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Measurement of the seismoelectric response from a shallow boundary

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ABSTRACT

Field experiments carried out at a site near Vancouver, Canada have shown that a shallow lithologic boundary can be mapped on the basis of its seismoelectric response. As seismic waves cross the boundary between organic-rich fill and impermeable glacial till, they induce electric fields that can be measured at the surface with grounded dipole receivers. Sledgehammer and blasting cap seismic sources, positioned up to 7 m away from the interface, have produced clear seismoelectric conversions.

Two types of seismoelectric signals are observed. The primary response is distinguished by near simultaneous arrivals at widely separated receivers. Its arrival time is equal to the time required for a seismic P-wave to travel from the shotpoint to the fill/till boundary. On the surface, its maximum amplitude (about 1 mV/m) is measured by dipoles located within a few meters of the shotpoint. At greater distances, the amplitude of the primary arrival decays rapidly with offset, and secondary seismoelectric arrivals become dominant. They differ from the primary response in that their arrival times increase with dipole offset, and they appear to be generated in the immediate vicinity of each dipole sensor.

Our studies show that the responses cannot be attributed to piezoelectricity or to resistivity modulation in the presence of a uniform telluric current. We infer that seismically induced electrokinetic effects or streaming potentials are responsible for the seismoelectric conversion, and a simple electrostatic model is proposed to account for the two types of arrivals. Although our experiments were small in scale, the results are significant in that they suggest that the seismoelectric method may be used to map the boundaries of permeable formations.

INTRODUCTION

The use of seismic sources to excite electromagnetic responses in the earth has been studied sporadically since the 1930s. However, the potential of seismoelectric methods remains largely unknown. This is due in part to the fact that the various conversion mechanisms are only partially understood. More importantly, the body of empirical evidence showing that seismoelectric signals can be clearly linked to geologic features is relatively small. The principal experimental difficulty is the identification of converted signals in the presence of much larger ambient electromagnetic noise. However, in recent years it has become possible to combat this problem through the use of recording systems with high dynamic range and digital processing techniques. At our Haney test site, near Vancouver, we have measured seismoelectric conversions that clearly originate at a shallow interface 1 to 3 m below the surface. In this paper, we show that the response exhibits the characteristics expected of a seismoelectric conversion, and that it can be used to map a subsurface boundary between road fill and glacial till.

There are at least four seismoelectric effects of interest in exploration geophysics: (1) the modulation, by seismic stress, of the resistivity of a volume of earth through which steady currents flow (Thompson, 1936); (2) seismically induced electrokinetic effects or streaming potentials (Ivanov, 1939; Thompson and Gist, 1993); (3) the piezoelectric effect (Volarovich et al., 1962; Russell et al. 1992); and (4) highly

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nonlinear processes that generate radio-frequency impulsive responses in sulfide-rich rocks (Sobolev et al., 1982; Kepic et al., 1995). Regardless of the conversion mechanism, seismoelectric exploration methods involve the use of a seismic source and electric or magnetic field receivers. In most earth materials, electromagnetic signals travel much faster than seismic waves. As a result, the delay between the shot moment and reception of an electromagnetic response is essentially equal to the time taken by the seismic wave to travel from shotpoint to target. The product of this delay with the seismic velocity gives an estimate of the distance from shotpoint to target. Data from different shotpoints can be used to locate or delineate the target.

Seismically induced electrokinetic effects appear to be the most reasonable explanation for the conversions we observe at Haney. Electrokinetic effects are the result of a coupling between fluid flow and electric current flow that arises because of the electric double layer that exists at a solid-liquid interface. The double layer consists of a layer of ions adsorbed on the solid matrix, and a parallel, diffuse layer of counterions in the pore fluid. At least part of the diffuse layer is free to move with the pore fluid. Thus, the flow of fluid relative to solid allows for the possibility of charge separation and the development of an electric field. Such relative motion is produced by the passage of seismic waves through porous media. Neev and Yeatts (1989) extended Biot's equations for elastic wave propagation in porous media (Biot, 1956) to calculate the electrokinetic effects accompanying compressional waves in the quasi-static case. Haartsen and Pride (1994) have developed a method recently for solving the general time-varying problem in which Maxwell's electromagnetic and Biot's mechanical wave equations are linked through electrokinetic coupling. They predict that seismic body waves in layered media produce two types of seismoelectric effects: (1) nonradiating fields that accompany body waves and are only observed when such waves pass by an EM receiver, and (2) EM fields that are generated by seismic waves traversing boundaries where there is a change in elastic or electric properties (including for example, porosity, permeability, and fluid chemistry).

In 1939 and 1940, Ivanov reported field measurements of the seismoelectric effect in sedimentary materials and proposed an electrokinetic mechanism for the conversion. He used the term "seismoelectric effect of the second kind" or "E-effect" to differentiate his observations from the modulation of earth resistivity by seismic stress. The responses recorded by Ivanov were not clearly linked to a subsurface interface. Rather, they were observed whenever a seismic wavefront arrived near the grounded dipole sensor. The delay between the moment of shot detonation and reception of an electromagnetic response depended on the shot-dipole separation rather than the distance from the shot to any particular interface or target.

Since the time of Ivanov, there have been several other reports of seismoelectric effects observed in the field and attributed to electrokinetic phenomena. Martner and Sparks (1959) documented clear seismoelectric responses from the base of the seismic weathered layer. Broding et al. (1963) measured the response along a 35 m profile in a borehole and found that it peaked opposite a sandy loam/shale interface. Borehole measurements by Parkhomenko and Gaskarov (1971) showed that seismoelectric responses in limestone were consistently stronger than those observed in clays. Migunov (1987) reported that electrical signals were generated when seismic waves reached the boundaries of kimberlite pipes. Maxwell et al. (1992) found that piezoelectric responses from a quartz vein were accompanied by other responses originating in the surrounding sediments and host rock. Large scale field experiments have been carried out by Thompson and Gist (1991, 1993). They reported that they were able to detect electrokinetic conversions from boundaries between impermeable rocks and permeable water-saturated sands at depths of up to at least 300 m.

The measurements discussed in this paper, and in two preliminary reports (Maxwell et al., 1992; Butler et al., 1994) have been made during numerous visits to the Haney site beginning in 1991. We have identified two types of seismoelectric arrivals in the Haney data. The primary response is generated when the seismic wave first reaches a boundary between road fill and glacial till. This is the dominant signal observed close to the shotpoint, and it is distinguished by the fact that it arrives simultaneously at widely separated sensors. As the shot-dipole offset increases, the amplitude of the primary arrival decays rapidly and secondary arrivals become evident. Unlike the primary response, the arrival times of these secondary signals tend to increase with dipole offset. Their apparent velocities are comparable to seismic velocities and they are likely caused by seismic waves generating electric fields in the immediate vicinity of each sensor. Our observations are similar to those reported in Martner and Sparks (1959). In some cases they recorded signals like our primary arrivals—a seismoelectric conversion occurred at the boundary immediately below the shot and was observed simultaneously by multiple dipole receivers. In other cases, the seismoelectric signal appeared to originate at the weathered layer boundary in the immediate vicinity of each dipole.

SITE DESCRIPTION

The Haney site, near Vancouver, Canada, is located on an unimproved dirt road that runs along the side of a steep slope in the Malcolm Knapp Research Forest of the University of British Columbia. The road fill, consisting of a permeable, organic-rich soil, overlies a highly impermeable, silty glacial till. There are four boreholes at the site, and drill cores show that the fill varies in thickness from 1 to 3 m across the width of the road. The deepest borehole, drilled to a depth of 10.4 m, penetrated 7.7 m of glacial till without encountering any obvious change in lithology. Attempts to reach bedrock were thwarted by difficult drilling conditions (abundant boulders) in the deeper part of the till layer. However, seismic refraction data indicates that bedrock lies at a depth of 10 to 15 m. The P-wave velocities of the fill, till, and bedrock are approximately 250, 2100, and 4000 m/s, respectively.

METHOD

The main components of the data acquisition system are a seismic source, geophones, grounded dipole sensors, amplifiers, and an eight-channel recorder. All instruments are battery powered to eliminate electrical noise produced by portable generators. Seismic arrivals are detected by conventional vertical component geophones. Electrical responses are measured by grounded dipole receivers consisting of two stainless steel stakes penetrating about 0.3 m into the ground and separated by 0.5 to 10 m. Each dipole is connected to a local differential preamplifier having a high input impedance (2 MΩ differential), a gain of 30, and a bandwidth of 2 Hz to 30 kHz.

Signals from each preamp and geophone are transmitted to the recording site along separate shielded twisted pair cables. Additional gain and band-pass filtering are then applied by passing each channel through a differential amplifier (Tektronix model AM502) prior to digitization. Dipole and geophone signals are filtered using corner frequencies of 0.1 to 10 Hz at the low end and 1000 Hz at the upper end. Data are digitized and recorded by a 12 bit A/D board mounted in a portable computer. Typically, we record eight channels of data with a sample interval of 0.05 ms and a record length of 400 ms. Records may be stacked using software that we have written to control the board.

A sledgehammer is a sufficiently energetic seismic source at Haney. Triggering is usually accomplished by monitoring the output of an accelerometer (piezoelectric transducer) mounted on the hammer. Experiments have shown that the impact of a hammer on an aluminum base plate, such as those commonly used for hammer seismic surveys, can generate an undesirable electromagnetic transient. This interference can be minimized by using a nonmetallic base plate made of plastic or hardwood. It is not necessary to replace the steel hammer.

We have also used blasting caps in water-filled boreholes to produce seismic sources at depth. Conventional zero-delay caps, or seismic caps are not suitable because the large initiation current can contaminate seismoelectric records at the time of detonation. Instead, we use fuse caps or electrical delay caps. Data acquisition is triggered by the flash of light that accompanies cap detonation. This flash is transmitted to the recording site by a fiber optic cable, one end of which is taped to the blasting cap (Kepic and Russell, 1996). Experience has shown that this method of triggering is reliable and accurate. Furthermore, it does not generate the electromagnetic noise associated with other triggering methods which rely on the termination or production of an electrical current at the time of detonation.

Borehole damage is an unfortunate side effect of the blasting cap source. The boreholes at Haney are lined with PVC or ABS pipe having inner diameters of 2 to 3 inches. Explosive pressures tend to destroy the borehole lining around the blasting cap and block the borehole at that depth. We attempted to eliminate this problem by enclosing the blasting cap in an open-ended steel cylinder having a diameter slightly less than that of the plastic liner and a length of about 20 cm. That technique was abandoned when we discovered that it produced an electromagnetic transient at the moment of detonation. The same problem would be expected for detonations in a steel-cased borehole. Apparently, one should avoid deforming metal, either by hammer or explosive, when making electrical measurements close to the shotpoint.

Powerline harmonic interference dominates the raw data we acquired at Haney. The peak-to-peak magnitude of noise at 60 Hz for example averages 5 mV/m which is about 5 to 500 times larger than the signals we measure. The use of a single-notch filter at 60 Hz would not solve the problem as several harmonics of 60 Hz are also quite strong. Instead, we suppress the harmonic noise routinely using a digital processing technique that involves subtracting from each trace sinusoids of the appropriate frequencies, amplitudes and phases. The amplitude and phase of each powerline harmonic are estimated using a least-squares approach over a portion of the trace that contains no signal [Butler and Russell, (1993); see Linville and Meek (1992) for a similar technique]. Unlike notch filtering, this procedure is capable of removing multiple harmonics without distorting or attenuating the signal. We have obtained improvements of up to 45 dB in the signal-to-noise ratio by subtracting 10–20 harmonics of 60 Hz from recordings made at Haney.

During some experiments, we have used a remote reference dipole to aid in noise suppression. The remote reference, located 60 to 70 m away from the shotpoint records the regional noise caused by powerlines and spherics, but not the seismoelectric response generated near the shotpoint. Regional noise is substantially reduced by subtracting scaled versions of the remote record from data acquired by the other dipoles. Major improvements in signal-to-noise may also be obtained by taking the difference between two dipoles placed symmetrically about the shotpoint. This method tends to cancel out regional noise and enhance the signal. Thompson and Gist (1993) have used it during large-scale field experiments. However, because the method yields an average of the signals measured by two dipoles, it is not ideal for studying how the seismoelectric response varies with dipole position, and has not been used on the data presented in this paper.

RESULTS AND DISCUSSION

Figure 1 shows seismic and electrical responses to a single sledgehammer blow on the road at Haney. The uppermost trace is from an accelerometer attached to the head of the hammer and simply gives the moment of hammer impact. Traces 2 and 3 are signals from the geophones located at the electrodes closest to the shotpoint. The final five traces show the (time-varying) potential difference across each dipole (i.e., the potential at the northern electrode minus that at the southern electrode). Powerline noise has been removed from the electrical traces by sinusoid subtraction

A clear seismoelectric response is visible on the dipole traces. It arrives simultaneously on all five dipole traces 6 ms after hammer impact. Since it precedes the seismic arrivals, the response cannot be attributed to shaking of the dipole electrodes, or to geophone cross talk. The simultaneous arrival at different dipoles is consistent with the idea that the response is generated at depth and propagates rapidly to the surface as an electromagnetic signal. Note that the polarity of the response is reversed on opposite sides of the shotpoint. This shows that electrodes near the shotpoint initially detected a drop in electric potential relative to the more distant electrodes. The peak to peak magnitude of the response is about 2.4 mV across dipoles D1–D4, and 0.8 mV across the more distant dipole (D5). The early cycles of both the electrical and the geophone signals exhibit similar dominant frequencies (70–100 Hz) but the electrical frequencies appear to be slightly higher. The later, low frequency cycles in the geophone data are surface waves.

Figure 2 shows the electrical responses measured by dipole D2 as the shotpoint was moved in 0.5 m increments across the road at Haney. The line of shotpoints lay between the 0 and 6 m marks in Figure 1b. The seismoelectric signal arrived earliest (about 4 ms after hammer impact) at the western edge of the road, and the delay gradually increased to about 14 ms as the shotpoint was moved to the east. This suggests that the distance to the seismoelectric source increased as the shotpoint was moved to the east.

The interpretation of the first arrival is also shown in Figure 2. At each shotpoint, an arc has been drawn with a radius equal to the product of the seismoelectric delay and the P-wave velocity of 250 m/s. Neglecting out-of-plane effects, the feature responsible for the seismoelectric conversion should be tangent to all of the arcs. The seismoelectric data therefore delineate a dipping interface 1 m deep in the west and 3.5 m deep at the eastern edge of the road. As shown in Figure 2, two shallow boreholes confirmed the existence of a dipping boundary between road fill and glacial till. The 40- to 50-cm discrepancy between the actual and estimated depths can be ascribed to uncertainty in the first-break picks (1 to 2 ms), and in the P-wave velocity of the road fill.

As a further check on the identity of the target, we measured the seismoelectric responses generated by the detonation of blasting caps at various depths in a borehole. Figure 3 shows the experiment layout and a typical shot record (the shot at 7.7 m depth). All measurements were made at the surface using one uphole geophone and six dipoles located 2, 4, 6, 8, 12, and 16 m north of the borehole. The uppermost trace is from the fiber optic trigger circuit and gives the time of detonation. The second trace shows the seismic arrival at the surface, and remaining traces are electrical responses measured by the dipoles. The responses differ in character but the first arrival occurs simultaneously at all six sensors 2 ms after detonation. This arrival is clear at near offsets but barely visible above the noise at the most distant dipole. The 2 ms delay corresponds to the time required for the seismic wave to travel 5 m up to the base of road fill where the seismoelectric conversion takes place. There is a further delay of 13 ms before the seismic wave reaches the geophone. This is the time required for the seismic wave to travel through 2.7 m of fill.

In Figure 4, we have plotted the seismoelectric responses measured by dipole D3 as the shotpoint was moved up from the bottom of the borehole. The most striking feature is the abrupt change in signal character that occurs opposite the fill/till interface. Blasting caps detonated below the boundary yielded higher amplitude and higher frequency responses. We speculate that this is indicative of better seismic coupling in

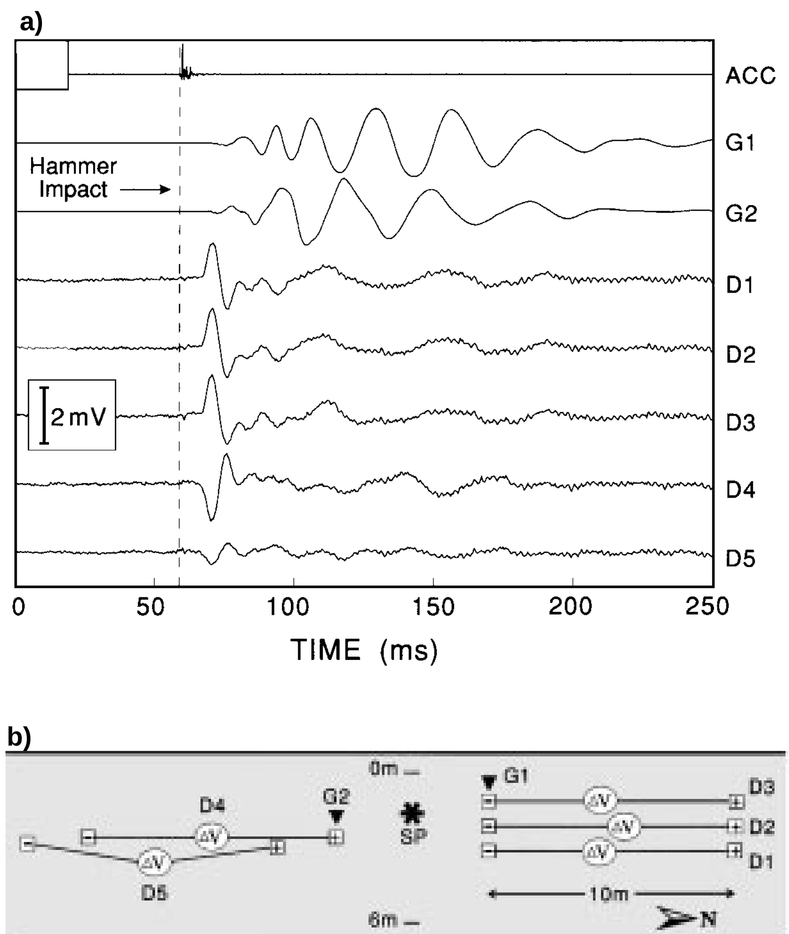


FIG. 1. (a) A shot gather showing the seismic (G1, G2) and electrical (D1–D5) responses to a single sledgehammer blow at the Haney site. The dotted line indicates the moment of sledgehammer impact as determined by the accelerometer trigger (ACC). Powerline harmonic noise has been removed from the dipole traces by sinusoid subtraction. (b) Plan view of the experimental layout on the road (shaded area) showing the shotpoint (SP), two geophones, and five 10 m dipoles. The geophones are located within a few centimeters of the electrodes closest to the shotpoint (i.e., 3 m from the shotpoint).

the dense, competent glacial till than in the loose, highly compressible road fill. The voltage spike at time zero on the trace at 8.7 m was caused by the use of a particularly narrow steel blast chamber. Blast chambers were not used for any of the shallower shots.

The main point illustrated by Figure 4 is that it shows clearly that seismoelectric conversion occurs at the fill/till interface. As expected, the delay between the instant of detonation and the reception of a response was proportional to the distance between the shot and the road fill/glacial till boundary. There was no delay when the shot was located directly opposite the boundary. The first arrivals can be fit well by two straight lines that intersect at time zero near the fill/till interface. The slopes of these lines, 200 and 2300 m/s, provide estimates of the seismic velocities in the road fill and glacial till respectively. These are in reasonable agreement with the average P-wave velocities of 250 and 2100 m/s derived from seismic measurements at the site.

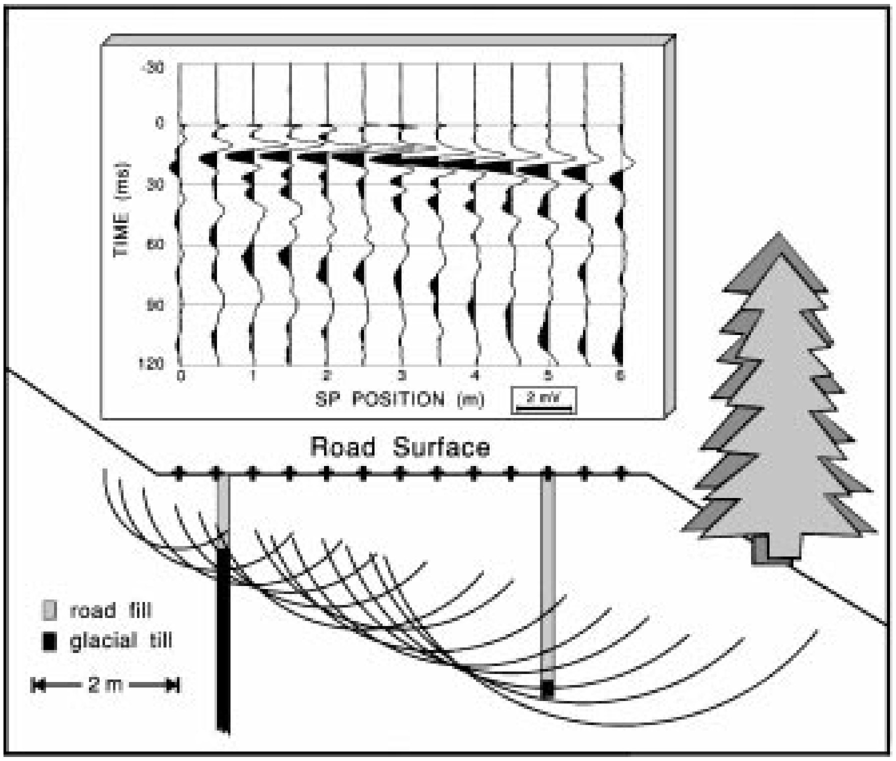


FIG. 2. Seismoelectric signals generated at 13 shotpoints distributed across the dirt road. Time zero indicates the moment of hammer impact. The cross-section below shows the estimated position of the seismoelectric target (a dipping interface tangent to the arcs), as well as the actual depths to the road fill/glacial till boundary in two boreholes.

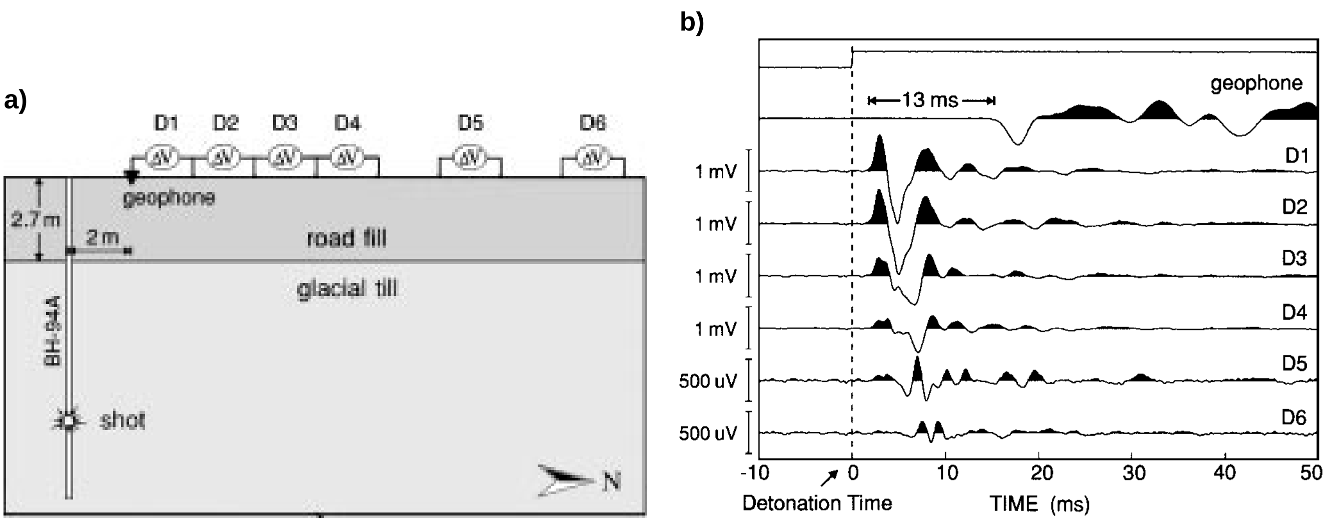


FIG. 3. (a) Illustration of the experimental layout used to measure seismoelectric responses to the detonation of blasting caps in a borehole. One geophone and six 2 m dipoles were deployed on the surface. (b) Response observed for a shot at 7.7 m depth.

Given that electrokinetic effects involve the motion of pore water, the state of water saturation in the subsurface is of obvious interest. We have monitored the position of the water table at Haney by measuring the natural water level in boreholes. These measurements ranged from about 0.5 m above, to a few meters below the fill/till boundary depending on the season and the road fill thickness at the point of measurement. Our studies indicate that the seismoelectric effect is present and originates at the same point (the base of fill) regardless of the water level. For example, the natural water level was 1 m below the fill/till interface during the experiment displayed in Figure 4, but the same type of data were acquired the previous year when the water table was 0.35 m above the interface. In both cases shots detonated at the boundary generated immediate electrical responses while those detonated at the water table yielded delayed signals.

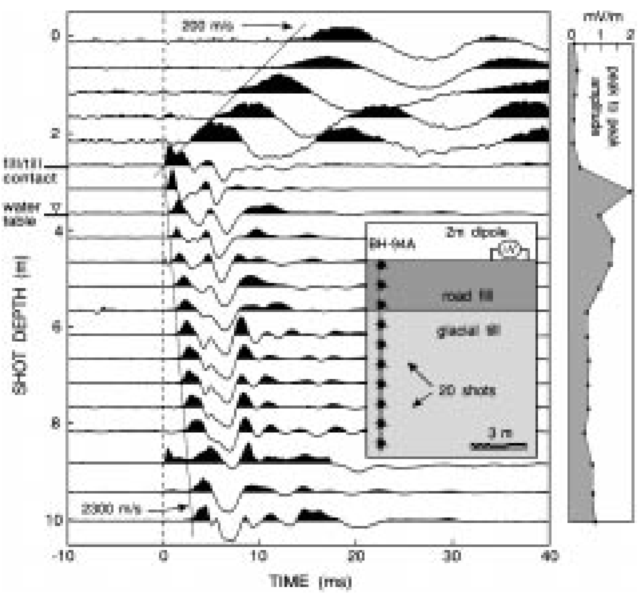


FIG. 4. Seismoelectric response versus shot depth in borehole 94A. These electrical signals were measured by a 2 m dipole on the surface 6 m north of the borehole. Note that the delay between detonation and response depends on the distance to the fill/till interface at 2.7 m depth. The peak to peak amplitude of each trace is indicated on the right.

We have also carried out experiments to determine how the seismoelectric response varies with dipole offset. Figure 5 shows a comparison of seismic and seismoelectric arrivals at offsets of 2 to 26 m from a shotpoint on the road surface, about 2 m above the fill/till interface. Twelve vertical geophones, and thirteen 2 m dipoles were laid out along the road to the north of the shotpoint as shown on the map. The two data sets were collected separately but with the same sledgehammer source.

The seismic data appear to be dominated by surface waves, and it is difficult to identify any meaningful arrivals apart from the first breaks. Beyond a distance of 6 m, the breaks are caused by a head wave traveling along the interface between road fill and glacial till. The rise in the first breaks at an offset of about 14 m is probably because of localized thinning of the road fill. In contrast, the seismoelectric record exhibits negligible surface wave interference and at least three coherent arrivals. The primary arrival, occurring 9 ms after hammer impact, appears simultaneously on all traces out to about 15 m offset. This is the seismoelectric response produced by the arrival of the seismic wave at the road fill/glacial till boundary below the shotpoint. Beyond 15 m, it is too weak to be identified clearly. However, two secondary subparallel and nonsimultaneous arrivals are evident. The secondary signals appear in a 40–50 ms time window beginning 2 to 3 ms before the first seismic arrival at each dipole. Furthermore, their apparent velocities are comparable to those of seismic body waves. This

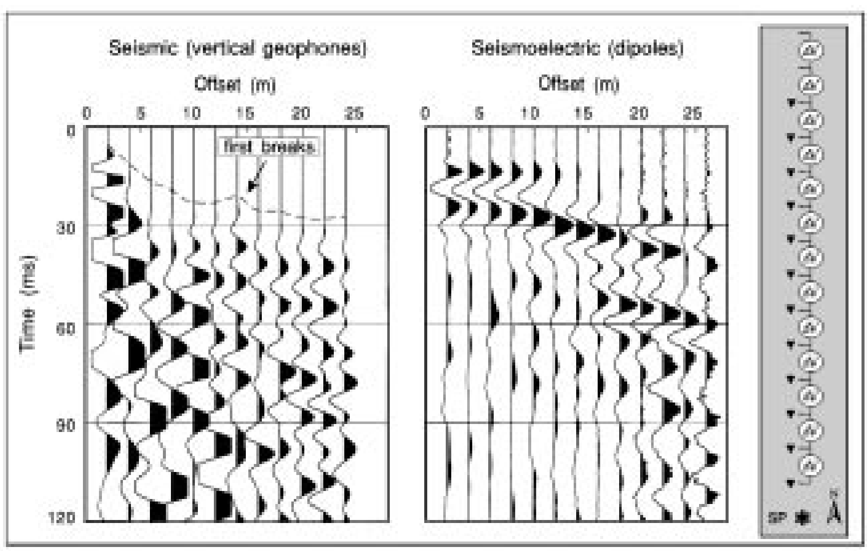


FIG. 5. Comparison of seismic and seismoelectric responses at offsets of 2 to 26 m from a shotpoint on the road. Unlike the seismic data, which is dominated by surface waves, the seismoelectric record shows three coherent arrivals. The first arrival, essentially simultaneous at all near offsets, is the electrical signal produced by a seismic wave crossing the fill/till interface 2 m below the shotpoint. Secondary events, probably related to the arrival of seismic waves near each dipole sensor, are evident at distant offsets. Trace amplitudes have been normalized for display purposes.

suggests that they may be associated with the arrival of seismic waves beneath each dipole beginning with the head wave that travels along the fill/till boundary. The two clear secondary events that appear between 30 and 60 ms exhibit moveout that is roughly hyperbolic. We speculate that they might be generated near each dipole by seismic waves that have been reflected from the till/bedrock interface. Unfortunately, any reflections of that type are obscured by ground-roll interference on the seismic record. At this time, the origin of the secondary arrivals is not well understood.

The seismoelectric traces in Figure 5 have been normalized for display purposes. As shown in Figure 6, the true peak to peak voltages measured across the 2 m dipoles varied by two orders of magnitude from a high of almost 3 mV at close range, to 20 μV at the farthest offsets. The amplitude of the primary seismoelectric response (the simultaneous first peak) is also plotted for the offset range where it is visible. Beyond an offset of 4 m, both amplitude curves can be approximated by straight lines in the log-log plot. The slopes of these lines indicate that peak to peak amplitudes decay approximately as $1/x^2$, while the amplitude of the primary signal falls off approximately as $1/x^4$, x being the dipole offset. The latter rate of decay is the same as would be exhibited by the horizontal component of the electrostatic field from a vertical electric dipole at the fill/till boundary beneath the shotpoint. The source of the primary seismoelectric response can be modeled therefore as a buried vertical dipole for offsets greater than 4 m.

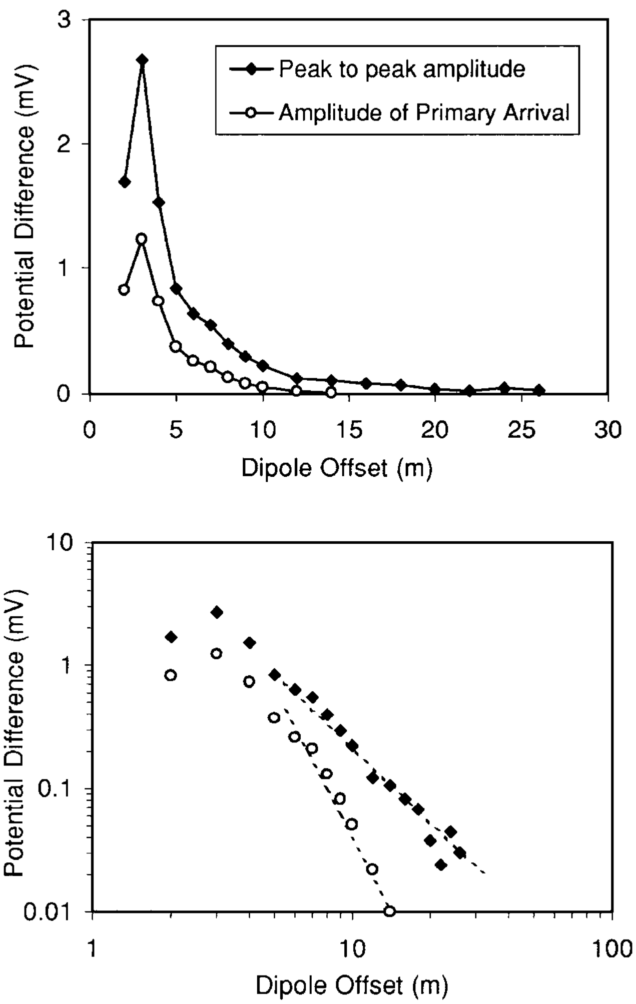


FIG. 6. Linear and log-log plots of seismoelectric amplitude versus offset for the experiment shown in Figure 5. The solid symbols indicate overall peak to peak trace amplitudes while the open symbols give the amplitude of the primary (simultaneous) arrival. The dashed lines in the log-log plot have slopes of -2 and -4.

THE SEISMOELECTRIC CONVERSION MECHANISM

The preceding experiments have served to identify the seismoelectric target and put some constraints on possible models for the conversion mechanism. First of all, we conclude that the conversion is not a piezoelectric effect. Although the glacial till contains quartz, there is no reason to expect that the quartz-rich grains would have been deposited with the alignment required to produce a measurable piezoelectric response. Furthermore, if the till was piezoelectric, blasting caps detonated within it should have produced immediate electrical responses regardless of the distance to the fill/till boundary.

Resistivity modulation is another mechanism that must be considered. According to this model, the resistivity of a volume of earth varies with stress during the passage of a seismic wave. Electric potentials caused by any natural currents flowing in that volume therefore vary as well. To investigate this possibility, we used an experiment devised by Ivanov (1939). We measured electrical responses to seismic sources located 3 m away from either end of a dipole sensor. To first order, we can assume that both shots changed the effective resistance between the dipole electrodes in the same way. Then, given that the ambient telluric currents were expected to be horizontal beneath the dipole, any telluric potential drop should have varied in the same way (had the same polarity) regardless of whether the shot was to the left or the right. However, the two shots actually yielded responses with opposite polarities, indicating that resistivity modulation is not the relevant mechanism at this site.

Figure 7 shows the results of a more detailed investigation of signal polarity. Seven dipoles were arranged in a radial pattern about the shotpoint. Each measured the potential at its outer electrode relative to that at its inner electrode. Apart from some early source-generated noise (the dipoles were very close to the shotpoint), the main feature is the seismoelectric arrival at 10 ms. The signal polarities indicate that the response (measured at the surface) begins with a flow of current radially inward toward the shotpoint. In the absence of a current source or sink, this net horizontal flow toward the shotpoint must be balanced by a net vertical flow downward beneath the shotpoint. Again, we cannot envisage any likely scenario by which resistivity modulation could cause horizontal telluric currents to change in this fashion.

Seismically induced electrokinetic effects appear to offer the best explanation for the conversion mechanism. The rigorous theoretical treatment of this problem is complicated, but a useful qualitative explanation can be given by a simple electrostatic model. The model accounts for the two types of arrivals present in our field data, and provides physical insight into the nature of the conversion process. We begin by accepting that the propagation of a P-wave through porous media involves relative motion between the solid matrix and the pore water. An inevitable consequence of this is the development of spatial variations in the water content per unit volume which are analogous to density variations that would accompany propagation in an elastic solid [Biot (1962) called this variation "the increment of fluid content"]. Because the mobile pore fluid carries a net charge (caused by the structure of the electrical double layer), variations in water content correspond to variations in electric charge density. Thus, the P-wave produces regions of excess positive and negative charge parallel to the seismic wavefronts, and electric fields develop between those regions. In the case of a P-wave radiating from a point source, one can show that electric fields are observed when seismic wavefronts pass by a sensor and when the spherical symmetry of the charged regions is broken by a boundary. The existence of these two types of electrical arrivals is also supported by the more rigorous analysis of Haartsen and Pride (1994).

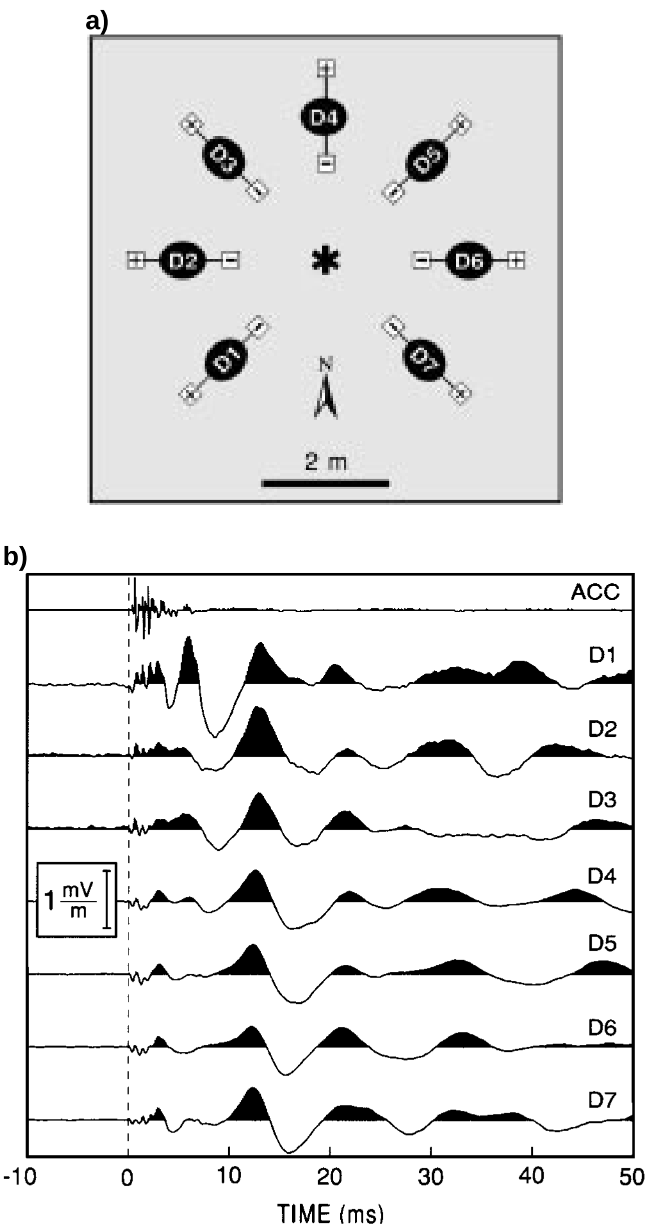


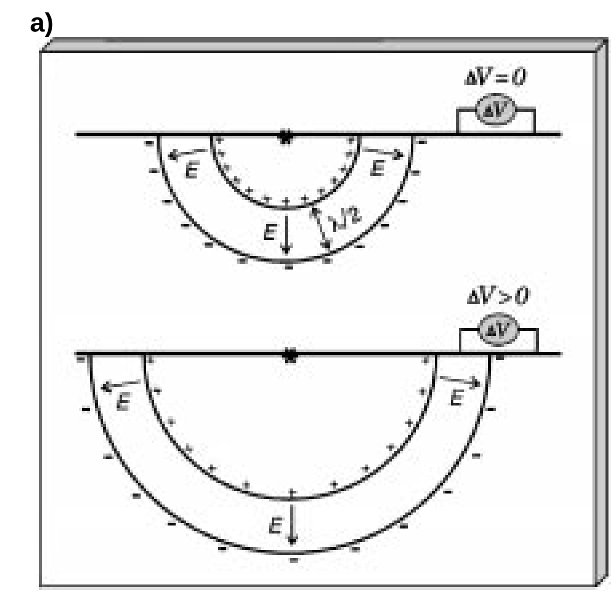
FIG. 7. (a) Plan view of seven 1.5 m dipoles arranged in a radial pattern to measure the directionality of the seismoelectric response. (b) This record, generated by a hammer blow at the center of the pattern, shows that the seismoelectric signal is approximately radially symmetric at the surface. The accelerometer trace (top) indicates the moment of hammer impact.

Our simplified model is illustrated in Figure 8. The regions of excess positive and negative charge are approximated by charged spherical shells separated by one-half of the dominant seismic wavelength. For a boundary condition, we require that the normal component of the electric field be zero at the earth's surface. Then, as long as the charged shells remain in a homogeneous medium, it is straightforward to show that the electrostatic field remains confined to the region between them. The contained electric field is not observed until the seismic wave reaches the dipole sensor (Figure 8a). This provides a qualitative explanation for the secondary seismoelectric signals measured at Haney. However, it does not explain why the *onset* of the secondary arrivals appears to precede the seismic first breaks by 2 to 3 ms at the far offsets in Figure 5.

Electric fields can also be observed when the spherical symmetry of the charged regions is broken by an inhomogeneity such as a boundary. In principle, the boundary could separate regions with differing permeabilities, elastic or electrokinetic properties (any property that affects the amount of charge transported by the seismic wave). Here, we consider the case of a perfectly reflecting boundary—a reasonable approximation for the case at Haney where loose fill overlies highly competent glacial till. As illustrated in Figure 8b, the reflected charge distribution can be modeled as the superposition of two simpler distributions: (1) a pair of hemispherical shells in homogeneous media, and (2) a pair of spherical caps joined at the boundary. As already discussed, the field associated with distribution (1) is confined to the region between the shells. The electrostatic field generated by distribution (2), however, is nonzero everywhere and can be observed simultaneously by widely separated sensors. For the case where the electrical properties are uniform across the boundary, it can be calculated analytically as the sum of the fields from two charged caps (e.g., see Jeans, 1948, 224). The horizontal component of this field has radial symmetry at the surface, and for sensor offsets significantly greater than the interface depth, it decays as $1/x^4$. Furthermore, the arrival time of this signal is equal to the seismic traveltime from shotpoint to boundary. The electrical response generated at a boundary in this way therefore provides an explanation for the primary seismoelectric arrival observed in our field data.

CONCLUSIONS

We have measured seismoelectric responses from a sedimentary boundary that are unusual in their clarity and detail. The boundary has been mapped by experiments involving multiple shotpoints, and a few stationary electric field receivers. Other experiments, involving several receivers and a stationary shot have revealed the existence of two types of seismoelectric arrivals. The success of our experiments hinged upon carefully designed instrumentation, effective methods for the removal of harmonic noise, and a test site with favorable geological conditions. Given the high signal-to-noise ratio obtained using a relatively weak seismic source, we expect the effective



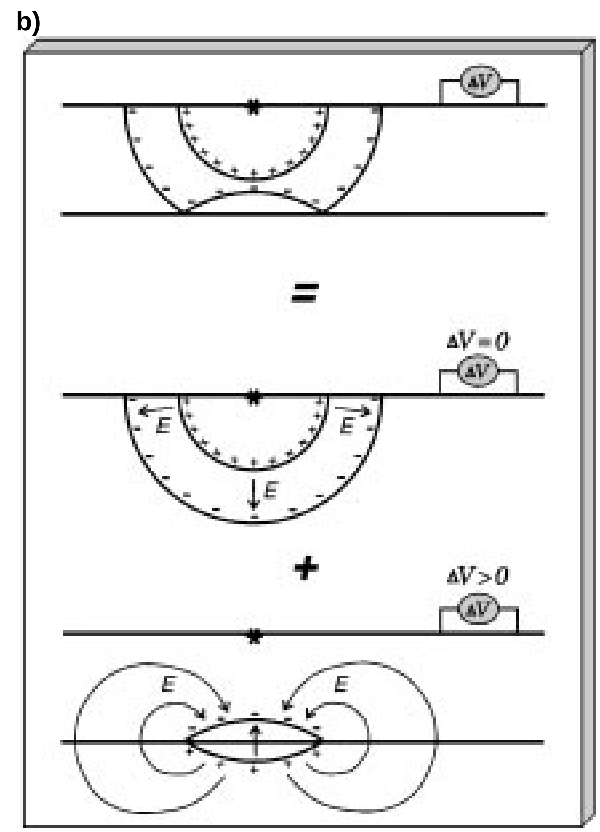


FIG. 8. Conceptual modeling of the two types of seismoelectric arrivals observed at Haney. Through electrokinetic coupling, regions of excess positive and negative charge form parallel to the peaks and troughs of a seismic *P*-wave. This charge separation produces electric fields that can be observed (a) when the *P*-wave passes by a receiver, and (b) when the spherical symmetry of the charged regions is broken by a boundary (here a perfect reflector).

depth of exploration to be much greater than the 3 m probed at Haney.

The boundary mapped at Haney is an interface between permeable, organic-rich road fill, and impermeable, silty glacial till, and seismically induced streaming potentials are believed to be responsible for the seismoelectric conversion. The experimental results, together with recent theoretical developments, indicate that seismoelectric effects may be used to map the boundaries of permeable formations.

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