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1 **The effect of rock permeability and porosity on seismoelectric conversion: experiment and**  
2 **analytical modelling**

3  
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**Abstract**

The seismoelectric method is a modification of conventional seismic measurements which involves the conversion of an incident poroelastic wave to an electromagnetic signal that can be measured at the surface or down a borehole. This technique has the potential to probe the physical properties of the rocks at depth. The problem is that we currently know very little about the parameters which control seismoelectric conversion and their dependence on frequency and permeability, which limits the development of the seismoelectric method. The seismoelectric coupling coefficient indicates the strength of seismoelectric conversion. In our study, we focus on the effects of the reservoir permeability, porosity and frequency on the seismoelectric coupling coefficient through both experimental and numerical modelling. An experimental apparatus was designed to record the seismoelectric signals induced in water-saturated samples in the frequency range from 1 kHz to 500 kHz. The apparatus was used to measure seismoelectric coupling coefficient as a function of porosity and permeability. The results were interpreted using a micro-capillary model for the porous medium to describe the seismoelectric coupling. The relationship between seismoelectric coupling coefficients and the permeability and porosity of samples were also examined theoretically. The combined experimental measurements and theoretical analysis of the seismoelectric conversion has allowed us to ascertain the effect of increasing porosity and permeability on the seismoelectric coefficient. We found a general agreement between the theoretical curves and the test data, indicating that seismoelectric conversion is enhanced by increases in porosity over a range of different frequencies. However, seismoelectric conversion has a complex relationship with rock permeability, which changes with frequency. For the low-permeability rock samples ( $0-100 \times 10^{-15} \text{ m}^2$ ), seismoelectric coupling strengthens with the increase of permeability logarithmically in the low frequency range (0-10 kHz); in the high frequency range (10-500 kHz), the seismoelectric coupling is at first enhanced, with small increases of permeability leading to small increases in size in electric coupling. However, continued increases of permeability then lead to a slight decrease in size and image conversion again. For the high-permeability rock samples ( $300 \times 10^{-15} \text{ m}^2 - 2200 \times 10^{-15} \text{ m}^2$ ), the seismoelectric conversion shows the same variation trend with low-permeability samples in low frequency range; but it monotonically decreases with permeability in the high frequency range. The experimental and theoretical results also indicate that seismoelectric conversion seems to be more sensitive to the changes of low-permeability samples. This observation suggests that seismic conversion may have advantages in characterizing low permeability reservoirs such as tight gas and tight oil reservoirs and shale gas reservoirs.

73 **Key words: porosity, permeability, seismoelectric conversion, coupling coefficients, frequency**  
74 **dependent**

75

## 76 **Introduction**

77

78 The passage of poroelastic waves through a porous media, such as a rock, sets up local fluid flows,  
79 and those flows lead to the generation of an electrical potential (e.g., Walker et al. 2014). This  
80 coupling of poroelastic waves and electrical potential generation is called the seismoelectric effects,  
81 and its size is characterised by the streaming potential coefficient,  $C_{sp}$  (also called the seismoelectric  
82 coupling coefficient,  $L(w)$ ). The seismoelectric effects have been investigated by experimental  
83 research and numerical modelling on various rock models (Glover & Jackson 2009; Bordes et al.  
84 2009; Schakel et al. 2011a, 2011b, 2011c, 2012; Sénéchal & Bordes 2012; Jougnot et al. 2013;  
85 Roubinet et al. 2016). The streaming potential coefficient  $L(w)$  depends on the microstructural and  
86 transport properties of the porous media, or in our case, the rocks that compose the reservoir (e.g.,  
87 Revil et al. 1999; Glover et al. 2010).

88

89 Different theoretical models of the streaming potential coefficient have been proposed. Packard (1953)  
90 presented a model for the frequency-dependent streaming potential coefficient for capillary tubes.  
91 Pride (1994) proposed the seismoelectric coupling coefficient for porous medium. Reppert et al. (2001)  
92 revised the Packard model by using the low- and high-frequency approximations of the Bessel  
93 functions, which is almost the same with the model of Packard. Since the streaming potential is  
94 induced at the solid-fluid phase, the surface conductivity of the solid surface affect this electric  
95 potential, and the surface conductivity of the rock has impact on the apparent effect of the  
96 permeability (Jouniaux et al. 2000). Jouniaux and Bordes (2012) summarized the experiments and  
97 theories of frequency-dependent streaming potential and indicated that the transition frequency was  
98 dependent of the permeability. It has also been recognised that the seismoelectric coupling coefficient  
99 is a function of permeability and porosity (Jouniaux & Pozzi 1995), thus the rock properties can be  
100 inferred from the seismoelectric conversion. However, the relationships between porosity,  
101 permeability and  $L(w)$  have not been studied in sufficient depth for their exact forms to be known. It is  
102 impossible to discuss permeability and porosity separately since these two parameters are  
103 interdependency properties of rocks. Permeability of porous media is usually expressed as function of  
104 some physical properties of the interconnected pore system such as porosity (Dullien, 1979). Because  
105 of complexity of the pore channels, few simple functions can exist. If the relationship between  
106 porosity, permeability and  $L(w)$  can be well understood, there is potential for using measurements of  
107  $L(w)$  from seismoelectric field measurements (Thompson et al. 2007; Dupuis et al. 2009) to obtain a  
108 reservoir's porosity and permeability, offering an alternative method for providing these critical  
109 values which does not depend on drilling wells or analysing cores.

110

111 The streaming potential and electro-osmotic pressure on sandstones, limestones and glass bead  
112 samples in the low frequency range (0-100 Hz) were measured by [Jouniaux and Pozzi \(1995\)](#) and  
113 [Pengra \(1999\)](#). Their results indicated that the fluid permeability of the material is also related to  
114 electro-kinetic effects. This has been confirmed for steady-state streaming potential and permeability  
115 measurements by [Glover et al. \(2006\)](#), and lead to the RGPZ model for permeability prediction. It was  
116 subsequently generalised by [Walker and Glover \(2010\)](#) in a study of characteristic length scales and  
117 scaling constants in fluid permeability prediction, and also to a link between time-dependent electro-  
118 kinetic coupling and fluid permeability.

119

120 Meanwhile, [Mikhailov et al. \(2000\)](#) detected the Stoneley-wave-induced electric field in a borehole  
121 (150 Hz centre frequency), and showed that borehole electro-seismic measurements could be used to  
122 characterize permeable zones. Since it was becoming clear that frequency-dependent  $L(\omega)$  was needed  
123 to process and understand seismoelectric field data, it was thought necessary to make well-controlled  
124 measurements of  $L(\omega)$  as a function of frequency in the laboratory. Several frequency-dependent  $L(\omega)$   
125 measurement apparatuses were designed and built for use in the frequency range 1 Hz to 1 kHz by  
126 applying a harmonically varying flow ([Tardif et al. 2011](#); [Glover et al. 2012a](#); [Glover et al. 2012b](#)),  
127 whose results have shown a clear relationship between transition frequency, grain size and steady-  
128 state fluid permeability. The laboratory measurements of the streaming potential properties of  
129 minerals was performed by [Morgan et al. \(1989\)](#), results indicated that the anomalies of the streaming  
130 potential can be used to predict the earthquake phenomena. [Schoemaker et al. \(2007, 2008\)](#) measured  
131 the streaming potential and dynamic permeability by using a Dynamic Darcy Cell (DCC) with a  
132 mechanical shaker in the range from 5 to 200 Hz. but the permeability is given by Darcy's law, which  
133 does not reflect the permeability through the potential acquired by seismoelectric measurements, and  
134 the relationship between permeability and seismoelectric coupling has not been studied.  
135 Independently, [Wang et al. \(2010\)](#) analysed the relationship between the permeability and streaming  
136 potential of rocks in low frequency range (0-70 Hz), indicating that  $L(\omega)$  has a strong predictive  
137 relationship with conventional gas permeability, while [Luong and Sprik \(2013\)](#) carried out streaming  
138 potential and electro-osmosis measurements (100 Hz -100 kHz) to characterize the zeta potential and  
139 the average pore size of porous materials on 6 unconsolidated samples. [Schoemaker et al. \(2012\)](#)  
140 experimentally validated a electrokinetic formulation of the streaming potential of [Pride \(1994\)](#). The  
141 streaming potential is measured in a frequency band ranging from 5Hz up to 150 Hz, and the  
142 numerical modelling use a frequency of 500 kHz. [Guan et al. \(2013\)](#) proposed a method of obtaining  
143 reservoir permeability using the seismoelectric well logging data of a fluid-saturated porous formation  
144 using data in the frequency range 0 Hz -2 kHz, finding that the amplitude ratio of the converted  
145 electric field to the pressure (what he called the REP, but which is formally the same as  $L(\omega)$  is  
146 sensitive to porosity, while the tangent of the REP's phase is sensitive to permeability. [Zhu et al.](#)

147 (2015) performed seismoelectric experiments on a porous quartz-sand sample with anisotropic  
148 permeability (20 kHz to 90 kHz) and showed that  $L(w)$  depends directly on permeability, inferring  
149 that the amplitudes of measured seismoelectric signals may, in principle, be used to determine the  
150 permeability in a given formation. Furthermore they also measured seismoelectric anisotropy and  
151 found that it is correlated with a sample's anisotropic permeability.

152

153 Though the rock properties (permeability, porosity, etc.) have been investigated based on  
154 seismoelectric conversion and qualitative results have been acquired from the previous research, the  
155 settled relationship to describe the interdependence of  $L(w)$  and permeability still needs to be  
156 identified more clearly. This is especially true at the higher frequencies because most research has  
157 been carried out at low-frequencies (less than 2 kHz in the above references). Luong and Sprik (2013)  
158 and Zhu et al. (2015) have made measurements of electro-kinetic parameters in the 'high' frequency  
159 range (but to less than 100 kHz). Luong and Sprik (2013) focused on determining the zeta potential  
160 and pore size by seismic methods and did not take account of porosity and permeability, while Zhu et  
161 al. (2015) pointed out that the rock permeability is related directly to seismoelectric signals, but did  
162 not provide the formal relationship between permeability and seismoelectric conversion coefficients.

163

164 In this study, we have performed seismoelectric measurements of rock samples as a function of  
165 porosity and permeability using natural and artificial samples at three different frequencies (1 kHz, 10  
166 kHz and 500 kHz, we define 0-10 kHz as the low frequency range and 10 kHz-500 kHz as the high  
167 frequency range in this paper). The full theoretical relationship between  $L(w)$ , permeability and  
168 porosity has been calculated based on a micro-capillary model of porous media. We give out the  
169 effect of permeability on transition frequency, which indicates that the transition frequency is  
170 inversely related to permeability. And the change of transition frequency leads to the complicated  
171 relationship between the permeability and  $L(w)$  divided into low frequency and high frequency ranges.  
172 In addition, we present the quantitative relationship between  $L(w)$  and permeability for a constant  
173 porosity with numerical and experimental data for the first time, including the dependence of  $L(w)$  on  
174 both high permeability and low permeability rocks. This work now gives us experimental evidence for  
175 the variation of seismoelectric coupling for rock samples as a function of frequency, porosity and  
176 permeability, as well as providing a validated theoretical model for seismoelectric coupling for rock  
177 samples as a function of frequency, porosity and permeability.

178

### 179 **Experimental method and samples**

180 A schematic of the apparatus for measuring  $L(w)$  is shown in Figure 1. The apparatus operates in a  
181 water tank (200×150×100 cm) (Zhu et al., 2008; Schakel et al., 2011a, 2011b, 2011c; Peng et al.  
182 2016; 2017). A compressional wave ( $P$ -wave) piezoelectric transducer is used as the seismic source.  
183 The resulting  $P$ -wave propagates through fully saturated rock samples causing a local travelling

184 pressure disturbance, which perturbs the electric double layer (EDL). This perturbation leads to a  
185 polarisation of charge which gives rise to a measurable macroscopic streaming potential across the  
186 porous medium, and which can be measured as an electric signal by the electrodes shown in [Figure 1](#).

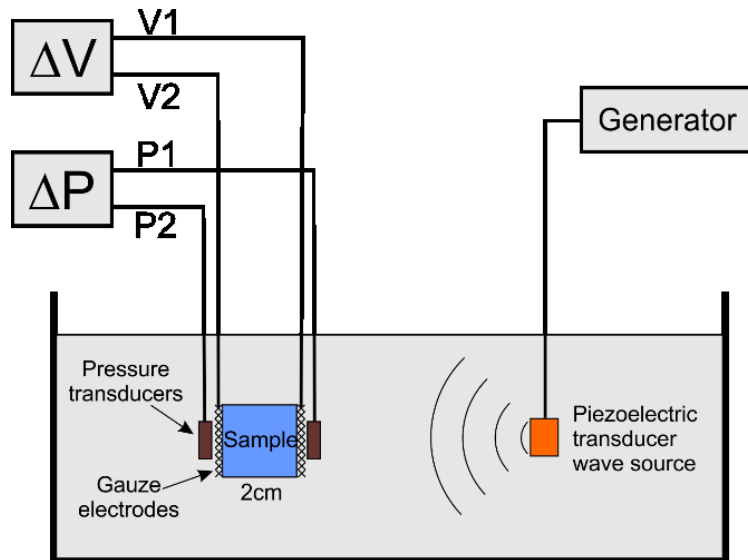
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188 In our study, we only consider the P-wave in our experiment and theory, and the P-wave converted S-  
189 waves when the P-wave hit the rock sample are considered to be ignored. Because (1) the P-waves in  
190 our experiment are nearly perpendicular incident, hence, the conversion to shear waves is not strong;  
191 (2) shear waves cannot propagate in fluids, so we think the converted shear wave is very weak even  
192 though the conversion happens; (3) the weak shear waves cannot induce the enough strong  
193 seismoelectric conversion which can be detected in our experiment. Because according to our results,  
194 the P-wave induced SE signals recorded in our experiment is not that strong, let along the S-wave  
195 induced SE signals.

196

197 The fluid in the water tank is an aqueous solution of NaCl with a concentration of  $1.5 \text{ g/dm}^3$ , the  
198 corresponding conductivity of the fluid is  $0.24 \text{ S/m}$ , which is measured with conductivity meter. Two  
199 gauze Ag/AgCl disc electrodes with a diameter of 2 cm (V1 and V2) and two pressure transducers (P1  
200 and P2) are placed on each side of the rock sample (touching the rock samples) to record the  
201 seismoelectric signal and the acoustic signal, respectively. Both of the two electrode potentials (V1  
202 and V2) are given relative to the common ground. And the two electrodes touch the rock samples  
203 when measurements are performed, so the measured SE signals are the coseismic fields in this paper  
204 ([Peng et al. 2017](#)). The instantaneous seismoelectric (streaming) potential  $\Delta V$  is obtained by taking  
205 the difference between the potentials of the electrodes on each side of the rock sample ( $\Delta V = V1 - V2$ ).  
206 The transient pressure difference,  $\Delta P$ , across the sample at any instant is obtained by taking the  
207 difference of the measurements made by transducers P1 and P2 ( $\Delta P = P1 - P2$ ). The ratio of  $\Delta V / \Delta P$  is  
208 the experimental value of the  $L(w)$  of the rock sample ([Zhu et al., 2008](#)).

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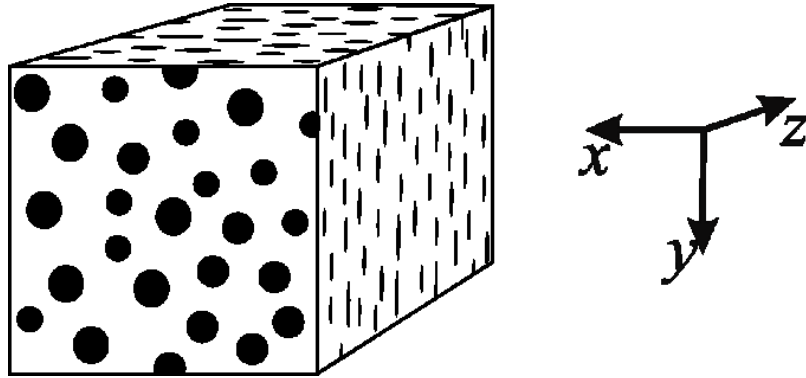


210  
211 **Figure 1.** Schematic diagram of the seismoelectric apparatus.  
212

213 Seismoelectric measurements were made on 28 natural rock samples (cylindrical core plugs with a  
214 length of 2 cm and a diameter of 2 cm) and 4 artificial fractured samples with controlled crack  
215 densities (cubic, 2×2×2 cm). The 28 natural samples include 6 granites and 22 sandstones and can be  
216 considered as isotropic samples, with a range of permeabilities from  $0.001 \times 10^{-15} \text{ m}^2$  to  $69.488 \times 10^{-15}$   
217  $\text{m}^2$  and porosities from 9.91% to 16.3%, as shown in [Table 1](#). The porosity is effective porosity  
218 acquired by helium porosimetry measurements. The gas volume method is used for determination of  
219 porosity according to Boyle's law. The artificial fractured samples were prepared using the  
220 manufacturing process described in [Ding et al. \(2013\)](#), [Ding et al. \(2014a; 2014b\)](#) and [Zhu et al.](#)  
221 [\(2015\)](#). The sand minerals and polymer materials as cracks are placed into the mould, and after the  
222 sample prepared in the mould, it was left to dry in a constant temperature oven for weeks. The sample  
223 was sintered in a furnace and the high molecular material discs were then decomposed and drained  
224 out, leaving voids as fractures. The densities of the four artificial samples with induced crack are 0%,  
225 3%, 6%, 9%, respectively. Each has a geometry represented in the schematic diagram shown in [Figure](#)  
226 [2](#). Since the volume of the induced cracks is small compared to the volume of the matrix porosity, all  
227 of the induced samples can be assumed to have the same porosity, which in our measurements was  
228  $(23.8 \pm 0.5)\%$ .

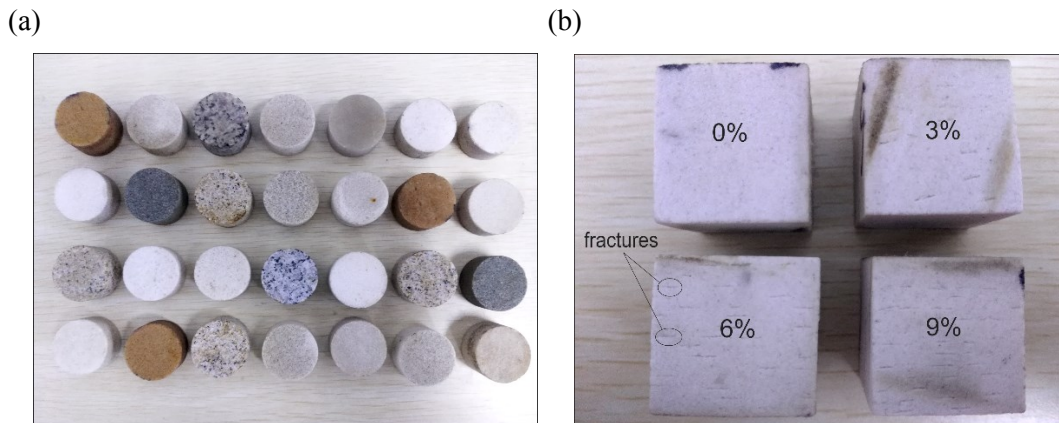
229  
230 Despite the change in porosity being insignificant, the induced cracks have a considerable effect on  
231 the size and anisotropy of the sample permeability. Measurements made in the three orthogonal  
232 directions on the artificial sample with 0% crack density all give the same value because the sample is  
233 isotropic. However, the other 3 artificial samples include induced fractures in a preferential direction.  
234 Measurements made on these samples in the three orthogonal directions provide three different  
235 seismoelectric coupling coefficients, one for each direction, due to the anisotropy of the samples.

236 Consequently, the 4 cubic samples, each with the same porosity, provide 10 independent  $L(w)$   
 237 measurements, each associated with a different permeability. **Figure 3** shows all of the experimental  
 238 samples used in this work, core samples on the left-hand side, and the induced crack samples on the  
 239 right-hand side. The presence of cracks can just be observed in these, as indicated in the figure.  
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**Figure 2.** Schematic of the artificial fractured sample. The porosity of all the fractured samples is 23.8%. The cracks are parallel to  $xy$  plane, and vertical to  $xz$  and  $yz$  planes.



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**Figure 3.** Photos of (a) the 28 natural samples, and (b) the 4 artificial fractured samples, in which the artificial fractures can just be seen.

**Table 1.** Porosity and permeability of the rock samples used in the study.

Sample Number	Porosity (%)	Permeability ( $\times 10^{-15}m^2$ )	Sample Number	Porosity (%)	Permeability ( $\times 10^{-15}m^2$ )
N1	10.84	0.001	N15	11.75	0.263
N2	13.44	0.002	N16	10.88	0.506
N3	12.40	0.004	N17	10.37	0.624
N4	12.15	0.007	N18	12.24	1.192
N5	15.98	0.020	N19	14.11	2.161
N6	10.96	0.023	N20	10.00	3.590
N7	12.16	0.025	N21	12.60	11.450
N8	10.93	0.034	N22	16.30	13.135

N9	13.03	0.079	N23	9.92	13.291
N10	9.91	0.090	N24	12.54	19.090
N11	12.92	0.117	N25	14.20	32.828
N12	13.62	0.122	N26	14.23	48.100
N13	11.29	0.134	N27	16.17	57.400
N14	15.05	0.259	N28	11.94	69.488
A1	23.8	310.79	A3y	23.8	1696.30
A2x	23.8	847.21	A3z	23.8	370.43
A2y	23.8	876.10	A4x	23.8	2053.27
A2z	23.8	325.53	A4y	23.8	2164.32
A3x	23.8	1509.34	A4z	23.8	427.39

252 **Note.** Samples labelled N are natural rock samples, and those labelled A are artificial fractured  
253 sandstones. A1 is the sandstone without cracks, while the labels  $x$ ,  $y$  and  $z$  represent the  
254 direction of wave propagation as defined in [Figure 3](#). The porosity is effective porosity  
255 acquired by helium porosimetry measurements. Permeability is Klinkenberg-corrected  
256 permeability ([Klinkenberg 1941](#)).  
257

## 258 **Theoretical development**

259 This section contains the major theoretical development in this paper. The first subsection references  
260 existing work and applies it to the capillary bundle model which we use. Subsequent sections develop  
261 the model further in order to obtain the frequency-dependent relationship between permeability,  
262 porosity and the seismoelectric coupling coefficient. And the theory we talk about in this paper is only  
263 the P-wave field. The seismoelectric measurements can be impacted by the conductivity of the rock  
264 surface ([Alkafeef & Alajmi 2006](#); [Wang et al. 2015](#)), in order to simplify the situation, we neglect the  
265 effects of surface conductivity on the seismoelectric coupling in our theory.  
266

### 267 *Theory of the electric double layer in porous media*

268 In this work we have modelled porous rock as a parallel aligned bundle of capillary of varying radius.  
269 The correspondence between the microscopic parameters of this capillary bundle model (the number  
270 of capillaries,  $n_0$ , and the capillary radius,  $r_0$ ) and the macroscopic reservoir parameters (the rock  
271 porosity,  $\phi$ , and permeability,  $k$ ) can be expressed as ([Kozeny 1927](#))

$$272 \quad \phi = n_0 \pi r_0^2, \text{ and} \quad (1)$$

$$273 \quad k = \frac{n_0 \pi r_0^4}{8}. \quad (2)$$

274 The ion concentration distribution of a multi-component electrolyte in a capillary satisfies the  
275 Boltzmann equation ([Harris 1971](#))

$$276 \quad C_i = C_0 \exp\left(\frac{-Z_i e \psi}{k_B T}\right), \quad (3)$$

277 where  $C_i$  is the concentration of component  $i$ ,  $C_0$  is the initial concentration of the solution,  $e=1.6 \times$   
278  $10^{-19}$  is the fundamental charge,  $n_A$  is Avogadro number,  $Z_i$  is the valence of component  $i$ ,  $\psi$  is a  
279 potential function of the distance between the wall of capillary and ions ( $\psi(r)$ ),  $k_B$  is Boltzmann

280 constant.  $T$  is absolute temperature (in K). The electric density  $\rho_e$  for a solution with a symmetric  
 281 cation-anion of valence  $Z$ - $Z$  in a capillary is

$$282 \quad \rho_e = en_A Z(C_+ - C_-) = 2FZC_0 \sinh\left(\frac{Ze\psi}{k_B T}\right). \quad (4)$$

283 The capillary charge satisfies the Poisson equation (Probstein 1994):

$$284 \quad \frac{1}{r} \frac{\partial}{\partial r} \left( r \frac{\partial \psi}{\partial r} \right) = -\frac{\rho_e}{\varepsilon} = -\frac{en_A}{\varepsilon} \sum_i Z_i C_0 \exp\left(\frac{-Z_i e \psi}{k_B T}\right), \quad (5)$$

285 where  $\varepsilon$  is the dielectric constant of the solution,  $r$  is the coordinate variable along capillary radius of  
 286 cylindrical coordinates, representing the distance between the center of capillary and ions.

287  
 288 In the  $Z$ - $Z$  valence symmetric salt electrolyte solution,  $Z_i$  is the valence of component  $i$ ,  $Z_i$   
 289 includes  $Z_+$  and  $Z_-$ ,  $Z_+ = Z$ ;  $Z_- = -Z$ , and the Poisson equation can be rewritten as

$$290 \quad \frac{1}{r} \frac{\partial}{\partial r} \left( r \frac{\partial \psi}{\partial r} \right) = -\frac{en_A}{\varepsilon} \sum_i Z_i C_0 \exp\left(\frac{-Z_i e \psi}{k_B T}\right) = \frac{2en_A Z C_0}{\varepsilon} \sinh\left(\frac{Ze\psi}{k_B T}\right). \quad (6)$$

291 If the thickness of the EDL,  $d$ , is much smaller than the capillary radius  $r_0$  ( $r_0 \gg d$ ),  $\Phi$  is very small  
 292 over the internal area of the capillary. Hence, we have

$$293 \quad \sinh\left(\frac{Ze\psi}{k_B T}\right) \approx \frac{Ze\psi}{k_B T}. \quad (7)$$

294 The Poisson equation can then be simplified as

$$295 \quad \frac{1}{r} \frac{\partial}{\partial r} \left( r \frac{\partial \psi}{\partial r} \right) = -\frac{\rho_e}{\varepsilon} = \frac{2en_A Z C_0}{\varepsilon} \sinh\left(\frac{Ze\psi}{k_B T}\right) = \frac{\psi}{(d^2)}, \quad (8)$$

$$296 \quad \text{where: } d = \sqrt{\frac{\varepsilon k_B T}{2e^2 Z^2 n_A C_0}}.$$

297 Substituting the following boundary conditions into equation 8 :  $r = 0, \frac{\partial \psi}{\partial r} = 0$ ;  $r = r_0, \psi = \zeta$ ,  $\zeta$  is  
 298 the zeta potential of electric double layer. Then the solution of the potential distribution in equation 8  
 299 can be obtained as

$$300 \quad \psi = \zeta \exp\left(-\frac{r_0 - r}{d}\right). \quad (9)$$

301 In this work we calculate the zeta potential using the empirical equation (Pride & Morgan, 1991). We  
 302 assume that the zeta potential could be constant with the concentration (Fiorentino et al. 2016), also  
 303 we assume that the zeta potential is the same for all the natural samples since the fluid concentration  
 304 is not changed in this study.

$$305 \quad \zeta = 0.008 + 0.026 \log_{10} C_0. \quad (10)$$

306 However, if a more analytical approach is required, a fully theoretical treatment of the zeta potential  
 307 has been published (Glover and Déry, 2010; Glover et al., 2012a, 2012b) and validated on a large  
 308 database of experimental measurements (Walker and Glover, 2018; Glover 2018).

309 The ion concentration distribution given as Equation (3), can be written as

$$310 \quad C_i = C_0 \exp\left(\frac{-Z_i e \psi}{k_B T}\right) = C_0 \sum_{n=0}^{\infty} \frac{(-Ze\zeta)^n}{(k_B T)^n n!} \exp\left[-\frac{n(r_0 - r)}{d}\right] \quad (11)$$

311 using Taylor series expansion.  $C_i$  includes  $C_+$  and  $C_-$  :

312 
$$C_+ = C_0 \exp\left(\frac{-Ze\psi}{k_B T}\right), \quad (12)$$

313 
$$C_- = C_0 \exp\left(\frac{Ze\psi}{k_B T}\right). \quad (13)$$

314

315 *Frequency-dependent hydraulic flow velocity with electro-kinetic influence*

316 Assuming that the porous fluid satisfies the convective form of the incompressible Navier-Stokes  
317 equation with an electro-kinetic external source (Deissler 1976)

318 
$$\eta \nabla^2 \mathbf{u} - \rho_f \left[ \frac{\partial \mathbf{u}}{\partial t} + (\mathbf{u} \cdot \nabla) \mathbf{u} \right] = \nabla P - \rho_e \mathbf{E}, \quad (14)$$

319 where  $\rho_f$  is the fluid density,  $P$  is the fluid pressure,  $\mathbf{u}$  is the flow velocity, and  $\eta$  is the fluid viscosity.

320 The electric field intensity  $\mathbf{E} = -\nabla U$  in the quasi-static limit of the Maxwell equations, where  $U$  is the  
321 streaming potential (Revil et al., 2015).

322

323 In our capillary model, fluid flow is incompressible and constrained to the  $z$ -direction. Hence,  $\nabla \cdot \mathbf{u} =$   
324  $0$ ,  $\mathbf{u} = u_z \mathbf{e}_z$ , where  $\mathbf{e}_z$  is the unit vector in the capillary axial direction, such that we may write  
325  $\frac{\partial \mathbf{u}(r,z)}{\partial z} = 0$ . Therefore,  $u_z$  is a function of  $r$ , and  $(\mathbf{u} \cdot \nabla) \mathbf{u}$  in the Navier-Stokes equation can be  
326 ignored.

327

328 In the following analysis, there are certain further assumptions: (i) that the pressure gradient of an  
329 pressure wave field in the  $z$ -direction of the porous medium is  $\frac{\partial P(z,t)}{\partial z}$ , (ii) that the pressure field is  
330 harmonic,  $P = P_0 e^{-i\omega t}$ ,  $\omega = 2\pi f$ , (iii) that the pressure induced streaming potential is also harmonic,  
331  $U = U_0 e^{-i\omega t}$ , (iv) that fluid flow is incompressible, so  $\nabla \cdot \mathbf{u} = 0$ , and (v) non-homogenous vibration  
332 of the solid-fluid interface caused by pressure waves can be ignored. Under these conditions, the  
333 Navier-Stokes equation can be expressed as

334 
$$\nabla^2 \mathbf{u} + K^2 \mathbf{u} = \frac{1}{\eta} (\nabla P + \rho_e \nabla U), \quad (15)$$

335 where  $K^2 = \frac{i\omega \rho_f}{\eta}$ .

336 Then substituting Equation (4) into this equation and multiplying both sides by  $r^2$  gives

337 
$$r \frac{\partial}{\partial r} \left( r \frac{\partial \mathbf{u}}{\partial r} \right) + K^2 r^2 \mathbf{u} = \frac{r^2}{\eta} \frac{\partial P}{\partial z} - \frac{2r^2 e n_A Z C_0}{\eta} \sinh\left(\frac{Ze\psi}{k_B T}\right) \frac{\partial U}{\partial z}. \quad (16)$$

338 When this equation is compared with the definition of a Bessel equation, when  $K^2 \neq 0$ , and taking  
339 account of the boundary condition: (at  $r = 0$ ,  $\mathbf{u}$  has a non-zero limited value and at  $r = r_0$ ,  $\mathbf{u} = 0$ )  
340 we can obtain the solution of Equation (16) as

341 
$$\mathbf{u}(r) = \frac{1}{K^2 \eta} \left[ 1 - \frac{J_0(Kr)}{J_0(Kr_0)} \right] \frac{\partial P}{\partial z} + \frac{\varepsilon \zeta}{\eta(1+K^2 d^2)} \left[ \frac{J_0(Kr)}{J_0(Kr_0)} - e^{\frac{r_0-r}{d}} \right] \frac{\partial U}{\partial z} \quad (17)$$

342 This is the distribution of the fluid velocity in the capillary, and  $J_0(Kr) = \sum_{m=0}^{\infty} \frac{(-1)^m}{(m!)^2} \left(\frac{Kr}{2}\right)^{2m}$  is the  
343 zero order Bessel function of the first kind.

344

345 *Frequency-dependent electrical flow with hydraulic flow influence*346 The molar flux of component  $i$ , in a dilute solution, is (Probstein 1994)

347 
$$q_i = -v_i Z_i C_i \frac{\partial U}{\partial z} + C_i \mathbf{u}, \quad (18)$$

348 where  $v_i$  is the ionic mobility of component  $i$ . The value of  $C_i$  is given by Equation (3). The  
 349 distribution of fluid velocity  $\mathbf{u}$  in the capillary is given by Equation (17). The molar flux is the sum of  
 350 the contribution of ionic migration and diffusion. The first term of Equation (18) is the molar flux  
 351 caused by ion migration in an electric field; and the second term indicates the molar flux caused by  
 352 ion diffusion. The second term relates to the fluid velocity in the capillary which mainly includes the  
 353 flow caused by pressure and the flow velocity caused by the forced diffusion of the fluid particles in  
 354 the electric field, i.e., Equation (17). Therefore, the flux of the ionic diffusion is mainly composed of  
 355 the flux generated by the diffusion of the fluid ions induced by the pressure and the molar flux  
 356 generated by the forced diffusion of the ions under the electric field.

357 
$$q_i = -v_i Z_i C_i \frac{\partial U}{\partial z} + \frac{C_i}{K^2 \eta} \left[ 1 - \frac{J_0(Kr)}{J_0(Kr_0)} \right] \frac{\partial P}{\partial z} + \frac{C_i \varepsilon \zeta}{\eta(1+K^2 d^2)} \left[ \frac{J_0(Kr)}{J_0(Kr_0)} - e^{-\frac{r_0-r}{d}} \right] \frac{\partial U}{\partial z}. \quad (19)$$

358 Integrating this equation along the capillary cross-section, the ion current density of component  $i$  for  
 359 one capillary is

360 
$$w_i = \int_0^{r_0} q_i(r) 2\pi r dr. \quad (20)$$

361 For a parallel aligned capillary bundle model, the ion current density per unit area is  $W_i = n_0 w_i$ . We  
 362 can obtain the total ion current density per unit area by integration over the radius of the capillary  
 363 tube, as

364 
$$W_i = n_0 \int_0^{r_0} q_i(r) 2\pi r dr, \quad (21)$$

365 to give

366 
$$W_i = 2\pi n_0 \left[ \int_0^{r_0} -v_i Z_i C_i r dr \frac{\partial U}{\partial z} + \int_0^{r_0} \frac{C_i}{K^2 \eta} \left[ 1 - \frac{J_0(Kr)}{J_0(Kr_0)} \right] r dr \frac{\partial P}{\partial z} \right. \\ \left. + \int_0^{r_0} \frac{C_i \varepsilon \zeta}{\eta(1+K^2 d^2)} \left[ \frac{J_0(Kr)}{J_0(Kr_0)} - e^{-\frac{r_0-r}{d}} \right] r dr \frac{\partial U}{\partial z} \right] \quad (22)$$

367 According to the definition of electric current density, the electric current density in capillary,  $I$ , is  
 368 (Bockris & Reddy 1970)

369 
$$I = \sum_i e n_A Z_i q_i. \quad (23)$$

370 Integrating this equation over the capillary cross-section, the electric current density of the whole  
 371 capillary bundle can be expressed as

372 
$$j = e n_A Z (W_+ - W_-), \quad (24)$$

373 or,

374 
$$j = e n_A Z (n_0 \int_0^{r_0} q_+(r) 2\pi r dr - n_0 \int_0^{r_0} q_-(r) 2\pi r dr) = F Z n_0 \int_0^{r_0} [q_+(r) - q_-(r)] 2\pi r dr. \quad (25)$$

375 This equation can be rewritten as

376 
$$j = -L(\omega) \frac{\partial P}{\partial z} - \sigma(\omega) \frac{\partial U}{\partial z}, \quad (26)$$

377 where

378 
$$L(\omega) = -2\pi en_A Z n_0 \left[ \int_0^{r_0} \frac{C_+}{K^2 \eta} \left[ 1 - \frac{J_0(Kr)}{J_0(Kr_0)} \right] r dr - \int_0^{r_0} \frac{C_-}{K^2 \eta} \left[ 1 - \frac{J_0(Kr)}{J_0(Kr_0)} \right] r dr \right], \quad (27)$$

379 
$$\alpha = 2\pi en_A Z n_0 \left[ \int_0^{r_0} v_{+Z} C_+ r dr - \int_0^{r_0} v_{-Z} C_- r dr \right], \text{ and} \quad (28)$$

381 
$$\beta = -2\pi en_A Z n_0 \left[ \int_0^{r_0} \frac{C_+ \varepsilon \zeta}{\eta(1+K^2 d^2)} \left[ \frac{J_0(Kr)}{J_0(Kr_0)} - e^{\frac{r_0-r}{d}} \right] r dr + \int_0^{r_0} \frac{C_- \varepsilon \zeta}{\eta(1+K^2 d^2)} \left[ \frac{J_0(Kr)}{J_0(Kr_0)} - e^{\frac{r_0-r}{d}} \right] r dr \right] \quad (29)$$

382 and, (30)

383

384 The term  $\sigma(\omega) \frac{\partial U}{\partial z}$  is the electric conducting current caused by the movement of charged particles  
 385 themselves (including ion migration and ion diffusion).  $\sigma(\omega)$  is defined as the composite  
 386 conductivity, with  $\alpha$  and  $\beta$  representing the coefficient of the electric current produced by ion  
 387 migration in the pore fluid and ion diffusion under the electric field force, respectively. The term  
 388  $L(\omega) \frac{\partial P}{\partial z}$  is the electric current produced by the pressure field in the pore fluid, and  $L(\omega)$  is defined as  
 389 the seismoelectric coupling coefficient ( $C_{sp}$ ).

390

391 The radius in the capillary model is  $r_0 = \sqrt{8k/\phi}$ , which is obtained by solving Equation (1) and  
 392 Equation (2) to eliminate  $n_0$  (i.e.,  $n_0 = \frac{\phi}{\pi r_0^2} = \frac{\phi^2}{8k\pi}$ ). Now, recalling,  $K = \sqrt{\frac{i\omega\rho_f}{\eta}}$ ,  $d = \sqrt{\frac{\varepsilon k_B T}{2e^2 Z^2 n_A C_0}}$ , and  
 393 the ion concentration distribution from Equation (12) and (13), combining Equation (28) and Equation  
 394 (29) we have

395 
$$\alpha = 2\pi en_A Z^2 \frac{\phi^2}{8k\pi} \left[ \int_0^{\sqrt{8k/\phi}} (v_+ C_+ - v_- C_-) r dr \right] \quad (31)$$

396 
$$\beta = -2\pi en_A Z \frac{\phi^2}{8k\pi} \left[ \int_0^{\sqrt{8k/\phi}} \frac{(C_+ - C_-) \varepsilon \zeta}{\eta(1+K^2 d^2)} \left[ \frac{J_0(Kr)}{J_0(K\sqrt{8k/\phi})} - e^{\frac{\sqrt{8k/\phi}-r}{d}} \right] r dr \right] \quad (32)$$

397 The composite conductivity is

398 
$$\sigma(\omega) = \alpha + \beta \quad (33)$$

399 The seismoelectric coupling coefficient can now be stated as

400 
$$L(\omega) = -2\pi en_A Z \frac{\phi^2}{8k\pi} \left[ \int_0^{\sqrt{8k/\phi}} \frac{C_+}{i\omega\rho_f} \left[ 1 - \frac{J_0(Kr)}{J_0(K\sqrt{8k/\phi})} \right] r dr - \int_0^{\sqrt{8k/\phi}} \frac{C_-}{i\omega\rho_f} \left[ 1 - \frac{J_0(Kr)}{J_0(K\sqrt{8k/\phi})} \right] r dr \right] \quad (34)$$

401 This equation is the expression of the streaming potential coupling coefficient in the frequency  
 402 domain (Yu et al., 2013).

403

404 **Table 2.** Parameters used in this work for calculating the seismoelectric coupling coefficient.

Parameters	Values	Parameters	Values
Fluid Concentration $c_0$ (kg/dm <sup>3</sup> )	$1.5 \times 10^{-3}$	Fluid viscosity $\mu$ (Pa·s)	$0.89 \times 10^{-3}$
Absolute temperature T (K)	298	Boltzmann constant $k_B$ (J·K <sup>-1</sup> )	$1.38 \times 10^{-23}$

Fluid density $\rho_0$ (kg/m <sup>3</sup> )	1000.0	Fundamental charge $e$ (C)	$1.6 \times 10^{-19}$
Dielectric constant of water $\epsilon$ (F/m)	$80 \times 8.854187817$ $\times 10^{-12}$	Ionic valence (NaCl) $Z$	1

**Note.** The value for  $c_0$  was measured in the laboratory. The values for  $\epsilon$  are from [Lide \(2010\)](#).

### Theoretical modelling results

The theoretical modelling carried out in this work uses a number of standard and experimental parameters which are given in [Table 2](#). [Figure 4](#) shows the seismoelectric coupling coefficient as a function of frequency, porosity and permeability derived from the theoretical treatment presented earlier and [Equation \(34\)](#) in particular.

We should know that there is a close relationship between porosity and permeability for real rocks. Permeability is dependent largely on the pore structure. Many empirical equations are proposed to estimate the relationship of permeability and porosity. One of the famous equations was presented as Kozeny-Carman equation (1937):

$$K = \frac{\phi^2}{c(1-\phi)^2 S^2} \quad (35)$$

Where  $\phi$  is porosity,  $c$  and  $S$  are the Kozeny-Carman constant and the specific surface area of solid phase, respectively. Hence, in fact, permeability cannot be discussed separately with porosity. In the capillary model, according to [Equation \(1\)](#) and [\(2\)](#), the relationship between the porosity and permeability is represented as

$$\frac{\phi}{K} = \frac{8}{r_0^2} \quad (36)$$

Hence, the theory in this paper takes the relationship between the porosity and permeability into consideration. And we focused on one parameter for one time (separated analysis of the porosity and permeability) to investigate the impact of a single variable.

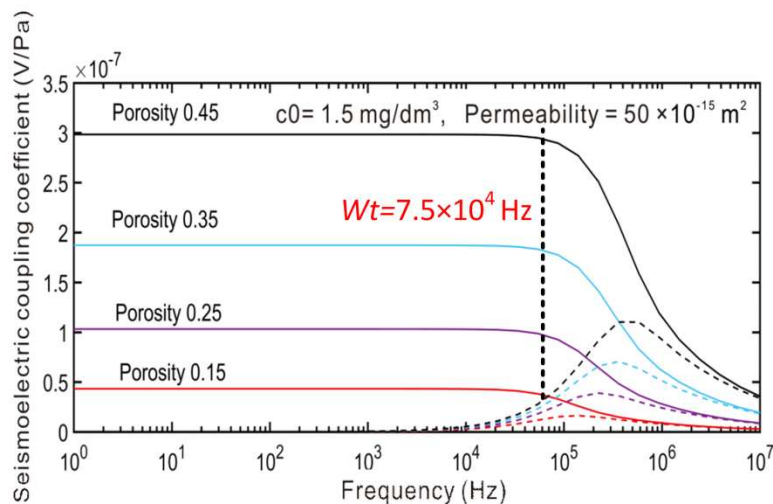
[Figure 4a](#) shows how the coupling coefficient is modelled to vary as a function of porosity and frequency. In the low frequency range, the coupling coefficient is independent of frequency, but decreases as the frequency increases in the high frequency range. The onset of the decrease (the transition frequency approximately at  $7.5 \times 10^4$  Hz) seems to be dependent upon porosity as evidenced by a shift in the peak of the out of phase component coupling coefficient to lower frequencies as porosity decreases. However, this shift is small. The low-frequency value of coupling coefficient increases with porosity by almost one order of magnitude as the porosity increases from 15% to 45%.

[Figure 4b](#) shows how the coupling coefficient is modelled to vary as a function of permeability and frequency. When the frequency is less than the transition frequency (as indicated in [Figure 4b](#)), the curve for the lowest permeability ( $1 \times 10^{-15}$  m<sup>2</sup>) shows the smallest value of coupling coefficient, and coupling coefficient increases with increases of permeability. Consequently, if the capillary bundle

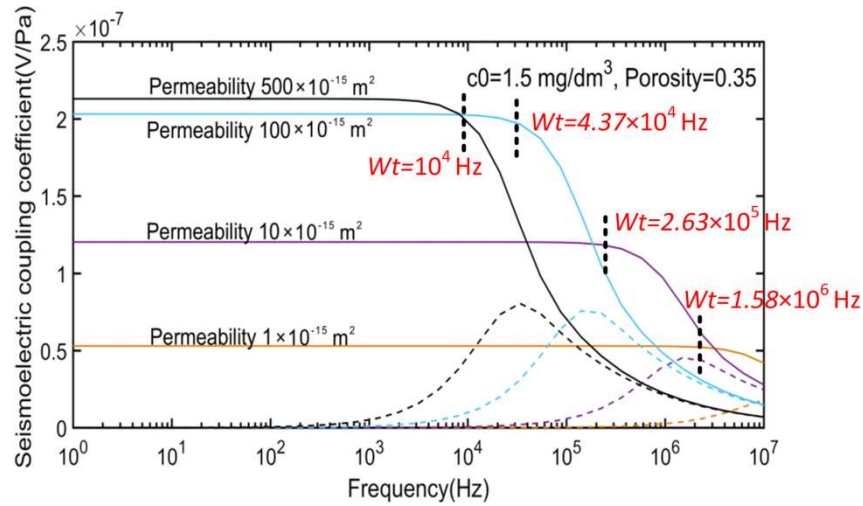
438 model represents real rocks well, we would expect high permeability rocks to present higher  
 439 seismoelectric coupling coefficients. In **Figure 4b**, we also observe that the increasing permeability  
 440 shifts the coupling coefficient curves to the left (i.e., to lower frequencies). The theoretical modelling  
 441 therefore clearly shows that the transition frequency of the coupling coefficient decreases with  
 442 increasing permeability.

443 The transition frequency indicates the beginning of the transition for the seismoelectric coupling. The  
 444  $L_{SE}(w)$  is constant up to the transition frequency above which it decreases, and the more permeable the  
 445 sample is, the lower the transition frequency is. When below the transition frequency, the viscous  
 446 force between the solid matrix and pore fluid are strong, so the relative movements of the pore fluid  
 447 and solid matrix are small, the attenuation mechanism in this case are usually an average motion of  
 448 the fluid relative to a solid phase, which is a macro-mode motion called global flow, and the  
 449 attenuation of the wave is small, hence, the elastic-wave induced seismoelectric coupling is in a quasi-  
 450 static state. Above the transition frequency, the inertia force of pore fluid makes its motion lag behind  
 451 that of solid matrix, thus causing the attenuation of the wave. In this case, the attenuation mechanism  
 452 that often occurs is the local flow or squirt flow mechanism. The non-equilibrium pore pressure  
 453 induced by acoustic wave forces the pore fluid to move locally. This mechanism is the main reason  
 454 for the attenuation of acoustic wave propagation in fluid-saturated media. Therefore, when the  
 455 frequency is higher than the transition frequency, the seismoelectric coupling coefficient decreases  
 456 rapidly.

457  
 458 (a)  
 459



460  
 461  
 462 (b)  
 463



464

465 **Figure 4.** The modelled seismoelectric coupling coefficient  $L(\omega)$  for the capillary bundle as a  
 466 function of frequency, and (a) porosity, and (b) permeability, resulting in the application of the  
 467 theoretical model for a capillary bundle developed earlier in this paper. The solid line and the dashed  
 468 line show the real and imaginary portions of coupling coefficient, respectively.

469

470 From Figure 4a and 4b, we can see that the  $L_{SE}(\omega)$  are independent with frequency (quasi-static) in a  
 471 certain frequency range. And this frequency range is about 75 kHz in Figure 4a and 10 kHz in Figure  
 472 4b. The permeability of natural rocks are almost less than  $500 \times 10^{-15} \text{ m}^2$ , so the quasi-static frequency  
 473 range can be considered as 0-10 kHz when we combine the results of these two figures. That's why  
 474 we define 0-10 kHz as the low frequency range and over 10 kHz as the high frequency range.

475

476 Figure 4a and 4b also present the imaginary parts of the seismoelectric coupling coefficient. We can  
 477 see that the magnitudes of imaginary parts (dashed line) of  $L_{SE}(\omega)$  change with the real parts (solid  
 478 line). And the sudden change point of imaginary parts is also changes with the critical frequency of  
 479 the real parts. However, the physical meaning of the imaginary contributions to the  $L_{SE}(\omega)$  is not sure.  
 480 But there are some indications. The generation of the streaming potential is based on the relative  
 481 motion of the solid matrix and the pore fluid (the electric double layer). This means that the fluid  
 482 velocity causes the streaming potential, which implies that the frequency dependence of the streaming  
 483 potential coupling coefficient depends on the frequency dependence of the dynamic fluid permeability.  
 484 The dynamic fluid permeability at low frequencies is controlled by viscous flow that is represented by  
 485 the real part of the dynamic permeability. And after reaching transition frequency, the inertial  
 486 acceleration begins to control the flow which is represented by the imaginary part of the dynamic  
 487 permeability. Hence, the real and imaginary parts of the streaming potential coupling coefficient are  
 488 affected by the same transition from viscous-dominated to inertial-dominated flow (Glover et al.,  
 489 2012). Also, the attenuation of acoustic wave caused by the relative motion between fluid and solid

490 can be described by the "resonance" mechanism. When the frequency is lower than the transition  
 491 frequency, the wave propagation causes the vibration of the solid matrix, the fluid vibrates with the  
 492 solid matrix only under the viscous force, which is almost the in-phase motion between fluid and solid  
 493 matrix. In this frequency range, the imaginary part of the coupling coefficient is nearly zero. When the  
 494 frequency increases more than the transition frequency, the effect of the inertial force of the fluid  
 495 increases gradually, and the relative displacement between the fluid and the solid matrix occurs, this  
 496 means that the vibration between the fluid and solid matrix is out of phase. The imaginary part of the  
 497 coupling coefficients increases gradually, and the real part of the coupling coefficients decreases  
 498 fastest when the imaginary phase of the coupling coefficients increases to the peak. This shows that  
 499 the out-of-phase vibration between solid and fluid is strongest at this time. When the frequency  
 500 continues to increase, the frequency of solid movement catches up with that of fluid, and the motion  
 501 of solid and fluid gradually converges, and the out-of-phase vibration weakens, resulting in the  
 502 imaginary phase of the coupling coefficient tending to zero, so the imaginary value of the coupling  
 503 coefficient may represent the intensity of out-of-phase vibration between solid and fluid. Large  
 504 imaginary phase of  $L_{SE}(w)$  indicates the strong out-of-phase vibration and large attenuation of  
 505 seismoelectric coupling.

506

507 A comparison of Figure 4a and Figure 4b shows that permeability seems to have larger influence on  
 508 transition frequency than porosity. For seismoelectric transition frequency, [Pride \(1994\)](#) defined the  
 509 transition frequency as

$$510 \quad \omega_t = \frac{\phi}{\alpha_\infty k} \frac{\eta}{\rho_f} \quad (37)$$

511 This transition frequency separates the low-frequency viscous flow regime from the high-frequency  
 512 inertial flow regime. Porosity is given by  $\phi$ , viscosity by  $\eta$ , tortuosity by  $\alpha_\infty$ , permeability by  $k$  and  
 513 fluid density by  $\rho_f$ .

514

515 The capillary model implies that  $\alpha_\infty = 1$ ,  $\phi = n_0 \pi r_0^2$ ,  $k = n_0 \pi r_0^4 / 8$  ([Equation \(1\)](#) and [Equation \(2\)](#),  
 516 respectively). The transition frequency then becomes

$$517 \quad \omega_t = \frac{8}{r_0^2} \frac{\eta}{\rho_f}, \quad (38)$$

518 which is consistent with the transition frequency of [Walker and Glover \(2010\)](#). Furthermore, a simple  
 519 substitution of  $\alpha_\infty = 1$  into [Equation \(37\)](#), the transition frequency  $\omega_t$  becomes

$$520 \quad \omega_t = \frac{\phi}{k} \frac{\eta}{\rho_f} \quad (39)$$

521 which is dependent on the porosity and permeability.

522 In addition, [Jouniaux and Bordes \(2012\)](#) analysed the transition frequency as a function of the  
 523 permeability on various samples with parameters (porosity  $\phi$ , intrinsic permeability  $k_0$ , formation

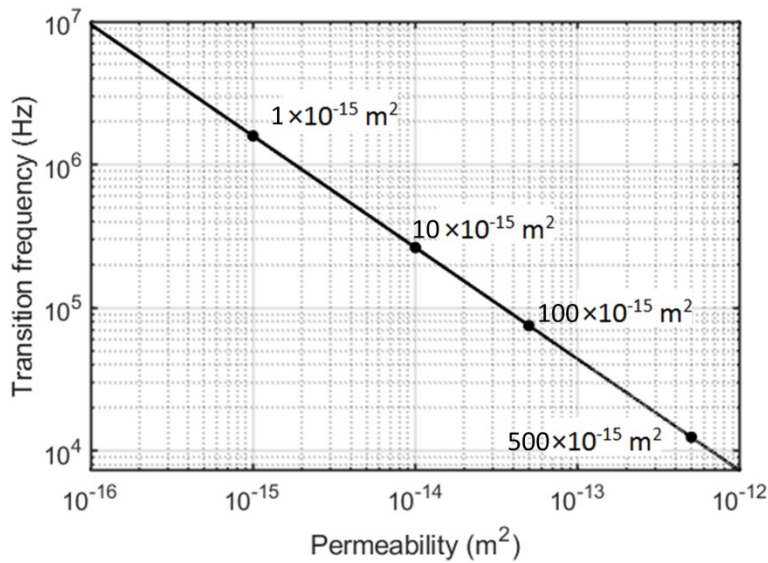
524 factor F ) measured from different authors, the predicted relationship between the transition frequency  
 525 and permeability is

$$526 \log_{10}(\omega_t) = -0.78 \log_{10}(k) - 5.5 \quad (40)$$

527 This equation is more practical since this relationship is based on actual measurements.

528  
 529 Because the transition frequency is a function of the capillary radius (Equation 38). Hence, we  
 530 interpret to arise from the greater sensitivity of the permeability to capillary radius (by a factor of 4)  
 531 than porosity (by a factor of 2) that can be confirmed by checking with Equation 1 and Equation 2.  
 532 The difference in the sensitivity of the transition frequency to permeability rather than to porosity  
 533 cannot in this case be attributed to the tortuosity of the pores as for our capillary model the tortuosity  
 534 is high, constant and equal to unity. Plotting the modelled seismoelectric coupling coefficient  
 535 transition frequency against permeability shows a very clear power law behaviour as evidenced by the  
 536 negative gradient straight line on the log-log plot shown in Figure 5 (Equation (40)).

537



538

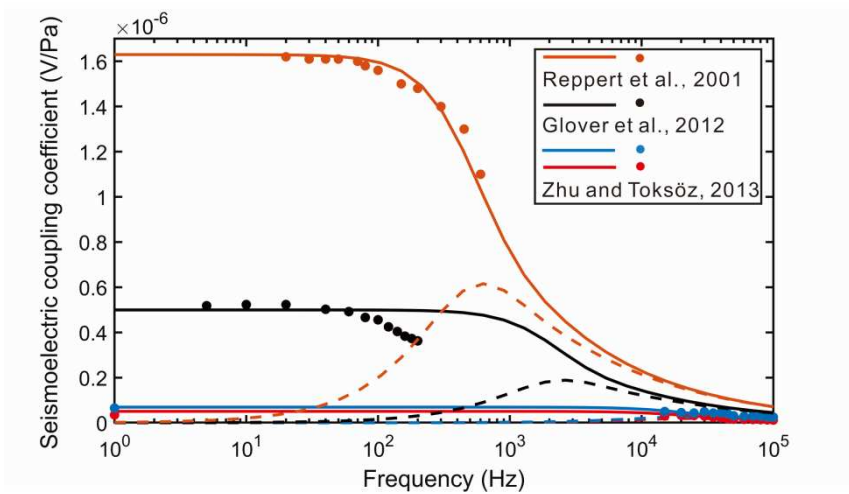
539 **Figure 5.** The modelled seismoelectric coupling coefficient transition frequency  $\omega_t$  as a function of  
 540 permeability.

541

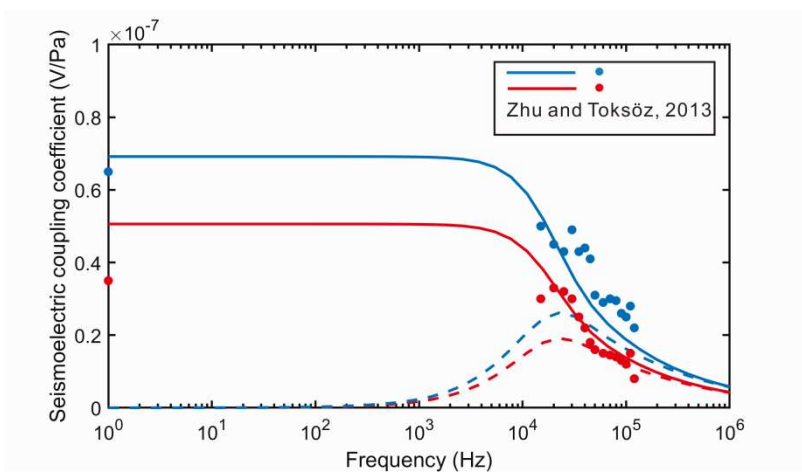
542 We can see that the changes in transition frequency lead to the changes of the quasi-static range of  
 543  $L_{SE}(w)$ , especially according to the results in Figure 4b. When the frequency is lower than the  
 544 transition frequency, the viscous force plays a leading role in the movement of fluid in porous media;  
 545 when the frequency is higher than the transition frequency, the inertial force dominate the movement  
 546 of fluid in porous media. The higher the permeability, the lower the transition frequency, the easier it  
 547 is for fluid in pore media to change from laminar to inertial flow. If the transition frequency is  
 548 invariable, the relationship between  $L_{SE}(w)$  and permeability will be simple (monotonically increasing

549 or decreasing), but the rapid change of conversion frequency with the increase of permeability results  
 550 in the complexity of the change of the dependency of  $L_{SE}(w)$  on permeability in frequency domain. As  
 551 can be seen from **Figure 7**, the change of frequency will present different variation curves between the  
 552  $L_{SE}(w)$  and permeability. At high frequencies, such as 100 kHz and 500 kHz, there will be a sudden  
 553 change in the  $L_{SE}(w)$  from increasing trend to decreasing trend. By knowing the relationship between  
 554 transition frequency and permeability, we know what kind of permeability range the transition  
 555 frequency of  $L_{SE}(w)$  will appear, and what frequency we should choose to do the experiment to  
 556 achieve the change of  $L_{SE}(w)$  covering both the quasi-static range and the range after the transition  
 557 frequency. Hence, for the experiment in the following part of this paper, we choose frequency in both  
 558 low frequency range (1 kHz, 10 KHz) and high frequency range (500 kHz). However, we have not got  
 559 the relationship of  $L_{SE}(w)$  and permeability at the transition frequency because of the equipment limit  
 560 (not enough experimental frequency range).

561  
 562 (a)



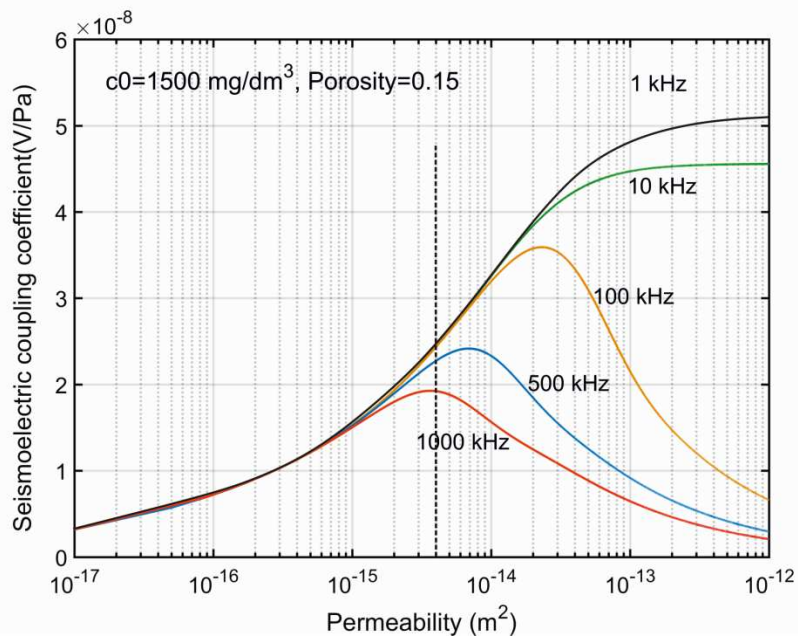
563  
 564 (b)



565

566 **Figure 6.** Comparisons between the measured published data of seismoelectric coupling coefficients  
 567 and the theoretical curves in this paper (a) and the enlarged figure of the measured data from Zhu and  
 568 Toksöz. The input parameters of the theoretical curves are from the corresponding published literature.  
 569

570 In Figure 6, the measured frequency-dependent seismoelectric coupling coefficients from Reppert et  
 571 al. (2001), Glover et al. (2012) and Zhu (2013) are presented. Comparing these measured data with  
 572 the theoretical curves in this paper, we found that in a wider frequency band (Glover et al., 5-200 Hz;  
 573 Reppert et al., 20-600 Hz; Zhu and Toksoz, quasistatic and 15k-20 kHz), the measured points from  
 574 Reppert's and Zhu's data are consistent with the theoretical curves, and there is a small deviation  
 575 between Glover's measured data and theoretical curves but within the error range. This indicates that  
 576 the seismoelectric coupling theory based on capillary model in this paper is reliable.  
 577



578  
 579 **Figure 7.** Modelled seismoelectric coupling coefficient  $L(\omega)$  for a capillary bundle as a function of  
 580 permeability for a range of different frequencies.  
 581

582 **Figure 7** shows how the modelled seismoelectric coupling coefficient varies with permeability for  
 583 different frequencies. One obvious characteristic of the curves is that the pattern of the variation of the  
 584 seismoelectric coupling coefficient with permeability changes with frequency. For frequencies of 1  
 585 kHz and 10 kHz with a porosity of 15%, the seismoelectric coupling coefficient increases as the  
 586 permeability increases, with the rate of increase being fast in the low permeability range (less than  
 587  $100 \times 10^{-15} \text{ m}^2$ ) and becoming slower in the high permeability range (over  $100 \times 10^{-15} \text{ m}^2$ ), but with no  
 588 sudden alteration in the rate of change of seismoelectric coupling coefficient with permeability. By  
 589 contrast, for higher frequencies (100 kHz, 500 kHz, 1000 kHz), while the seismoelectric coupling

590 coefficient increases for low permeabilities as before, the rate is the same until a particular value of  
591 permeability is reached whereupon the seismoelectric coupling coefficient changes its behaviour and  
592 decreases with any further increase in permeability. Consequently, there is a sudden alteration in the  
593 rate of change of seismoelectric coupling coefficient with permeability from positive permeabilities to  
594 negative at high permeabilities. The permeability values at which this occurs for all 3 frequencies  
595 modelled and shown in Figure 7 are  $2.5 \times 10^{-14} \text{ m}^2$ ,  $7 \times 10^{-15} \text{ m}^2$  and  $4.05 \times 10^{-15} \text{ m}^2$ , respectively. The  
596 presence of a threshold between 2 different regimes of behaviour is, however, interesting. Below the  
597 permeability threshold shown by the vertical dashed line we have what are generally termed tight  
598 clastic and carbonate rocks and gas and oil shales, while above the threshold exist more conventional  
599 clastic and fractured carbonate reservoirs. The threshold represents the peak  $L(w)$  for a frequency of  
600 1000 kHz, and occurs at a permeability of  $4.05 \times 10^{-15} \text{ m}^2$  (4.05 mD). Since the frequency of seismic  
601 sources in field explorations are generally less than 1000 kHz, and becoming lower with propagation,  
602 it follows that rocks with permeabilities less than  $4.05 \times 10^{-15} \text{ m}^2$  (4.05 mD) will see an increase in  
603  $L(w)$  with permeability and those above this permeability will experience a decrease in  $L(w)$  with  
604 permeability. Consequently, there is significantly different behaviour for unconventional reservoir  
605 rocks and conventional ones.

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## 607 **Experimental results**

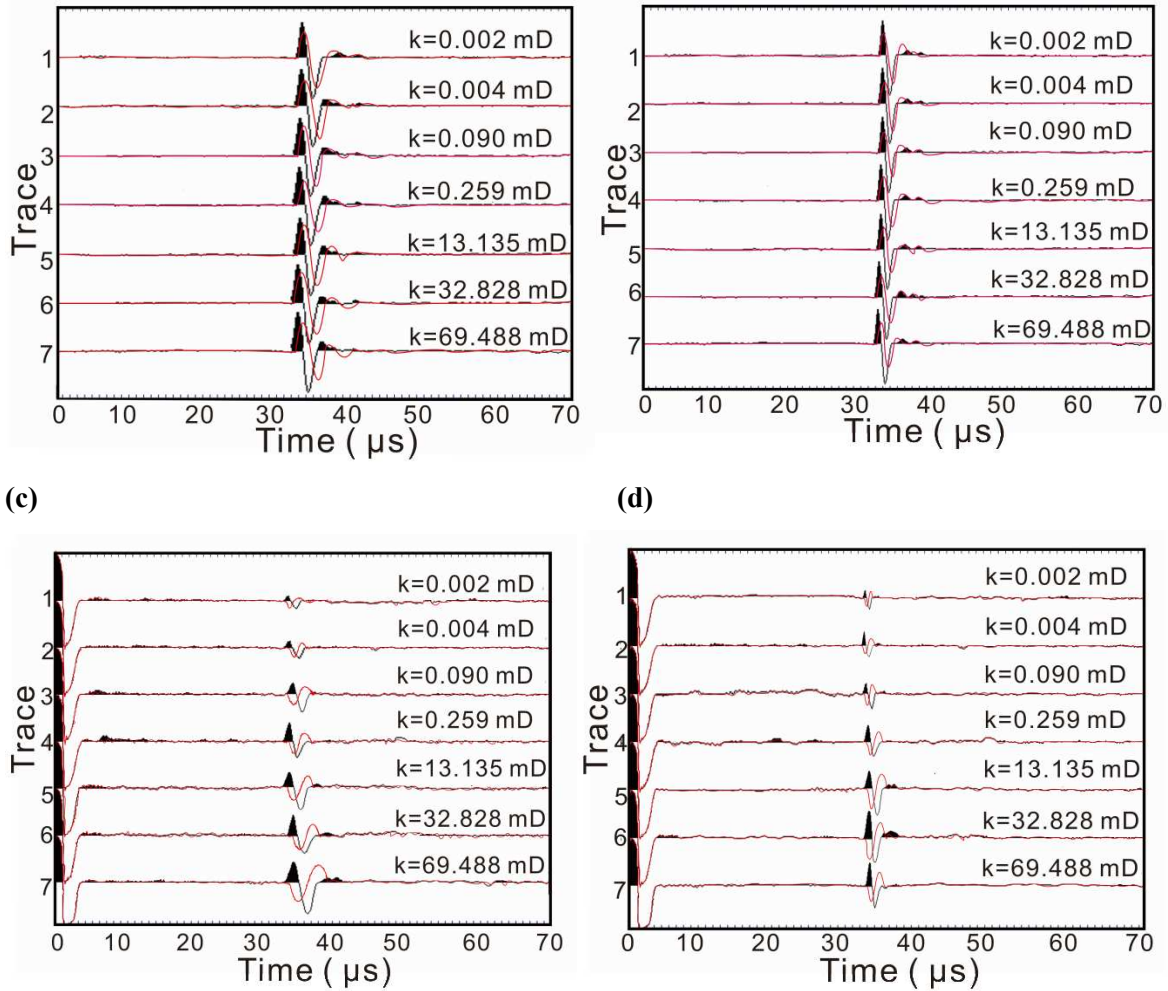
### 608 *Seismoelectric pressure and voltage signal measurement*

609 We acquired the seismoelectric conversion signals and acoustic signals of natural samples using the  
610 experimental apparatus shown in Figure 1. Figure 8 shows the measurements made in the experiment.  
611 Figure 8a and 8b shows the acoustic signals received at the two measurement ends of a selection of 7  
612 sandstone samples with typical behaviour, each having a different permeability and each being  
613 measured at 10 kHz and 500 kHz, respectively. Sandstone samples have little or no effect on the  
614 acoustic signal acquired at the front interface of samples (P1), hence, the P1 of each rock sample is  
615 the same and are shown as the in-filled black line. After the  $P$ -waves propagate through the rock  
616 sample, there is both an amplitude decrease and a travel time delay on P2 due to the thickness of rock  
617 sample as shown by the red curves.

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619 (a)

(b)



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**Figure 8.** The acoustic signals and seismoelectric signals measured at the two sides of 7 sandstone samples from a total of 28 natural samples. Figure 8a and 8b represents the acoustic signals ( $P_1$  = black in-filled and  $P_2$  = red), recorded at frequencies of (a) 10 kHz, and (b) 500 kHz. Figure 8c and 8d illustrates the seismoelectric potential signals of two electrodes ( $V_1$  = black in-filled and  $V_2$  = red), recorded at frequencies of (c) 10 kHz, and (d) 500 kHz.

629 **Figure 8c and 8d** shows the electric signals of the same 7 natural sandstones recorded by two electrodes at the frequencies of 10 kHz and 500 kHz, respectively. We observe that the polarity of high voltage pulses at the start is the same, but the seismoelectric responses between 30  $\mu$ s and 40  $\mu$ s exhibit a polarity reversal of the seismoelectric signals received by the two electrodes. This phenomenon is in agreement with the experimental results of [Zhu et al. \(2008\)](#). The amplitudes of the seismoelectric signals received by  $V_2$  (red) is smaller than that collected by  $V_1$  (black); this is caused by the intrinsic attenuation of the wave propagation from the front side ( $V_1$ ) to the back side ( $V_2$ ) of the rock sample. Hence, the induced SE signals at the back side ( $V_2$ ) is smaller. There is also a short time delay between the signals because of the travel time in the propagation of P-waves in the rock sample. It can be seen that the amplitudes of the seismoelectric signals change with rock permeability.

639 It is also possible to note that the seismoelectric amplitude increases as the permeability increases for  
640 the frequency of 10 kHz (Figure 8c), but such an association is not clear for a frequency of 500 kHz in  
641 Figure 8d.

642

643 In the following subsections we present experimental results together with associated modelling. The  
644 results are presented first for the natural samples with porosities between 9.91% and 16.3%, and then  
645 for the artificial samples which have a tightly controlled, higher porosity of 23.8%. This has not been  
646 done because there is a particular difference in their behaviour, but to convenience the comparison of  
647 experimental results with experimental modelling.

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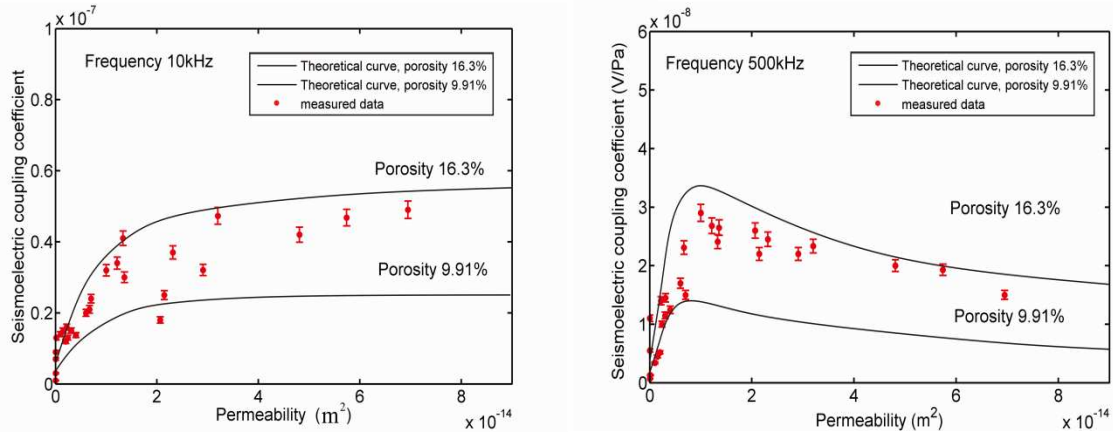
649 *Seismoelectric coupling coefficient of natural rock samples as a function of permeability*

650 Figure 9 presents the relationship between the seismoelectric coupling coefficient and permeability  
651 for those samples with porosities varying between 9.91% and 16.3% (the porosities are showed in  
652 Table 1). Each of the parts of Figure 9 includes both measured data as symbols with error bars as well  
653 as bounding theoretical curves for the highest and lowest porosity in the dataset, which were obtained  
654 by applying our capillary bundle model for the bounding porosities. Figure 9a represents behaviour in  
655 the low frequency range and the measurements were made at 10 kHz. The theoretical value of the  
656 seismoelectric coupling coefficient increases with permeability throughout the range of  
657 permeabilities. The rate of increase of seismoelectric coupling coefficient is fast in the very low  
658 permeability range and becomes smaller at higher permeabilities. It is instructive to note that almost  
659 all the experimentally measured values fall within the envelope formed by the theoretical curves. This  
660 indicates that the experimental results are generally consistent with the theoretical trend. Figure 9b  
661 shows the theoretical and experimental data in the high frequency range (500 kHz). In this case the  
662 seismoelectric coupling coefficient first increases as permeability increases, reaches a peak, then  
663 decreases as permeability increases further. Once again, the experimental data generally falls within  
664 the envelope provided by the modelling curves.

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666 (a)

(b)



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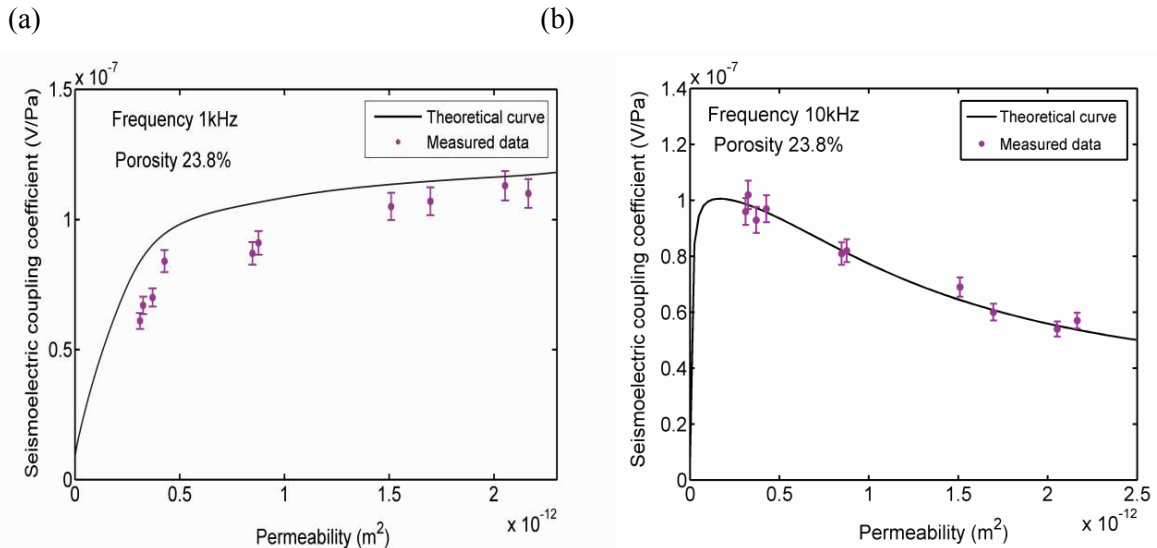
**Figure 9.** Seismoelectric coupling coefficient as a function of rock permeability for natural rock samples falling in the porosity range 9.91% to 16.3%, for measurements made at (a) 10 kHz, and (b) 500 kHz. The red symbols represent experimentally measured values of seismoelectric coupling coefficient ( $L(\omega)$ ), and have error bars which are all about  $\pm 5\%$ , thanks to the stability of both the pressure and potential measurements. The black curves represent modelled seismoelectric coupling coefficient using the capillary bundle model developed in this work and represented by Equation (34).

*Seismoelectric coupling coefficient of artificial samples as a function of permeability*

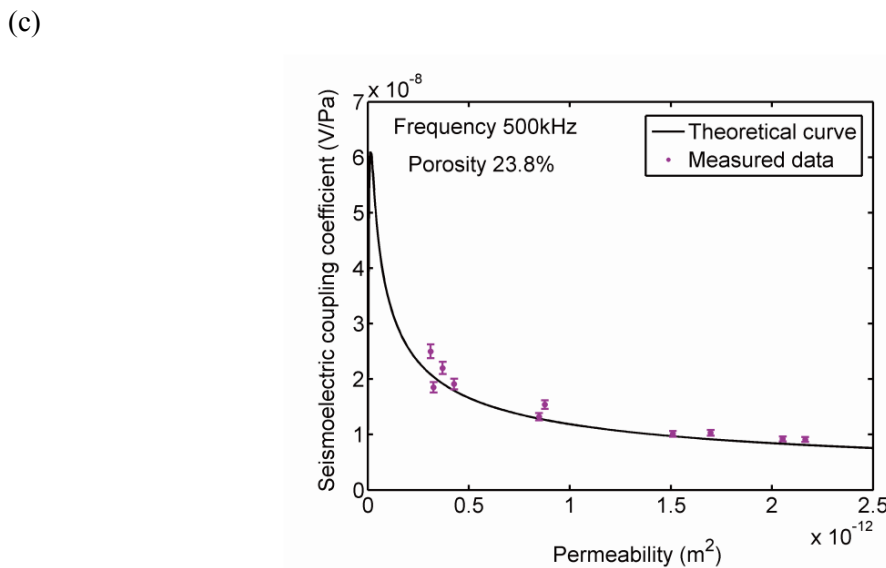
[Figure 10](#) shows the modelled seismoelectric coupling coefficient curve and experimental data for the artificial sandstones which all have a tightly controlled porosity equal to 23.8%. Once again we find that the seismoelectric coupling coefficient increases with the permeability at the frequency of 1 kHz ([Figure 10a](#)), the increase being fast at low permeabilities and becoming slower in the high permeability range. This behaviour is the same as for the natural rock samples at 10 kHz. For the data at 10 kHz ([Figure 10b](#)), the behaviour of the seismoelectric coupling coefficient on the 4 artificial sandstones is similar to that of natural rock samples measured at 500 kHz. The modelling curves would predict a very sharp increase of seismoelectric coupling coefficient with increasing permeability in the low permeability range, but these artificial samples have such a high porosity that none of them have such low permeabilities. However, in the higher permeability range, where we saw decreases in seismoelectric coupling coefficient with increasing permeability for the natural samples, we also get the same behaviour for the artificial samples. This same behaviour is continued when measurements are made at 500 kHz ([Figure 10c](#)), where once again there is no experimental data fitting into the initial very low permeability increases in seismoelectric coupling coefficient with permeability, but the higher permeability behaviour, which includes a significant drop in seismoelectric coupling coefficient with increasing permeability, sees the modelled curve once again matching the experimental data very well.

695 Interestingly, because all the samples share the same porosity, we can compare the experimental data  
 696 with just one theoretical curve for that single porosity value, and all of the samples fall on or very  
 697 close to the theoretical line, and in all cases within the error bars of the experimental measurements.  
 698 This is a further validation of the use of a capillary bundle model for modelling the seismoelectric  
 699 coupling coefficient of rocks, but also attests to the accuracy with which the experimental  
 700 measurements were made.

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 706 **Figure 10** Seismoelectric coupling coefficient as a function of rock permeability for artificial rock  
 707 samples with the porosity constrained to be equal to 23.8%, four measurements made at (a) 1 kHz, (b)  
 708 10 kHz, and (c) 500 kHz. The purple symbols represent experimentally measured values of  
 709 seismoelectric coupling coefficient  $L(w)$ , and have error bars which are all about  $\pm 5\%$ , thanks to the  
 710 stability of both the pressure and potential measurements. The black curves represent modelled

711 seismoelectric coupling coefficient using the capillary bundle model developed in this work and  
712 represented by Equation (34) for a porosity of 23.8%.

713

## 714 **Discussion**

715 The low permeability rock samples (natural rock samples) we used in this paper had a varying  
716 porosity. By contrast, the artificial sandstones all have the same, tightly controlled porosity. The  
717 permeability of these samples is very high. Consequently, we do not have experimental data of  
718 seismoelectric coupling for low-permeability samples, where the porosity is tightly controlled. Access  
719 to such data would allow us to confirm the analytical modelling curves of the  $L(w)$  variation in the  
720 low permeability range as we have done for the high permeability artificial samples in Figure 10. This  
721 is an important goal because the  $L(w)$  peak as a function of permeability and frequency occurs in the  
722 low permeability range.

723

724 The question arises why there is a sudden decrease of  $L(w)$  with further increasing permeability at  
725 high frequency (higher than 10 kHz). Most researches had previously indicated that seismoelectric  
726 conversion is enhanced as permeability increases (Jouniaux & Pozzi, 1995, 4 kHz; Mikhailov et al.,  
727 2000, 150 Hz; Zhu et al., 2015, 30 kHz). This is because a high permeability leads to a better flow,  
728 which can induce a larger relative displacement between the charges at the solid-fluid interface, and  
729 can then produce strong seismoelectric effects (Shaw, 1992; Pride, 1994; Haartsen & Pride, 1997;  
730 Haines et al., 2007). However, we found that the effect of permeability is to decrease the  $L(w)$  in the  
731 high frequency regime after it reaches a certain value. The higher the permeability, the smaller the  
732  $L(w)$  in high frequency ranges. Shatilo et al. (1998) evaluated the ultrasonic attenuation on a set of  
733 rocks and found that the attenuation coefficient of  $P$ -waves in the water-saturated sandstones  
734 increases with permeability under high frequency (750 kHz). This indicates that in the case of high  
735 frequency, the higher the permeability, the greater the attenuation of  $P$ -waves. Therefore, the  $P$ -wave  
736 induced seismoelectric conversion in high-permeability rocks becomes weaker in high frequency  
737 range (Figure 7, Figure 9b, Figure 10b and Figure 10c).

738

739 In Figure 5, we can see that the transition frequency decreases with the increase of permeability.  
740 When the frequency is larger than the transition frequency, the inertia force plays a dominant role in  
741 fluid flow. This shows that when the transition frequency decreases, leading a decrease in the  
742 corresponding frequency at which the inertial forces start to play a dominant role in fluid. And this  
743 also means that the inertia force is easier to dominate fluid flow. In this case, the fluid flow caused by  
744 inertia force at high frequency belongs to squirt flow, which is the main cause of large-scale  $P$ -wave  
745 attenuation (Mavko & Jizba., 1991; Mavko et al., 2009; Dovorkin et al., 1994, 1995). Hence,  
746 permeability, transition frequency and attenuation are inseparable. With the increase of permeability,

747 the transition frequency decreases, and the attenuation of P-wave increases, which weakens the  
748 seismoelectric conversion.

749

750 As can be seen from the comparison between the experimental results and the theoretical results in  
751 Figure 9 and 10, the experimental data points and the theoretical curves are not completely consistent,  
752 mainly because of the errors in both experiments and theoretical simulation. Although we want to  
753 minimize the errors in the study, the measurement errors of the experimental results are unavoidable,  
754 and the simulation of the theoretical model is very difficult to achieve complete agreement with the  
755 actual rock. Therefore, we consider that the small deviation between theoretical and experimental  
756 results is reasonable.

757

758 The theory of seismoelectronic coupling based on capillary model in this paper is only applied to the  
759 medium of single-phase saturated fluid, without considering the case of multi-phase fluid. [Jackson](#)  
760 [\(2008, 2010\)](#) used the capillary model to study the seismoelectric coupling theory of multi-phase fluid,  
761 but the results were quasi-static theory analysis, and the seismoelectric coupling coefficient is a  
762 function of saturation. The relation of  $L(w)$  and permeability was not discussed in detail. For  
763 unsaturated case, [Boorders et al. \(2015\)](#) also studied the effect of water saturation on seismoelectric  
764 conversion based on the amplitudes ratio of seismic and seismoelectric waves. Although considering  
765 the situation of unsaturated media, they did not focus on the effect of permeability on seismoelectric  
766 coupling.

767

768 In the expression of seismoelectric coupling coefficient in Pride theory ([Pride, 1994](#)), the tortuosity  
769 ( $\alpha_\infty$ ) and a porous-material geometry term ( $\Lambda$ ) which contained in the dimensionless number  $m$  are  
770 difficult to determine, and the relationship between seismoelectric coupling coefficient and  
771 permeability is not intuitive. The theory in this paper is not compared with the seismic-electric  
772 coupling coefficient in Pride theory because of the differences of input parameters. The error of  
773 parameter determination may lead to great divergence between the two theories. The comparison of  
774 the two theories still needs to be further discussed in detail in future work.

775

776 Seismoelectric phenomena reveal the coupled properties that link the passage of seismic waves, fluid  
777 flow, porosity and permeability of reservoir rocks. And pore fluid permeability is commonly used in  
778 the characterization of reservoir rocks, any relationship between them would be useful ([Glover &](#)  
779 [Jackson, 2009](#)). The relationship might calculate the permeability of a rock from a seismoelectric  
780 measurement without recourse to empirical data-fitting ([Glover et al. 2006](#)). The derived  
781 seismoelectric coefficient in this paper reflects the relationship between the intensity of seismoelectric  
782 conversion and rock properties (porosity and permeability). In addition, measurements on natural and  
783 artificial samples, low permeability and high permeability samples have confirmed the validity of the

784 theoretical relationship presented in this work. Consequently, this theoretical model may provide a  
785 template for seismoelectric exploration to predict the reservoir permeability, that is, the reservoir  
786 permeability could be deduced if the seismoelectric coupling coefficients can be measured downhole.

787

## 788 **Conclusions**

789 We have investigated the effects of permeability and porosity on the seismoelectric conversion using  
790 both experimental measurements and theoretical analysis. We measured the seismoelectric conversion  
791 using 28 rock sample samples with porosities in the range 9.91% to 16.3% and 4 artificial sandstones  
792 with constant porosity equal to 23.8%. We have also developed and implemented the capillary bundle  
793 model to calculate the seismoelectric coupling coefficient as a function of porosity, permeability and  
794 frequency theoretically. Experimental and theoretical analyses show that both porosity and  
795 permeability affect seismoelectric conversion and present a quantitative dependence between  
796 permeability and the seismoelectric coupling.

797 Both experimental and theoretical analyses of seismoelectric coupling indicate that  
798 seismoelectric conversion is stronger for high porosity rocks across a wide frequency range. But the  
799 effects of permeability on seismoelectric coupling are complex and can be divided into two  
800 permeability regions where  $4.05 \times 10^{-15} \text{ m}^2$  (4.05 mD) is the permeability demarcation point, below  
801 this permeability value (unconventional reservoir), the seismoelectric coupling enhances with the  
802 increase of permeability; and over this value (conventional reservoir), the seismoelectric coupling  
803 increases first and then decreases with the increase of permeability. In addition, the dependency of  
804 permeability on the seismoelectric coupling is different for different frequency range. At low  
805 frequencies (1 kHz) both the natural and artificial samples show that seismoelectric conversion is  
806 enhanced by increases in permeability, with the greatest sensitivity in the lower frequency range. At  
807 higher frequencies (10 kHz-500 kHz) there is a great increase in seismoelectric conversion with  
808 increasing permeability, but the seismoelectric conversion reaches a peak and then declines rapidly,  
809 especially at the higher frequencies. The quantitative relationship between permeability and the  
810 seismoelectric coupling is dependent on the frequency and permeability range, based on this  
811 quantitative relationship, the permeability can be inferred by the seismoelectric conversion.

812 The comparison of all the theoretical curves and measured data indicate that theoretical capillary  
813 bundle model developed in this work has been implemented and found to match the experimental data  
814 of natural and artificial samples very well. The curve was in very good agreement with the  
815 experimental data at all three frequencies from 1 kHz to 500 kHz. The sensitivity of seismoelectric  
816 coefficient increasing in permeability at low permeability raises the possibility that these changes  
817 might lead to a method for obtaining the permeability of reservoirs with low permeabilities such as  
818 tight oil in tight gas and shale gas reservoirs.

819

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823

824 **References**

825 Alkafeef, S. & Alajmi, A., 2006. Streaming potentials and conductivities of reservoir rock cores in  
826 aqueous and non-aqueous liquids, *Colloids Surf. A*, **289**, 141–148.

827 Bockris, J. O. & Reddy, A. K. N., 1970. Modern electrochemistry: An introduction to an  
828 interdisciplinary area. *Plenum Press*, Vol. 1.

829 Bordes, C., Jouniaux, L., Garambois, S. & Dietrich, M., 2009. Seismoelectric and seismomagnetic  
830 measurements: original experiments within the low noise underground laboratory of rustrel (france)  
831 (invited), *American Geophysical Union, Fall Meeting 2009*, abstract id. NH31C–1123.

832 Bordes, C., Sénéchal, P., Barrière, J., Brito, D., Normandin, E., & Jougnot, D., 2015. Impact of water  
833 saturation on seismoelectric transfer functions: a laboratory study of coseismic phenomenon,  
834 *Geophys. J. Int.*, **200**(3), 1317–1335.

835 Carman, P., 1937. Fluid flow through a granular bed, *Transactions of the Institution of Chemical*  
836 *Engineers*, **15**, 150–167.

837 Deissler, R. G., 1976. Derivation of the Navier–Stokes equation, *American Journal of Physics*,  
838 **44**(11), 1128–1130.

839 Ding, P., Di, B., Wei, J., Di, X. & Li, X., 2013. Construction and experiments of synthetic sandstones  
840 with controlled fracture parameters, *75th EAGE Conference & Exhibition Incorporating SPE*  
841 *EUROPEC*.

842 Ding, P., Di, B., Wei, J., Li, X. & Deng, Y., 2014a. Fluid-dependent anisotropy and experimental  
843 measurements in synthetic porous rocks with controlled fracture parameters, *Journal of*  
844 *Geophysics and Engineering*, **11**(1): 4905–4909.

845 Ding, P., Di, B., Wang, D., Wei, J. & Li, X., 2014b. P and S wave anisotropy in fractured media:  
846 experimental research using synthetic samples, *Journal of Applied Geophysics*, **109**, 1–6.

847 Dullien, F., 1979. Porous Media, Fluid Transport and Pore Structure, Elsevier, New York.

848 Dupuis, J. C., Butler, K. E., Kopic, A.W. & Harris, B. D., 2009. Anatomy of a seismoelectric  
849 conversion: measurements and conceptual modeling in boreholes penetrating a sandy aquifer,  
850 *Journal of Geophysical Research Solid Earth*, **114**(B10).

851 Dvorkin, J., Nolen-Hoeksema, R. & Nur, A., 1994. The squirt flow mechanism: macroscopic  
852 description, *Geophysics*, **59**, 428–438.

853 Dvorkin, J., Mavko, G. & Nur, A., 1995. Squirt flow in fully saturated rocks, *Geophysics*, **60**(1), 97–  
854 107.

855 Fiorentino, E. A., Toussaint, R. & Jouniaux, L., 2016. Lattice boltzmann modelling of streaming  
856 potentials: variations with salinity in monophasic conditions, *Geophysical Journal International*,  
857 **205**(1), ggw041.

858 Glover, P. W. J., Zadjali, I. I. & Frew, K. A., 2006. Permeability prediction from MICP and NMR  
859 data using an electrokinetic approach, *Geophysics*, **71**(4), F49–F60.

860 Glover, P. W. J. & Jackson, M. D., 2009. Borehole electrokinetics, *The Leading Edge*, **29**(6), 724–728.

861 Glover, P. W. J. & Déry, N., 2010. Streaming potential coupling coefficient of quartz glass bead  
862 packs: Dependence on grain diameter, pore size, and pore throat radius, *Geophysics*, **75**(6), F225–  
863 F241.

864 Glover, P. W. J., Walker, E., Ruel, J. & Tardif, E., 2012a. Frequency dependent streaming potential of  
865 porous media—Part 2: Experimental measurement of unconsolidated materials, *Int. J. Geophys.*,  
866 **2012**, 1–17.

867 Glover, P. W. J., Walker, E. & Jackson, M.D., 2012b. Streaming-potential coefficient of reservoir  
868 rock: A theoretical model. *Geophysics*, **77**(2), D17–D43.

869 Glover, P. W. J., 2018. Modelling pH-dependent and microstructure-dependent streaming potential  
870 coefficient and zeta potential of porous sandstones, *Transp Porous Med.*, 1573–1634.

871 Guan, W., Hu, H. & Wang, Z., 2013. Permeability inversion from low-frequency seismoelectric logs  
872 in fluid-saturated porous formations, *Geophysical Prospecting*, **61**(1), 120–133.

873 Haartsen, M. W. & Pride, S. R., 1997. Electro seismic waves from point sources in layered media, *J.*  
874 *Geophys. Res.*, 102(1022), 24745–24784.

875 Haines, S. H., Pride, S. R., Klemperer, S. L. & Biondi, B., 2007. Seismoelectric imaging of shallow  
876 targets, *Geophysics*, 72(2), G9–G20.

877 Harris, S., 1971. An introduction to the theory of the Boltzmann equation, *Journal of Theoretical*  
878 *Biology*, **234**(1), 123–131.

879 Jackson, M. D., 2008. Characterization of multiphase electrokinetic coupling using a bundle of  
880 capillary tubes model, *J. Geophys. Res.*, **113**, B04201.

881 Jackson, M. D., 2010. Multiphase electrokinetic coupling: Insights into the impact of fluid and charge  
882 distribution at the pore scale from a bundle of capillary tubes model, *J. Geophys. Res.*, **115**,  
883 B07206.

884 Jougnot, D., Rubino, J. G., Carbajal, M. R., Linde, N. & Holliger, K., 2013. Seismoelectric effects due  
885 to mesoscopic heterogeneities, *Geophysical Research Letters*, **40**(10), 2033–2037.

886 Jouniaux, L. & Pozzi, J. P., 1995. Permeability dependence of streaming potential in rocks for various  
887 fluid conductivities, *Geophys. Res. Letters.*, **22**, 485–488.

888 Jouniaux, L., Bernard, M. L., Zamora, M. & Pozzi, J. P., 2000. Streaming potential in volcanic rocks  
889 from Mount Peleé, *J. Geophys. Res.*, **105**, 8391–8401.

890 Jouniaux, L., & Bordes, C., 2012. Frequency-dependent streaming potentials: a review. *International*  
891 *Journal of Geophysics*, **2012**, 1–11.

892 Klinkenberg, L. J., 1941. The permeability of porous media to liquids and gases. *Drilling &*  
893 *Production Practice*, **2**(2), 200-213.

894 Kozeny, J., 1927. Über kapillare Leitung der Wasser in Boden, *Sitzungsber.Akad. Wiss. Wien*,  
895 **136**(2a), 271–306.

896 Lide, D. R., 2010. *CRC Handbook of Chemistry and Physics*, 90th edn (Internet), CRC Press.

897 Luong, D. T. & Sprik, R., 2013. Streaming potential and electroosmosis measurements to characterize  
898 porous materials, *ISRN Geophys.*, **2013**, 1–8.

899 Mavko, G. & Jizba, D., 1991. Estimating grain-scale fluid effects on velocity dispersion in rocks,  
900 *Geophysics*, **56**(12), 1940–1949.

901 Mavko, G., Mukerji, T. & Dvorkin, J., 2009. *The rock physics handbook : tools for seismic analysis of*  
902 *porous media*, granular media.

903 Mikhailov, O. V., Queen, J. & Toksöz, M. N., 2000. Using borehole electroseismic measurements to  
904 detect and characterize fractured (permeable) zones, *Geophysics*, **65**(1), 1098–1112.

905 Morgan, F. D., Williams, E. R & Madden, T. R., 1989. Streaming potential properties of westerly  
906 granite with applications, *Journal of Geophysical Research Solid Earth*, **94**(B9), 12449-12461.

907 Pengra, D. B., Li, S. X. & Wong, P. Z., 1999. Determination of rock properties by low-frequency ac  
908 electrokinetics, *Journal of Geophysical Research Solid Earth*, **104**(B12), 29485–29508.

909 Peng, R., Wei, J. X., Di, B. R., Ding, P. B. & Liu, Z. C., 2016. Experimental research on  
910 seismoelectric effects in sandstone, *Applied Geophysics*, **13**(3), 425–436.

911 Peng, R., Di, B., Wei, J., Ding, P., Zhao, J., Pan, X. & Liu, Z. C., 2017. Experimental study of the  
912 seismoelectric interface response in wedge and cavity models, *Geophysical Journal International*,  
913 **210**, 1703–1720.

914 Pride, S. & Morgan, F. D., 1991. Electrokinetic dissipation induced by seismic waves, *Geophysics*,  
915 **56**(7), 914–925.

916 Pride, S., 1994. Governing equations for the coupled electromagnetics and acoustics of porous media,  
917 *Physical Review B Condensed Matter*, **50**(21), 15678.

918 Probst R. F., 1994. *Physicochemical Hydrodynamics: An Introduction*, 1–8. Chemie Ingenieur  
919 Technik CIT.

920 Reppert, P. M., Morgan, F. D., Lesmes, D. P. & Jouniaux, L., 2001. Frequency-dependent streaming  
921 potentials, *Journal of Colloid & Interface Science*, **234**(1), 194–203.

922 Revil, A., Schwaeger, H., Iii, L. M. C. & Manhardt, P. D., 1999. Streaming potential in porous media:  
923 2. Theory and application to geothermal systems, *Journal of Geophysical Research Solid Earth*,  
924 **104**(B9), 20033–20048.

925 Revil, A., Jardani, A., Sava, P. & Haas, A., 2015. The seismoelectric method. Chapter 2, John Wiley  
926 & Sons, Ltd.

- 927 Roubinet, D., Linde, N., Jougnot, D. & Irving, J., 2016. Streaming potential modeling in fractured  
928 rock: insights into the identification of hydraulically active fractures, *Geophysical Research*  
929 *Letters*, **43**(10).
- 930 Schakel, M. D., Smeulders, D. M. J., Slob, E. C. & Heller, H. K. J., 2011a. Laboratory measurements  
931 and theoretical modeling of seismoelectric interface response and coseismic wave fields, *J. Appl.*  
932 *Phys.*, 109(7), 074903, doi:10.1063/1.3567945.
- 933 Schakel, M. D., Smeulders, D. M. J., Slob, E. C. & Heller, H. K. J., 2011b. Seismoelectric interface  
934 response: experimental results and forward model, *Geophysics*, **76**(4), N29-N36.
- 935 Schakel, M. D., Zhu, Z. & Smeulders, D. M. J., 2011c. Seismoelectric interface response—full-  
936 waveform model and measurements, in 73rd Eage Conference and Exhibition Incorporating Spe  
937 Europec, doi:10.3997/2214-4609.20149491.
- 938 Schakel, M. D., Smeulders, D. M. J., Slob, E. C. & Heller, H. K. J., 2012. Seismoelectric  
939 fluid/porous-medium interface response model and measurements, *Transp. Porous Media*, 93(2),  
940 271–282.
- 941 Schoemaker, F., Smeulders, D. & Slob, E., 2007. Simultaneous determination of dynamic  
942 permeability and streaming potential, *SEG expanded abstracts*, **26**, 1555–1559.
- 943 Schoemaker, F., Smeulders, D. & Slob, E., 2008. Electrokinetic effect: Theory and measurement,  
944 *SEG Technical Program Expanded Abstracts*, 1645–1649.
- 945 Schoemaker, F., Grobbe, N., Schakel, M., Ridder, S. de., Slob, E. & Smeulders, D., 2012.  
946 Experimental validation of the electrokinetic theory and development of seismoelectric  
947 interferometry by crosscorrelation, *International Journal of Geophysics*, **2012**, 7-25.
- 948 Sénéchal, P. & Bordes, C., 2012. On the relevance of the acquisition configuration for seismoelectric  
949 recording, *EGU General Assembly Conference Abstracts*, **14**.
- 950 Shatilo, A. P., Sondergeld, C. & Rai, C. S., 1998. Ultrasonic attenuation in Glenn Pool rocks,  
951 Northeastern Oklahoma, *Geophysics*, **63**(2), 465–478.
- 952 Shaw, D. J., 1992. The solid-liquid interface, in *Introduction to Colloid and Surface Chemistry*, 4th  
953 edn, pp. 151–173, Butterworth-Heinemann.
- 954 Tardif, E., Glover, P. W. J. & Ruel, J., 2011. Frequency-dependent streaming potential of Ottawa  
955 sand, *Journal of Geophysical Research Solid Earth*, **116**(B4).
- 956 Thompson, A. H., Hornbostel, S., Burns, J., Tom, M., Robert, R. et al., 2007. Field tests of  
957 electroseismic hydrocarbon detection, *Geophysics*, **72**(72), 565.
- 958 Walker, E. & Glover, P. W. J., 2010. Permeability models of porous media: characteristic length  
959 scales, scaling constants and time-dependent electrokinetic coupling, *Geophysics*, **75**(6), E235–  
960 E246.
- 961 Walker, E., Glover, P. W. J. & Ruel, J., 2014. A transient method for measuring the DC streaming  
962 potential coefficient of porous and fractured rocks, *Journal of Geophysical Research: Solid Earth*,  
963 **119**(2), 957–970.

964 Walker, E. & Glover, P. W. J., 2018. Measurements of the Relationship between Microstructure, pH,  
965 and the Streaming and Zeta Potentials of Sandstones, *Transp Porous Med.*, **121**, 183–206.

966 Wang, J., Hu, H. S., Xu, X. R. et al., 2010. Experimental measurement study on rock permeability  
967 based on the electrokinetic effect, *Chinese J. Geophys.*, **53**(8), 1953–1960.

968 Wang, J., Hu, H., Guan, W. & Li, H., 2015. Electrokinetic experimental study on saturated rock  
969 samples: zeta potential and surface conductance, *Geophys. J. Int.*, **201**(2), 869–877.

970 Yu, H., Guan, J. T., Chen, H., Zheng, H. X. & Fang, W. J., 2013. Mathematical simulation on the  
971 frequency dispersion characteristics of the streaming potential in reservoir rocks, *Chinese Journal*  
972 *of Geophysics*, **56**, 676–687 (in Chinese).

973 Zhu, Z. Y., Toksöz, M. N. & Daniel, R. B., 2008. Electroseismic and seismoelectric measurements of  
974 rock samples in a water tank, *Geophysics*, **73**(5), E153–E164.

975 Zhu, Z. Y., & Toksöz, M. N., 2013. Experimental measurements of the streaming potential and  
976 seismoelectric conversion in berea sandstone, *Geophysical Prospecting*, **61**(3), 688–700.

977 Zhu, Z. Y., Toksöz, M. N. & Zhan, X., 2015. Seismoelectric measurements in a porous quartz-sand  
978 sample with anisotropic permeability, *Geophys. Prospect.*, **64**, 700–713.

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