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Seismic reflections from depths of less than two meters

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Abstract. Three distinct seismic reflections were obtained from within the upper 2.1 m of flood-plain alluvium in the Arkansas River valley near Great Bend, Kansas. Reflections were observed at depths of 0.63, 1.46, and 2.10 m and confirmed by finite-difference wave-equation modeling. The wavefield was densely sampled by placing geophones at 5-cm intervals, and near-source nonelastic deformation was minimized by using a very small seismic impulse source. For the reflections to be visible within this shallow range, low seismic *P*-wave velocities (<300 m/s) and high dominant-frequency content of the data (~450 Hz) were essential. The practical implementation of high-resolution seismic imaging at these depths has the potential to complement ground-penetrating radar (GPR), chiefly in areas where materials exhibiting high electrical conductivity, such as clays, prevent the effective use of GPR. Potential applications of these results exist in hydrogeology and environmental, Quaternary, and neotectonic geology.

Introduction

In oil exploration, seismic reflection has been the geophysical method of choice for more than 60 years. However, until about 15 years ago, the reflection method was not used routinely to extract geologic and hydrologic information from the earth's subsurface at depths of less than 30 m. Due to interference caused by the nearly simultaneous arrival of several wave types near the seismic source, reports of successful imaging at depths of less than 10 m have been limited to a very small number of refereed publications, including papers by Birkelo et al. (1987), Miller et al. (1989), Bachrach et al. (1998). and Steeples (1998). To date only Birkelo, Bachrach, and Steeples (1998) have confirmed observable seismic reflections originating less than 3 m below the surface.

We report here developments in near-surface seismic-reflection surveying that have led to the detection and evaluation of seismic reflections at depths of less than 2 m. We obtained three distinct seismic reflections from the upper 2.1 m of flood-plain alluvium in the Arkansas River valley, an area in which only a single reflection was observed from the water table at a depth of 2.6 m with a two-way traveltime of 22 ms during shallow seismic surveys undertaken in 1986 (Birkelo et al., 1987). Based on these recent findings, we offer evidence that ultrashallow seismic-reflection data can be obtained with levels of vertical and horizontal resolution potentials usually associated only with GPR surveys.

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Geologic Setting

The Arkansas River alluvial valley near Great Bend, Kansas, served as the test site for our experiments. Near-surface material there consists of unconsolidated, medium- to coarse-grained sand and medium to coarse gravel interspersed with thin, discontinuous paleosols and clay stringers associated with the Arkansas River. The near-surface stratigraphy varies rapidly on the scale of meters to tens of meters. A hand-augured test hole 5 m from the seismic line revealed the presence of sand to a depth of about 1.5 m, where a hard layer was encountered but not sampled. This layer may be composed of partially calcified paleosol, coarse gravel, or clay. During our seismic survey, the water table was located 2.1 m below the surface, based on a measurement taken at a test well located 25 m from the seismic line. A well drilled about 40 m from the seismic line struck bedrock, a fine- to medium-grained Cretaceous-age sandstone, at 29 m (Sophocleous et al., 1987).

Seismic Data and Interpretation

Improving our capacity to measure the near-source wavefield and using a very small impulsive source were instrumental in obtaining the ultrashallow seismic reflections reported here. We collected data using single, Mark Products L-40A 100-Hz geophone receivers with a receiver-group interval of 5 cm and a spike length of 14.5 cm. Typically, "shallow" seismic surveys have receiver-group intervals of 1 m or more. Because we increased receiver coverage by at least an order of magnitude, our ability to delineate and improve the coherence of ultrashallow reflections at the expense of other interfering phases was enhanced. We used a 96-channel Bison Model 24096 seismograph with 24-bit A/D conversion. Data were recorded with a 1/4-ms sample interval, a 4-Hz pre-A/D low-cut filter, and an antialias filter down 60 dB at 2000 Hz.

The seismic-energy source was a single shot from a .22-caliber rifle using subsonic solid-point short ammunition. More powerful shallow seismic exploration sources such as seisguns, hammers, and small explosives—and, indeed, the .22-caliber rifle when using long rifle ammunition—generate sufficient near-field nonlinear deformation to preclude the detection of ultrashallow reflections. The tip of the rifle barrel was placed in a prepunched hole 2 cm in diameter and 12 cm deep. The rifle was fired at a depth of 10 cm, providing a 2-cm expansion zone at the tip of the barrel. Each subsequent shot was fired 10 cm farther along the seismic line.

Three adjacent common-shotpoint gathers with shotpoints separated horizontally by 10 cm constitute Fig. 1. We know from uphole traveltime measurements made concurrently with the seismic field experiment that the reflection from the water table should arrive at the geophones with a two-way traveltime

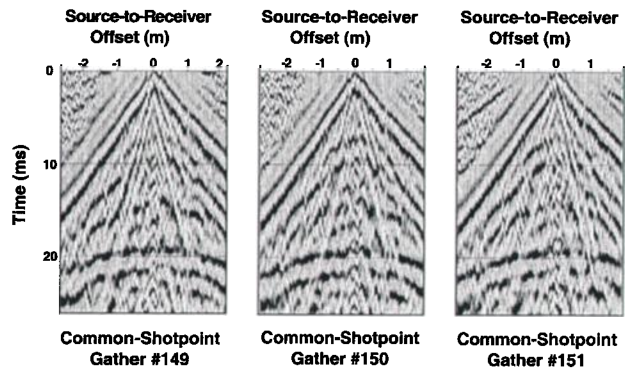

Figure 1. Three common-shotpoint gathers from test site with shotpoint separation of 10 cm and 5-cm geophone interval. Digital scaling (4-ms AGC window) and band-pass filtering (500-900 Hz, with 12 dB/octave slopes) were applied.

of 20 ms ± 1 ms. Hence, we interpret the prominent reflection at 19 ms on the common-shotpoint gathers as having originated at the water table. Birkelo et al. (1987) found a reflection event at a similar location in space and time which was also interpreted as the water table, and this reflection moved downward in time in response to the pumping of a well a few meters away. Water-table measurements made in 23 test wells within 200 m of the seismic line by Sophocleous et al. (1987) suggest that the water table is very nearly flat on the scale of tens of meters. The Arkansas River, with a gradient of less than 1.5 m/km, flows freely less than 100 m from the seismic line, and the river level is very close to the water table both at the test site and in the surrounding area (Sophocleous et al., 1987).

In addition to the prominent reflection event visible at 19 ms, two shallower reflections (at 8 ms and 15 ms) can be seen on each of the three common-shotpoint gathers in Fig. 1. Our interpretation is that these reflections were generated by alluvial gravel deposits interfingered with alluvial sand deposits. The peak-to-peak period of all three reflections is slightly more than 2 ms, indicating that the dominant frequency of the reflection information is ~ 450 Hz. Thus, using the commonly applied quarter-wavelength criterion (Widess, 1973) and *P*-wave velocities of 180 m/s to 255 m/s, the data should have a vertical-resolution potential of 10-15 cm.

Analyzing the seismograms yielded constraining information about the physical properties and geometry of the subsurface. The *P*-wave velocity in the surface layer was confirmed by using both the direct *P*-wave, whose measured velocity was 180 m/s, and the velocity determined by the hyperbolic moveout of the reflection from the base of the surface layer, also 180 m/s. Given the *P*-wave velocity of the surficial material and the zero-offset two-way traveltime of the shallowest reflection, we determined that the interface generating the reflection is about 63 cm deep, thus fixing the thickness of the surface layer. Using Dix's formula (Dix, 1955) to obtain interval velocity from the hyperbolic-moveout velocity of the second reflection, along with the zero-offset two-way traveltime of the reflection and the surface-layer information, we determined that the thickness of the second layer is about 83 cm and its *P*-wave velocity is about 255 m/s. We used the preceding information about the water table as well as the depth of the water table (2.1 m) to determine that the thickness and *P*-wave velocity of the third layer is about 64 cm and 205 m/s, respectively. The probable error limits of these observations are discussed subsequently in conjunction with the modeling results.

Finite-Difference Modeling

Our interpretation of the subsurface (Fig. 2A) was independently constrained by synthetic seismograms generated for direct comparison to the field data. The synthetic seismograms in Figs. 2B and 2C were produced using a stress-velocity, finite-difference elastic wave-equation model that was fourth-order in space and second-order in time. A free-surface boundary condition was implemented at the top of the model, and absorbing boundaries were used at the bottom and sides. Grid dispersion was minimized by maintaining at least 8.8 grid points per wavelength in the slowest velocity region (V_S = 60 m/s). The courant number was 0.45 using the fastest *P*-wave velocity of the model. The model was density normalized (i. e., $\rho = 1$); acoustic-impedance changes are represented by the velocity profile in Fig. 2A. Other important model parameters were time-step (2.5 μs), vertical and horizontal node spacing (both 1 cm), number of nodes in the horizontal direction (2400), and number of nodes in the vertical direction (1700). The model required about 47 hours to complete 16,000 iterations (40 ms) using a Sun Sparc Ultra 30 station with 192 MB RAM. Modeling details and the source code of the model are currently available under GNU General Public License (GPL) at http://www.geo.ukans.edu/~brian/fdmod.html.

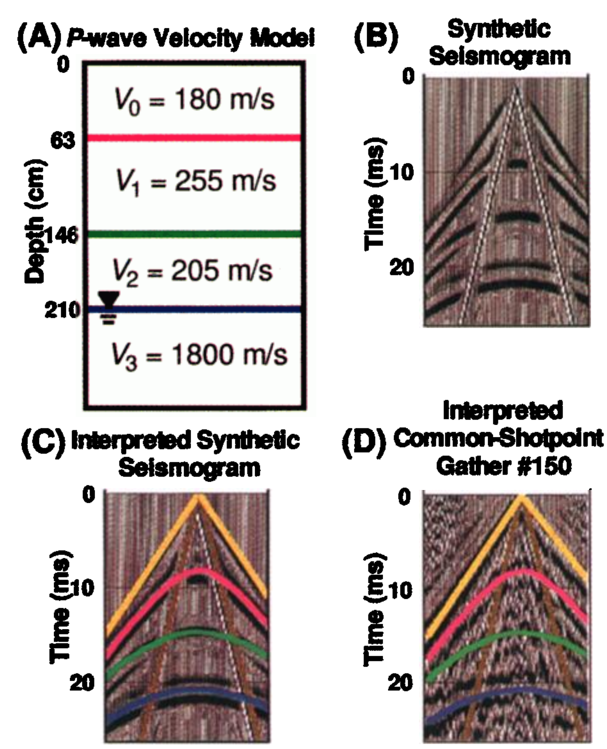

Figure 2. (**A**) *P*-wave velocity model used to generate finite-difference synthetic seismogram in (**B**). Interpreted synthetic seismogram (**C**) shows reflections from the three layer boundaries in the velocity model (red, green, blue), the direct *P*-wave (orange), and the Rayleigh wave (brown). Interpretation of common-shotpoint gather #150 (**D**) was used as input to the forward model to create (**A**), (**B**), and (**C**).

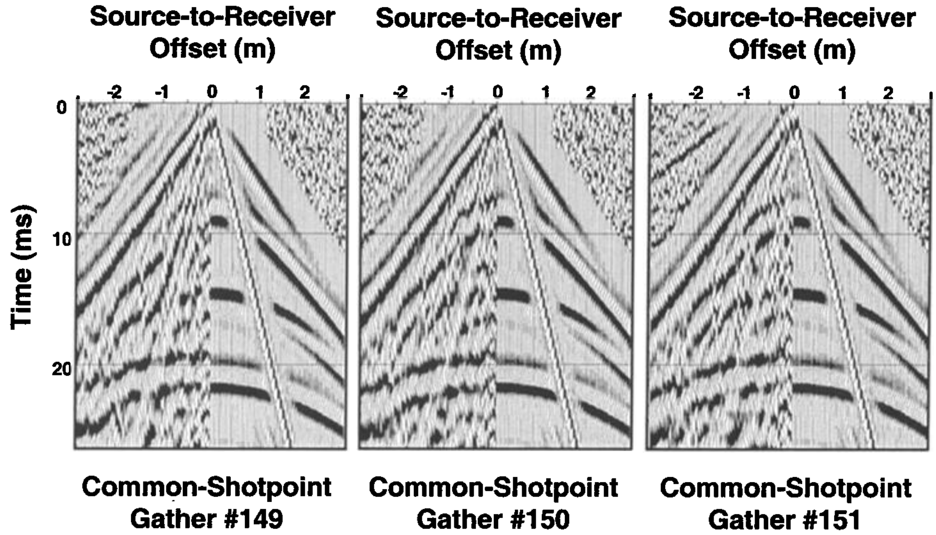

Figure 3. Spliced side-by-side comparisons of the three common-shotpoint gathers in Fig. 1 with the synthetic seismogram in Fig. 2B. The left portions of the panels are real data; the right portions are synthetic seismograms.

The shooting geometry for the model was arranged to mimic common-shotpoint gather #150 (Fig. 2D). Observations were made at every fifth node of the model, where nodes were 1 cm apart; thus, observations were made at 96 nodes, with 5-cm spacing. The energy source was a 450-Hz Ricker-wavelet located 10 cm below the surface of the model.

Our initial calculations of subsurface geometry and *P*-wave velocities from the field data served as a starting point for the forward modeling; we then systematically varied layer thicknesses and velocities to obtain the best visual fit to the real data. Input parameters that remained fixed throughout the process were surface-layer velocity and thickness and the two-way traveltime to the water-table interface. Below the water table, a *P*-wave refraction velocity of 1800 m/s (calculated from a field pseudowalkaway) was used as input to the model.

When compared with the real seismograms, the synthetic seismograms showed noticeable discrepancies when the final model boundaries (Fig. 2A) were moved more than 3 cm or the *P*-wave velocities were altered by more than 5 m/s. When viewed side-by-side (Fig. 3), a close correlation exists between the synthetic seismograms and the real data, supporting our final interpretation. Hence, to the extent that the near-surface layers are isotropic seismic media, our estimates of layer thicknesses are within about 3 cm of being correct.

Discussion

High-quality GPR walkaway data have been obtained along the same profile as the seismic data (Baker et al., 1998). Using the measured velocity to the water table of 0.12 m/ns and assuming the use of 225 to 450 MHz antennae, the quarter-wavelength approximation (Widess, 1973) for vertical resolution yields a vertical resolution potential of 13 to 7 cm, respectively. Hence, the seismic data have a resolution potential comparable to that expected from GPR data.

Three characteristics of the Great Bend site allow imaging shallower than 2 meters: First, the site contains favorable impedance contrasts within the upper 2 m of the subsurface. Subsurface layers must have differences in their geoacoustical properties (i.e., seismic velocity and density) large enough to ensure that the energy reflected at the boundaries will be sufficient to be detected at the surface. Second, the seismic velocities of the material within the upper 2 m at the Great Bend site are very low (<300 m/s). Third, the frequency content of the seismic data is greater than 400 Hz.

At any geographic location, the importance of seismic velocity and the dominant-frequency content of the data to the detection of shallow reflections can be observed by invoking the

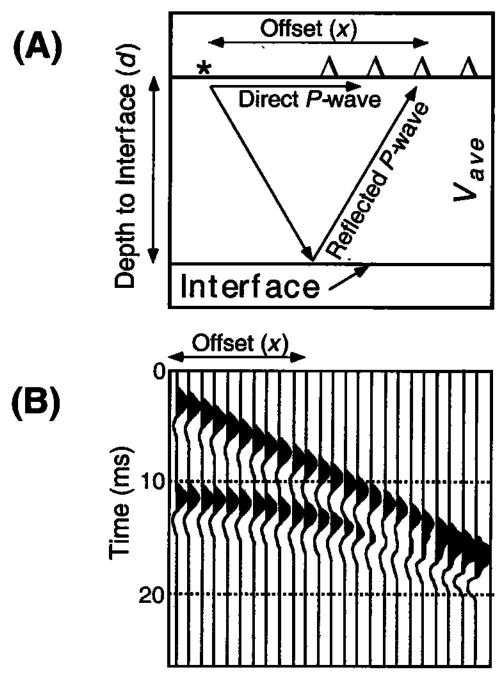

Figure 4. (A) There is a source-to-receiver offset (x) at which energy in the form of single-cycle sinusoidal source pulses from the direct *P*-wave and the reflected *P*-wave remain undistorted for a given frequency (f), a specific depth to an interface (d), and the average velocity to the interface (V_{ave}). (B) Note that for source-to-receiver offsets greater than x the reflected energy is distorted by interference and that this distance is frequency dependent.

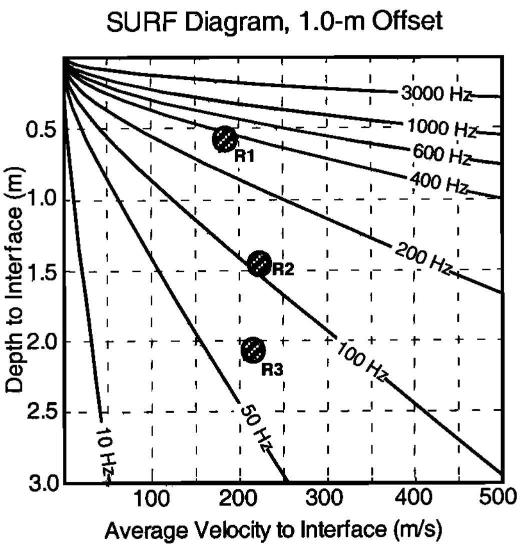

Figure 5. SURF diagram of the three reflecting interfaces (shallow to deep: R1, R2, R3) at the Great Bend site, with a source-to-receiver offset of 1.0 m. The interfaces are plotted using the observed average velocity to the interface. Using the diagram, we see that for all three interfaces the *minimum* data frequency necessary to prevent a reflecting *P*-wave from interfering with the direct *P*-wave is about 400 Hz. This supports the conclusion that the shallow reflections were visible because of low *P*-wave velocities within the material (<300 m/s) and high (~450 Hz) frequency content of the data.

relationships among velocity, frequency, and depth in the arriving traveltimes of seismic energy (Fig. 4). Assuming that the source wavelet is a simple, single-cycle sinusoid, a specific minimum critical frequency (f) exists above which no interference between direct *P*-wave energy and reflected *P*-wave energy occurs, given any arbitrary source-to-receiver offset (x), depth to an interface (d), and average *P*-wave velocity above the interface (V_{ave}):

$$f = \frac{V_{ave}}{(x^2 + 4d^2)^{1/2} - x} .$$

By rearranging this equation, one can also determine the shallowest undistorted reflection obtainable when the velocity of the material and the frequency content are known. Using the equation, we generated a family of figures that can be used to determine Shallow Underground Reflection Feasibility (SURF). Given the average *P*-wave velocity and depth to each of the three reflecting interfaces in the Great Bend data, the frequency content necessary to see reflections without distortion from the direct *P*-wave is about 400 Hz (Fig. 5). On the SURF diagram note that because the frequency content of the seismic field data is 450 Hz, undistorted reflections from depths shallower than 0.5 m cannot be obtained at offsets of 1 m or more, even when an interface with a high impedance contrast exists at that depth. Also, if the average velocity to the interface at 63 cm were above 300 m/s, an undistorted reflection would not be recorded at offsets of 1 m or more. Thus the SURF diagram shows the conditions under which undistorted ultra-shallow reflections can be observed.

Conclusions

Interpreting the subsurface by comparing real and synthetic seismograms yielded reflecting boundaries at depths of 0.63, 1.46, and 2.1 m, with seismic *P*-wave velocities in each layer of 180, 255, and 205 m/s. These findings show that seismic-reflection techniques can be used to image the upper 2 m of the earth's subsurface with a vertical resolution potential of the order of 10-15 cm, where geoacoustical conditions are favorable, which is comparable to the typical resolution potential of GPR. Factors critical to obtaining seismic-reflection information from depths shallower than 2 m are low seismic *P*-wave velocities (< 300 m/s) and high data frequency content (> 400 Hz). Such imaging capability could have a variety of applications to geology, hydrology, mining, engineering, and environmental site-characterization problems.

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