Seismoelectric interface response: Experimental results and forward model

M. D. Schakel¹, D. M. J. Smeulders², E. C. Slob¹, and H. K. J. Heller¹

ABSTRACT

Understanding the seismoelectric interface response is important for developing seismoelectric field methods for oil exploration and environmental/engineering geophysics. The existing seismoelectric theory has never been validated systematically by controlled experiments. We have designed and developed an experimental setup in which acoustic-to-electromagnetic wave conversions at interfaces are measured. An acoustic source emits a pressure wave that impinges upon a porous sample. The reflected electric-wave potential is recorded by a wire electrode. We have also developed a full-waveform electrokinetic theoretical model based on the Sommerfeld approach and have compared it with measurements at positions perpendicular and parallel to the fluid/porous-medium interface.

INTRODUCTION

When the grain surfaces of soils and rocks are in contact with a fluid electrolyte, they typically acquire a chemically bound surface charge that is balanced by mobile counterions in a thin fluid layer surrounding the grains. The bound charge is immobile, whereas the counterions can move with the fluid. At the interface between the immobile and counterions, the so-called zeta potential is defined. A comprehensive theory for coupled seismic and electromagnetic (EM) wave propagation with the zeta potential as a key parameter is derived by Pride (1994). From research based on this electrokinetic theory, one can distinguish two seismoelectric effects: (1) a coseismic (electric) field is coupled to seismic waves and therefore propagates with seismic wave velocity (e.g., Pride and Haartsen, 1996) and (2) a seismic wave that traverses an interface with a contrast in electrical or mechanical properties produces EM waves that propagate outside the support of the seismic waves with much higher EM wavespeeds (e.g., Pride and Haartsen, 1996; Haartsen and Pride, 1997). These effects have been verified experimentally in the laboratory (e.g., Zhu et al., 1999, 2000; Zhu and Toksoz, 2003; Block and Harris, 2006; Bordes et al., 2006) and in the field (e.g., Long and Rivers, 1975; Butler et al., 1996; Mikhailov et al., 1997; Beamish, 1999; Garambois and Dietrich, 2001; Dupuis et al., 2007; Haines et al., 2007). Also, the coseismic magnetic fields associated with a Stoneley wave (Zhu and Toksoz, 2005) and with a shear wave have been measured (Bordes et al., 2006, 2008).

Electrokinetic theory combines poroelastic and EM theory through coupling in flux/force relations. The so-called dynamic electrokinetic coupling coefficient that occurs in the flux/force relations has been validated experimentally (Reppert et al., 2001). Seismoelectric exploration can combine seismic resolution and the sensitivity to pore fluids of EM methods (Haines et al., 2007). Garambois and Dietrich (2001), Block and Harris (2006), and Dupuis and Butler (2006) compare measured and theoretically predicted coseismic electric field amplitudes.
The numerically predicted amplitude-decay pattern of the seismolectric conversion at an interface (e.g., Haartsen and Pride, 1997; Garambois and Dietrich, 2002; Haines and Pride, 2006) has been observed by several workers (e.g., Butler et al., 1996; Mikhailov et al., 1997; Dupuis et al., 2007). Mikhailov et al. (1997) compare field measurements of this pattern as a function of horizontal distance with a numerical model and find reasonable agreement, and Charara et al. (2009) model and measure seismolectric signals at a fluid/porous-medium interface in a laboratory experiment. However, Charara et al. do not consider the spatial variation of reflected waveform and amplitude.

In this article, we test the predictive power of electrokinetic theory for the seismolectric interface response. Laboratory measurements and modeling on the basis of the Sommerfeld approach and Pride’s electrokinetic theory are presented. Measurements are conducted at several positions in a water tank in which a fluid-saturated porous sample is immersed. Excellent agreement between theory and experiments for waveforms and spatial amplitude pattern is found. Measured amplitudes are shown to be dependent on salinity. Only a single amplitude field scaling factor is necessary to find absolute agreement. Our research validates electrokinetic theory for the seismolectric interface response.

SEISMOELECTRIC EXPERIMENTAL SETUP

The experimental setup consists of a 58 × 39 × 28 cm water tank in which a 1.125 inch diameter P-wave Panametrics source transducer (model V3638, 500-kHz center frequency), a 380 μm diameter A-M Systems silver/silver chloride (Ag/AgCl) electrode, and a Technoglas porous sample (P3 glass filter) are mounted (see Figure 1). The sample is made of sintered 40–100 μm glass particles that are carefully saturated with and immersed in a 10^{-3} M sodium chloride (NaCl) solution. A 500 kHz single sine pulse with a 500 mV peak-to-peak amplitude from an Agilent waveform generator (model 33220A) is fed into an ENI amplifier (model 2100L RF Power Amplifier) set at 50 dB gain. The output signal is coupled into a piezoelectric source transducer. The electrode is clamped in a computer-controlled mounting bracket so it can be positioned in the water layer with high accuracy. An Analogic preamplifier (model D1000) and a Yokogawa oscilloscope (model DL4200) are used for recording. Parameter values are given in Table 1.

The cylindrical coordinate system (r, z) is also indicated in Figure 1. The z-axis connects the centers of the sample and the transducer. At several positions along the z-axis (Figure 1), electric potential measurements are performed. These on-axis recordings, averaged over 8192 source pulses, are shown in Figure 2a. A 144–896 kHz numerical band-pass filter is applied in all measurements. The amplifier is calibrated with respect to amplitude and waveform using typical time signals. An electric pulse arrives at approximately 0.1 ms at each position, corresponding to the traveltime of the acoustic wave from the source to the sample surface. Because of the large EM wavespeed (approximately 1.1 × 10^{7} m/s), the electric pulses arrive almost simultaneously at each position. Note that the pulse decays in

![Figure 1. Schematic of the experimental setup. The electrode is clamped in a computer-controlled mounting bracket. Electrode recordings are with respect to ground level. The r- and z-axes coincide with those of Figure 3.](image-url)

Table 1. Parameters of the porous medium. The values for $K_s$, $K_f$, $\eta$, $\rho_f$, $\varepsilon_f$, and $\mu$ are taken from the literature (see below). Values for $K_s$ and $G$ are derived from experiments, and $\phi$, $\rho_s$, $k_0$, and $\tau_\infty$ are measured directly. When applicable, fluid parameters are the same as for the porous medium. The temperature is 293.15 K.

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Parameter</th>
<th>Value</th>
<th>Units</th>
</tr>
</thead>
<tbody>
<tr>
<td>$K_s$</td>
<td>Bulk modulus skeleton grains$^a$</td>
<td>49.9 × 10^{9}</td>
<td>Pa</td>
</tr>
<tr>
<td>$K_f$</td>
<td>Bulk modulus (pore) fluid$^b$</td>
<td>2.2 × 10^{9}</td>
<td>Pa</td>
</tr>
<tr>
<td>$\eta$</td>
<td>Pore fluid viscosity$^b$</td>
<td>1 × 10^{-3}</td>
<td>Pa s</td>
</tr>
<tr>
<td>$\rho_f$</td>
<td>(Pore) fluid density$^b$</td>
<td>998</td>
<td>kg/m^{3}</td>
</tr>
<tr>
<td>$\Lambda$</td>
<td>Weighted pore-volume-to-surface ratio</td>
<td>1.229 × 10^{-5}</td>
<td>m</td>
</tr>
<tr>
<td>$\varepsilon_f$</td>
<td>Relative permittivity of the (pore) fluid$^b$</td>
<td>80.1</td>
<td>—</td>
</tr>
<tr>
<td>$\varepsilon_s$</td>
<td>Relative permittivity of the solid$^b$</td>
<td>4</td>
<td>—</td>
</tr>
<tr>
<td>$\mu$</td>
<td>(Fluid) magnetic permeability ( = $\mu_0$)</td>
<td>$4\pi \times 10^{-7}$</td>
<td>H/m</td>
</tr>
<tr>
<td>$K_o$</td>
<td>Bulk modulus framework of grains</td>
<td>6.6 × 10^{9}</td>
<td>Pa</td>
</tr>
<tr>
<td>$G$</td>
<td>Shear modulus framework of grains</td>
<td>5.5 × 10^{9}</td>
<td>Pa</td>
</tr>
<tr>
<td>$\phi$</td>
<td>Porosity of the porous medium</td>
<td>0.345</td>
<td>—</td>
</tr>
<tr>
<td>$\rho_s$</td>
<td>Solid density</td>
<td>2.212 × 10^{3}</td>
<td>kg/m^{3}</td>
</tr>
<tr>
<td>$k_0$</td>
<td>Permeability</td>
<td>3.1 × 10^{-12}</td>
<td>m^{2}</td>
</tr>
<tr>
<td>$\tau_\infty$</td>
<td>Tortuosity of the porous medium</td>
<td>2.1</td>
<td>—</td>
</tr>
</tbody>
</table>

$^a$Johnson and Plona (1982).

$^b$Lide (2010).

$^c$Pride (1994); Johnson et al. (1994); Jocker and Smeulders (2009).
amplitude as the distance from the interface increases. The measured (true) peak-to-peak amplitude at \( (r, z) = (0, -0.3) \text{ cm} \) is approximately 0.25 mV. In a separate experiment, acoustic pressure was measured along the \( z \)-axis. In Figure 2b, we show the acoustic waveforms measured with a Specialty Engineering Associates needle hydrophone (model PZT-Z44-1000-S/N), calibrated over the frequency range of interest. From these measurements, we can extrapolate that the acoustic pulse will indeed arrive at the interface around 0.1 ms.

FORWARD MODELING OF THE SEISMOELECTRIC INTERFACE RESPONSE

The acoustic source is modeled by (e.g., Hall, 1987)

\[
\hat{\rho}(\omega, R, \theta) = \frac{A(\omega)}{R} e^{-i k R} D(\theta),
\]

(1)

where \( \omega \) is the radial frequency, \( R = \sqrt{(r - r_0)^2 + (z - z_0)^2} \) is the distance to the source, \( \theta \) is the angle of incidence (see Figure 3), \( A(\omega) \) is the amplitude spectrum, and \( k \) is the fluid wavenumber. The directivity function \( D(\theta) \) is given by

\[
D(\theta) = \frac{J_1(ka \sin \theta)}{ka \sin \theta},
\]

(2)

Here, \( J_1 \) is the Bessel function of the first kind and first order, and \( a \) is the radius of the transducer. At \( (r, z) = (0, 0) \), \( A(\omega) \) is determined directly from the measurement (see Figure 4). As an additional test, the source response as a function of \( r \) is measured for a 15.3 cm on-axis distance from the transducer. A single sine 500 kHz source pulse with a 500 mV peak-to-peak amplitude is used to excite the source transducer. Results are shown by the asterisks in Figure 5. The dashed line is the theoretical value obtained from equation 1. We notice that the wavefield of the transducer is well described by equations 1 and 2, although the theory predicts interference patterns that are not recorded in the experiment.

The modeled source pressure wavefield is now expanded into conical waves by means of the Sommerfeld integral (see e.g., Brekhovskikh [1960] and Aki and Richards [2002]). The total reflected electric potential \( \hat{\varphi}(\omega, r, z_r) \) can then be expressed in terms of a Sommerfeld integral (\( r_0 = 0; z_0, z_r < 0 \)):

\[
\hat{\varphi}(\omega, r, z_r) = -iA(\omega) \int_0^{\infty} \frac{k_r}{k} D(k_r) J_0(k_r r) e^{ik z_0} e^{ik k_2 z_r} R^E(k_r) dk_r,
\]

(3)

where \( k_r = k \sin \theta \) and \( k_z = k \cos \theta \) are the radial and vertical components of \( k \), respectively, and \( k_2 \) is the vertical component of the fluid EM wavenumber. The seismoelectric reflection coefficient \( R^E(k_r) \) relates the incident pressure wavefront to a reflected EM signal as derived by Schakel and Smeulders (2010), adopting full electrokinetic theory (Pride, 1994). Note that their coefficient is displacement normalized, whereas \( R^P \) is pressure normalized (in V/Pa).

Expressing \( k_r \) and \( k_z \) in terms of \( k \) and \( \theta \), equation 3 is written as

\[
\hat{\varphi}(\omega, r, z_r) = -iA(\omega) \int_0^{\pi/2 + \infty} \frac{D(\theta) k \sin \theta J_0(k_r \sin \theta)}{\theta} \times e^{ik_2 z_0} e^{ik k_2 z_r} R^E(\theta) d\theta,
\]

(4)

Figure 2. (a) Electric potential and (b) acoustic pressure as a function of vertical distance \( z \) at \( r = 0 \).

Figure 3. Geometry of seismoelectric model and experiment. A pressure (P-) wave is converted to an EM wave at the interface.
where \( k_E^2(\theta) = \omega \sqrt{1/c_E^2(\omega)} - (\sin^2 \theta/c_P^2) \) and \( \text{Im}[k_E^2(\theta)] < 0 \).

Fluid EM and pressure wave velocities are given by, respectively, 
\[ c_E(\omega) = \sqrt{(\mu \varepsilon_0 - (i \sigma \varepsilon_0/\omega))^{-1}} \quad \text{and} \quad c_P = \sqrt{K_0/\rho}, \]

where \( \varepsilon_0 \) is the vacuum permittivity and \( \sigma \) is the fluid conductivity.

The path of integration is along straight lines from zero to \( \pi/2 \) and from \( \pi/2 \) to \( \pi/2 + i \infty \) in the complex \( \theta \) plane.

The second integral over complex \( \theta \) is simplified further using the substitution \( \theta = \pi/2 + i \ln \left( \sqrt{\gamma^2 + 1} \right) \):

\[
\hat{\phi}(\omega, r, z_r) = -\frac{ia(\omega)}{a_r} \int_0^{\pi/2} J_0(k r_r \sin \theta) J_1(k a \sin \theta) \times e^{i k z r_r \sin \theta} e^{i k E^2(\theta)z_r} R^E(\theta) d\theta
\]

\[
+ \frac{A(\omega)}{a_r} \int_0^{\infty} J_0(k r_r \sqrt{\gamma^2 + 1}) J_1\left(k a \sqrt{\gamma^2 + 1}\right) \sqrt{\gamma^2 + 1} d\gamma
\]

\[
\times e^{i k z r_r \sin \theta} e^{i k E^2(\gamma)z_r} R^E(\gamma) d\gamma,
\]

where equation 2 is used for \( D(\theta) \). Equation 5 can be readily evaluated numerically and yields the total reflected EM wavefield through the inverse fast Fourier transform. The source is located at \( (r_0, z_0) = (0, -15) \) cm (Figure 3). For each frequency, we use a recursive adaptive Simpson quadrature algorithm implemented in Matlab. A 144–896 kHz numerical band-pass filter is applied, and the input parameters of Tables 1 and 2 are used. Figure 6 shows the resulting modeled reflected electric potential and its frequency spectrum for a \( 10^{-3} \) M NaCl salinity.

**COMPARISON BETWEEN EXPERIMENT AND FORWARD MODEL**

Figure 7 compares measured waveforms (solid lines) at various positions in the fluid along the negative \( z \)-axis (Figure 3) with modeled waveforms (dashed lines). The agreement between measured and modeled waveforms is excellent. The modeled amplitudes are scaled to the measured amplitudes by the amplitude field scaling factor \( A_C = 0.19 \), which is obtained by comparing the theoretical Fourier spectrum maxima of the entire field to the measured maxima. Thus, \( A_C < 1 \) means that the theory overpredicts the actual measured electric potentials.

In Figure 8, the scaled theoretical Fourier spectra are compared with the experiments. Note that the measured pulses have

<table>
<thead>
<tr>
<th>Salinity (M NaCl)</th>
<th>Conductivity (S/m)</th>
<th>( \zeta ) (mV)</th>
<th>( A_C )</th>
</tr>
</thead>
<tbody>
<tr>
<td>( 10^{-4} )</td>
<td>( 1.27 \times 10^{-3} )</td>
<td>51.7</td>
<td>0.03</td>
</tr>
<tr>
<td>( 10^{-3} )</td>
<td>( 1.20 \times 10^{-2} )</td>
<td>61.5</td>
<td>0.19</td>
</tr>
<tr>
<td>( 10^{-2} )</td>
<td>( 1.01 \times 10^{-1} )</td>
<td>58.1</td>
<td>0.41</td>
</tr>
</tbody>
</table>
their energy distributed over slightly smaller frequencies than predicted by theory. Figure 9 compares measured waveforms (solid lines) at various positions in the fluid along the r-axis at $z = -1.3$ cm (Figure 3) with modeled waveforms (dashed lines). Again, $A_C = 0.19$ is used. As in the measurements along the z-axis, good agreement in the waveforms as well as in the spatial amplitude pattern is observed. We notice a slight asymmetry in the measured data. This is most likely caused by a small misalignment in the setup. In Figure 10, the corresponding frequency spectra are shown. We recall that the $A_C = 0.19$ value is obtained from Fourier spectra of the entire field shown in Figures 8 and 10.

To compare predicted amplitudes with measurements more directly, peak-to-peak amplitudes $V_{pp}/C_0/C_1$ are plotted in Figure 11 and compared with our model (dashed lines). Again, the spatial amplitude pattern is excellently predicted by theory for the measurements along the negative z-axis (Figure 11a) and parallel to the r-axis at $z = -1.3$ cm (Figure 11b). We also check the repeatability of the experiment. There is a small amplitude decrease over two days. The waveform is preserved on these longer timescales. In Figure 11 (gray lines), we also plot the vertical electric dipole approximation $A_D e^{-ikd}/(r^2 + z^2)^{3/2}$ (see also, e.g., Dupuis et al. [2007] and Haines et al. [2007]), where $d$ is the distance from the origin and the coefficient $A_D$ is...
obtained from a best-fit procedure. The vertical electric dipole approximation overestimates the spatial amplitude decay. This means that the full model (equation 5) is needed to explain the amplitude measurements.

Figures 12 and 13 show measured and full-waveform modeled \( V_{pp} \)-values where \( C = 10^{-4} \) and \( C = 10^{-2} \) M NaCl solutions are used, respectively. We determine the \( A_C \)-values in the same way as for the \( 10^{-3} \) M NaCl solution. Resulting values are given in Table 2. The spatial amplitude pattern is also excellently predicted for these salt solutions.

For higher salinities, the \( A_C \) values are closer to (or approach) unity. This observation is in agreement with Block and Harris (2006), who find that the discrepancy between electrokinetic theoretical predictions and measurements decreases with increasing salinity. Block and Harris (2006) also suggest that when low-salinity electrolytes are used to saturate the pores of sandstones, surface conductivity effects other than those predicted by Pride’s electrokinetic theory may become dominant. The Pride theory used here incorporates diffuse-layer excess conduction associated with electromigration and electroosmosis. When the pore width is much larger than the Debye length, which is a measure of the diffuse-layer thickness, surface conductivity effects become negligible with increasing salinity. Our sample has a 16–40 \( \mu \)m pore width, whereas the Debye length is on the order of 3–30 nm.

We also observe that for all salinities, the \( A_C \) values differ from unity. Several possible reasons can be identified. First, the zeta potentials are obtained from crushed sample material suspensions. A Zetasizer is used for this. Probably, the zeta potential measured on ground-up material has a much higher value than that measured on intact (aged) surface material. This observation alone could be responsible for \( A_C \) being different from

Figure 10. Measured (solid lines) and modeled (dashed lines) electric potential spectra at various positions in the fluid parallel to the \( r \)-axis at \( z = -1.3 \) cm for \( C = 10^{-3} \) M NaCl. The black and gray solid lines correspond with measured spectra for negative and positive \( r \)-coordinates, respectively.

Figure 11. Measured (symbols) and modeled (dashed lines) peak-to-peak amplitudes \( V_{pp} \) as a function of (a) \( z \) at \( r = 0 \) and (b) \( r \) at \( z = -1.3 \) cm for \( C = 10^{-3} \) M NaCl. Experiments were repeated several times to check repeatability. Here, \( A_C = 0.19 \). Also, the electric dipole approximation is plotted (gray lines).

Figure 12. Measured (symbols) and modeled (dashed lines) peak-to-peak amplitudes \( V_{pp} \) as a function of (a) \( z \) at \( r = 0 \) and (b) \( r \) at \( z = -1.3 \) cm for \( C = 10^{-4} \) M NaCl. Experiments were repeated several times to check repeatability. Here, \( A_C = 0.03 \).
CONCLUSIONS

The electrokinetic conversion of an acoustic wave to an EM wave at an interface was systematically investigated in a laboratory setup. We independently determined the mechanical and electrical parameters. It was demonstrated that this effect is well predicted by electrokinetic theory in terms of waveform and spatial amplitude pattern. At $C = 10^{-3}$ and $C = 10^{-2}$ M NaCl, amplitude field scaling factors of 0.19 and 0.41 were used. At $C = 10^{-4}$ M NaCl, however, the value was only 0.03, which may be related to surface conductivity effects other than those present in electrokinetic theory. The magnitude of acoustic-to-EM conversion was shown to depend on the salinity of the saturating fluid, which is promising for applications in the oil and gas industry because the zones of interest are saturated with (resistive) hydrocarbons instead of water. A careful characterization of the source radiation pattern was part of the forward model, indicating that electrokinetic interface response predictions require proper modeling of the source. The results of this study indicate that the acoustic-to-EM wave responses of these rock/ fluid combinations can be predicted by a full-waveform model for general subsurface exploration purposes.

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