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Seismoelectric imaging of the vadose zone of a sand aquifer

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Seismoelectric imaging of the vadose zone of a sand aquifer

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ABSTRACT

We have acquired a 300 m seismoelectric section over an unconfined aquifer to demonstrate the effectiveness of interfacial signals at imaging interfaces in shallow sedimentary environments. The seismoelectric data were acquired using a 40 kg accelerated weight drop source and a 24-channel seismoelectric recording system composed of grounded dipoles, preamplifiers and seismographs. Interfacial signals were remarkably clear in the shot records, arriving simultaneously at offsets up to 40 m from the seismic source. The most prominent signal was generated at the water table at a depth of approximately 14 m and had peak amplitudes on the order of 1 μV/m. A weaker response was generated at a shallower interface that is interpreted to be a water retentive layer. The validity of these two laterally continuous events, and of other discontinuous events indicative of vadose zone heterogeneity, is corroborated by the presence of reflections exhibiting similar characteristics in a ground penetrating radar profile acquired along the same line.

INTRODUCTION

Mechanical wave propagation through porous media can generate electromagnetic signals, known as seismoelectric effects, by electrokinetic coupling mechanisms that involve the motion of charge in the electrical double layer at the solid-liquid interface (Pride, 1994). Such signals, and reciprocal phenomena (Thompson et al., 2007), are of interest for the information they may be able to provide on pore fluid type and porous medium properties such as porosity and permeability (e.g. Thompson and Gist, 1993; Garambois and Dietrich, 2002).

Compressional waves in poroelestic media cause pore fluid to move relative to the solid matrix thereby moving the excess electrical charge in the outer, mobile portion of the electrical double layer. These streaming currents result in charge separations and hence electrical fields arising between zones of compression and rarefaction. In a homogenous medium, this phenomenon gives rise to a co-seismic electric field that is confined within the compressional wave (Neev and Yeatts, 1989; Dupuis and Butler, 2006). When a compressional wave encounters heterogeneity such as an interface that changes the streaming currents and distorts the resulting charge distribution, it generates an unbounded electric field, which we call an interfacial seismoelectric effect (Haartsen and Pride, 1997). These effects are expected to propagate (diffuse) through the earth as electromagnetic signals and therefore appear nearly simultaneously at widely separated receivers with an arrival time essentially equal to the one-way seismic traveltime from shotpoint to interface.

Conceptual models (e.g. Butler et al., 1996) and rigorous theoretical modeling (e.g. Haartsen and Pride, 1997; Garambois and Dietrich, 2002) indicate that the interfacial effect should be a multipole electrical source that develops over a Fresnel zone having a diameter that increases with depth and seismic wavelength. Higher order terms will diminish more rapidly with distance leaving the dipole term to dominate. Thus, an interfacial seismoelectric signal emanating from a horizontal boundary is expected to exhibit symmetry and amplitude characteristics similar to that of a vertical electrical dipole centered on the interface directly below the shot.

While the existence of interfacial seismoelectric effects in porous media has recently been confirmed by several investigators (e.g. Butler et al., 1996; Mikhailov et al., 1997; Russell et al., 1997; Garambois and Dietrich, 2001; Haines et al., 2007, Strahser et al., 2007) only a handful of studies have shown that the method can be used to map interfaces. Martner and Sparks (1959) mapped lateral variations in seismic traveltimes through the weathered layer by exploiting the co-seismic effect associated with a seismic P-wave critically refracted at the base of that layer. Thompson and Gist (1993) were the first to attempt seismoelectric profiling making use of interfacial electrokinetic seismoelectric effects, and inferred that they were able to image high permeability water sands and low permeability shales at depths of up to 300 m. Butler et al. (1996) used interfacial seismoelectric effects to map variations in the depth to a layer of heavily compacted, impermeable glacial till underlying 1 - 3 m of organic-rich fill.

In this paper, we present measurements of remarkably clear interfacial effects obtained over an unconfined sand aquifer. The results prove that seismoelectric methods can be used to trace subsurface interfaces in a manner analogous to multi-channel seismic reflection surveying.

SITE DESCRIPTION

The survey site was situated within the Gnangara Mound region on the northern fringes of Perth, Western Australia – a region hosting important groundwater resources including a sandy 'superficial' aquifer typically 50 m thick. The data were collected along Cypress Rd., 1.7 km west of groundwater production well P-90. Regional hydrogeological studies (Davidson, 1995) and geological log from borehole P-90 indicate that the superficial aquifer at this site is composed of a series of fine to coarse-grained quartz sands underlain by a siltstone layer at 58 m depth. Shallow discontinuous water retentive layers exert control over aquifer recharge and help to maintain near-surface moisture needed to support local ecosystems (M. Martin, personal communication, 2007). The geological log from P-90 identifies such a layer between 6 and 8 m depth and refers to it as "coffee rock" – a friable, limonite-cemented sand (Davidson, 1995) that is also evident as a zone of slightly elevated counts on the borehole's gamma ray log.

We chose to survey a 300 m segment of the road where an earlier GPR survey had indicated the presence of a water retentive layer together with , an increase in vadose zone heterogeneity and a shallowing of the water table in the approach to a topographic low. The objective was to determine whether seismoelectric conversions measured previously in boreholes at two nearby sites (Dupuis et al., 2007) could be measured on the surface and used to map lateral variations in shallow subsurface interfaces.

METHOD

Our recording spread (Figure 1) consisted of 26 electrodes at 4 m intervals connected to form 24 dipoles arranged end-to-end except for a four metre shot gap at the centre. Three 12-channel, 24-bit seismographs (Geometrics Geodes) with associated seismic cables were used to record the data after it was buffered by custom-built differential preamplifiers. Four shotpoints spaced 1 m apart were placed in the shot gap (offset about 2 m from the line for convenience) and three to five impacts from a 40 kg accelerated weight drop source were recorded at each point..The array advanced towards the west for 300 m as illustrated in Figure 1 with shot records collected at every metre.

The site chosen for this traverse was within 200 m of a power line. Electrical noise at 50 Hz and its harmonics, measured 0.1 – 0.4 mV/m peak-to-peak. A harmonic subtraction algorithm applied during data processing (Butler and Russell, 2003) and band-pass filtering (60-375 Hz, minimum phase) proved effective in reducing this noise to a manageable level. Furthermore, the shot redundancy at each shotpoint allowed us to discard any records that exhibited excessive residual harmonic noise prior to stacking.

During preliminary tests, we found that high contact impedances between our 40 cm stainless steel rod electrodes and the dry surficial sands made our data more susceptible to contamination by demodulated AM radio broadcasts (Kepic and Butler, 2002). To alleviate this problem we augured shallow holes (~50 cm deep) and either hammered one of our stainless steel rods into the bottom or inserted a sheet of aluminium foil before backfilling with sand and pouring on a mixture of water and soil-wetting agent.

RESULTS AND DISCUSSION

To interpret the various arrivals in the shot records, we combined data from the four shotpoints in each shot gap following an approach suggested by Kepic and Rosid (2004) to form composite shot gathers, or "super gathers" with very dense spatial sampling (96 traces at 1 m intervals). This approach guards against spatial aliasing and therefore facilitates identification of various seismoelectric arrivals as well as wavefield separation techniques such as f-k filtering (although such filtering was not necessary to reveal the shallow seismoelectric events measured at this site).

Figures 2(a) and (b) show seismic and seismoelectric super gathers centred at the 128 m mark on the survey line. In order to remove any risk of cross-talk, the seismoelectric data were acquired first before placing geophones at the dipole midpoints and repeating the shots. The two gathers exhibit many similarities. There are direct arrivals, ground roll and one or two shallow seismic reflections, which appear as hyperbolas. In the seismoelectric record, these events represent co-seismic signals. The two gathers differ at early time however where a remarkably clear seismoelectric signal, (1), can be seen arriving simultaneously at offsets up to 40 m from the shot. The signal is inverted in polarity on opposite sides of the shot and arrives 35 ms after impact, well before the arrival of co-seismic signals over most of the receiver spread. Both of these characteristics are consistent with the model of a vertical electric dipole-like source and we conclude that the signal is most likely an interfacial seismoelectric effect of electrokinetic origin.

The arrival time of this prominent interfacial effect is one-half the arrival time of the reflection hyperbola appearing at 70 ms in the seismic data. This indicates that the same interface is responsible for the seismoelectric conversion and the seismic reflection. Based on the local geology and borehole experiments at nearby sites, we anticipate that this interface is the water table which is the strongest near-surface acoustic impedance contrast and commonly found at depths ranging from 10 to 20 m. The seismic data in

Figure 2(a) provides two ways for us to estimate it's depth at this site. Refraction modelling suggests a two layer model consisting of 10.5 m of unsaturated sediments with a velocity of 320 m/s underlain by saturated sediments with a velocity of $\approx$ 1780 m/s. Alternatively, the two-way time to the onset of the reflection hyperbola, 70 ms, and the observed normal moveout velocity of 400 m/s suggest a depth of 14 m. We suspect that our refraction interpretation underestimates the depth because it is unable to resolve an increase in velocity through the vadose zone that would be expected due to increased sediment compaction and water saturation with depth.

This depth estimate allows us to compare the amplitude versus offset characteristics of the measured interfacial signal (1) to the amplitude variations that would be expected using the approximate model of a vertical dipole source located 14 m below the shotpoint. Figure 3 indicates that the dipole model is reasonable as a first order approximation although the measured amplitudes decay slightly more gradually than predicted. The difference may be attributed to the shallow depth of the interface (14 m) which is not much larger than the radius of the first Fresnel zone (approximately 8 m) over which the source is distributed (Garambois and Dietrich, 2002). The amplitudes best follow the trend between 12 and 35 m where they can be reliably measured.

Although the water table provides a strong interfacial signal, it is not the only interface detected. Figure 2 (c) illustrates two additional events, (2) and (3), which are seen in other super gathers along the profile. In spite of its weak amplitude, event (2) has the phase reversal expected for an electrokinetic interfacial signal; it appears more clearly in the stacked section presented later. The lack of polarity reversal on the shallowest event, (3), suggests that it is not electrokinetic in origin and not a "direct field" signal of the type reported by Haines et al. (2007). We are uncertain of its origin but speculate that it could be a result of strong downgoing seismic waves modulating the resistivity of a shallow layer through which telluric currents are flowing, thereby modulating the voltage drop from those currents across the dipole receivers. This 'resistivity modulation' mechanism has been recognized for some time (Thompson, 1936; Long and Rivers, 1975; Russell et al., 1997) but has not been extensively studied.

Creation and Interpretation of a Seismoelectric Section

Our survey was designed to yield a stacked seismoelectric section that would be analogous to a common depth point stack in multi-channel seismic reflection surveying. The approach was similar to that used by Thompson and Gist (1993) for their larger scale experiment but did not require wavefield separation filtering since the near-surface velocity structure at this site naturally provided good separation between the interfacial and co-seismic signals over a wide range of offsets.

The first step in the processing flow was to assemble vertical stacks of the shot records at each shotpoint and reverse the polarity of the traces at negative offsets. A tapered mute was used to remove the portion of each shot record dominated by co-seismic noise and mean scaling was then applied to the data before traces with offsets between 14 and 40 m were stacked to form a single trace which was plotted at the shot location. Since seismoelectric conversions from a near-horizontal interfaces are expected to be anti-symmetric about the shotpoint, the polarity reversal and stacking process enhances any interfacial effects relative to noise from distant sources which would be of the same polarity on either side of the shot. It would also tend to cancel event (3) in Figure 2(c). This process was repeated for each of the 300 shots spaced 1 m apart. Each stacked trace was then averaged with six neighbouring traces (three from each side) to enhance coherency and yield the stacked seismoelectric section shown in Figure 4.

Figure 4 also shows a 50 MHz GPR profile collected at the same site although at a different time and on the other side of the road, approximately 5 m away. The resolution of the GPR data is better at this site because dry sandy conditions limited the seismic pulse bandwidth. However there are many similarities between the two profiles, including the indications of sedimentary heterogeneity that appear in the form of discontinuous events at 35m, 128 m and 280 m along the line.

The strong coherent signal (1) related to the water table appears clearly in both profiles. The depth estimate of 14 m given above is consistent with the signal's arrival time on the GPR profile if we assume a radar wave velocity of 0.14 m/ns - a reasonable value for partially saturated sands. We note however that the GPR data was collected several months prior to the seismoelectric survey. A second interfacial signal (2), identified as a weak event in the super gather of Figure 2(c), can also be traced across most of the seismoelectric section. We speculate that it originates at the same interface as a shallow GPR reflection exhibiting similar morphology and represents a water retentive layer. Depth estimates from the seismoelectric and GPR profiles place this interface at a depth between 6 and 7 meters – consistent with the depth of 6 m reported for the 'coffee rock' layer in borehole P-90. The difference in the separation of events (1) and (2) on the seismoelectric and GPR time sections can be attributed to the tendency for seismic velocity to increase with depth due to sediment compaction and increased water saturation; in contrast radar wave velocity decreases with increasing moisture content.

Finally, we note that there are some differences between the seismoelectric and GPR profiles, particularly beneath the topographic low at the west end of the line where the GPR data exhibits more complexity. This suggests that the two methods provide complementary information given differences in their sensitivities to various physical parameters. More work is required to ascertain which physical properties variations are most important in the seismoelectric case.

CONCLUSIONS

The results of this experiment demonstrate that it is possible to use seismoelectric profiling to map subsurface interfaces within partially and fully saturated sediments. In particular, it allowed us to image the water table as well as a shallower interface interpreted as a water retentive layer, which was not resolved by seismic reflection or refraction. The observed variations in interfacial signal amplitude with offset provide a first order fit to the simple, approximate model of a vertical electrical dipole-like source, thereby supporting the interpretation that the signal is of electrokinetic origin. The physical property changes most important for the generation of the observed interfacial signals are not known conclusively. However, we suspect that the strong response from the water table is likely related to significant changes in acoustic impedance and electrical conductivity accompanying the relatively abrupt increase in water saturation that would be expected in coarse grained sediments such as sands. The signal generated at the water retentive layer is expected to be related to similar physical parameters but may also include variations in porosity or permeability.

Relatively dry, sandy near-surface conditions such as those found on the Gnangara Mound are challenging for seismic surveying because they typically exhibit high seismic absorption coefficients with a resultant decrease in high frequency content and resolution. In the case of this seismoelectric field trial however, the disadvantages was offset somewhat by the fact that (i) co-seismic signals associated with direct P-waves were slow to spread across the receiver array, and (ii) the water table was sufficiently deep to allow for clear separation between different interfacial signals within the vadose zone.

Our results demonstrate that it is possible to measure interfacial seismoelectric effects from depths exceeding 10 m and show that the method may become a valuable tool, sensitive to the presence of pore water and complementary to GPR, for the characterization of aquifers. It is also foreseeable, that the method could be useful at much greater depths as it continues to evolve and more concerted efforts are made to separate interfacial effects from co-seismic interference.

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FIGURE CAPTIONS

Figure 1: Illustration of the seismoelectric array geometry and shooting progression.

Figure 2: Seismic (a) and seismoelectric (b) super gathers centred at the 128 m mark along the line. A second seismoelectric super gather (c), from the 36 m mark, reveals additional shallower interfacial effects. The signal to noise ratio in (b) has been improved further by stacking five adjacent super gathers. Trace spacing is 1 m and the RMS amplitude of each trace has been normalized to be the same value.

Figure 3: Predicted (solid line) and measured amplitude vs offset for interfacial seismoelectric signal (1) in Figure 2(b) emanating from 14 m depth. Crosses and triangles represent measurements at positive and negative offsets respectively. The dipole moment was adjusted to best fit the measured data.

Figure 4: GPR (top) and seismoelectric (bottom) sections acquired along a 300 m traverse. Variable time delays (elevation statics) have been applied to the traces in each profile to account for topography relative to arbitrary datums.

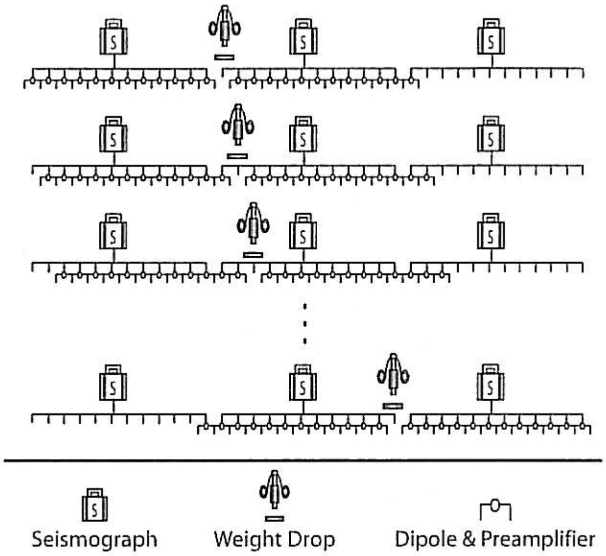

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Figure 1

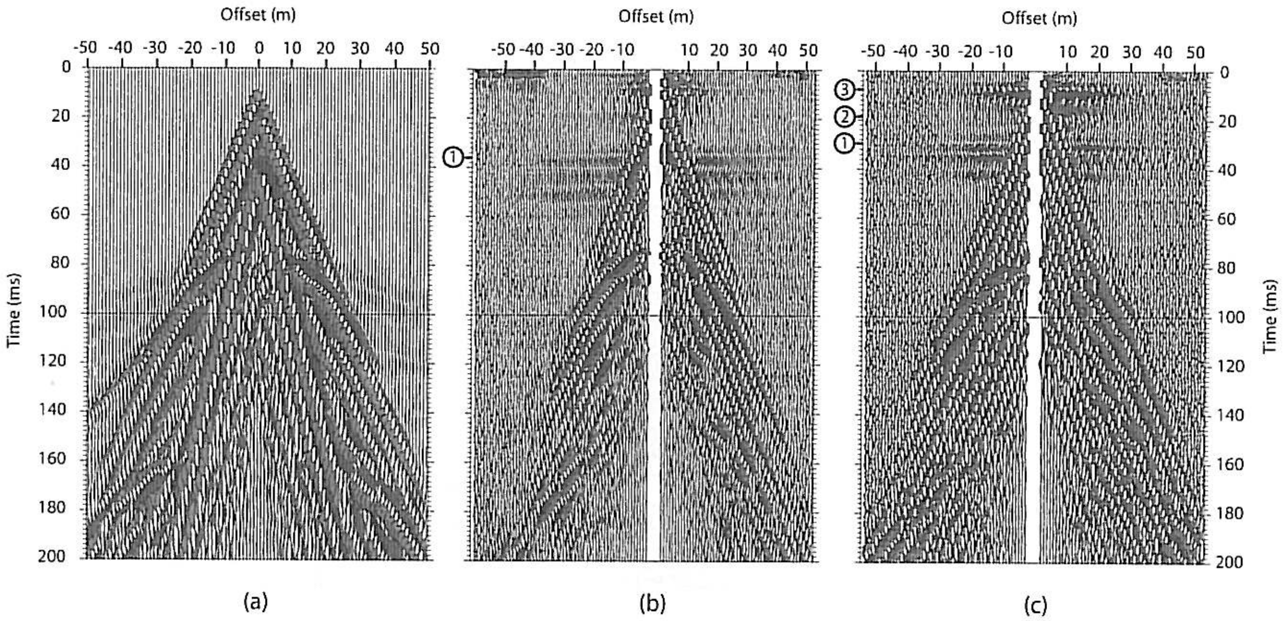

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Figure 2

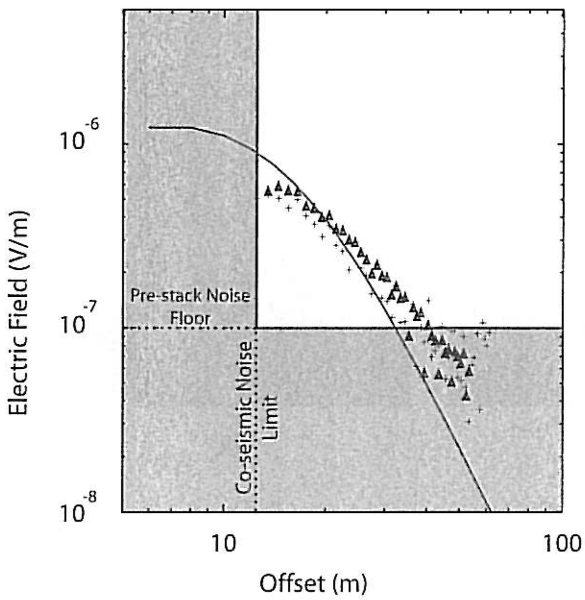

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Figure 3

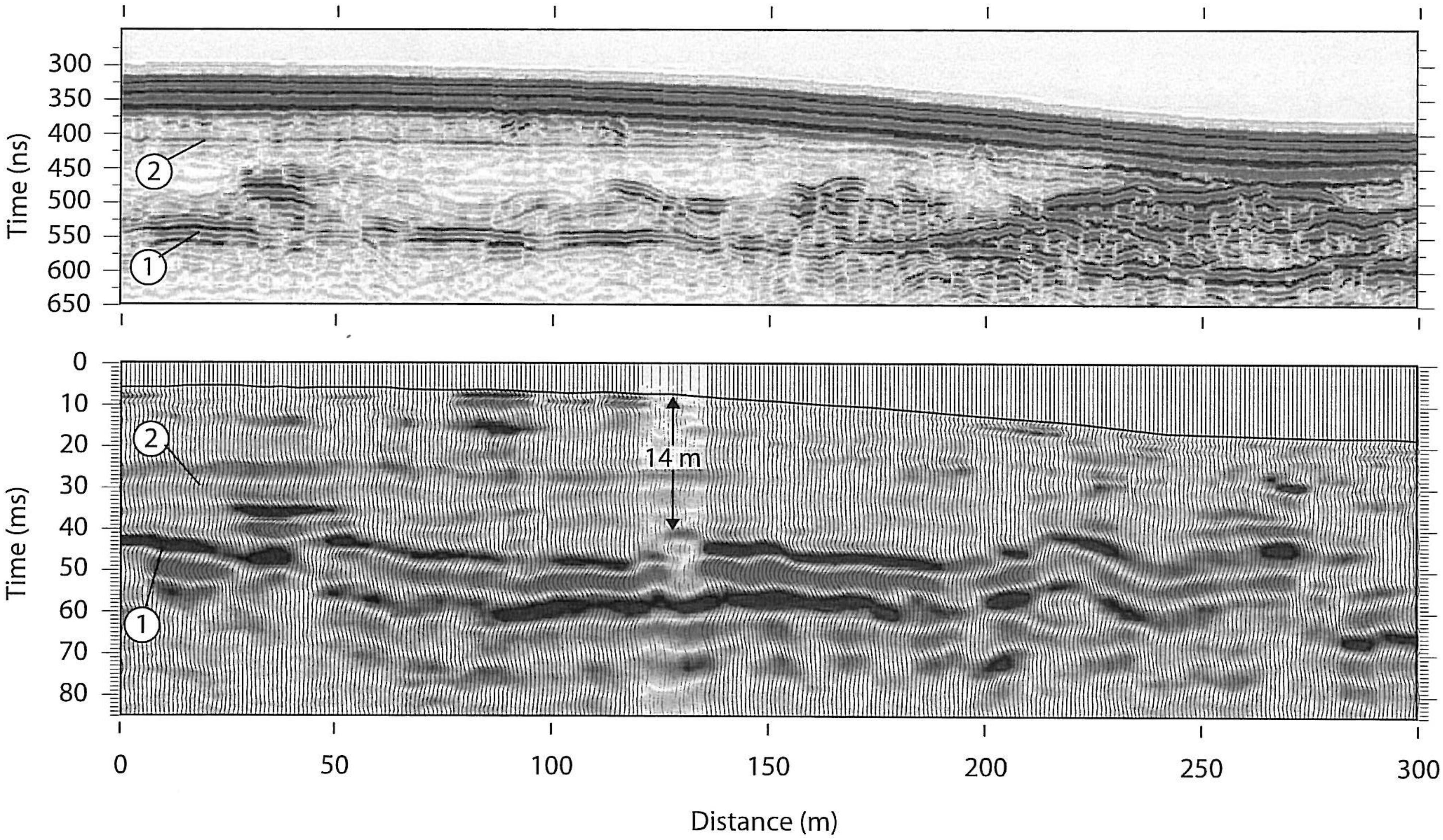

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Figure 4

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Seismoelectric imaging of the vadose zone of a sand aquifer

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Additional Information