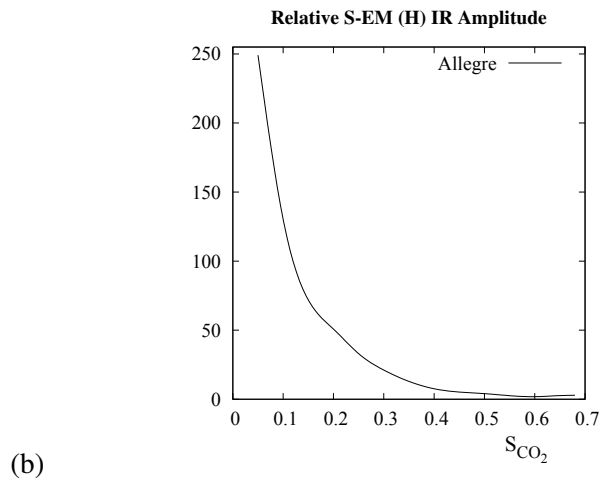
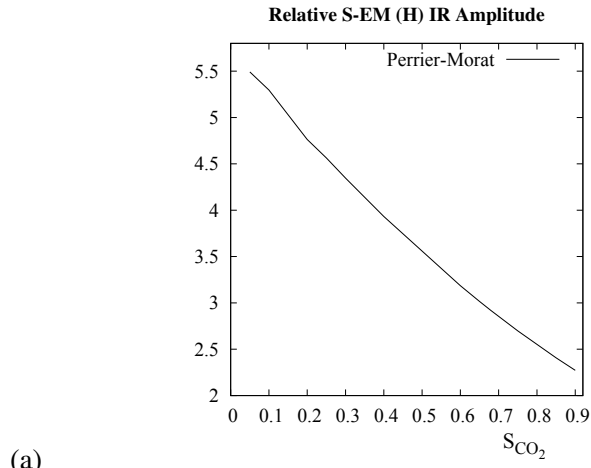


Figure 9: *Relative amplitude of the Interface Response magnetic field for (a) Perrier-Morat model and (b) Allègre model; both cases are compared to the amplitude of the S-EM (H) response when using the electrokinetic coupling  $L_0$  assuming  $S_w^n S(w) = 1$  in Eq.13.*



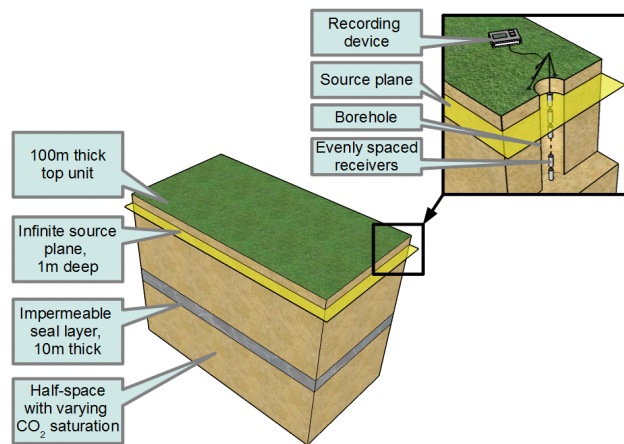


Figure 10: *Model with a seal layer.*

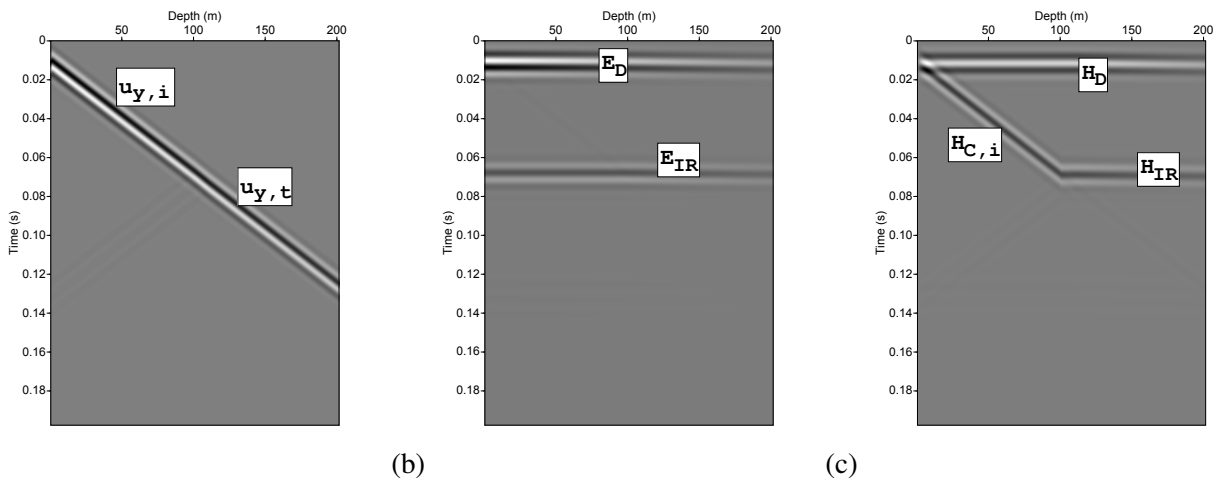


Figure 11: Results for the model of CO<sub>2</sub> reservoir with a seal layer. CO<sub>2</sub> saturation is 65% within the semi-space (below depth 100m). Horizontal displacement  $u_y$  (a), horizontal electric field  $E_y$  (b) and horizontal magnetic Field  $H_x$  (c) obtained with the FE-algorithm using an infinite source along  $y$ , as described in Section 4.1.

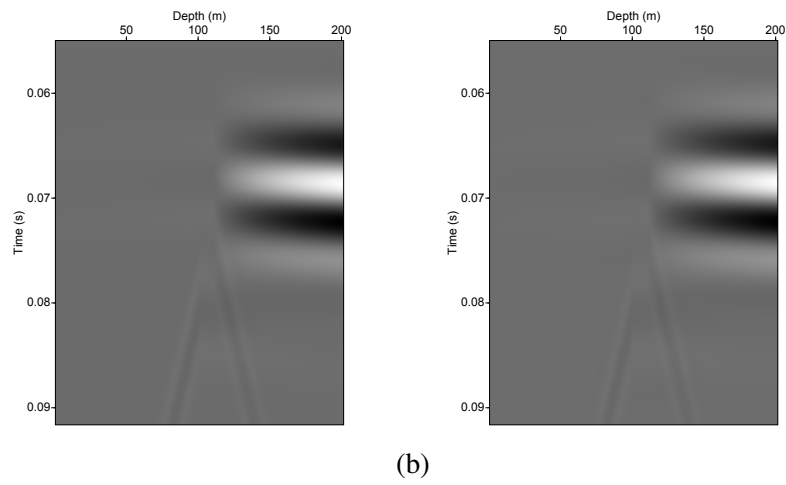


Figure 12: From the magnetic field well gather, -Figure 11(c)- we select a time window, displaying the difference of the magnetic field IR response between two different  $CO_2$  saturations, namely 5% and 65% (a); while in (b) the same result as in (a) is displayed, but the difference is taken considering  $CO_2$  saturations of 25% and 65%.

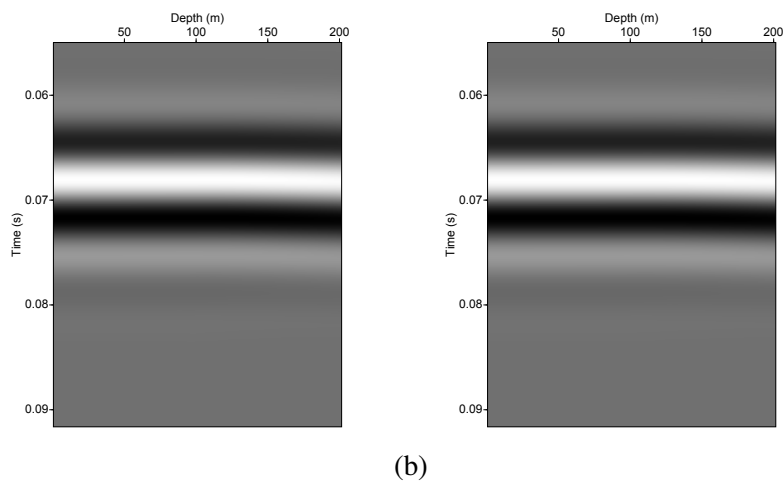


Figure 13: From the electric field well gather, -Figure 11(b)- we select a time window, displaying the difference of the electric field IR response between two different  $CO_2$  saturations, namely 5% and 65% (a); while in (b) the same result as in (a) is displayed, but the difference is taken considering  $CO_2$  saturations of 25% and 65%.

	<b>Sand</b>	<b>Sandstone I</b>	<b>Sandstone II</b>
$\phi$ (%)	35	20	12
$m$	2.05	1.70	1.7
$k_0$ (m <sup>2</sup> )	$10^{-11}$	$10^{-13}$	$4.32 \times 10^{-13}$
$k_s$ (Pa)	$36 \times 10^9$	$36 \times 10^9$	$36 \times 10^9$
$k_f$ (Pa)	$2.27 \times 10^9$	$2.27 \times 10^9$	$2.27 \times 10^9$
$k_{fr}$ (Pa)	$2.92 \times 10^9$	$14.40 \times 10^9$	$18.45 \times 10^9$
$G_{fr}$ (Pa)	$2.49 \times 10^9$	$14.08 \times 10^9$	$20.38 \times 10^9$
$v_S$ (m/s)	1104	2485	2909
$v_P$ (m/s)	2344	4017	4529
$v_{EM}$ (m/s)	$8.69 \times 10^5$	$1.03 \times 10^6$	$3.1 \times 10^6$
$\eta_w$ (Pa.s)	$1 \times 10^{-3}$	$1 \times 10^{-3}$	$1 \times 10^{-3}$
$\eta_g$ (Pa.s)	$1.8 \times 10^{-5}$	$1.8 \times 10^{-5}$	$1.8 \times 10^{-5}$
$\rho_s$ (Kg/m <sup>3</sup> )	$2.6 \times 10^3$	$2.6 \times 10^3$	$2.6 \times 10^3$
$\rho_w$ (Kg/m <sup>3</sup> )	$1 \times 10^3$	$1 \times 10^3$	$1 \times 10^3$
$\rho_g$ (Kg/m <sup>3</sup> )	1	1	1
$C_0$ (mol/L)	$1 \times 10^{-3}$	$1 \times 10^{-3}$	$1 \times 10^{-3}$
$\sigma$ (S/m)	$1.59 \times 10^{-3}$	$1.14 \times 10^{-3}$	$1.2 \times 10^{-4}$
$\zeta$ (V)	-0.07	-0.07	-0.07
$\kappa_w$	80	80	80
$\kappa_s$	4	4	4
$\kappa_g$	1	1	1
$T$ (K)	298	298	298

Table 1: *First two columns correspond to properties of the model described in Section 4.1, whilst the third one to the model used in Section 5. Seismic and electromagnetic velocities are calculated at the source peak frequency of 120 Hz.*

$\phi$ (%)	2	4	6	8	10	12
$k_0$ ( $m^2$ )	0	$1.09 \times 10^{-15}$	$2.79 \times 10^{-14}$	$1.28 \times 10^{-13}$	$2.50 \times 10^{-13}$	$4.32 \times 10^{-13}$
$\phi$ (%)	14	16	18	20	22	24
$k_0$ ( $m^2$ )	$6.86 \times 10^{-13}$	$1.02 \times 10^{-12}$	$1.46 \times 10^{-12}$	$2.00 \times 10^{-12}$	$2.66 \times 10^{-12}$	$3.46 \times 10^{-12}$

Table 2: Permeability values associated with porosity values ranging from 2 to 24 % computed using the empirical laws of Bourbié et al. (1987).

	<b>Top layer</b>	<b>Seal</b>	<b>Semispace</b>
$\phi$ (%)	20	20	20
$m$	1.7	1.7	1.7
$k_0$ (m <sup>2</sup> )	$0.5 \times 10^{-13}$	$0.4 \times 10^{-17}$	$0.5 \times 10^{-13}$
$k_s$ (Pa)	$36 \times 10^9$	$36 \times 10^9$	$36 \times 10^9$
$k_{fr}$ (Pa)	$5.40 \times 10^9$	$5.40 \times 10^9$	$5.40 \times 10^9$
$G_{fr}$ (Pa)	$5.80 \times 10^9$	$5.80 \times 10^9$	$5.80 \times 10^9$
$\rho_s$ (Kg/m <sup>3</sup> )	$2.6 \times 10^3$	$2.6 \times 10^3$	$2.6 \times 10^3$
$\kappa_s$	4	4	4
$k_w$ (Pa)	$2.27 \times 10^9$	$2.27 \times 10^9$	$2.27 \times 10^9$
$\eta_w$ (Pa.s)	$1 \times 10^{-3}$	$1 \times 10^{-3}$	$1 \times 10^{-3}$
$\rho_w$ (Kg/m <sup>3</sup> )	$1 \times 10^3$	$1 \times 10^3$	$1 \times 10^3$
$\kappa_w$	80	80	80
$C_0$ (mol/L)	$1 \times 10^{-3}$	$4 \times 10^{-2}$	0.4
$k_{CO_2}$ (Pa)	—	—	$25 \times 10^6$
$\eta_{CO_2}$ (Pa.s)	—	—	$1.5 \times 10^{-5}$
$\rho_{CO_2}$ (Kg/m <sup>3</sup> )	—	—	505
$\kappa_{CO_2}$	—	—	1
$\sigma$ (S/m)	$3.41 \times 10^{-4}$	$1.55 \times 10^{-3}$	* 0.12
$\zeta$ (V)	-0.07	-0.027	-0.0023
$T$ (K)	298	298	298

Table 3: *Properties of the model analysed in Section 7, the \* value for the electrical conductivity for the semispace corresponds to the water saturated case; the value for an effective fluid mixture with  $S_{CO_2}=65\%$  is equal to  $1.59 \times 10^{-2}$*

744 **Appendix A. One dimensional SHTE equations**

745 Consider Pride's equations in the equivalent form given in Zyserman et al. (2012)

$$746 \quad (\sigma + i\epsilon\omega)E - \nabla \times H + L(\omega)\eta_f k^{-1}(\omega) \left[ i\omega u^f - L(\omega)E \right] = -J_e^{ext}, \quad (\text{A.1})$$

$$747 \quad \nabla \times E + i\omega\mu H = -J_m^{ext}, \quad (\text{A.2})$$

$$748 \quad -\omega^2 \rho_b u^s - \omega^2 \rho_f u^f - \nabla \cdot \tau(u) = F^{(s)}, \quad (\text{A.3})$$

$$749 \quad -\omega^2 \rho_f u^s + \eta_f k^{-1}(\omega) \left[ i\omega u^f - L(\omega)E \right] + \nabla p_f = F^{(f)}, \quad (\text{A.4})$$

$$750 \quad \tau_{lm}(u) = 2G_{fr} \varepsilon_{lm}(u^s) + \delta_{lm} \left( \lambda_c \nabla \cdot u^s + \alpha K_{av} \nabla \cdot u^f \right), \quad (\text{A.5})$$

$$751 \quad p_f(u) = -\alpha K_{av} \nabla \cdot u^s - K_{av} \nabla \cdot u^f. \quad (\text{A.6})$$

752 Here  $\tau$  and  $\varepsilon$  are the stress and strain tensors,  $\rho_b = \phi\rho_f + (1 - \phi\rho_s)$  the bulk density,  $K_{av}$  is the fluid-storage  
753 coefficient and  $\alpha K_{av}$  is the Biot coupling coefficient, with  $\alpha = 1 - K_{fr}/K_s$ . In the seismic frequency regime,  
754 for most fluid saturated rocks, displacement currents can be safely neglected against conduction currents in the  
755 factor multiplying the electric field  $E$  in the first term of Eq.(A.1). Besides, as it has been demonstrated (Haines  
756 and Pride, 2006), if  $\eta L^2(\omega)/(\sigma k(\omega)) \ll 1$ , the electroosmotic feedback can be neglected in Biot's equations,  
757 and the latter decouples from Maxwell's equations. As we are dealing with just seismic shear sources, we set the  
758 electromagnetic sources, and the seismic source acting upon the fluid to zero,  $J_m^{ext} = J_e^{ext} = F^{(f)} = 0$ . Moreover,  
759 as  $F^{(s)}$  is considered to be a shearing force parallel to the  $y$  axis acting on a horizontal infinite plane upon a  
760 horizontally layered Earth, only solid displacements  $u^s = u_y^s(z, \omega)$  and fluid displacements  $u^f = u_y^f(z, \omega)$   
761 are possible. Therefore, we have  $E = E_y(z, \omega)$  and  $H = H_x(z, \omega)$ . Under these considerations, Eq.(A.1)-  
762 Eq.(A.6) can be rewritten as

$$763 \quad \sigma E - \partial_z H = i\omega\eta_f k^{-1}(\omega)L(\omega)u^f, \quad (\text{A.7})$$

$$764 \quad \partial_z E + i\omega\mu H = 0 \quad (\text{A.8})$$

$$765 \quad -\omega^2 \rho_b u^s - \omega^2 \rho_f u^f - \partial_z (G_{fr} u^s) = F^s, \quad (\text{A.9})$$

$$766 \quad -\omega^2 \rho_f u^s + i\omega\eta_f k^{-1}(\omega) u^f = 0, \quad (\text{A.10})$$



767 where we used  $\partial_z$  to denote the derivative with respect to  $z$ . Notice that in, although we have not considered  
768 it, in the seismic frequency range it is usual to take the low frequency limit for the electrokinetic coupling  
769 coefficient and the dynamic permeability; in this case we would have  $i\omega(\eta_f/k_0)L_0u^f$  as the r.h.s. of Eq. (A.7)  
770 and  $i\omega\eta_f k^{-1}(\omega) \approx (-\omega^2 g_0 + i\omega\eta/k_0)$  in Eq. (A.10); in this last expression  $g_0$  is the mass coupling coefficient.  
771 Finally, following Santos et al. (2004), we express  $F^{(s)}$  as

$$772 \quad F^{(s)}(z, \omega) = F(\omega)\partial_z\delta(z - z_f). \quad (\text{A.11})$$

773 Here  $F(\omega)$  is the Fourier transform of the source time signature, and the Dirac's delta derivative must be  
774 understood in the distributional sense; by  $z_f$  we denote the depth at which the source is located. The set of  
775 equations (A.7)-(A.10) must be completed with appropriate boundary conditions. For Maxwell's equations we  
776 use absorbing boundary conditions (ABC's) at the top and bottom boundaries, for Biot's equations we use the  
777 free boundary condition for the air-soil interface and again ABC's at the bottom boundary; see Zyserman et al.  
778 (2010, 2012) and references therein for details. In the present case they read, for Maxwell's equations and  
779 Biot's equations respectively:

$$780 \quad (1 - i)\sqrt{\frac{\sigma}{2\omega\mu}}E - \nu H = 0, \quad (\text{A.12})$$

781 where  $\nu$  takes the value -1 at the top boundary and the value 1 at the bottom one,

$$782 \quad \partial_z u^s = 0 \quad \text{Top boundary,} \quad (\text{A.13})$$

$$783 \quad -G_{fr}\partial_z u^s = i\omega(\rho_b - \rho_f^2/g(\omega))\sqrt{\frac{G_{fr}}{\rho_b - \rho_f^2/g(\omega)}}, \quad \text{Bottom boundary,} \quad (\text{A.14})$$

784 where  $g(\omega) = (1/\omega)\text{Im}(\eta/k(\omega))$ . As already stated, the system of equations (A.7)-(A.14) is solved by means  
785 of a finite elements procedure.