Geophysical applications of electrokinetic conversion

By A.H. THOMPSON and G.A. GIST
Exxon Production Research Company
Houston, Texas

In Volume 1 of GEOPHYSICS (1936), R.R. Thompson of Humble Oil and Refining Company proposed that the coupling of seismic and electrical energy can be used as an exploration tool. Subsequent efforts were directed toward using electric-field antennas as geophones. Small-scale studies concluded that antennas are not as sensitive as conventional geophones and their response is highly variable, depending on the local soil conditions. By 1959, Mariner and Sparks of Pan American Petroleum reported in GEOPHYSICS a systematic study of seismoelectric coupling using explosives at various depths. Their results were the first to show that the conversion of seismic to electromagnetic energy at the water table could be detected using surface antennas. No studies have systematically explored the full potential of electroseismic methods using the same level of experimental sophistication used in seismic exploration.

We have been successful in detecting seismic-to-electromagnetic energy conversion at a depth of 300 m in siliciclastics of the Texas Gulf Coast. This article summarizes the main features of our method and describes the successful field test. We observed the related conversion from electromagnetic to seismic energy in a second experiment. The results suggest the feasibility of using electrokinetic coupling for near surface measurements of aquifers, including detection of pollutant migration. Because electrokinetic coupling is related to the rock permeability and to the properties of the pore fluids (such as electrical conductivity), electroseismic effects provide tools that are complementary to seismology for the characterization of the subsurface.

The principal uncertainty in electrokinetic exploration is the potential for detection of signals from greater depths. Our model calculations indicate that gas-water contacts and high permeability zones should produce particularly strong signals, and that it may be possible to see such structures at depths of more than 1000 m.

Most of our effort has been on the conversion of seismic to electromagnetic energy, a process we call *electroseismic prospecting* or ESP (US Patent 4904942). We briefly discuss the inverse process, *electro-osmotic surveying,* or EOS, and point out that the two processes are complementary: ESP is most sensitive to high permeability formations while EOS is dominated by low permeability formations.

Description of ESP. A conventional seismic source generates seismic waves (Figure 1). The active area is illustrated to be a gas-water contact but it could be any high permeability zone containing water or other polar liquid. At the gas-water contact, or some other contrast in acoustic impedance, a portion of the incident seismic wave is converted to the Biot slow wave. The Biot slow wave at seismic frequencies is a diffusive pressure wave in the pore fluid. The slow wave generates relative displacement between the pore fluid and the rock grains. (For a review of Biot theory, see "Sediment Acoustics" by R.E. Stoll in the series, Lecture Notes in Earth Sciences.) The relative motion between the pore fluid and the rock grains distorts the electric dipoles on the mineral surfaces. The distortion of the surface dipoles produces an electric field known as the streaming potential, which has the same time-dependent behavior as the incident seismic wave. The time-dependent streaming potential produces an electromagnetic wave that propagates to the surface of the earth where it can be detected with electric field antennas.

We have developed a forward modeling simulation of ESP to estimate the magnitude of the expected signals and to guide data acquisition and processing. The first step in the modeling is the seismic source, which is described with standard empirical equations. The conversion of P-wave energy to the relative velocity between the fluid and the rock matrix is calculated using Biot theory. Direct laboratory tests of this conversion in the seismic frequency range are not feasible because of sample size limitations, making this the most uncertain part of the signal estimation. Other researchers have

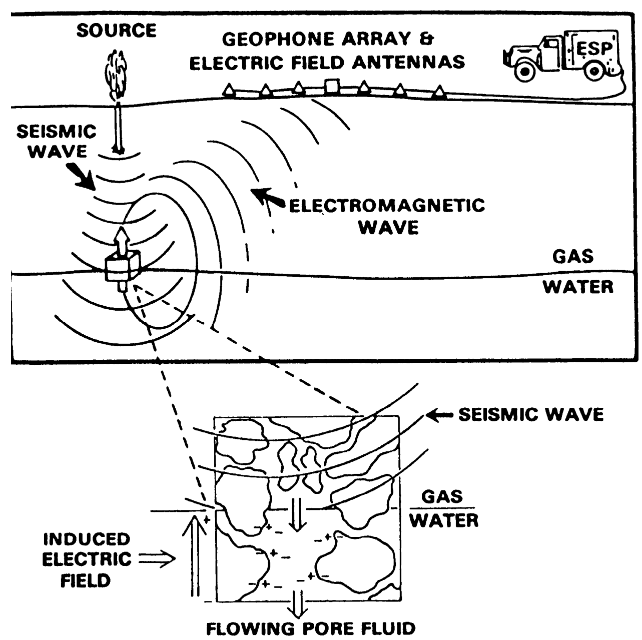

Figure 1. Schematic illustration of ESP.

recently been successful in detecting conversion from extensional and flexural modes in a bar to an electric field, as well as the inverse process of converting an applied electric field to extensional and flexural modes (private communication, Z. Zhu, MIT). Further work will be required to see if conversion from acoustic to electrical energy at interfaces can be observed in the laboratory.

We use laboratory measurements of the streaming potential to calculate the conversion between the relative fluid-rock motion and the induced electric field. This step is reliable, since the streaming potential has been measured for many rocks in the frequency range of seismic waves, and the variations in the streaming potential from one rock to another are small compared to the other uncertainties in the model. The propagation of the EM wave to the surface of the earth is well understood in the context of electromagnetic exploration *(Electromagnetic Methods in Applied Geophysics*, SEG, 1989). The detection of the signal involves the use of dipole antennas of the type used on controlled-source audiofrequency magnetotellurics (CSAMT). In summary, all of the components of the propagation and conversion are well understood in seismic and electromagnetic exploration except for the step of converting the P-wave energy into relative motion of the pore fluid and the rock matrix.

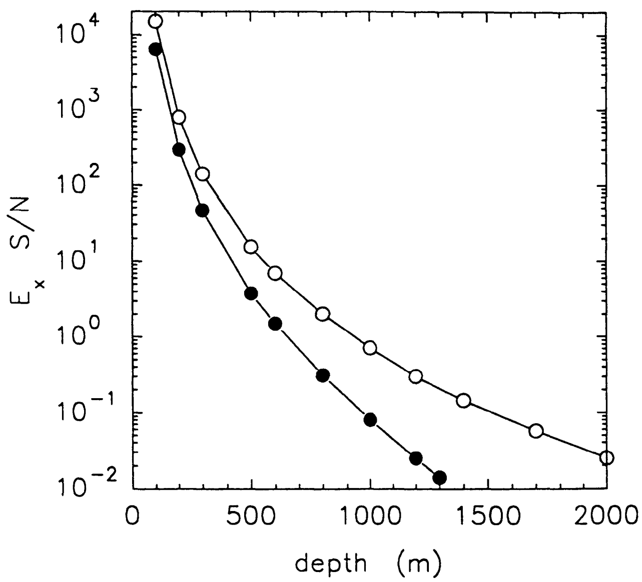

Figure 2. Horizontal electric field signal-to-noise ratio as a function of depth. Solid circles = 50 Hz; open circles = 10 Hz. The calculation is for a single shot point above a gas/water contact.

Components of the ESP amplitude. Biot theory suggests two different mechanisms for converting the seismic wave amplitude into relative fluid-rock motion: the relative fluid-rock motion that accompanies a seismic wave in a high-permeability formation, and mode-conversion from the seismic wave to a slow wave at an interface.

The relative fluid-rock motion induced by a seismic wave can be calculated using Biot theory (Dutta and Ode, GEOPHYSICS 1979). The fluid motion is described by a Darcy particle velocity, defined in analogy with Darcy's law for fluid flow in a porous medium. The Darcy velocity for a seismic wave is determined primarily by permeability and fluid viscosity, so the largest Darcy velocities are associated with gas-saturated, high-permeability formations.

Biot theory shows that a seismic wave is partially converted into the slow wave at an interface. This interface can be a change in acoustic impedance, or a change in pore fluid, porosity, or even permeability. Of the many possible types of interfaces, the largest amount of mode conversion into a slow wave occurs at a gas-water contact in a high-permeability formation. Once created, the slow wave is rapidly attenuated, with an attenuation length of 0.15 m in a typical brine-saturated Gulf Coast sand at 50 Hz. Within this distance, the slow wave generates a large relative fluid-rock motion (the Darcy particle velocity). The Darcy velocity from the slow wave is typically larger than that generated by passage of a P-wave through a high-permeability formation without slow wave mode conversion.

We have measured the streaming potential as a function of frequency on Berea sandstone to confirm the frequency dependence and amplitude of this component of the ESP response. We used two techniques, a constant pressure gradient and a constant flow velocity method. We find that the streaming potential is nearly independent of frequency in the range 1- 100 Hz. A typical streaming potential for a low-salinity brine in Berea sandstone, glass or quartz is 5 x 10^{-8} V/Pa.

To model the complete ESP response, we assume that a spherical seismic wave is incident on a horizontal layer where the Darcy particle velocity is converted by the &reaming potential coupling to an electric field. As the seismic pulse impinges on the horizontal layer, circular regions of positive and negative displacement move outward along the layer from a point directly beneath the seismic source. These circular regions are the Fresnel zones of the seismic wave. The first Fresnel zone is the portion of the horizontal layer reached by the seismic wave within one-half wavelength from the initial arrival. At later times, the seismic wave exhibits successive Fresnel zones. The EM radiation from each Fresnel zone can be represented by an electric multipole. The first Fresnel zone has dipole symmetry. The lowest order multipole of the combined first and second Fresnel zones is an electric octupole when the areas of the two Fresnel zones are the same. The electric field falls off with distance r from an octupole as $1/r^5$, compared to $1/r^3$ for a dipole, so the electric field from the higher-order Fresnel zones can be neglected compared to that from the first Fresnel zone.

In calculating the surface fields, we assume that the source of the EM radiation is a vertical electrical dipole corresponding to the first Fresnel zone. The dipole moment is proportional to electric field times the effective volume, which differs for the two Biot mechanisms. If the Darcy particle velocity is due to fluid-rock motion in a high-permeability formation, then the effective volume is the area of the first Fresnel zone times the formation thickness. If the Darcy particle velocity is due to mode conversion to a slow wave at a gas contact, then the effective volume is the area of the first Fresnel zone times the slow wave attenuation length.

We present the results of the ESP model calculations as a signal-to-noise ratio. We expect to achieve noise levels after data processing of 10^{-10} V/m for the electric field, and 10^{-12} T for the magnetic field, typical of modem EM surveys. Since the vertical component of the electric field is smaller than the horizontal and is much more difficult to measure, we will consider only the horizontal component.

We first calculate the expected ESP signal-to-noise ratio for a gas-water contact in a high-permeability (1 Darcy) sand formation typical of the shallow Gulf Coast. The contact is taken to be circular with a radius of 100 m, and the seismic source is a 2.3 kg dynamite charge. The dependence of the horizontal electric field S/N on the depth of the gas-water contact is shown in Figure 2 for seismic pulse center frequencies of 10 Hz and 50 Hz. The detector offset is assumed to be one half of the depth which is the position of maximum electric field. For 10 Hz, the S/N is greater than one for depths up to 920 m. At the higher frequency of 50 Hz, this depth is 620 m. The maximum magnetic field has S/N = 8×10^{-3}. The magnetic signal is too small for a useful ESP measurement.

In the absence of Biot slow waves generated at interfaces, an electric field is generated by the relative fluid-rock motion that accompanies the seismic P-wave. We calculated the radiation from a disk-shaped 1 Darcy sand formation 10 m thick and 100 m in radius surrounded by an impermeable shale. The ESP response has S/N > 1 for depths up to 615 m. There is a strong dependence of the ESP signal on the formation permeability. The ESP method will respond most strongly to high-permeability formations.

Principles of data processing. Six principles of data processing enhance our ability to increase the signal-to-noise ratio and to distinguish ESP events from background noise and from seismic arrivals at the antennas:

1. The ESP signal arrives at the antennas at virtually the same time, independent of offset
2. The signal arrives in approximately one half the time required for a seismic arrival
3. The signal originates from the first Fresnel zone centered directly below the shot point
4. The signal changes polarity on opposite sides of the shot point
5. Dipping beds can be characterized by the symmetry of the signal about the shot point
6. Electromagnetic attenuation that causes distortion of the pulse as it travels to distant antennas is just large enough to make resolution of the electromagnetic moveout difficult

Acquisition. We designed a field test to demonstrate the feasibility of using surface sources and antennas for measuring the ESP response from formations as deep as 300 m. The test was conducted at the Friendswood test site, a property about 48 km south of Houston, Texas and owned by Exxon Production Research Company. The survey was recorded over an existing research well that was 300 m in depth. The logs from this well give detailed geologic information. A 3 m thick gas-saturated sand at a depth of 234 m was of particular interest as a target zone for ESP.

Shot holes were spaced 12 m apart (Figure 3). Each shot hole was double-loaded (two 0.5 kg charges in each shot hole) at depths of 12 and 9 m. Seismic data were recorded with 12-geophone inline arrays, spaced 12 m apart along a line 6 m to one side of the shot holes. The horizontal electric field is recorded with 12 m long antennas. Each antenna consists of two stainless steel rod electrodes planted about 0.7 m deep in clay-rich soil. These electrodes are connected to battery-powered custom amplifiers located in the field. These units provide ground isolation, filtering, and amplification. The amplified output was connected through standard seismic cable to an 84-channel EG&G recording system. Each shot was recorded with a 2 s record and a sample rate of 0.5 ms.

The number of channels of ESP data recorded for a single shot was limited by the 28 available battery-powered custom amplifiers used to amplify the antenna signals. At the start of the survey, we used 26 antennas, each 12 m long and spaced 24 m apart. Shot records for the 26 antenna channels and 52 geophone channels were recorded using the 12 m deep charges. We then displaced each antenna by 12 m and recorded data using the 9 m deep charges. By shooting the line twice, we were able to record antenna data at 52 stations spaced 12 m apart using 26 preamplifiers.

Noise measurements were made at various locations. Our field noise measurements are consistent with the commonly reported environmental noise at seismic frequencies. The measured noise voltage scales with the antenna length. One consequence of this is that there will be an optimum length for an antenna. The noise level for a short antenna (length less than 1 m) may be limited by electrochemical electrode noise. Noise in a very long antenna (length greater than 100 m) may be dominated by environmental noise because the noise voltage increases in proportion to antenna length while the signal falls off away from the source. These ideas lead us to conclude that the optimum arrangement will have the largest possible number of relatively short antennas, to optimize dynamic range, and a large number of sources, or a repetitive source like a vibrator, to stack out the effect of noise spikes.

Interpretation. There are several problems in analyzing the ESP data: high-amplitude coherent noise spikes; high levels of cultural background noise; seismic effects; and low signal levels. The following processing procedures address these problems and enhance the signal-to-noise ratio. Wiener and bandpass filters were used to reduce high-frequency noise. To suppress coherent noise spikes, a noise-stacking and subtraction process combined with a sign flip on the "high" side

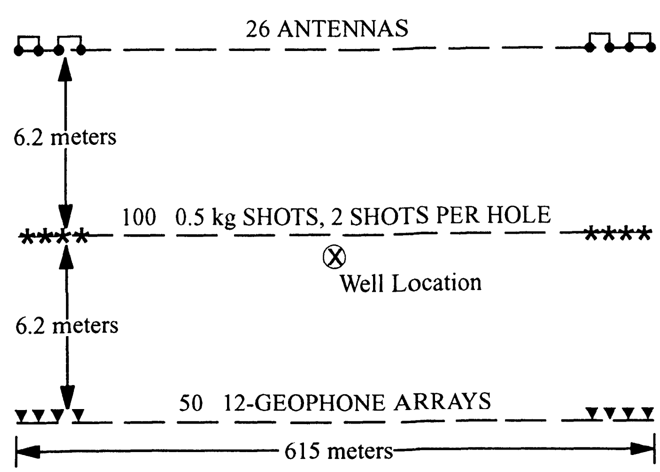

Figure 3. Layout of the ESP experiment at Friendswood.

of the shot, phase-difference filtering, and stacking were used. The seismic energy can be suppressed with a velocity filter. The signals were then enhanced through adaptive coherence filtering. Results for the ESP, seismic, and well log data are shown in Figure 4.

The processed ESP section shows a strong event at a depth of approximately 68 m. This event can be identified on a shot gather by the change in polarity on opposite sides of the shot point. "The peak antenna voltage for this event from a single shot is 700 nV, which gives a surface horizontal electric field of 60 nV/m. Our model calculation predicts a peak electric field from a single shot of 5 nV/m, which is 12 times smaller than the measured signal. The estimation of the electric field amplitude to within an order of magnitude of the field result is encouraging. The result suggests that our model of ESP signal generation includes the important physical processes. The factor of 12 discrepancy between the model calculation and the amplitude measured at Friendswood can be attributed to uncertainties in the model input parameters, and to experimental uncertainties in absolute voltage measurements.

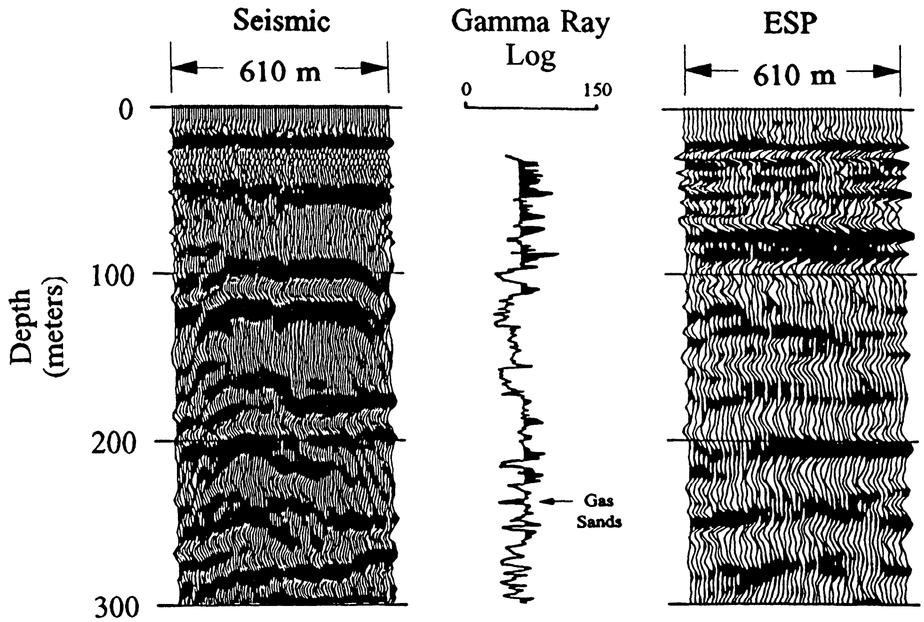

Figure 4. Final stacked depth sections plotted with the gamma ray log (shaded to highlight shaly intervals).

There are good ties between the seismic and ESP depth sections at approximately 25, 55, 205, and 246 m. Some events in the ESP data do not tie to a reflection in the seismic data (for example, at 86 m).

Well log data indicate the presence of a sequence of high permeability water sands and low permeability shales throughout the depth interval. In the first 123 m, these beds are thin, 1- 12 m in thickness. Antenna ESP data resolve several of these shallow features while seismic data do not. ESP data are well correlated with features in the gamma ray log. The gas sand, identified from density and neutron logs, is also shown on the gamma ray log. This gas sand was discovered when the well was drilled several years before the ESP test. Unfortunately, the gas from this small trap appears to have leaked off between the time the well was drilled and the time of the ESP test. The seismic reflection from 230 m is washed out at the center of the line at the high point of the structure compared to a stronger reflection in a previous seismic line recorded at the same location.

Several analysis procedures were run to verify that the continuous events on the stacked ESP sections are not a processing artifact. First, the seismic data were run through the same processing stream as the ESP data. The resulting stacked section showed no continuous events at short times. The antenna data were then reprocessed without accounting for the change in antenna polarity on opposite sides of the shot point. This procedure also resulted in no continuous events. The results of these two tests give confidence that the ESP events are not easily created by processing artifacts.

ESP events at depths of less than 275 m seem to be robust and well established through ties to the seismic data. We have studied weaker coherent events in the ESP data at greater depths by observing changes in signal amplitude when the processing conditions were changed. We conclude that real ESP sources are detected down to at least 450 m, which is approximately two electromagnetic skin depths. Detection of events from this depth is consistent with the theoretically estimated signal strength.

The high resolution in the ESP events at depths less than 125 m is consistent with the frequency content of the signal. The center frequency of the seismic data is between 30-40 Hz with little energy above 70 Hz. Coherent events in the ESP data were observed in the frequency band between 110-120 Hz. It should be noted that no pulse deconvolution has been applied to the ESP data. This is particularly important since the ESP data have been stretched in time by a factor of two compared to the seismic data in order to plot both as a depth section. (Recall that ESP events arrive after the one-way seismic time compared to seismic events arriving at the two-way traveltime.) Applying a pulse deconvolution to the ESP data should further improve the apparent depth resolution.

Seismic data have a higher signal-to-noise ratio than the ESP data (Figure 4). We attribute this, in part, to the difference in fold. There were 624 geophones and 26 antennas. There is ample room to improve the quality of ESP data.

In addition to the Friendswood test, we attempted larger scale tests in south Texas. These experiments were not successful because of instrumental problems. A distributed seismic acquisition system resulted in a higher than expected background noise level. Great care must be taken to electrically isolate antennas from one another and from the recording system, and to disable any geophone test circuits that can apply a voltage to the antennas. A second south Texas test using acquisition equipment with a smaller noise level was unable to detect an ESP response because the antennas were too long. A small number of long antennas gives incomplete noise cancellation (analogous to having low CDP fold) and a low signal-to-noise ratio for shallow ESP events. The optimal arrangement is a large number of short antennas.

EOS. In electro-osmotic surveying, or EOS (Figure 5), a source antenna on the surface generates a time-dependent electric field at depth. The source field creates an electro-osmotic pressure gradient which in turn transfers stress to the rock matrix at locations where there are discontinuities in the rock-fluid properties. The time-dependent stress on the matrix induces a seismic wave that is detected at the surface with geophones.

The generalization of Biot theory to include electrokinetic effects is described by Neev and Yeatts *(Physical Review B* 1989). This theory provides the fundamental equations for understanding both ESP and EOS. The theory of Neev and Yeatts is based on the Onsager relations, which provide a theoretical relationship between the streaming potential and the electroosmotic coefficient. We have verified this predicted relationship by measuring the streaming potential and electroosmotic coefficient at seismic frequencies on several sandstones. The results support the use of the Neev and Yeatts theory to describe the EOS process.

The theory predicts that the most efficient EOS coupling of an applied electric field to a compressional wave will occur where the permeability changes to a small value over a small distance. The strain amplitude created in the earth by the EOS coupling mechanism for reasonable experimental parameters is predicted to be larger than that generated from a 0.5 kg dynamite charge or a vibrator truck. The EOS signal generated by large permeability gradients should be easily detectable using existing seismic detection technology. Neev and Yeatts also conclude that there is no coupling between electrokinetic effects and a shear wave for either ESP or EOS.

The model calculations show that EOS (EM source with seismic detection) is not a true reciprocal to ESP (seismic source with EM detection). The largest EOS signals will be produced in low permeability rock compared to the largest ESP signals from high permeability rock.

We conducted field tests of EOS at the same location as the ESP test. The source antenna was two electrodes of aluminum foil 3 x 5 m in size buried 0.5 m below the surface. Electrode separation was 300 m. Pulsed or swept frequency signals were applied to the antenna with a 20 kW audio power amplifier. Typical antenna currents were 150 A. The direct electromagnetic pick-up on geophones made the use of swept frequency methods impossible. We used hydrophones down a cased hole located between the two source electrodes as detectors. The casing shielded the hydrophones from electromagnetic pick-up (a suggestion from T.R. Madden), reducing the pick-up by about 100 dB. With this arrangement, we were successful in detecting seismic energy generated by a pulsed electric field. We have not yet collected sufficient data or developed the theory needed to generate an EOS image of the subsurface.

Applications. Our experiments have produced data from depths on the order of 300 m. The electromagnetic skin depth will set the scale for the useful depth of exploration. At a depth of five skin depths, the signal is attenuated by approximately a factor of 100. We take five times the skin depth as a criterion for the ultimately achievable depth of investigation.

In the Gulf Coast, the skin depth at seismic frequencies is on the order of 300 m. The potential useful depth of investigation is then near 1.5 km. In more typical sedimentary basins, the electrical conductivity is a factor of 10 lower than in the Gulf Coast. The lower conductivity gives a skin depth near 1 km and a useful depth of investigation of 5 km. Greater depth penetration is possible at lower frequencies.

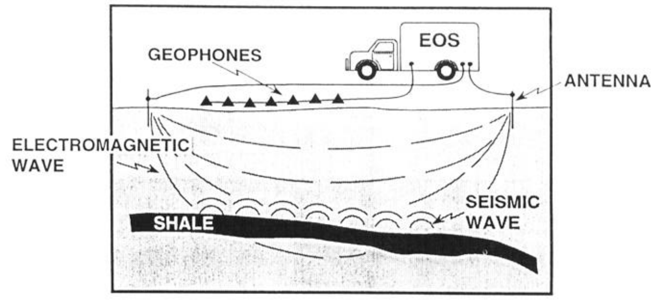

Figure 5. Schematic illustration of EOS.

ESP's physical principles can be incorporated into a logging-tool configuration, but the electromagnetic fields are probably too small to be measured. Our experiments with downhole sources and surface antennas showed that ESP events arc easily detected in this reverse-VSP type configuration.

Offshore surveys of ESP and EOS may be both practical and, in some cases, more desirable than land surveys. The conducting sea water shields antennas from cultural and atmospheric noise. The electromagnetic skin depth in sea water at seismic frequencies is greater than 30 m. This means that electrodes placed within a few meters of the sea floor will detect electric potentials essentially equal to those. on the sea floor.

Some of the earliest proposed uses of electrokinetic effects were as electromagnetic geophones that would be less sensitive to mechanical coupling to the surface weathered layer. This application remains attractive.

In both ESP and EOS, one half of the round trip to the horizon of interest is by seismic propagation while the other half is by electromagnetic propagation. Since the frequency-dependent attenuations of the seismic and electromagnetic energy are different and the EM attenuation depends sensitively on rock conductivity, the depth-dependent frequency content of the signal may be useful in estimating conductivity as a function of depth. The seismic traveltime determines the depth of the measurement and the spatial resolution.

Conclusions. We have demonstrated than an electroseismic ESP signal can be detected from a depth of 300 m. From our experience, we suggest that the greatest potential for succcssful application of electrokinetic effects is in shallow exploration. This could involve exploration for aquifers, cultural artifacts, or for pollution monitoring. Lessons learned from shallow exploration will be directly applicable to future efforts to detect electrokinetic conversion at greater depths. LE

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