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ANALYSIS AND EVALUATION OF INTERWELL SEISMIC LOGGING
TECHNIQUES FOR HYDROCARBON RESERVOIR CHARACTERIZATION

Final Report

By
Jorge O. Parra
Brian J. Zook
Vernon R. Sturdivant

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Chandra Nautiyal, Project Manager
Bartlesville Project Office
P.O. Box 1398
Bartlesville, OK 74005

Prepared by
Southwest Research Institute
6220 Culebra Road
P.O. Box 28510
San Antonio, TX 78228-0510

MASTER

FOREWORD AND ACKNOWLEDGEMENTS

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INTRODUCTION AND SUMMARY OF PROJECT

Background

The oil and gas industry is presently emphasizing the production and recovery of established reserves rather than exploration of new fields. This has brought into focus the need for a better understanding of reservoir rock properties and geologic structures. Of particular interest are factors that directly affect the distribution of the reservoir fluids, namely the permeability of the rock and the presence and locations of fractures and impermeable boundaries that channel or constrain the movement of fluids. A new conceptual extension of sonic logging is offered in this project for specific application in heterogeneous reservoirs. This concept is one of interwell seismic logging which is made possible by the prevalence of the relatively large number of wells located in the reservoir. In this application, high resolution seismic transmission measurements between a pair of boreholes can provide information equivalent to a seismic section which uniquely is oriented in the plane of the boreholes. These interwell seismic data is analyzed to determine various petrophysical properties of the reservoir under investigation.

A thorough study of high-resolution interwell seismic measurements techniques and data analysis at the BP Exploration Devine test site was a topic of research for the first year of the present project (Parra *et al.*, 1991). Experiments were conducted to evaluate an experimental interwell seismic system and analyzed high-resolution seismic data. The data were acquired to examine full waveforms recorded in the pulsed mode and in the swept mode, as well as to perform traveltimes inversion tomography for mapping lithology changes and geological structure between wells. The signal enhancement of the sweep seismograms was obtained by crosscorrelating sweep data with bandpass-filtered and tapered pilot signals. The cross-correlated sweep seismogram exhibits coherent seismic events and compressional wave moveout consistent with the corresponding 2-cycle and 5-cycle seismic sections. The analysis of 1-D traveltimes inversion tomography data together with sonic logs inferred the presence of anisotropic conditions in Del Rio clay formation. The 2-D traveltimes inversion results demonstrated that when anisotropic conditions in the formation exist, distortions in the tomograms produced by assuming isotropy are reduced by allowing the medium to be transversely isotropic. The horizontal velocity tomogram represents a more realistic imaging of the layered structure indicated by well logs. In particular, the uniform velocity distribution corresponding to the Del Rio clay formations connects the wells with similar signatures and documents a more homogeneous layer structure.

In addition, in the first year a forward modeling solution was generalized to be representative of two-phase media so as to include the effect of permeability and porosity on seismic wave propagation through heterogeneous fluid-filled formations (Parra, 1991). Thus, in addition to delineating the structure of reservoirs, interwell full waveform logs, interpreted by model analysis, can be used to estimate rock physical properties of direct interest to hydrocarbon recovery and to track injected fluids during secondary and tertiary extractions.

In the second year of the present research project, interwell seismic and reverse VSP seismic experiments were conducted at the Buckhorn test site in Illinois, to analyze high-

resolution seismic waveforms and to perform traveltime inversion tomography for mapping the Silurian Kankakee limestone formation which is the more prolific oil producer in the Mt. Sterling area. Preliminary results of reverse VSP and interwell seismic data and seismic modeling (Parra et al., 1993; Parra, 1993) inferred that those techniques can delineate the hydrocarbon reservoir and geological structures at the Buckhorn test site.

The topics of research proposed for the third year of this project involves a full-scale field test of interwell seismic logging techniques, model studies, and data interpretation as well as data processing methods designed to yield information on the reservoir continuity and to determine *in situ* rock physical parameters. In particular, delineate reservoir architecture using seismic guided waves and determine whether shales within a sand reservoir are continuous, and as a consequence forming permeability barriers. To facilitate the field experiments we selected the Gypsy test site in Oklahoma, where BP Exploration drilled six wells of 1,310 ft. deep over a water saturated sand to related sedimentary formations to reservoir flow units and to provide a data base for assessing reservoir characterization technology.

Summary of Project Efforts

The objective of this project is to investigate interwell seismic logging techniques for indirectly interpreting oil and gas reservoir geology and rock physical properties. The work involves a balanced study of theoretical and numerical modeling of seismic waves transmitted between pairs of wells combined with experimental data acquisition and processing at controlled field conditions. The field applications of this reservoir probing concept are aimed at demonstrating high resolution measurements and detailed interpretation of heterogeneous hydrocarbon-bearing formations.

The first part of this third year project efforts was devoted to thoroughly evaluating interwell seismic logging and reverse VSP in a hydrocarbon-bearing formation at the Buckhorn test site in Illinois. Specifically, the data from the experiments conducted in the second year of this project were analyzed to delineate geological structures and to extract rock physical parameters. In particular, to determine attenuation and anisotropy from seismic waveforms to characterize the Silurian Kankakee limestone formation which is a prolific oil producer in the area. The data analysis and interpretation infers the presence of a low quality factor heterogeneity which correlate accurately with a low-velocity and high-porosity zone delineated with traveltime tomography and well log information. These results indicate that the high attenuation affecting the seismic pulses is caused by the heterogeneous conditions of the porous and fractured partially saturated rock matrix. In addition, the agreement between synthetic and field data confirms the interpretation of well logs and seismic measurements recorded in the hydrocarbon-bearing formation at the Buckhorn test site. These results have suggested that reverse VSP (using several 3-component detectors in shallow boreholes) and interwell seismic data integrated with logs and seismic modeling can delineate the hydrocarbon reservoir and geological structures at the test site. These results are described in the first chapter of this report. The following chapter is devoted to numerical modeling to investigate the wave field radiated by a seismic source in an anisotropic layered medium. The results of this work suggest that the three-component particle motion associated with a dipping fracture zone is sensitive to the source polarization relative to the plane containing the

source and detector borehole. In addition, the agreement between synthetic and field data validates the layered earth transversely isotropic solution and confirms the interpretation of the well logs and crosswell seismic measurements recorded in the hydrocarbon-bearing formation at the Buckhorn test site in Illinois.

The second part of this project is devoted to the evaluation of continuity logging techniques for hydrocarbon reservoir continuity. Specifically, this part of the project includes the evaluation of methods of measurements, modeling and data processing to delineate the reservoir architecture and relate dispersion and attenuation measurements to rock physical properties. The third chapter includes the theoretical analysis of acoustic wave propagation in fluid-filled porous media (based on the theory of poroelasticity which was developed in the first year of this project) to calculate dispersion and attenuation of guided waves trapped in low-velocity layered media. The sensitivity of phase and group velocity as well as attenuation to input parameters (e.g., permeability, porosity, layer thickness, etc.) were analyzed to characterize guide modes produced by acoustic waves propagating in fully saturated layer media. The results of this work indicated that the acoustic guided wave attenuation is sensitivity to the *in situ* permeability. In particular, the attenuation changes significantly as the *in situ* permeability of the low-velocity layer is varied at the frequency corresponding to the minimum group velocity (Airy phase). Alternatively, the attenuation of the wave modes are practically unaffected by those permeability variations in the layer at the frequency corresponding to the maximum group velocity.

The next chapter deals with the analysis of wave propagation in layered fluid-filled porous media based on Biot theory to calculate synthetic seismograms containing guided waves trapped in low-velocity poroviscoelastic regions. The numerical simulations were conducted to demonstrate seismic guided wave propagation characteristics in poroviscoelastic media for planning continuity logging surveys. The results suggested that the maximum amplitude normal modes can be detected at the interfaces with impermeable media. Also the results demonstrated that the Airy phase wave groups of the normal modes carry useful information on the fluid-flow characteristics of the formation such as permeability.

The last three chapters include a study of rock physical properties and interwell seismic data to evaluate interwell seismic logging techniques for hydrocarbon reservoir continuity. In particular, to evaluate methods of measurements and processing to delineate the reservoir architecture and to relate dispersion and attenuation with rock physical properties. The architecture includes structure of flow units, sealing surfaces and boundary surfaces that result from deposition and diagenesis. Several of these surfaces may have elastic properties that will support the propagation of guided waves, which may contain valuable information related to the continuity and rock physical properties of thin layers. These layers may serve either as conduits or barriers to hydrocarbon flow. The principal application of this study was to identify and analyze the propagation characteristics of guided waves in the Gypsy-fluvial sandstone formation at the Gypsy site in Oklahoma. The correlation of the lithology with well logs and interwell seismic data indicated the origin of the waveforms associated with continuous and discontinuous geological units. The continuity of low-velocity clean sands in the Gypsy sandstone interval using guided waves was confirmed and it was consistent with previous results of crosswell and pulse test data. Leaky mode dispersion characteristics were

associated with thin continuous mudstones and sands within the Gypsy interval. In addition, the attenuation and dispersion of direct P-waves associated with the water-saturated Gypsy sandstone formation was related to permeability using the unified Biot and squirt-flow mechanism. As a consequence, realistic seismic responses were generated using Gypsy reservoir parameters which were determined from logs, core information and interwell seismic data. The attenuation and dispersion results from real and synthetic waveforms suggested that the loss of energy of the seismic waves propagating within the low-velocity clean sands in the Gypsy sandstone formation are caused by the permeable conditions of the fully saturated sandstone unit.

CHAPTER 1

A FEASIBILITY STUDY OF HIGH-RESOLUTION REVERSE VSP AND INTERWELL SEISMIC METHODS FOR HYDROCARBON RESERVOIR CHARACTERIZATION

Reverse VSP and interwell seismic measurements together with log data have yielded information on the anisotropic characteristic of the shale and heterogeneities associated with the lateral changes in porosity within the Silurian Kankakee limestone formation in the Mt. Sterling area. High-resolution seismic experiments were conducted to delineate geological structures and obtain a detailed knowledge of the rock physical properties between wells at the Western Kentucky Petroleum Buckhorn test site near Quincy, Illinois. Specifically, experiments were conducted to analyze seismic data for mapping geological structures and to extract rock physical parameters from seismic waveforms to characterize the Kankakee which is a prolific oil producer in the area. A significant reduction of amplitude for waves traveling in the limestone formation infers the presence of a heterogeneity which correlates accurately with a low-velocity and high-porosity zone delineated with traveltimes tomography and well log information. These results indicate that the high attenuation affecting the seismic pulses is caused by the heterogeneous conditions of the porous and fractured partially saturated rock matrix. In addition, the agreement between synthetic and field data confirms the interpretation of well logs and seismic measurements recorded in the hydrocarbon-bearing formation at the Buckhorn test site. This suggests that reverse VSP (using several 3-component detectors in shallow boreholes) and interwell seismic data integrated with logs and seismic modeling can delineate the hydrocarbon reservoir and geological structures at the test site.

The main goal of high-resolution subsurface seismic measurements together with well log and core information is to delineate geological structures and obtain a detailed knowledge of the rock physical properties between wells as part of a well planned reservoir characterization program. This chapter demonstrates the feasibility of using reverse VSP and interwell seismic techniques and well logs to characterize structure and rock physical parameters in a hydrocarbon-bearing formation at the Buckhorn test site in Illinois. The reverse VSP method is used in the experiments to detect and map the Silurian Kankakee limestone formation which is less than 8 m thick and is horizontally distributed at a depth of about 200 m. The concepts and applications of RVSP and cross-well seismic measurements are given in Hardage (1992).

The Kankakee formation is formed by porous zones partially saturated with oil and water. The well logs and core information suggest significant lateral changes in porosity within the Kankakee which is a prolific oil producer in the Mt. Sterling area in Illinois. The interwell seismic method was applied to determine other rock physical properties between wells such as rock quality factors (Jones, 1986) which may be associated with the pore-fluid viscosity and permeability of the porous zones of the Kankakee limestone formation (Parra and Xu, 1994). In fact, a porous zone within the Kankakee was delineated as a low-velocity anomaly using travel time tomography at the Buckhorn test site by Saito (1991). In addition, the same oil-bearing porous zone was delineated and characterized as a conductive target

within a more resistive limestone using resistivity tomography by Shima and Imamura (1992).

First the experiments conducted at the Buckhorn test site are described, followed by discussions in the processing of reverse VSP and interwell seismic data to predict the changes in subsurface velocity using a layered earth inversion approach. Next, zero vertical offset full waveforms are analyzed using time-frequency representation to extract direct P waves from which phase velocity and attenuation curves are determined by the spectral ratio method (Ganley and Kanasewich, 1980). Finally, the rock physical parameters determined from reverse VSP data, well logs, and interwell seismic data are used to produce a synthetic seismogram which ties very well with the corresponding observed field seismic data.

REVERSE VSP AND INTERWELL SEISMIC EXPERIMENTS

The Experiments and Well Layout

Reverse VSP and interwell seismic measurements were conducted at the Buckhorn test site in Illinois, where five 270 m deep open boreholes were drilled in an oil-producing field for the purpose of various borehole geophysical experiments, using seismic and electrical methods. According to the exploration and production histories of the field, the oil reservoir is the porous zone of Silurian Kankakee limestone formation which is horizontally distributed at the depth of about 200 m, and has less than 8 m of thickness. In fact, two out of the five boreholes drilled produce oil from the Kankakee formation. The results of several logging measurements conducted just after drilling the boreholes indicated a good correlation of the formation from well-to-well (Saito, 1991). However, the logs also suggest significant lateral changes in porosity within the Kankakee formation. The Kankakee formation is considered to be a thin high-velocity layer in the low-velocity background shale. Figure 1a shows the lithologic column depicting the limestone and shale formations, and Figure 1b illustrates a three-dimensional view of the boreholes at the site.

A pneumatically coupled three-component seismic detector (Owen and Parra, 1993) and two 24-element hydrophone arrays were used to acquire RVSP and cross-well seismic data respectively, using a 1000-Joule arc discharge borehole seismic source (Owen and Schroeder, 1987). This source was placed in well A, and the pneumatically coupled detector was placed in wells D, B, E, and C. For example, the pneumatically coupled detector was clamped in borehole B at a depth of 27.4 m to record RVSP measurements between wells A and B. A 24-channel data acquisition system was used for the experiments, to record three-component reverse VSP data at several source depths ranging from 140 to 240 m. The data were recorded with a 62 μ s sample rate, a user controlled fixed gain, 2048 samples per trace, and a variable time delay depending on the source-receiver offset.

In general, reverse VSP and interwell seismic measurements conducted in existing oil and gas fields involving impulsive source (e.g., explosives, air gun, sparker, etc.) must be carefully planned in order to avoid damaging the casing and cement bond in the source well. The main concern of the reservoir engineers and geologists regarding the use of downhole seismic sources is that the source-generated seismic disturbances can potentially destroy the cement bonds. The issue of source-induced borehole damage is given by Winbow (1989).

He suggested that the safest downhole source is the source that applies a low-level stress over a large surface area of the casing. For example, piezoelectric and vibrator controlled frequency sources have low-amplitude near-field displacements and as a consequence they are less likely to damage the source well. Also, explosive configurations such as explosive cords distribute pressure over a length of the casing rather than being concentrated at a point (Omnes, 1990). That is, the local stress can be small and produce a strong far-field signal (Hardage, 1992).

Rotation of Reverse VSP Seismograms

The reverse VSP data records consisted of three traces corresponding to the x, y and z axis accelerometers in the pneumatic receiver. In this pneumatic probe the accelerometer transducers were fixed with respect to the housing and the orientation of the probe was reflected in the response of the two horizontal transducers. The orientation of the pneumatic probe was determined from the data itself by displaying the x and y trace on a particle motion plot. Since the accelerometers are well isolated from one another, the first arrival response on the particle motion plot was aligned in the direction of the source with respect to receiver. For the purpose of picking first arrival times in the x-axis seismograms, the data from each of the four borehole pairs A-B, A-C, A-D, and A-E was rotated to align the x-axis data with the plane formed by the source and receiver boreholes. The rotation angle of each seismogram was determined graphically using several sets of particle motion plots from various source depths. As a result of the rotation process the seismograms exhibit an increase in amplitude of the x-component and a relative decrease in the y-component compared with their pre-rotation counterparts. As an example, Figures 2 and 3 show x- and y-component seismograms (from boreholes A and B) produced before and after the rotation, respectively.

Figure 4 shows the z-component seismogram which was unchanged during the rotation process. This seismogram illustrates a strong tube wave associated with a damage zone in well A below the Kankakee limestone formation at about a depth of 205 m. This is a change in borehole diameter which was observed in the caliper log at that depth (Parra et al., 1991). The source generated a tube wave (inside the fluid-filled borehole) which was reflected at the interface at the depth of 205 m, where unconsolidated clay was found to produce a tight plug. The analysis of tube wave propagation in fluid-filled boreholes in the presence of interfaces and washed-out zones can be reviewed in White (1983), and Balch and Lee (1984).

Reverse VSP Processing and Travel Time Inversion

In order to characterize the subtle reflections that are hidden by the stronger tube-waves f-k filtering analysis was performed (Hardage, 1992; Balch and Lee, 1984). The f-k filtered x- and z-component seismograms are shown in Figures 5 and 6. As a result of this processing some of the tube-wave events were removed. In particular, the strong tube wave event associated with the borehole damage zone was reduced by enhancing the reflection effects associated with structural features of the Kankakee limestone formation. As a consequence, the reflections at the shale/Kankakee limestone interface become stronger and more easily identified in the vertical component seismogram rather than the horizontal component seismogram. In this case, the rotated and filtered vector response associated with

each source position accurately minimizes the y-axis amplitudes and emphasizes the near-vertical particle motions corresponding to the source and detector geometry. These results suggest that the reflections observed in the vertical seismogram associated with the shale/Kankakee interface can be used as indicators for predicting the presence of the limestone formation using reverse VSP data.

A layered inversion approach that resolves only vertical changes in the subsurface velocity field was used. This type of inversion is reasonable at the Buckhorn test site because the geological section penetrated by boreholes A, B, C, D, and E is a horizontal limestone-shale sequence. Figures 7a and 7b presents velocity profiles generated from the data sets acquired in single borehole pairs. The shape of each profile is consistent with the others and there is only a small variation in the velocities themselves. This variation is attributable to noise in the first break picks and possibly laterally variable P-wave velocities near boreholes. Given the relatively small number of time picks used in each inversion, however, the results are very encouraging concerning the utility of the reverse VSP method for accurately resolving velocity profiles.

In general, good agreement is observed between the log data and the inversion results. The tie between the velocity logs and the vertical velocity profiles obtained from the inversion of the direct compressional wave times of arrivals demonstrates that high-frequency reverse VSP measurements can delineate geological structures in layered areas. In particular, the results show that the thin Kankakee limestone layer was resolved and the vertical velocity distribution of the shale unit was determined by the reverse VSP method. This suggests that the reverse VSP method can be used to map the 3-D velocity distribution and geological structures at distances of about 100 m away from borehole A by placing 3-component detectors in shallow boreholes at different radial orientations from the source borehole A.

DATA PROCESSING AND INTERPRETATION OF INTERWELL SEISMIC DATA

One-Dimensional Traveltime Seismic Inversion of Interwell Seismic Data

Interwell seismic measurements have become an extremely efficient technique for determining rock physical parameters and to detect heterogeneities in hydrocarbon-bearing formations. In particular, traveltime tomography inversion was successful in estimating the compressional wave velocity distribution within the Kankakee limestone formation which is considered to be a thin high-velocity layer in the low-velocity background shale. The results of several logging measurements conducted just after drilling five boreholes indicate a good well-to-well correlation with the lithology (Saito, 1991).

In the interwell seismic experiments two 24-element hydrophone arrays (streamers) were used to acquire the data. The source was placed in well A and the streamers were placed in wells D and B. The pressure interwell seismic data was recorded by moving the source in increments of 2 m in well A. A common source gather consisted of 24 traces corresponding to the 24 receivers in one of the 24-element hydrophone arrays. For example, one of the multiple offset datasets recorded between wells A and B consists of 24 source locations and 24 receiver locations.

Figure 8 shows a comparison between the velocity inversion profiles A-D and A-B and the compressional wave velocity log A. The velocity distribution profiles were obtained using the straight-ray inversion method assuming a layered earth model. Since an isotropic model was assumed, the horizontal velocity distribution of the anisotropic shale was approximately determined. The comparison between the velocity inversion profiles and the log in Figure 8 shows that the horizontal velocity distribution (from the inversion) is greater than the vertical velocity distribution derived from the sonic log (or the reverse VSP inversion) in the shale geological unit. In addition, comparisons between interwell velocity inversion profiles and velocity logs illustrate the isotropic characteristics of the Keokuk-Burlington formation (between 75-135 m of depth), and the Kankakee formation. These results suggested that the interwell seismic method and the reverse VSP method of seismic measurements can resolve the thin Kankakee limestone layer and determine the velocity distribution of the formations.

Dispersion and Attenuation Parameters from Interwell Seismic Data

Zero-vertical offset seismograms recorded in wells D and B were selected to extract dispersion and attenuation curves from interwell seismic data at source depths ranging from 154 m to 200 m, respectively, in well A. Figure 9 illustrates zero-vertical offset seismograms produced at 20 m and 46 m separation distances from the source well A. Both seismic sections show strong first arrivals for waves traveling in the shale formation, and weak first arrivals for waves traveling in the Silurian Kankakee formation. To compare the vertical velocity distribution with zero vertical offset waveforms, compressional wave velocity logs recorded in wells A, D and B are also illustrated in Figure 9. The logs show a relatively constant P-wave velocity distribution in the shale and a higher velocity in the limestone. However, a low-velocity zone within the limestone is observed in the logs. This zone starts in well D and extends to well B, and is located along well D at a depth between 196 m and 198 m.

The shale/Kankakee limestone interface is clearly defined by a sharp reflection event observed in both seismograms. The horizontal compressional wave velocity in the shale varies between 2.7 km/s and 3.5 km/s, and the compressional wave velocity in the Kankakee limestone formation varies between 4.0 km/s and 6.0 km/s (Parra and Zook, 1992). Several waveforms were selected to determine attenuation and dispersion effects in the shale and Kankakee limestone formation. Figures 10 and 11 illustrate in detail the waveforms that are used to extract quality factors, Q , and dispersions between wells D and B. The direct P waves associated with the limestone formation are clearly shown by the detectors in well D. Alternatively, the direct P waves observed in well B have been decreased in amplitude and broadened by the attenuated medium between wells D and B.

The compressional waves were determined from the seismic traces using time-frequency analysis (Cohen, 1989; Hlawatsch and Boudreaux-Bartels, 1992). As an example, Figures 12a and 12b show time-frequency plots for 30 ms, and 10 ms window lengths, respectively, of a zero-vertical offset seismic trace which was recorded at a depth of 186 m within the shale formation. The plots were produced using short-term Fourier transform with a Gaussian weighted analysis window and they represent spectral energy (i.e., squared amplitude spectra). In figure 12a the first contour plot represents a P-wave having a peak-

frequency of 750 Hz, the second peak is a reflection which arrives after 15 ms, and the last peak is a low-frequency tube wave which arrives after 27 ms at the detector in well D. In a similar manner time-frequency plots were produced for the corresponding zero-vertical offset seismic trace recorded in well B at the depth of 186 m. In this second example, the P-wave time-frequency plot for the first 10 ms window also exhibits a peak-frequency of 750 Hz as shown in Figure 13b. Time-windows of 10 ms were used to extract the P-wave pulses recorded at 20 m and 46 m as indicated by the time-frequency plots in Figures 12b and 13b. The peak-to-peak amplitude of the compressional waveform observed in well B is approximately four times smaller than the amplitude of the P waves recorded by the detector in well D as illustrated in Figure 14. In a similar manner, time-frequency analysis was applied to the zero-vertical offset traces recorded within the Kankakee limestone formation at the depth of 198 m between wells A-D and A-B, respectively. The direct P-wave pulses extracted from the zero-vertical offset traces are shown in Figure 15. These waveforms indicate a reduction of amplitude about nine times for the waves traveling within the target zone of interest in the limestone.

The next step in the interpretation was to calculate attenuation and dispersion from the selected seismic waveforms using the spectral ratio method (Ganley and Kanasewich, 1980). The waveforms given in Figure 14 associated with the shale formation are used to extract the quality factor and phase velocity at a depth of 186 m. The quality factor was determined to be $Q = 28$ and the phase velocity was found to reach its asymptotic value at 3 km/s, corresponding to the horizontal velocity of the shale, which was estimated using travel-time tomography. The dispersion characteristics are clearly observed in the phase-velocity curve by showing an increase of the velocity between 200 Hz and 500 Hz (see Figure 16). In addition, attenuation curves were calculated from seismic traces recorded at the depths of 180 m, 184 m, and 190 m. These attenuation curves, with corresponding quality factors $Q = 50$, 29, and 28, are shown in Figure 17. The highest quality factor of $Q = 50$ is associated with the Grassy Creek shale and the quality factors $Q = 29$ and 28 correspond to the Sweetland Creek shale.

Similarly, attenuation and phase velocities were calculated for the Kankakee limestone formation from the traces recorded at 196 m, 198 m, and 200 m and are given in Figures 18 and 19 together with the corresponding quality factors $Q = 4.5$, 7, and 4.4. As was expected, the low-quality factors and low-phase velocities of the limestone formation are associated with the high-porosity and low-velocity heterogeneity delineated by Saito (1991). In addition, core information from well D has indicated that the Kankakee limestone is formed by vertical fractures, vugs, and a very porous matrix rock which is partially saturated with oil and water. As a consequence, the high-attenuation of the seismic pulses traveling in the Kankakee formation is caused by the heterogeneous conditions of the porous and fractured rock matrix which is partially saturated with fluids. The presence of a gas phase is also possible at these shallow depths. Since the limestone zone is partially saturated with hydrocarbons and gas/water, the absorption mechanism may be caused by the interaction of porous fluid with the rock frame rather than absorption within the rock frame or the fluids (Jones, 1986). This suggests that the quality factor Q can be more accurately determined using a frequency dependent viscosity model.

In order to verify the RVSP and the interwell seismic interpretation, an attempt was made to estimate the changes in the internal fabric of the Kankakee using the variations in P- and S-wave velocity from logs. For this purpose, the compressional-wave velocity/shear-wave velocity ratios (v_p/v_s) were calculated from the well logs recorded in wells A, D, and B, as shown in Figure 20. The corresponding velocity ratios are illustrated in Figure 21. The v_p/v_s values calculated from well logs recorded in well A exhibit a high velocity ratio (v_p/v_s) of about 2.6 at the depth of 205 m below the Kankakee limestone formation. Similarly, a value of 3.1 associated with well D is obtained at a depth of 203 m, and a v_p/v_s ratio of about 3.65 was obtained at a depth of 200 m from the velocity logs in well B. These unusual v_p/v_s values observed in wells A, D, and B corresponds to an unconsolidated zone below the limestone formation which correlates with the tube wave reflection developed at the depth of 205 m in the source well A. This tube wave is observed in the reverse VSP seismograms shown earlier in Figures 5 and 6. For comparison, the average v_p/v_s value is 1.9 between 196 m to 199 m in the Kankakee limestone formation associated with wells A and B. This is a typical velocity ratio for a clean limestone without fractures and porosity. However, the v_p/v_s value in the same region within the limestone in well D decreases to a value of 1.5 which is typical of a porous limestone rock matrix.

SEISMIC MODELING

A layered earth model was constructed using rock physical parameters estimated from the logs and the velocity inversion and attenuation results. This model consists of a limestone-shale sequence formed by a thick isotropic limestone layer followed by an anisotropic low-velocity shale of thickness 60 m and an isotropic thin high-velocity limestone layer representing the Kankakee formation. This thin layer overlays the bottom half-space formed by a low-velocity shale having the same characteristics as those of the upper shale layer. The rock physical parameters of this model are given in Table 1.

To simulate synthetic seismograms, a Ricker wavelet source pulse signal was assumed. A pressure seismogram was calculated for a horizontal point force signal centered at 600 Hz and located at a horizontal distance of 46 m from the detectors. The traces in the seismogram shown in Figure 22a correspond to outputs of detectors spaced 2 m apart. The shale formation is characterized by large P-wave amplitudes (first arrival) followed by converted compressional and shear waves. The seismic section also shows the presence of the Kankakee limestone layer which attenuates the seismic waves observed by detectors within the thin layer. To check the synthetic seismogram, the corresponding observed field data recorded in well B is illustrated in Figure 22b. The tie of synthetic data (using the layered geological model) and field data confirms the interpretation of well logs and crosswell seismic data which suggest a predominantly horizontal geology over the surveyed zone and an attenuated limestone. In general, the results of the simulation compare very well with the raw field data except for the P-waves traveling in the Kankakee (a highly heterogeneous formation) and for the large amplitude tube waves, which cannot be produced with the present program. However, the computer code can be modified for simulating seismic data containing tube waves. This can be done through the use of equivalent line mechanism representations of the source and receiver wells (Kurkjian, *et al*, 1992).

SUMMARY AND CONCLUSIONS

The high-resolution reverse VSP field experiments and data analysis have yielded important results which demonstrate that a pneumatic triaxial detector probe responds to high-frequency seismic waves produced by a 1000-Joule arc-discharge borehole source in the sedimentary formations at the Buckhorn test site in Illinois. The layer inversion (1-D) results show good agreement with the log data. This suggests that the high-resolution reverse VSP method can resolve small features, in particular, the thin Kankakee limestone layer. Reverse VSP and interwell seismic inversions together with log data have yielded information on the velocity distribution of the shale formation. These results inferred that the reverse VSP (using several 3-component detectors in shallow boreholes) and interwell seismic measurements integrated with log data can detect and delineate the geological structures at the Buckhorn test site. The presence of the low-velocity zone previously detected and delineated using travel-time tomography has been characterized by performing attenuation measurements (using hydrophone detectors). The low-velocity zone is located in the Kankakee formation along borehole D at a depth between 196 m to 198 m where the compressional and shear logs also showed low velocities. This zone having low-quality factor within the limestone correlated accurately with a high-porosity heterogeneity delineated with travelttime inversion tomography and well log information. In addition, the core information from well D indicates that the Kankakee limestone contains vertical fractures, vugs, and a very porous matrix which is partially saturated with oil and water. Thus, the high-attenuation affecting the seismic pulses traveling in the Kankakee formation is caused by the heterogeneous conditions of the porous and fractured partially saturated rock matrix.

High-resolution RSVP and interwell seismic techniques can assist the reservoir engineers and geologists to obtain detail information about the heterogeneous nature of oil and gas fields. During primary production the goal is to position wells at correct locations, and during secondary production the goal is to maximize the volume of hydrocarbon contacted by injected fluids. Thus, RVSP and interwell seismic techniques together with seismic tomography can be potentially powerful diagnostic and monitory techniques for enhanced oil recovery operations. The main effort to conduct the seismic measurements in an existing oil field is the well preparation logistic which is an unavoidable element associated with borehole related measurements. In addition, the interwell seismic data acquisition costs are high for multiple offset measurements when several source and detector levels are required. The charges typically range from 10 K to 20 K per day with additional cost of pulling tubing and preparation of the well for the survey. In spite of the presently high acquisition costs and well preparation logistics, the RVSP and interwell seismic, emerging technologies, when fully developed and commercialized, will have a large impact on reservoir management.

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TABLE 1. Formation Parameters

Region	Thickness (m)	Density (g/cm ³)	C ₁₁ (GPa)	C ₁₃ (GPa)	C ₃₃ (GPa)	C ₄₄ (GPa)	C ₆₆ (GPa)	Q
Limestone	134	2.87	65.7	18.4	65.7	23.7	23.7	40
Shale	60	2.55	23.9	11.7	16.6	3.7	6.1	28
Kankakee	7.5	2.10	50.0	14.0	50.0	18.0	18.0	4.5
Shale	∞	2.55	23.9	11.7	16.6	3.7	6.1	28

Figure 1. a) Lithologic column, and b) a three-dimensional view of five boreholes at the Buckhorn test site near Quincy, IL.

Figure 2. Reverse VSP seismograms recorded using the pneumatically-coupled detector at a depth of 27.4 m. Source in well A and detector in well B. a) x-component seismogram, and b) y-component seismogram.

Figure 3 Rotated reverse VSP seismograms. The x-axis accelerometer aligned with plane defined by source-receiver wells. Source in well A and detector in well B. a) x-component seismogram, and b) y-component seismogram.

Figure 4. z-component seismogram recorded using the pneumatically-coupled detector at a depth of 27.4 m. Source in well A and detector in well B.

Figure 5. f-k filtered x-component seismograms. Source in well A and detector in well B.

Figure 6. f-k filtered z-component seismogram. Source in well A and detector in well B.

(a)

(b)

Figure 7. Compressional wave velocity profiles derived from rotated x-component reverse VSP seismograms and sonic logs. The dashed-dotted lines represent the compressional wave velocity logs and the solid lines represent the velocity inversion profiles from RVSP data.

The source is in well A and detectors are in wells (a) D and B; (b) C and E.

Figure 8. Compressional wave velocity profiles derived from interwell seismic data and sonic logs. Source is in well A and hydrophone streamers are in wells a) D and b) B.

Figure 9. Zero-vertical offset seismograms produced between wells A, D, and B.

Figure 10. Zero-vertical offset waveforms recorded between wells A and D.

Figure 11. Zero-vertical offset waveforms recorded between wells A and B.

Figure 12. Time-frequency analysis of zero-vertical offset waveform for the shale formation between wells A and D. Source and detector at a depth of 186 m. a) Time-window of 30 ms, and b) Time-window of 10 ms.

Figure 13. Time-frequency analysis of zero-vertical offset waveform for the shale formation between wells A and B. Source and detector at a depth of 186 m. a) Time-window of 30 ms, and b) Time-window of 10 ms.

**Detector
in Well D**

**Detector
in Well B**

Figure 14. Zero-vertical offset P-waves for the shale formation source and detectors at a depth of 186 m.

**Detector
in Well D**

**Detector
in Well B**

Figure 15. Zero-vertical offset P-waves for the Kankakee limestone formation. Source and detectors at a depth of 198 m.

Figure 16. Dispersion and attenuation curves for the shale formation between wells D and B at a depth of 186 m.

Figure 17. Attenuation curves for the shale formation between wells D and B at several depths.

Figure 18. Attenuation waves for the Kankakee formation between wells D and B.

Figure 19. Dispersion waves for the Kankakee formation between wells D and B at a depth of 198 m.

Figure 20. Compressional and shear wave velocity logs recorded between 190 m and 210 m.

Figure 21. Compressional wave/shear wave velocity ratio (v_p/v_s) in well A, D, and B.

Figure 22. a) Pressure synthetic seismogram for a point-force located in the shale formation at a depth of 160 m. b) Common-source seismogram produced by an arc discharge source stationed in well A at a depth of 160 m. Wells spaced 46 m.

CHAPTER 2

NUMERICAL SIMULATION OF ELASTIC WAVEFIELD RADIATED BY A POINT FORCE IN A LAYERED TRANSVERSELY ISOTROPIC MEDIUM: MODEL STUDIES

The characterization and detection of fracture zones have become important for storage of nuclear toxic waste, geothermal, hydrocarbon reservoir characterization and hydrological applications as well as to investigate stress-related phenomena in underground mines. In particular, to recognize the alignment of stress-induced cracks, and principal stress directions of these cracks (from high-resolution elastic wave propagation data) has application in almost every subsurface investigation concerned with the physical properties of rock formation. This information should enable fractured rock to be identified and investigated and the internal structure of known geology to be characterized in more detail. For example, information on the orientations of natural *in-situ* microfractures may be expected to accurately predict the orientation of artificially created hydraulic fractures as well as indicate the preferred direction of fluid movement in hydrocarbon reservoir characterization applications (Crampin, 1980). In addition, a detailed knowledge on the interpretation of seismic velocity structure together with the distribution of mining-induced seismicity can be used to characterize the stress rock mass which may be highly stressed in high-velocity zones and highly fractured in low-velocity zones. That is, in an underground mine, the low-velocity seismic zone can be of low rock burst potential, while the high-velocity zone can be an active seismic zone having the potential for significant seismic energy release and rock bursting (Young and Maxwell, 1992)

In this chapter, the analysis of elastic wave propagation in transversely isotropic media is developed for a point force at an arbitrary angle with respect to the axis of symmetry of a horizontally layered structure. As an example, the seismic response of a dipping low-velocity anisotropic layer is analyzed as a fracture zone in an unfractured host medium. In addition, rock physical parameters previously determined from reverse VSP, well logs, and interwell seismic data (i.e., acquired at the Buckhorn test site in Illinois) are used to validate the layered transversely isotropic solution by producing a synthetic seismogram which agrees with the corresponding observed seismic data.

FORMULATION

Equation of Motion in Cylindrical Coordinates

In a cylindrical coordinate system (r, θ, z) , consider a medium characterized by five elastic constants c_{11} , c_{12} , c_{13} , c_{33} , and c_{44} , and rock material density ρ . The stress tensor components for the transversely isotropic medium in terms of the displacement vector $u = (u_r, u_\theta, u_z)$, with the axis of symmetry coinciding with the z -axis are given by,

$$\sigma_{rr} = c_{11} \frac{\partial u_r}{\partial r} + c_{12} \left(\frac{1}{r} \frac{\partial u_\theta}{\partial \theta} + \frac{u_r}{r} \right) + c_{13} \frac{\partial u_z}{\partial z},$$

$$\begin{aligned}
\sigma_{\theta\theta} &= c_{12} \frac{\partial u_r}{\partial r} + c_{11} \left(\frac{1}{r} \frac{\partial u_\theta}{\partial \theta} + \frac{u_r}{r} \right) + c_{13} \frac{\partial u_z}{\partial z}, \\
\sigma_{zz} &= c_{13} \frac{\partial u_r}{\partial r} + c_{13} \left(\frac{1}{r} \frac{\partial u_\theta}{\partial \theta} + \frac{u_r}{r} \right) + c_{33} \frac{\partial u_z}{\partial z}, \\
\sigma_{\theta z} &= c_{44} \left(\frac{1}{r} \frac{\partial u_z}{\partial \theta} + \frac{\partial u_\theta}{\partial z} \right), \\
\sigma_{zr} &= c_{44} \left(\frac{\partial u_z}{\partial r} + \frac{\partial u_r}{\partial z} \right),
\end{aligned}$$

and

$$\sigma_{r\theta} = c_{66} \left(\frac{\partial u_\theta}{\partial r} - \frac{u_\theta}{r} + \frac{1}{r} \frac{\partial u_r}{\partial \theta} \right), \quad (1)$$

where $c_{66} = \frac{1}{2} (c_{11} - c_{12})$.

These stress tensor components given by equation (1) are substituted into the steady state

equation of motion $\nabla \cdot \vec{\sigma} + \rho \omega^2 \mathbf{u} + \mathbf{F} = 0$, to yield the coupled partial differential equations for the displacement components u_r , u_θ , and u_z in the medium:

$$\begin{aligned}
a \left[\nabla^2 u_r - \frac{1}{r} \left(2 \frac{\partial u_\theta}{r \partial \theta} + \frac{u_r}{r} \right) \right] + b \frac{\partial \Delta}{\partial r} + c \frac{\partial^2 u_z}{\partial z \partial r} + d \frac{\partial^2 u_r}{\partial z^2} + F_r + \rho \omega^2 u_r &= 0, \\
a \left[\nabla^2 u_\theta - \frac{1}{r} \left(\frac{u_\theta}{r} - \frac{2}{r} \frac{\partial u_r}{\partial \theta} \right) \right] + b \frac{\partial \Delta}{r \partial \theta} + c \frac{1}{r} \frac{\partial^2 u_z}{\partial r \partial \theta} + d \frac{\partial^2 u_\theta}{\partial z^2} + F_\theta + \rho \omega^2 u_\theta &= 0, \\
e \nabla^2 u_z + f \frac{\partial \Delta}{\partial z} + g \frac{\partial^2 u_z}{\partial z^2} + F_z + \rho \omega^2 u_z &= 0,
\end{aligned} \quad (2)$$

$$e \nabla^2 u_z + f \frac{\partial \Delta}{\partial z} + g \frac{\partial^2 u_z}{\partial z^2} + F_z + \rho \omega^2 u_z = 0,$$

where:

$$\begin{aligned}
a &= c_{66}, \\
b &= c_{11} - c_{66}, \\
c &= c_{13} + c_{44} + c_{66} - c_{11}, \\
d &= c_{44} - c_{66}, \\
e &= c_{44}, \\
f &= c_{13} + c_{44}, \\
g &= c_{33} - c_{13} - 2c_{44},
\end{aligned}$$

and
$$\Delta = \frac{\partial u_r}{\partial r} + \frac{u_r}{r} + \frac{1}{r} \frac{\partial u_\theta}{\partial \theta} + \frac{\partial u_z}{\partial z}.$$

The vectors (u_r, u_θ, u_z) and (F_r, F_θ, F_z) correspond to components of the displacement and body force per unit of volume in the r, θ, z directions, respectively.

Equation of Motion in the Wave Number Domain

Expanding the displacement and body force in a Fourier series with respect to the azimuth θ , a system of equations is obtained in terms of the vector displacement $(u_{rn}, u_{\theta n}, u_{zn})$ and body force $(F_{rn}, F_{\theta n}, F_{zn})$ wave coefficients. This resulting system of equations is reduced (in the wavenumber domain) using the Hankel transform method to one set of two coupled partial differential equations and a separate uncoupled partial differential equation. In the wavenumber domain, k , the coupled system of differential equations is given by

$$\left[c_{44} \frac{d^2}{dz^2} - k^2 c_{11} + \omega^2 \rho \right] u_{1n} - k(c_{13} + c_{44}) \frac{du_{2n}}{dz} + F_{1n} = 0, \quad (3a)$$

$$-k(c_{13} + c_{44}) \frac{du_{1n}}{dz} - \left[c_{33} \frac{d^2}{dz^2} - c_{44} k^2 + \omega^2 \rho \right] u_{2n} - F_{2n} = 0, \quad (3b)$$

and the uncoupled partial differential equation is given by

$$\left[c_{44} \frac{d^2}{dz^2} - c_{66} k^2 + \omega^2 \rho \right] u_{3n} + F_{3n} = 0. \quad (4)$$

The functions $u_{1n}(z, k)$ and $u_{2n}(z, k)$ are coupled through equations (3a) and (3b) and are independent of $u_{3n}(z, k)$ which must satisfy equation (4). The functions u_{1n} and u_{2n} are associated with quasi-compressional and quasi-shear waves, respectively, and the function $u_{3n}(z, k)$ in equation (4) is associated with waves whose particle motion is polarized in the horizontal plane.

A Point Force in an Unbounded Medium

To derive the Green's function in the wavenumber domain using the method given by Kupradze (1979), the coupled system of partial differential equations in equation (3) is expressed in matrix form as

$$\mathcal{L}u = F, \quad (5)$$

with the following vector notation:

$$u = \begin{pmatrix} u_{1n} \\ u_{2n} \end{pmatrix} \quad \text{and} \quad F = \begin{pmatrix} -F_{1n} \\ F_{2n} \end{pmatrix}.$$

$\mathcal{L}$ is the differential operator of components $\mathcal{L}_{ij}$, given by

$$\mathcal{L}_{ij} = \begin{pmatrix} \left[c_{44} \frac{d^2}{dz^2} - c_{11} k^2 + \omega^2 \rho \right] & -k(c_{13} + c_{44}) \frac{d}{dz} \\ -k(c_{13} + c_{44}) \frac{d}{dz} & -\left[c_{33} \frac{d^2}{dz^2} - c_{44} k^2 + \omega^2 \rho \right] \end{pmatrix}. \quad (6)$$

Since the Green's function for a point force in a transversely isotropic unbounded medium was developed by Kazi-Aoual *et al.* (1988), only the final expression for the vector wave displacement is presented in the wavenumber domain. Thus, displacement components due to a horizontal point force $f_1(\omega)$ (for $z \geq z_s$) are given by

$$(u)_h = \frac{f_1(\omega)}{c_{33}c_{44}(v^2 - \chi^2)} \begin{pmatrix} -c_{33} \frac{(v^2 - \alpha^2)}{k} & c_{33} \frac{(\chi^2 - \alpha^2)}{\chi} \\ -v(c_{13} + c_{44}) & k(c_{13} + c_{44}) \end{pmatrix} \begin{pmatrix} \frac{k}{v} e^{-v(z-z_s)} \\ e^{-\chi(z-z_s)} \end{pmatrix}, \quad (7)$$

and for a vertical point force $f_2(\omega)$ they are

$$(u)_v = \frac{f_2(\omega)}{c_{33}c_{44}(v^2 - \chi^2)} \begin{pmatrix} -k(c_{13} + c_{44}) & (c_{13} + c_{44})\chi \\ c_{44} \frac{(v^2 - \beta^2)}{v} & c_{44} \frac{(\chi^2 - \beta^2)}{k} \end{pmatrix} \begin{pmatrix} e^{-v(z-z_s)} \\ \frac{k}{\chi} e^{-\chi(z-z_s)} \end{pmatrix}, \quad (8)$$

where χ and v are wavenumbers associated with quasi-compressional and quasi-SV waves traveling in a transversely isotropic medium. These parameters can be obtained from the following relationship:

$$\begin{aligned} \begin{pmatrix} \chi^2 \\ v^2 \end{pmatrix} = \frac{S}{4c_{44}c_{33}} \left[\left(k^2 - \frac{\omega^2 \rho}{S_a} \right) + \left(k^2 - \frac{\omega^2 \rho}{S_b} \right) \right] \\ \pm \frac{1}{2} \left\{ \frac{S^2}{4c_{44}^2 c_{33}^2} \left[\left(k^2 - \frac{\omega^2 \rho}{S_a} \right) + \left(k^2 - \frac{\omega^2 \rho}{S_b} \right) \right]^2 - \frac{4c_{11}}{c_{33}} \left(k^2 - \frac{\omega^2 \rho}{c_{11}} \right) \left(k^2 - \frac{\omega^2 \rho}{c_{44}} \right) \right\}^{1/2}, \end{aligned} \quad (9)$$

in which

$$S_a = \frac{S}{2c_{33}}$$

$$S_b = \frac{S}{2c_{44}}$$

$$S = c_{33}c_{11} + c_{44}^2 - (c_{13} + c_{44})(c_{13} + c_{44}).$$

$$\alpha^2 = \frac{c_{44}}{c_{33}} k^2 - \frac{\omega^2 \rho}{c_{33}}$$

and
$$\beta^2 = \frac{c_{11}}{c_{44}} k^2 - \frac{\omega^2 \rho}{c_{44}}.$$

The vector displacement solution associated with quasi-SH waves is obtained directly from equation (4), i.e.,

$$u_{3n} = \frac{f_1(\omega)}{c_{44}} \frac{e^{-s(z-z_1)}}{s},$$

where
$$s = \left(\frac{c_{66}}{c_{44}} k^2 - \frac{\omega^2 \rho}{c_{44}} \right)^{1/2}.$$

WAVE PROPAGATION IN A STRATIFIED TRANSVERSELY ISOTROPIC MEDIUM

Formulation

To develop the general solution of a point force in a layered transversely isotropic medium, the displacement-stress matrix is required to be expressed in terms of upgoing and downgoing waves and unknown wave coefficients η_j (for $j = 1, 4$). The displacement-stress matrix in the wavenumber domain for the j^{th} layer of the transversely isotropic medium is expressed by

$$u_j = J_j(k) g_j \eta_j, \quad (10)$$

where $u_j = \left[u_1^{(j)}, u_2^{(j)}, \sigma_{21}^{(j)}, \sigma_{22}^{(j)} \right]^T,$

$$\eta_j = \left[\eta_1^{(j)}, \eta_2^{(j)}, \eta_3^{(j)}, \eta_4^{(j)} \right]^T,$$

and $g_j = \text{diag} \left(\exp(-v_j z^-), \exp(-\chi_j z^+), \exp(v_j z^+), \exp(\chi_j z^+) \right)$

in which

$$z^- = z - z_{j-1}, \text{ and } z^+ = z - z_j,$$

and the matrix $J_j(k)$ is written as

(11)

$$\begin{bmatrix} -kv_j & k\chi_j & -kv_j & k\chi_j \\ -\gamma_j^2 f_j & \Omega_j^2 f_j & \gamma_j^2 f_j & -\Omega_j^2 f_j \\ -c_{44}^{(j)} k [v_j^2 - \gamma_j^2 f_j] & c_{44}^{(j)} k [\chi_j^2 - \Omega_j^2 f_j] & c_{44}^{(j)} k [v_j^2 - \gamma_j^2 f_j] & -c_{44}^{(j)} k [\chi_j^2 - \Omega_j^2 f_j] \\ \left[-c_{13}^{(j)} k^2 + c_{33}^{(j)} \gamma_j^2 f_j \right] v_j & \left[c_{13}^{(j)} k^2 + c_{33}^{(j)} \Omega_j^2 f_j \right] \chi_j & \left[-c_{13}^{(j)} k^2 + c_{33}^{(j)} \gamma_j^2 f_j \right] v_j & \left[c_{13}^{(j)} k^2 + c_{33}^{(j)} \Omega_j^2 f_j \right] \chi_j \end{bmatrix},$$

where $\gamma_j^2 = v_j^2 - \beta_j^2,$

$$\Omega_j^2 = \chi_j^2 - \beta_j^2,$$

$$f_j = \frac{c_{44}^{(j)}}{c_{13}^{(j)} + c_{44}^{(j)}},$$

$$\text{and } \beta_j^2 = \left[\frac{c_{11}^{(j)}}{c_{44}^{(j)}} k^2 - \frac{\omega^2 \rho_j}{c_{44}^{(j)}} \right].$$

For quasi-SH waves, the displacement stress-matrix takes the form

$$\begin{pmatrix} u_3^{(j)} \\ \sigma_{23}^{(j)} \end{pmatrix} = \begin{pmatrix} 1 & 1 \\ -c_{44}^{(j)} s_j & c_{44}^{(j)} s_j \end{pmatrix} \begin{pmatrix} \eta_5^{(j)} e^{-s_j(z-z_{j-1})} \\ \eta_6^{(j)} e^{s_j(z-z_j)} \end{pmatrix}, \quad (12)$$

where the parameter s_j is given by

$$s_j = \left(\frac{c_{66}^{(j)}}{c_{44}^{(j)}} k^2 - \frac{\omega^2 \rho_j}{c_{44}^{(j)}} \right)^{1/2}.$$

As a check on the derivation given by equation (11), the corresponding solution for a layered isotropic medium is deduced. Replacing in equation (11)

$$\begin{aligned} c_{13}^{(j)} &= \lambda_j \\ c_{11}^{(j)} &= c_{33}^{(j)} = \lambda_j + 2\mu_j \\ \text{and } c_{44}^{(j)} &= c_{66}^{(j)} = \mu_j, \end{aligned}$$

yields the displacement-stress matrix in the wavenumber domain. For this case, the matrix $J_j(k)$ becomes

$$J_j(k) = \begin{pmatrix} -k & \chi_j & -k & \chi_j \\ -v_j & k & v_j & -k \\ 2\mu_j k v_j & -\mu_j(2k^2 - k_{sj}^2) & -2\mu_j k v_j & \mu_j(2k^2 - k_{sj}^2) \\ \mu_j(2k^2 - k_{sj}^2) & -2\mu_j k \chi_j & \mu_j(2k^2 - k_{sj}^2) & -2\mu_j k \chi_j \end{pmatrix}, \quad (13)$$

where $k_{sj}^2 = \omega^2 \rho_j / \mu_j$, $v_j = (k^2 - \omega^2 \rho_j / (\lambda_j + 2\mu_j))^{1/2}$, and $\chi_j = (k^2 - k_{sj}^2)^{1/2}$.

The coefficients of the stress-displacement matrix given by equation (13) are the same as those coefficients associated with the Green's function for an elastic isotropic medium derived by Apsel (1979).

Boundary Conditions

The analysis of wave propagation in anisotropic layered media is a boundary value problem in which the motion is governed by the elastic wave equation in each layer, and the displacement and stress fields must satisfy the conditions of continuity of displacements and tractions across each interface, and the radiation conditions above the layered medium $z \leq 0$ and in the underlying half-space $z \geq 0$. To avoid precision problems caused by the growing exponential in the individual matrices for large wavenumbers, the global matrix method (Schmidt and Tango, 1986; Mal, 1988) is implemented, as an alternative to the traditional propagation approach to calculate the vector wave displacement and the pressure. To facilitate the global matrix analysis, equation (10) is written in its vector and matrix form

$$\{u_m(z)\} = [J_m][g_m(z)]\{\eta_m\} \quad (14)$$

In equation (14) the vector $\{\eta_m\}$ contains the unknown wave coefficients of wave propagation in the positive and negative z -directions.

In the global matrix formulation the boundary conditions at the layer interface are satisfied in the form of difference of field quantities across the layer interface. These conditions can be represented by using equation (14) as follows:

$$[J_{m+1}][g_{m+1}(z_m)]\{\eta_{m+1}\} - [J_m][g_m(z_m)]\{\eta_m\} = \{u_{m+1}(z_m)\} - \{u_m(z_m)\} \quad (15)$$

where z_m is the coordinate of the interface between m^{th} and $(m+1)^{\text{th}}$ layers. For example, the radiation condition associated with the upper half-space is enforced to equation (15) for $m = 1$ by making the downgoing part of $\{\eta_1\} = 0$. In a similar manner the radiation condition is applied to the lower half-space by making the upgoing part of $\{\eta_{N+1}\}$ vanish.

An assembly of equation (15) for $m = 2, 3, \dots, N$, and reduced version of equation (15) for $m = 1$, and $N + 1$ forms a linear system of equations given by

$$[Q]\{\eta\} = \{B\} \quad (16)$$

In equation (16), the global matrix $[Q]$ is composed of submatrices obtained from $[J_m]$, $[g_m(z_m)]$, and $[g_{m+1}(z_m)]$, for $m = 1, 2, \dots, N + 1$. These submatrices contain coefficients associated with the physical properties of the anisotropic layered medium as well as the coordinates of the interfaces. The global vector $\{B\}$ represents the external excitations (e.g., point forces or pressure sources) in the form of jump conditions at the interfaces z_m , $m = 1, 2, \dots, N + 1$, which are obtained from the right hand side of equation (15). The global vector $\{\eta\}$ is formed by the unknown wave coefficients with downgoing and upgoing waves in each layer.

NUMERICAL EXAMPLES

The vector wave displacement solution of a point force in the presence of an isotropic dipping layer in an unbounded medium developed by Parra *et al.* (1993) is generalized using

the present development to simulate the response of dipping multilayer transversely isotropic structure with respect to the z-axis symmetry. The integral solution of the vector displacement in terms of Bessel functions (Hankel transforms) contains kernels in the wavenumber domain which are derived directly from equations (8), (9), and (11). Alternatively, to simulate the pressure, an expression is derived from equations (1) in terms of the displacements given by

$$p = \frac{(c_{11} + c_{12} + c_{13})}{3} \left(\frac{\partial}{\partial r} + \frac{1}{r} \right) u_r + \frac{(2c_{13} + c_{33})}{3} \frac{\partial u_z}{\partial z} .$$

The numerical evaluation of the resulting wavenumber integrals of highly irregular oscillatory kernels was accomplished using the modified and extended version of the Clenshaw-Curtis method, given in Parra *et al.* (1993). The vector displacement and pressure time responses in terms of borehole separation r , rock physical properties and the number of layers were obtained numerically using FFT.

The Response of a Dipping Fracture Zone

Since strain build up by tectonic forces introduces changes in the fractured rock such as stress concentrations, creep, changes in porosity, and stress-induced anisotropy, a fracture zone can be represented as a thin anisotropic layer having a lower velocity and more lossy material than its boundary rock host medium. The host rock is assumed to be isotropic and unbounded. The response of a low-velocity transversely isotropic layer dipping 40 degrees with respect to the horizontal separation between source and detector boreholes is calculated using the present analysis. The geometry of this fracture zone of thickness 5 m, which includes the source and detector locations, is illustrated in Figure 1, and the physical properties of the model are given in Table 1. The horizontal and vertical displacement synthetic seismograms for a horizontal force pulse signal centered at 500 Hz were calculated at 12 different detector positions for the model shown in Figure 1. The horizontal force was applied at the depth of 5 m (below the dipping layer) and polarized in the xy horizontal plane at an angle of 45 degree with respect to the plane containing the boreholes, and detectors were located at a horizontal distance of 35 m from the source. The three-component detectors were placed 7.5 m apart above the layer, in the layer, and below the layer.

Figure 2 illustrates the vector displacement seismograms using the physical parameters given in Table 1. The same figure shows as a reference the corresponding seismograms associated with an isotropic dipping low-velocity layer in an isotropic unbounded medium. Although, the sensitivity of the shear waves to the internal structure of the fracture zone (anisotropic conditions) is small for wavelengths comparable to its thickness, the low-velocity layer has a distinctive signature which can be better characterized using hodogram plots. As an example, particle motion plots projected on the plane yz are illustrated in Figure 3(a) and (b). The plots exhibit shear-wave splitting which effect is expected to diminish as the angle between the horizontal point force and the plane containing the boreholes decreases (Parra *et al.*, 1993). The split shear-waves are reflection events produced at the dipping low-velocity layer interface. The faster split shear-wave is polarized parallel to the fracture zone over the

range of angles associated with detectors 3, 4, 5, 6 and 7. Alternatively, the wavefields observed at detectors 9 through 12 follow the same patterns of those wavefields radiated by the point force in an unbounded medium. These observations suggest that the use of several source polarizations and several three-component detectors will be appropriate for mapping and detecting fracture zones and their orientations in the 3-D space using crosswell seismic techniques.

The Response of a Thin Layer in an Anisotropic Shale

In this section I apply the present solution to calculate the seismic response of a hydrocarbon bearing formation at the Buckhorn test site in Illinois (Parra *et al.*, 1991 and Saito, 1991). Reverse VSP and interwell seismic experiments were conducted at the site for mapping geological structures and to determine rock physical parameters to characterize the Silurian Kankakee limestone formation, which is an oil producer in the Mt. Sterling area in Illinois. Reverse VSP and interwell seismic measurements together with log data have yielded information on the anisotropic characteristics of the shale and heterogeneities within the limestone associated with lateral changes in porosity. According to the exploration and production histories of the field, the oil reservoir is the porous zone of the Silurian Kankakee which is horizontally distributed at a depth of about 200 m, and has less than 8 m of thickness. The results of several logging measurements conducted just after drilling five boreholes indicate a good correlation of the formation from well to well. The Kankakee is considered as a thin high-velocity layer in the low-velocity background shale.

A three-dimensional view of the wells is shown in Figure 4. Two 24-element hydrophone arrays were used to conduct attenuation measurements and traveltime tomography using a 1000-joule arc discharge borehole seismic source. The source was placed in well A, at a depth of 180 m and hydrophone arrays were placed in wells D, B, and E. The rock physical parameters and structure determined from reverse VSP, interwell seismic data and well logs were used to construct a layered model. This model consists of a limestone-shale sequence formed by a thick isotropic limestone layer followed by an anisotropic low-velocity shale of thickness 60 m and an isotropic thin high-velocity limestone layer representing the Kankakee formation. This thin layer overlays the bottom half-space formed by a low-velocity shale having the same characteristics as those of the upper shale section. The rock physical parameters of this model are given in Table 2.

Figure 5 shows a pressure seismogram calculated for a horizontal point force signal centered at 600 Hz and located at a horizontal distance of 46 m from the source. The shale formation is characterized by large P-wave amplitudes followed by converted compressional and shear waves. The seismogram also shows the presence of the Kankakee limestone layer which attenuates the seismic waves observed by detectors within the thin layer. To check the synthetic seismogram, the corresponding observed field data recorded in well B is illustrated in Figure 6. The agreement between synthetic data and field data confirms the interpretation of well logs and crosswell seismic data which suggest a predominantly horizontal structure and an attenuated limestone. The model results compare very well with the raw field data except for the P-waves traveling in the Kankakee (a highly heterogeneous formation) and for the large amplitude tube waves which cannot be produced with the present program. How-

ever, the code can be modified for simulating seismic data containing tube waves. This can be done through the use of equivalent line mechanism representations for the source and receiver wells (Kurkjian *et al.*, 1992).

CONCLUSIONS

The formulation of elastic wave propagation in layered transversely isotropic media for a point force at an arbitrary angle with respect to the vertical axis of symmetry of layered structure was developed. The results of this work demonstrate that the three-component particle motion associated with a dipping low-velocity fractured zone is sensitive to the orientation of the source relative to the plane containing the source and detector borehole. This suggests that a dipping fracture zone can be detected by analyzing shear-wave polarization diagrams (constructed from waveforms recorded at different detector positions in the borehole) for a source polarized at different angles with respect to the plane containing the boreholes. In addition, the agreement between synthetic and field data validates the computer model and confirms the interpretation of well logs and crosswell seismic data recorded at the Buckhorn test site in Illinois. The model results compare very well with the raw data except for the large amplitude tube waves which can not be simulated with the current modeling program. However, this code can be modified for modeling seismic data acquired with a borehole hydrophone, and a borehole pressure source using equivalent line mechanism representations for the source and detector wells.

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TABLE 1. Formation Parameters for a Dipping Low-Velocity Layer

Region	Thickness (m)	Density (g/cm ³)	C ₁₁ (Ga)	C ₁₃ (Ga)	C ₃₃ (Ga)	C ₄₄ (Ga)	C ₆₆ (Ga)	Q
Layer	5	2.42	17.0	6.3	11.7	2.7	5.4	10
Upper and lower medium	∞	2.63	37.9	15.2	37.9	11.3	11.3	40

TABLE 2. Formation Parameters for a Limestone-Shale Layer Sequence

Region	Thickness (m)	Density (g/cm ³)	C ₁₁ (Ga)	C ₁₃ (Ga)	C ₃₃ (Ga)	C ₄₄ (Ga)	C ₆₆ (Ga)	Q
Limestone	134	2.87	65.7	18.4	65.7	23.7	23.7	40
Shale	60	2.55	27.7	15.52	16.6	3.7	6.1	28
Kankakee	7.5	2.10	50.0	14.0	50.0	18.0	18.0	4.5
Shale	∞	2.55	27.7	15.52	16.6	3.7	6.1	28

Figure 2. Comparisons of vertical-, radial-, and transverse-component seismograms produced by a 500-Hz horizontal point force pulse signal in the presence of a dipping low-velocity layer. The solid lines represent waveforms produced by a transversely isotropic layer.

Figure 3a. Two-axis hodogram plots for a force in the presence of a dipping low-velocity layer. Hodograms were produced from waveforms calculated for detector depths 7.5 m apart from -35 m through 2.5 m.

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Figure 3b. Two-axis hodogram plots for a force in the presence of a dipping low-velocity layer. Hodograms were produced from waveforms calculated for detector depths 7.5 m apart from 10 m through 47.5 m.

Figure 4. A three-dimensional view of the wells at the Buckhorn test site in Illinois.

Figure 5. A pressure synthetic seismogram produced by a point force in the shale formation at a depth of 180 m.
Wells spaced 46 m.

Figure 6. Common-source seismogram produced by an arc discharge source stationed in well A at a depth of 180 m.
Wells spaced 46 m.

CHAPTER 3

ATTENUATION AND DISPERSION OF SEISMIC GUIDED WAVES IN LAYERED PERMEABLE MEDIA FOR CONTINUITY LOGGING APPLICATIONS: MODEL STUDIES

Cross-hole acoustic tomography has become an extremely efficient technique for determining rock physical parameters and to detect heterogeneities in oil-bearing formations as well as discontinuities and dirt bands in coal layers. In particular, discontinuities like clay dikes, or faults can drastically slow down the production of a coal mine, and heterogeneities like shales can form permeability barriers to fluid flow in sand gas reservoirs. Evidence for mode conversion of borehole Stoneley waves (tube waves) to stratigraphically guided waves has been discovered in data from cross-hole acoustic experiments conducted by DOE in Colorado (Johnson and Albright, 1987). The results of these field experiments have shown that guided waves couple strongly to tube waves in both source and receiver boreholes. These guided waves traveling in low-velocity coal layers were found to be the primary mechanism for tube-wave coupling between the boreholes. In addition, physical model results have shown that these guide waves and the resulting tube waves disappear in the presence of a fault in the layers (Krohn, 1990). Cross-hole acoustic in comparison with other geophysical sensing techniques, holds the promise to confirm the continuity of heterogeneous formations between boreholes including the potential for inferring physical rock properties and determining whether shales within a sand gas reservoir are continuous.

We present the analysis of wave propagation in fluid-filled porous media based on Biot's theory (Biot, 1956a, 1956b, 1962) modified in accordance with the homogenization theory of Auriault (1980), and Auriault *et al.* (1985). Biot theory has been used in several applications of acoustic wave propagation in porous media such as porous rocks (Geertsma and Smit, 1961; and Johnston *et al.*, 1979), poroelastic soils (Attenborough *et al.*, 1986), and marine sediments (Stoll and Bryan, 1969; and Yamamoto, 1983). Turgut and Yamamoto (1990) conducted cross-hole acoustic pulse transmission experiments in saturated beach sand. They showed that the Biot theory predicts porosity and permeability using measured acoustic wave velocities (P and S) and attenuation.

In this chapter, the analysis of acoustic wave propagation is developed to calculate dispersion and attenuation curves of guided waves trapped in layered low-velocity formations based on the boundary value problem given by Parra (1991). The sensitivity of phase and group velocity as well as attenuation to input parameters (e.g., permeability, porosity, layer thickness, etc.) are analyzed to characterize guided modes produced by acoustic waves propagating in fully saturated layer media.

THEORY

Modified Biot Equations

The theory of propagation of acoustic waves in fluid-filled saturated porous unbounded media have been modified in accordance with the homogenization theory of Auriault (1980). The constitutive relations and balance equations of the modified Biot's theory in the frequency-domain are:

- Constitutive Stress/Displacements Laws:

- (i) Total stress of the porous medium

$$\tau_{ij} = 2\mu e_{ij} + \lambda e \delta_{ij} - \alpha p \delta_{ij} \quad (64)$$

- (ii) Stress in the pore fluid

$$\sigma_{ij}^{(f)} = -\phi p \delta_{ij} = \frac{\phi}{\beta} [\alpha \nabla \cdot u + \phi \nabla \cdot (U - u)] \delta_{ij} \quad (65)$$

- Momentum Balance Equation for total stress:

$$\frac{\partial \tau_{ij}}{\partial x_j} = \nabla \cdot \tau = -\omega^2 [\rho_s (1 - \phi) u + \phi \rho_f U] , \quad (66)$$

- Generalized Darcy Law:

$$w = \phi (U - u) = \theta [\omega^2 \rho_f u - \nabla p] , \quad (67)$$

where:

e	=	total volumetric strain of saturated porous medium
u	=	displacement in the solid frame
U	=	displacement in the pore fluid
τ_{ij}	=	total stress tensor of the saturated porous medium
$\sigma_{ij}^{(f)}$	=	stress in pore fluid
p	=	fluid pressure
ϕ	=	porosity
λ	=	$(1 - \phi) \bar{\lambda}$
μ	=	$(1 - \phi) \bar{\mu}$
α	=	$1 - K_b/K_s$
K_b	=	$(1 - \phi) (3\bar{\lambda} + 2\bar{\mu}) / 3$
ρ	=	$\rho_s (1 - \phi) + \rho_f \phi$
β	=	$(\alpha - \phi) / K_s + \phi / K_f$
K_f	=	pore fluid modulus

$$\text{and } \theta = -\kappa(\omega)/\omega,$$

in which $\kappa(\omega)$ is the frequency-dependent generalized Darcy coefficient assuming $\exp(-j\omega t)$ (Auriault *et al.*, 1985) and is given by

$$\kappa(\omega) = -\frac{j\phi}{\omega\rho_f} \frac{J_2[a(\omega\rho_f/\eta)^{1/2}]}{J_0[a(\omega\rho_f/\eta)^{1/2}]},$$

where η is the bulk viscosity of the saturant fluid and J_0 and J_2 are Bessel functions of order 0 and 2. The radius of the porous is given by $a = (8\bar{k}/\phi)^{1/2}$, where $\bar{k}$ is the intrinsic permeability of the material. The parameters $\bar{\lambda}$ and $\bar{\mu}$ are the Lamé coefficients of the porous skeleton or matrix, and K_g and K_b are the bulk grain modulus and the bulk frame modulus, respectively.

Eliminating between Eqs. (1)-(4) the vector wave displacement of the liquid, U , and the total stress tensor of the porous medium, τ_{ij} , the following coupled system of differential equations is obtained:

$$\mu \frac{\partial}{\partial x_j} \left(\frac{\partial u_i}{\partial x_j} + \frac{\partial u_j}{\partial x_i} \right) + \lambda \frac{\partial^2 u_j}{\partial x_i \partial x_j} + \omega^2 \rho u_i - \alpha \frac{\partial p}{\partial x_i} = 0 \quad (5a)$$

$$\theta \frac{\partial^2 p}{\partial x_j^2} - \alpha \frac{\partial u_j}{\partial x_j} - \beta p = 0; \quad (5b)$$

for $i, j = 1, 2, 3$,

where

$$\rho = \rho + \rho_f^2 \omega^2 \theta$$

$$\text{and } \alpha = \alpha + \rho_f \omega^2 \theta.$$

In Eq. (5) the direction of the layering is assigned to the vertical coordinate $z = x_1$, whereas the homogeneous (horizontal) directions are $x = x_1$, and $y = x_2$.

Solution of the Coupled Equations and the Theoretical Model

We seek solutions to be represented by plane harmonic waves of the form

$$(u_x, u_y, u_z, p) = (a_x, a_y, a_z, b) \exp j [(\xi z + kx) - \omega t]. \quad (6)$$

The solution given by Eq. (6) simplifies the analysis by suppressing the terms containing partial derivatives with respect to y in Eq. (5). Thus, substituting Eq. (6) into the system of Equation (5) yields a coupled system of equations given by

$$\begin{pmatrix} (\lambda + 2\mu)k^2 + \mu\xi^2 - \omega^2\rho & (\lambda + \mu)k\xi & j\alpha k \\ (\lambda + \mu)k\xi & (\lambda + 2\mu)\xi^2 + \mu k^2 - \omega^2\rho & j\alpha\xi \\ j\alpha k & j\alpha\xi & \theta k^2 + \theta\xi^2 + \beta \end{pmatrix} \begin{pmatrix} a_x \\ a_z \\ b \end{pmatrix} = 0, \quad (7a)$$

and the uncoupled equation given by

$$[\mu(-\xi^2 - k^2) + \omega^2\rho] a_y = 0. \quad (7b)$$

The functions a_x , a_z , and b are coupled through Eq. (7a) and are independent of a_y which must satisfy Eq. (7b). The functions a_x , a_z , and b are associated with fast compressional, slow compressional and vertically polarized shear waves. The function a_y in Eq. (7b) is associated with waves whose particle motion is polarized in horizontal plane (SH-waves).

The existence of nontrivial solutions for a_x , a_z , and b demands the vanishing of the determinant in Eq. (7a), i.e.

$$\text{Det}(k, \xi) = \theta(\lambda + 2\mu) \left(\xi^2 + k^2 - \lambda_1^2 \right) \left(\xi^2 + k^2 - \lambda_2^2 \right) \left(\xi^2 + k^2 - k_s^2 \right) = 0. \quad (8)$$

The solution of Eq. (8) yields

$$\xi_1^2 = \lambda_1^2 - k^2, \quad \xi_2^2 = \lambda_2^2 - k^2 \quad \text{and} \quad \xi_3^2 = k_s^2 - k^2,$$

where λ_1 and λ_2 are wavenumbers associated with the fast and slow compressional waves traveling in a fluid-filled porous medium, and k_s is the wavenumber associated with vertical polarized shear waves. These parameters can be obtained from the relations

$$\lambda_1^2 + \lambda_2^2 = k_p^2 - \alpha^2/\theta(\lambda + 2\mu) - \beta/\theta, \quad (9a)$$

$$\text{and} \quad \lambda_1^2 \lambda_2^2 = -k_p^2 \beta/\theta, \quad (9b)$$

where

$$k_s^2 = \omega^2 \rho / \mu, \quad (9c)$$

$$\text{and} \quad k_p^2 = \omega^2 \rho / (\lambda + 2\mu). \quad (9d)$$

For horizontal polarized shear wave motion the propagation characteristics are obtained directly from Eq. (7b) and is given by

$$\xi_{SH} = \left(\frac{\omega^2 \rho}{\mu} - k^2 \right)^{1/2}. \quad (10)$$

The theoretical model to be studied consists of a fluid-filled porous medium formed by N parallel horizontal layers overlying a uniform half-space. The j^{th} layer has a thickness $h_j = z_j - z_{j-1}$, and is bounded by the planes $z = z_{j-1}$, and $z = z_j$; its matrix properties are denoted by $\rho_s^{(j)}$, $K_s^{(j)}$, $K_b^{(j)}$, λ_j , μ_j , $\bar{k}_j$ and ϕ_j ; and its saturant fluid parameter are denoted by $\eta^{(j)}$, $\lambda_f^{(j)}$ and $\rho_f^{(j)}$.

The excitation is an incident plane P-SV wave formed by upgoing and downgoing waves, respectively. We assume the displacement vector and the pressure within the j^{th} layer is given by

$$u_j = I_j e_j \chi_j, \quad (11a)$$

where

$$u_j = \begin{bmatrix} u_x^{(j)} & u_z^{(j)} & p^{(j)} \end{bmatrix}^T, \quad (11b)$$

$$e_j = \begin{bmatrix} \exp(-\delta_j z^-), \exp(-\epsilon_j z^-), \exp(-\gamma_j z^-), \\ \exp(\delta_j z_j^+), \exp(\epsilon_j z_j^+), \exp(\gamma_j z_j^+) \end{bmatrix} \exp j(kx - \omega t) \quad (11c)$$

$$\text{and } \chi_j = \begin{bmatrix} \chi_1^{(j)} & \chi_2^{(j)} & \chi_3^{(j)} & \chi_4^{(j)} & \chi_5^{(j)} & \chi_6^{(j)} \end{bmatrix}, \quad (11d)$$

in which $z^- = z - z_{j-1}$; $z^+ = z - z_j$, $\delta_j = j\xi_1^{(j)}$, $\epsilon_j = j\xi_2^{(j)}$ and $\gamma_j = j\xi_3^{(j)}$.

The coefficients of the matrix I_j are obtained by applying the conditions for compressional wave motion and shear wave motion: $\nabla \times u_{\text{fast}} = 0$

$$\nabla \times u_{\text{slow}} = 0$$

$$\nabla \cdot u_s = 0$$

$$\theta \nabla^2 p - \beta p - \alpha \nabla \cdot u_k = 0; \quad (12)$$

$$\text{for } k = \begin{cases} 1 & \text{fast} \\ 2 & \text{slow} \end{cases}.$$

The vector displacements u_{fast} and u_{slow} are associated with fast compressional and slow compressional, and the vector wave displacement u_s is associated with vertical polarized shear waves. The resulting matrix of coefficients is given by

$$I_j = \begin{pmatrix} jk & jk & \gamma_j & jk & jk & -\gamma_j \\ -\delta_j & -\epsilon_j & jk & \delta_j & \epsilon_j & jk \\ t_j & s_j & 0 & t_j & s_j & 0 \end{pmatrix}, \quad (13)$$

where the new parameters are

$$\delta = \left(k^2 - \lambda_1^2 \right)^{1/2},$$

$$\epsilon = \left(k^2 - \lambda_2^2 \right)^{1/2},$$

$$\gamma = \left(k^2 - k_s^2 \right)^{1/2},$$

$$t = \alpha \lambda_1^2 / (\theta \lambda_1^2 + \beta),$$

$$\text{and } s = \alpha \lambda_2^2 / (\theta \lambda_2^2 + \beta).$$

Introducing the total stress components of the porous medium and the volumetric flux component of the fluid in its relative motion versus the solid, w , the stress-displacement matrix is obtained:

$$u_j = J_j e_j \chi_j, \quad (14)$$

$$\text{where } u_j = \left[u_x^{(j)}, u_z^{(j)}, p^{(j)}, \tau_{xz}^{(j)}, \tau_{zz}^{(j)}, w_z^{(j)} \right]^T.$$

The vectors e_j and χ_j are given by Eqs. (11c) and (11d), respectively. The matrix J_j is given by

$$\begin{pmatrix} jk & jk & \gamma_j & jk & jk & -\gamma_j \\ -\delta_j & -\epsilon_j & jk & \delta_j & \epsilon_j & jk \\ t_j & s_j & 0 & t_j & s_j & 0 \\ -2jk\delta_j\mu_j & -2jk\epsilon_j\mu_j & -\tilde{d}_j^2\mu_j & 2jk\delta_j\mu_j & 2jk\epsilon_j\mu_j & -\tilde{d}_j^2\mu_j \\ \tilde{r}_j^2\mu_j - \alpha_j t_j & \tilde{q}_j^2\mu_j - \alpha_j s_j & -2jk\gamma_j\mu_j & \tilde{r}_j^2\mu_j - \alpha_j t_j & \tilde{q}_j^2\mu_j - \alpha_j s_j & 2jk\gamma_j\mu_j \\ -(a_j + b_j t_j) & -(a_j + b_j s_j)\epsilon_j & jka_j & (a_j + b_j t_j)\delta_j & (a_j + b_j s_j)\epsilon_j & jka_j \end{pmatrix},$$

where

$$c_j^2 = \lambda_1^{(j)2} (\lambda_j + 2\mu_j) / \mu_j$$

$$\begin{aligned}
a_j &= \theta_j \omega^2 \rho_f^{(j)} \\
b_j &= -\theta_j \\
\tilde{d}_j^2 &= 2k^2 - k_s^{(j)2} \\
\tilde{r}_j^2 &= 2k^2 - c_j^2 \\
\text{and } \ell_j^2 &= \lambda_2^{(j)2} (\lambda_j + 2\mu_j) / \mu_j \\
\tilde{q}_j^2 &= 2k^2 - \ell_j^2.
\end{aligned}$$

In a similar fashion as the above analysis the displacement in the y-direction and the corresponding tangential stress component, $\sigma_t^{(j)}$, associated with SH-waves are obtained as a function of the wavenumber and frequency. That is, the displacement stress matrix for an incident plane SH-waves is given by

$$\begin{pmatrix} u_y^{(j)} \\ \sigma_t^{(j)} \end{pmatrix} = \begin{pmatrix} 1 & 1 \\ -\mu_j q_j & \mu_j q_j \end{pmatrix} \begin{pmatrix} \zeta_1^{(j)} e^{-q_j z^-} \\ \zeta_2^{(j)} e^{q_j z^+} \end{pmatrix}, \quad (15)$$

where

$$q_j = \left(k^2 - \frac{\omega^2 \rho_j}{\mu_j} \right)^{1/2}.$$

Dispersion Relations for Guided Waves in Layered Fluid-Filled Porous Media

The analysis of wave propagation in multilayer media is a mixed boundary value problem in which the motion is governed by the coupled wave equations in each layer, and the displacement and stress fields must satisfy the conditions of continuity of displacements and tractions across each interface, and the radiation conditions above the layered medium $z \leq 0$ and in the underlying half-space $z \geq 0$. To avoid precision problems caused by the growing exponential in the individual matrices for large wavenumbers, the global matrix method (Schmidt and Tango, 1986; Mal, 1988) is implemented, as an alternative to the traditional propagation approach, to determine dispersion and attenuation relations.

Using m as the label of layers, consider a full space consisting of an upper half space ($m = 1$), $(N - 1)$ layers of finite thickness ($m = 2, 3, \dots, N$) and a lower half space ($m = N + 1$). To facilitate the global matrix analysis, Eq. (14) is written in its vector and matrix form

$$\{u_m(z)\} = [J_m][e_m(z)]\{\chi_m\}, \quad (16)$$

where the factor $\exp(jkx - j\omega t)$ has been omitted. For $m = 2, 3, \dots, N$, the vector $[\chi_m]$ contains the unknown wave coefficients of wave propagation in both the positive and negative

z-directions in the layer. On the other hand, due to the radiation, the upper half space $m = 1$ contains only upgoing part of the wave coefficients and the lower half space $m = N + 1$ contains only the downgoing part. As a result, the corresponding $[J_1]$, $[J_{N+1}]$, $[e_1(z)]$ and $[e_{N+1}(z)]$ should take their reduced form.

In the global matrix formulation the boundary conditions at the layer interfaces are satisfied in the form of difference of field quantities across the interface. These conditions for the m^{th} interface can be represented by using Eq. (16) as follows:

$$[J_{m+1}][e_{m+1}(z_m)]\{\chi_{m+1}\} - [J_m][e_m(z_m)]\{\chi_m\} = \{u_{m+1}(z_m)\} - \{u_m(z_m)\}, \quad (17)$$

where z_m is the coordinate of the interface between m^{th} and $(m+1)^{\text{th}}$ layers.

An assembly of Eq. (17) for $m=2,3,\dots,N$, each consisting of six rows, and the above mentioned reduced version of Eq. (17) for $m = 1$, and $N + 1$, each consisting of three rows, forms a linear system of equations given by

$$[G]\{\chi\} = \{B\}. \quad (18)$$

In Eq. (18), the global matrix $[G]$ is composed of submatrices obtained from $[J_m]$, $[e_m(z_m)]$, and $[e_{m+1}(z_m)]$, for $m=1,2,\dots,N+1$. These submatrices contain coefficients associated with the physical properties of the skeleton and fluid of the layered poroelastic medium as well as the coordinates of the interfaces. The global vector $\{B\}$ represents the external excitations (e.g., incident plane waves or point sources) in the form of jump conditions at the interfaces z_m , $m=1,2,\dots,N+1$, which are obtained from the right hand side of Eq. (17). The global vector $\{\chi\}$ is a collection of the unknown wave coefficients with downgoing and upgoing waves in the layers. The dimension of the global matrix $[G]$ is $6N \times 6N$.

The next part of the analysis is to obtain dispersion and attenuation relations associated with the inplane guided waves in the layered fluid-filled porous medium. This is accomplished by setting the determinant of the global matrix $[G]$ in Eq. (18) equal to zero, which represents a free excitation state, $\{\chi\} = \{0\}$, i.e.

$$\det[G(k, \omega)] = 0, \quad (19)$$

and implicitly determines the wavenumber, k , of the guided modes at a specified frequency (see, e.g. Mal, 1988). A pole of the periodic equation (19) corresponds to a complex wave-number root, $k = \text{Re}(k) + j \text{Im}(k)$ at a given frequency ω . The phase velocity of the mode is equal to $\omega/\text{Re}(k)$, while its attenuation Q^{-1} is equal to $2 \text{Im}(k)/\text{Re}(k)$. In this case, the attenuation corresponds to the spatial Q^{-1} (i.e., a dimensionless parameter which describes the energy loss) given by Aki and Richards (1980). We have developed a method to search for the k values on the complex wavenumber plane that satisfies Eq. (19). This method consists of a combination of two-dimensional bracketing and minimization techniques, and is described below.

The Root-Finding Method of the Dispersion Equation

A well known numerical algorithm to find real roots of an equation is the bisection method. We have successfully extended this concept to find complex roots of Eq. (19). Our proposed algorithm consists of two steps. First, we divide the complex plane into grids and identify the grids that contain a root by a systematic bracketing scheme. Then we consider the magnitude of the function given by the left hand side of Eq. (19) and turn the root finding into a minimization problem.

Figure 1 illustrates a complex wavenumber plane which has been divided into grids arranged in rows and columns. Similar to the bisection method, the signs of both the real and imaginary parts at the four corners, A, B, C, and D, of a grid, are examined. A necessary condition for the existence of a complex root is that both the real and imaginary parts experience a sign change among the four nodes, generating two curves in the grid on which the real and imaginary parts vanish respectively. Clearly, a root resides in the grid if and only if these two curves intersect. Whether this condition is satisfied can be checked by book keeping the sign-change sequence along the contour of the four sides of the grid. When the sign change occurs alternately for the real and imaginary parts along this contour we are assured that a root does exist in the grid.

If a root is expected in the grid, minimization of the magnitude of the complex determinant given in Eq. (19) will lead to the location of the root. This is accomplished by the use of a standard downhill SIMPLEX method (Nelder and Mead, 1965). This procedure always converges and needs function evaluations only, not requiring derivatives. The two-dimensional version of SIMPLEX starts with a set of three arbitrary points from the boundary of the grid as guessed values. A series of steps is then taken by automatic reflection and contraction to modify the triangle. As a result, the vertex corresponding to the highest value of the magnitude of the determinant is replaced by a new point with a lower value in each step. This iteration continues until the location of the root is found within specified accuracy. Comparison between the complex roots obtained using these algorithms and the real roots calculated using standard root searching routine have demonstrated that the porosity and permeability of the skeleton and the fluid viscosity always cause shift of the roots in the complex wavenumber plane to the negative direction of the real and imaginary axes.

NUMERICAL EXAMPLES

The theory of guided waves is now applied to calculate dispersion and attenuation of acoustic waves trapped in a low-velocity layer. In particular, we calculate Rayleigh waves which are excited by a P-SV plane wave. In all the cases, a low-velocity sand surrounded by a high-velocity sandstone of lower permeability and porosity is analyzed. The physical properties of the solid matrix and fluid (saturant) are given in Table I. The first numerical example that we investigate is associated with the effect of the layer thickness of the low-velocity sand on the guided modes. Figures 2 and 3 show that as the layer thickness is increased the number of modes increases and they tend to overlap in the lower frequency range. The phase velocities are represented by dotted-dashed lines and the group velocities by dashed lines for guided waves propagating in a low-velocity layer. The solid lines represent the phase and group velocity of the fundamental Rayleigh mode. The guided-modes are

strongly coupled in the formation shear wave characteristics and have a low cut-off frequency which increases with each higher mode. The phase velocity of the guided waves in the low-velocity permeable zone has an initial value equal to the shear wave velocity of the host rock and decreases with frequency toward the vertically polarized SV-wave velocity layer. The attenuation variations as a function of frequency emphasize the predominant role of the host rock shear-wave attenuation at and around the cut-off frequency and the role of the low-velocity zone shear-wave attenuation at higher frequencies.

Figure 4 illustrates normal mode responses for varying low-velocity layer thickness. In this case, we select the mode 3 to demonstrate that as the layer thickness is decreased this mode becomes a higher-frequency mode, which can be excited within the 8-m thick layer at the Airy phase frequency of approximately 400 Hz (group velocity minimum). In order to excite the wave mode 3 within a layer of 2-m thickness it would require a source center frequency of about 1450 Hz at the spectral enhancement of such mode. Since the guided mode dispersion characteristics are associated with the layer thickness, these model results may be used to predict layer thickness changes within a low-velocity zone and to infer whether irregular bed thicknesses damp an acoustic signal (or reduce the number of guided wave modes).

Figures 2 through 4 show complicated dispersion characteristics of the Rayleigh-type waves, including two group velocity minima in each mode in nearly all cases. The minima occur at relatively constant separation in frequency and tend to always overlap the low group velocity segment of the next lowest mode. These Airy phases correspond to local maximum attenuations, characteristic of the fluid-flow interaction of the modes in the low-velocity permeable layer. A minimum attenuation occurs at the frequency associated with a maximum group velocity in all the modes, where enhanced transmission can occur. The attenuation corresponding to the Airy phase (which emphasizes fluid-flow interaction within the low-velocity permeable layer) may reduce the guided wave transmission characteristics in the permeable layer. For example, transmission characteristics of acoustic waves can be analyzed using the normal mode information for planning subsurface acoustic measurements. The most suitable frequency bands can be determined directly from the attenuation and group velocity of the normal modes. Specifically, the source center frequency of 1040 Hz can be used to excite the wave mode 2 (as shown in Fig. 2) associated with a low-velocity layer having those rock physical parameters given in Table I. This source center frequency in the frequency band of 500-2000 Hz corresponds to the minimum group velocity (Airy phase) of mode 2. Although normal modes having enhanced transmission characteristics can be excited at a source center frequency corresponding to the Airy phase, the attenuation curves in this case exhibit a local maximum which may reduce the amplitude of the seismic guided wave as it propagates in the low-velocity layer between large borehole separations. However, if the attenuation of the medium is not too high and source-detector separations are typical of those used for hydrocarbon reservoir characterization, acoustic guided waves can be detected by exciting the modes at the frequency associated with the Airy phase which value emphasizes fluid-flow interaction within the low-velocity permeable layer. This suggests that processing methods can be explored for extracting *in situ* rock physical properties (such as permeability and porosity) from group velocity and attenuation data.

Next, in order to analyze the permeability effect on the normal mode responses, we calculate dispersion and attenuation curves using the rock formation and saturant fluid properties given in Table I, as well as the permeabilities 0.05, 0.1, and 0.2 darcy for the low-velocity layer having a thickness of 4 m. The result given in Fig. 5 shows that the attenuation curves change significantly as the permeability is varied at the frequency corresponding to Airy phase. Especially, when the low-velocity layer becomes more permeable than the host rock the attenuation at the Airy phase increases. On the other hand, the attenuation of the guided mode is practically unaffected by those permeability values at the frequency corresponding to the maximum group velocity curve.

Figure 6 illustrates dispersion and attenuation curves for various layer porosities. As expected, the poroelastic dispersion and attenuation decrease when the porosity decreases. In particular, Fig. 6 shows that the Airy phase is slightly shifted toward the lower frequency as the layer becomes more porous than the host rock. In addition, there is a small increase in the attenuation curves as the porosity increases at frequency values greater than those associated with the minimum group velocity. That is, a decrease in porosity may decrease the efficiency of fluid-flow in the permeable layer.

CONCLUSIONS

The formulation of wave propagation in layered fluid-filled porous media, based on Biot and homogenization theories, has been presented that allows the study of the dispersion and attenuation characteristics of acoustic guided waves trapped in low-velocity regions using the global matrix method. The sensitivity of dispersion and attenuation to input parameters such as porosity and permeability were analyzed to characterize wave mode responses which may carry useful information associated with rock physical properties of the formation. In particular, the results demonstrate that the attenuation changes significantly as the permeability of the low-velocity layer is varied at the frequency corresponding to the Airy phase. Alternatively, the attenuation of the guided modes are practically unaffected by those permeability variations (in the low-velocity region) at the frequency corresponding to the maximum group velocity. The results of this work also show that the normal mode information (i.e., dispersion and attenuation) can be used for planning subsurface seismic measurements and for developing processing methods to predict *in situ* rock physical properties (such as permeability and porosity) from group velocity and attenuation data.

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TABLE I. Formation and Saturant Parameters

NAME	K_s (P_s)	ρ_s (kg/m^3)	v_p (m/s)	v_s m/s	v_f m/s	ρ_f kg/m^3	η (poise)	ϕ (%)	$\tilde{k}$ darcy
Upper and Lower Medium	37.9×10^9	2650	3550	2240	1500	1000	.01	10	.01
Layer	35×10^9	2600	1920	1270	1500	1000	.01	20	.2

Figure 1. Two-dimensional bracketing in a complex wavenumber plane.

Figure 2. Dispersion and attenuation of Rayleigh waves of a low-velocity permeable layer having a layer thickness equal to 2 m.

Figure 3. Dispersion and attenuation curves of normal Rayleigh waves of a low-velocity permeable layer having a layer thickness equal to 4 m.

Figure 4. Dispersion and attenuation of Rayleigh waves illustrating the normal mode 3 response for varying low-velocity layer thickness when the layer thickness is equal to 2, 4, and 8 m.

Figure 5. Dispersion and attenuation curves of Rayleigh waves illustrating the normal mode 3 response for varying in situ layer permeability. The low-velocity layer is water saturated with permeabilities successively equal to .05, .1, and .2 darcy.

Figure 6. Dispersion and attenuation curves of Rayleigh waves illustrating the normal mode 3 responses for varying layer porosity. The low-velocity layer is water saturated with porosities successively equal to 15, 20, and 30 percent.

CHAPTER 4

SEISMIC GUIDED WAVES IN LAYERED POROVISCOELASTIC MEDIA FOR CONTINUITY LOGGING APPLICATIONS: MODEL STUDIES

The gas industry is continuing to emphasize research and development efforts on new and advanced technology to improve reservoir descriptions through the producing life and development history of heterogeneous gas reservoirs. A very important aspect of this need is the ability to reduce the uncertainty of estimating probable reserves and lower the operating costs to recover incremental reserves in producing and depleted gas fields. Established methods for reducing uncertainty in heterogeneous reservoir compartments, such as VSP and crosswell techniques may enhance resolution, but they are currently not economically justifiable in onshore gas fields. Continuity logging using guided waves is an alternative approach to analyzing interwell seismic data to confirm the continuity of heterogeneous gas reservoir compartments; in particular, the continuity of sand and shale stratigraphy in gas reservoirs.

In the present work, the solution of a coupled system of differential equations based on Biot and homogenization theories is adapted to calculate seismic guided waves trapped in low-velocity layers. A unified representation of the medium that includes fluid-solid interactions and viscoelastic losses is incorporated in the general solution. The guided wave part of the vector wave field and fluid-pressure of the complete wave motion in layered dissipative media is verified and used to simulate dispersion and attenuation of seismic guided waves for continuity logging applications. The results of this work suggest that the multimode wave solution is appropriate to simulate seismic guided wave signatures to indicate continuity of layered earth structures in poroviscoelastic reservoirs. In particular, the normal mode information can be used for planning continuity logging surveys and for interpreting the corresponding seismic data. Further, fluid-pressure waveforms show that maximum amplitude normal modes can be detected at layer interfaces in fluid-filled porous media, and the corresponding Airy phase wave groups may carry information on the formation permeability.

Continuity logging using guided waves is a method commonly used in coal geophysics to determine discontinuities like clay dikes and faults in coal seams (Buchanan, 1983), and is currently being evaluated for reservoir continuity in hydrocarbon fields. In heterogeneous geological formations, continuity logging can be used to correlate lithology and stratigraphy as well as to map the continuity of producing horizons and to delineate the reservoir architecture. In this case the architecture includes structure of flow units, sealing surfaces and boundary surfaces that result from deposition and diagenesis. Several of these surfaces may have elastic properties that support the propagation of guided waves. As a consequence, guided waves may contain valuable information related to the continuity and rock physical properties of thin layers (Breitzke et al. 1987, and Breitzke 1992), and surfaces which serve either as conduits or barrier to fluid flow (Turpening et al. 1992, Krohn 1993; and Parra and Xu 1994). The continuity logging method of measurement consists of locating a seismic source in one borehole in a low-velocity layer and deploying a detector array in a second borehole. Detection of dispersive guided waves transmitted between wells indicates continuity of the layered structure in a hydrocarbon reservoir. These guided wave signatures are

either leaky modes or normal modes (or both) trapped in low-velocity layers by critical refraction of seismic waves at the layer interfaces.

In the current evaluation of the seismic guided wave technology for determining reservoir continuity in oil and gas fields, several issues are being addressed. One important issue is the detectability of guided waves in attenuated media at large well separation for continuity logging applications. A second issue is whether guided waves carry useful information associated with rock physical properties of the formation for reservoir characterization. In this paper we will address these two issues by examining transmission and detection of seismic guided waves in layered media. First, we present a brief theoretical discussion describing the basic Biot equations and analysis to calculate the guided wave motion in dissipative media. Second, dispersion and attenuation curves are calculated for simple low-velocity layers to produce synthetic seismograms for analyzing guided-wave propagation between boreholes in poroviscoelastic media. In particular, verification of the guided wave solution is presented.

FORMULATION

The analysis of seismic wave propagation for point sources in porous and permeable layered media based on Biot and homogenization theories requires the solution of the coupled system of differential equations (Parra 1991):

$$\mu \nabla^2 \mathbf{u} + (\lambda + \mu) \nabla \nabla \cdot \mathbf{u} + \omega^2 \rho \mathbf{u} - \alpha \nabla p = -F \delta(\mathbf{r} - \mathbf{r}_0) \quad (1a)$$

$$\theta \nabla^2 p - \beta p - \alpha \nabla \cdot \mathbf{u} = -V \delta(\mathbf{r} - \mathbf{r}_0), \quad (1b)$$

where

$\mathbf{u}$	=	displacement of the bulk rock matrix
p	=	the fluid pressure
λ	=	$K_b - 2\mu/3$
μ	=	$(1 - \phi) \bar{\mu}$
ρ	=	$\rho + \rho_f^2 \omega^2 \theta$
α	=	$\alpha + \rho_f \omega^2 \theta$
α	=	$1 - K_b/K_s$
ρ	=	$(1 - \phi) \rho_s + \phi \rho_f$
β	=	$(\alpha - \phi)/K_s + \phi/K_f$
K_f	=	pore fluid modulus
and θ	=	$-K(\omega)/j\omega$.

In this last equation $K(\omega)$ is the frequency-dependent generalized Darcy coefficient (Auriault, 1985) which is function of the fluid viscosity, η , the fluid density, ρ_f , and the intrinsic

permeability of the material, $\tilde{k}$. The parameter $\bar{\mu}$ is the rigidity of the porous matrix, and K_s and K_p are the bulk grain modulus and the bulk rock matrix modulus, respectively. In addition, ρ_s is the grain density, and ϕ is the porosity.

On the right hand side of (1a) we have a point force, and a volume source in the right hand side of (1b). The first equation becomes the elastic wave equation as the porosity approaches to zero. The advantage of using the above system of equations is for its compact form which is easy for determining the Green's function. The third important equation is the generalized Darcy law given by

$$w = \theta [\omega^2 \rho_f u - \nabla p] . \quad (2)$$

Equation (2) relates the displacement of the fluid relative to the solid with the displacement of the bulk solid matrix and the pressure gradient.

In order to approach the present boundary value problem for a point source in a layered medium, the Green's tensor in the wavenumber domain is used. This tensor contains simple coefficients associated with the physical characteristics of the saturated porous medium and leads directly to express the displacement-stress matrix, the fluid pressure, and the volumetric flux component, w , in terms of upgoing and downgoing waves in each layer (Parra 1991). To avoid precision problems caused by growing exponential in individual matrix for large wavenumbers, the global matrix method is implemented in the solution as an alternative to the traditional propagation approach. In the global matrix the boundary conditions are satisfied in the form of difference of field quantities across each layer interface (Parra and Xu 1994). Thus, the general solution is obtained in the form of a linear system of equations given by

$$[G]\{\chi\} = \{B\} , \quad (3)$$

where the main matrix $[G]$ contains the physical properties of the rock bulk matrix and fluid of the layered poroelastic medium. The vector $\{B\}$ contains the external excitations, and $\{\chi\}$ represents the unknown wave coefficients. These wave coefficients are used to determine the vector wave displacements and the fluid pressure. Alternatively, to obtain dispersion and attenuation relations, the determinant of the global matrix $[G]$ must be equal to zero, i.e.,

$$\det[G(k, \omega)] = 0 . \quad (4)$$

This equation is the period equation and its solution determines the complex wave number, k , of a normal mode at a specified frequency. A complex wavenumber root at a given frequency, ω , is

$$k = \text{Re}(k) + j\text{Im}(k) , \quad (5)$$

from which the phase velocity and attenuation are given by

$$c(\omega) = \omega/\text{Re}(k) \quad (6a)$$

and

$$Q^{-1} = 2 \operatorname{Im}(k) / \operatorname{Re}(k) . \quad (6b)$$

In order to simulate seismic guided waves, the guided wave part of the vector wave field and the fluid-pressure of the complete wave motion in a layered dissipative medium is used (Parra 1991). This general solution may be expressed as

$$I_m = \int_0^\infty T(k, \omega) J_n(kr) dk ; \quad (7)$$

for $n = 0, 1, 2$.

The kernel $T(k, \omega)$ of the above integral contains complex poles, which are slightly off the real axis toward the positive direction of the imaginary axis associated with the poroviscoelastic characteristics of the medium. The complete motion given by (7) can be separated into two main integrals. The first integral represents the body wave part of the solution (having a non-singular kernel), and the second integral represents the guided wave part. This second integral is evaluated analytically through contour integration and results in a sum of the residues (Kundu and Mal 1985). Thus, the guided wave part of the total motion is given by

$$I_g = \sum_{m=1}^M F[\operatorname{Re}(p_m)] H_m^{(1)}(p_m r) , \quad (8)$$

where

$H_m^{(1)}$	=	Hankel function
p_m	=	complex pole
M	=	total number of poles
r	=	source detector separation
and F	=	products of reflection and transmission matrices containing the physical properties and the geometry of the layered poroviscoelastic media.

The advantage of using the multimode wave solution lies in the fact that once accurate normal modes and leaky modes are obtained from (4), the guided wave part of the response for various source-detector locations can be calculated in an extremely efficient manner.

NUMERICAL EXAMPLES

The earth layered medium modeled as a lossless rock matrix saturated with a viscous fluid may only be applicable in exceptional cases. The porous medium must be considered as a whole rather than as a medium which takes into account the separate physical properties of fluid and solid (Biot 1962). As a consequence, a unified representation of the medium that may include the fluid-solid interactions and the viscoelastic losses associated with other physical mechanism (e.g., anelasticity of cracks, squirt flow, bubble squeezing, etc.) should be more appropriate. The analysis of a unified poroviscoelastic model was proposed by Biot (1962) and it was applied by Rasolafosaon (1991). The model given by Rasolafosaon con-

sists of a synthesis between Biot's poroelastic model and viscoelastic model in the context of plane wave propagation in porous viscoelastic media. We have incorporated this concept in the general layered poroelastic displacement and pressure solution by introducing viscoelasticity in the solid rock matrix. The dispersion and absorption of the unified model includes the quality factors of Q_p and Q_s for compressional and shear waves, respectively.

The analysis of guided waves is now applied to calculate dispersion and attenuation as well as synthetic seismograms of seismic waves trapped in a poroviscoelastic layered medium. In the following examples, a low-velocity sand having a thickness of 8 m, a permeability of 200 mD and a viscoelastic quality factor of 50 is analyzed. The physical properties of the solid rock matrix and fluid are given in Table 1.

Verification of the Guided Wave Solution

The vector particle displacement solution of the complete wave motion is formed by several integrals which must be solved numerically to produce synthetic seismograms for several source and detector positions. This task requires several hours of computer time, and in some cases may require several days of computer time. Therefore, an analytic solution is desirable to simulate layered earth responses. These goals can be accomplished using the closed form solution (8) to produce fluid-pressure and vector wave displacement seismic waveforms. This forward solution may be appropriate for predicting the behavior of guided waves that are observed in real seismic data to confirm continuity in gas reservoirs. For example, in gas fields the well separation may exceed 1/2 mile, as a consequence guided waves may be the only events present in the continuity logging seismic data.

To verify the guided wave part of the solution we first calculate the horizontal and vertical particle displacement components using the general solution given by Parra (1991) for a horizontal point force pulse signal centered at 400 Hz as shown in Figs. 1a and 1b. The source is located at 20 cm below the center of a fluid-filled poroviscoelastic layer of 8 m thickness and the detectors are located at a horizontal distance of 100 m from the source. The second detector is in the top layer interface at $z = 0$, and the eighteenth detector is in the bottom layer interface at $z = 8$ m. The traces in the seismograms correspond to outputs of detectors spaced 1/2 m apart. The seismograms show full waveforms extremely rich in modal seismic wave phenomena including body compressional waves, body shear waves, head waves, and reflections, as well as interface waves and guided waves. As was expected, the horizontal component waveforms exhibit large body compressional and shear waves and guided waves for detectors within the layer. Alternatively, the amplitude wavetrains associated with detectors at the layer boundaries and above and below the layer are smaller than those at the center of the waveguide. In addition, seismic events observed in the vertical component seismogram exhibit large amplitudes above and below the center of the layer in all the traces, except for the trace in the center where the source and detector are located practically at the same depth. This observation is consistent with the radiation pattern characteristics of a horizontal point force (White 1983; Balch and Lee 1984).

Next, we produce the guided wave part of the wave motion (vector wave field synthetic seismograms) using the same source and detector array configuration as well as the same

geometry and physical parameters as those used in the previous example. This requires to calculate dispersion and attenuation of normal modes trapped in the water-saturated poro-viscoelastic medium using (4). Figure 2 shows the phase velocities represented by dotted-dashed lines and the group velocities by dashed lines. The guided waves are strongly coupled in the formation shear wave characteristics and have a low cutoff frequency which increases with each higher mode. The phase velocity of the guided mode has an initial value equal to the shear wave velocity of the host rock and decreases with frequency toward the vertical polarized SV-wave velocity layer. Thus, multimode synthetic seismograms are produced using the dispersion and attenuation curves given in Fig. 2. A comparison between the guided wave seismograms given in Figs. 3a and 3b with the full waveform seismograms given in Figs. 1a and 1b indicates that the guided waves (normal modes) can be easily identified in the full waveform seismograms. In particular, the group velocity maximum and group velocity minimum (Airy phase) wavetrains can be clearly recognized in the horizontal and vertical component seismograms. The fast strong amplitude wave groups observed at a travel time delay of about 60 ms belong to the low-frequency part of the Rayleigh-type mode. On the other hand, the slow high-frequency wave groups observed at a time delay of about 92 ms belong to the Airy phase part of the Rayleigh-type modes.

In order to thoroughly verify the guided wave solution, the near zero-vertical offset seismic trace is selected from the horizontal component seismogram in Fig. 3a. In this case time-frequency representation is applied to the selected trace given in Fig. 4 to produce the contour plot given in Fig. 5. Superimposed on the contours is a theoretical Rayleigh-type wave dispersive curve of the second normal mode given in Fig. 2. This plot represents the group velocity signature for the normal mode observed in the trace associated with the low-velocity layer of thickness 8 m. That is, the group velocity of this normal mode starts at a velocity about the shear-wave velocity of the surrounding medium and decreases as a function of frequency to reach the group velocity minimum Airy phase, and then increases slowly towards the shear-wave velocity of the waveguide. Following this analysis, the corresponding full waveform selected from the horizontal component seismogram given in Fig. 1a is illustrated in Fig. 6.

The time-frequency plot shown in Fig. 7 exhibits two main events. These events are a direct P-wave arriving at 35 ms and having a peak frequency of about 380 Hz, and a dispersive event arriving at 75 ms, which appears to be a normal mode. The corresponding group velocity contour plot in Fig. 8 illustrates the normal mode event starting about at the shear-wave velocity of the surrounding medium at a cutoff frequency of 110 Hz, and decreases as a function of frequency to a group velocity minimum (Airy phase) corresponding to 260 Hz, and then increases towards the shear-wave velocity of the low-velocity layer. To examine the mode content and to verify this group identification we superimposed on the contour plot the theoretical Rayleigh-type wave dispersion curve of the second normal mode given in Fig. 2. These results indicate that the normal mode signature extracted from the full waveform trace and the guided wave calculated using the multimode wave solution are identical. Both signatures have the same Airy phase and dispersion characteristics. We expect that the only events observed at the far field will be guided waves and their presence will confirm the continuity of structural features in oil and gas reservoirs. As a consequence, the guided wave solution is appropriate to simulate normal modes and leaky modes. Since guided waves are

characterized by their dispersions, group velocity maximum and minimum wavetrains, these wave groups are the key elements to determine if guided waves are present in the structure of a reservoir.

Dispersion and Attenuation of Guided Waves

The guided wave comprising the normal mode is the result of constructive interference between upgoing and downgoing wave fronts internally reflected within the low-velocity layer. The most distinctive characteristics of seismic waves guided in the low-velocity layer is their dispersion. That is, the various spectral components of a broadband source signal propagate at different velocities so that the source waveform changes with distance away from the source. In particular, the most efficient propagation in the layered waveguide is associated with spectral zones where the phase velocity is relatively constant with frequency at a group velocity maximum or minimum (Airy phase).

In Fig. 2 several group velocity minimums are observed and their velocities varied between 920 m/s to 960 m/s. Alternatively, group velocity maximums are observed between about 1800 m/s to 1200 m/s. That is, if all the modes were excited by a broadband source, the group velocity maximum wavetrain would be much dispersive than the group velocity minimum wavetrain (Airy phase). The wave groups in the vicinity of group velocity minima are the most coherent components of the transmitted guided wave and, therefore result in enhanced signal amplitudes.

Figure 2 also presents attenuation curves of each Rayleigh-type mode as a function of frequency, illustrating the predominant role of the host rock shear-wave attenuation at and around the cutoff frequency and the role of the low-velocity zone shear-wave attenuation at higher frequencies. In addition, attenuation curves are characterized by relaxation peaks which correlate with minimum group velocity Airy phases. The relaxation peaks are associated with the poroviscoelastic characteristics of the formation. The usefulness of the unified poroviscoelastic model can be demonstrated by calculating a fluid-pressure synthetic seismogram using the dispersion and attenuation curves given in Fig. 2. Here the normal mode that will be excited the most by a broadband source (having a peak frequency of 1000 Hz) is represented by solid lines at the Airy phase frequency of 1000 Hz.

Figure 9 shows the fluid-pressure synthetic seismogram produced by a horizontal point force located in the middle of the water-saturated layer of 8 m thickness. A 24-detector array having a detector spacing of 0.5 m is placed at 100 m from the source well. The seismic traces in the middle of the layer exhibit coherent group velocity maximum wave trains at 72 ms (i.e., a group velocity of 1.38 km/s). In addition, a coherent Airy phase wave train is observed in the traces at 107 ms. The average amplitude of the group velocity maximum wavetrain is greater than that of the Airy phase wave group. The decrease in amplitude is caused by the dissipation of the medium between wells. Furthermore, stronger wave groups are observed at detectors near and at the layer interfaces together with high reflection amplitudes. The wave groups are associated with the permeability contrast between the layer and the surrounding medium (Parra and Xu 1994). In this case, the macroscopic flow is not negligible, the discontinuity (layer interface) induces pressure gradients which modify the

global flow. That is, the fluid-pressure in the layered medium is due to the Biot mechanism included in the unified poroviscoelastic model. Since the wave group amplitudes observed in the fluid-pressure seismogram increase at the layer interfaces, where fluid-flow interaction occurs, the Airy phase wave group may contain information associated with the permeability of the low-velocity layer. These results suggest that detectors placed in the middle and at the layer boundaries can be used to measure the best normal mode guided wave amplitudes using a common source crosswell configuration in a fluid-filled saturated porous medium. For comparison, Fig. 10 illustrates the synthetic seismogram of the horizontal component of the vector wave displacement for the dry bulk rock matrix. In this case, the wave group amplitude and reflections do not increase at the layer interfaces.

Since a source center of frequency of 1000 Hz is used in the examples given in Figs. 9 and 10, the modes excited at this frequency are attenuated. In particular, the Airy phase wave group amplitudes are much smaller than the amplitudes of group velocity maximum wavetrains. The wave field response formed by modes having less attenuation are those generated by a low-frequency source. For example, the synthetic seismogram produced by a point force having a source center frequency of 400 Hz (see Fig. 3a) exhibits a coherent Airy phase wavegroup at the well separation of 100 m. However, the Airy phase wavetrain becomes strongly attenuated at larger well separations. This phenomena can be illustrated by synthetic seismograms at well separations of 200 m and 400 m which are depicted in Figs. 11 and 12, respectively. The waveforms detected at 200 m have coherent wave groups showing the effect of the source radiation pattern, and more dispersive Airy phase wave groups than those detected at 100 m well separation (see Fig. 3a). On the other hand, at 400 m well separation, we observe a group-velocity maximum event which does not exhibit the effect of the source radiation pattern. The amplitudes of the Airy phase wavetrains decrease faster than the amplitudes of the group velocity maximum wavetrains for large well separations. In fact, to identify the presence of the Airy phase wave group in the 400 m well separation seismogram the waveforms were amplified about 70 times. These examples demonstrate how the Airy-phase amplitude decreases as the well separation increases for waves propagating in a poroviscoelastic medium having a low quality factor of $Q_v = 50$.

Energy losses of seismic guided waves propagating between detector wells two and three can be predicted using the spectra of two zero-vertical offset waveforms given in Figs. 11 and 12. The detector wells two and three are located at the horizontal distances of 100 m and 400 m from the source well, respectively. The spectra and the attenuation curve are shown in Figs. 13 and 14. The spectra suggest that at frequencies less than 200 Hz there are 10 to 15 dB difference which correspond to a wave amplitude reduction of 5 to 6 times. Alternatively, at frequencies greater than 200 Hz the energy losses are much greater. In this example the attenuation was calculated using the amplitude ratio method and the quality factor was obtained using least-square curve fitting. The value of $Q_v = 47$ is closed to the value of 50 used to generate the seismogram. This analysis illustrates the type of guided wave amplitude and signatures that may be expected in a medium having an attenuation of 15 dB/m in continuity logging surveys. In addition, these results confirm the multimode wave solution and its applicability for modeling continuity logging data.

CONCLUSIONS

The analysis of wave propagation in layered fluid-filled porous media based on Biot theory has been presented to calculate dispersion and attenuation of normal modes and synthetic seismograms containing guided waves trapped in low-velocity poroviscoelastic regions. The numerical simulations were conducted to demonstrate seismic guided wave propagation characteristics in poroviscoelastic media for planning continuity logging surveys. The results demonstrate that the normal mode signature extracted from a full waveform trace using time-frequency analysis and the corresponding guided wave calculated with the guided wave solution are identical. This suggests that the guided wave solution is appropriate to simulate normal mode signatures to confirm continuity of layered structures in hydrocarbon reservoirs. In addition, the results of this work suggest that maximum amplitude normal modes can be detected at the interfaces (boundaries) in permeable media and Airy phase wave groups may carry useful information on the fluid-flow characteristic of the formation such as permeability.

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NAME	K_s (GP_a)	ρ_s (kg/m^3)	v_p (m/s)	v_s m/s	ϕ (%)	$\tilde{k}$ darcy	v_f m/s	ρ_f kg/m^3	η (poise)
UPPER & LOWER MEDIUM	37.9	2650	3550	2240	10	.01	1500	1000	.01
LAYER	35	2600	1900	1270	20	.2	1500	1000	.01

Table 1. Formation and saturant parameters for a water saturated medium.

Figure 1. Common-source full waveforms for a water saturated poroviscoelastic medium. Source center frequency = 400 Hz; well separation = 100 m. The quality factor $Q_v = 50$ and the waveguide permeability is 200 mD. (a) Horizontal displacement component-seismogram. (b) Vertical displacement component-seismogram.

Figure 2. Dispersion and attenuation curves for Rayleigh-type modes in a water saturated poroviscoelastic layered medium. The quality factor $Q_v = 50$ and the waveguide permeability is 200 mD.

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Figure 3. Common-source seismograms for Rayleigh-type modes in a water saturated poroviscoelastic medium. Source center frequency = 400 Hz; well separation = 100 m. The quality factor $Q_v = 50$ and the waveguide permeability is 200 mD. (a) Horizontal displacement component. (b) Vertical displacement component.

Figure 4. U_x component seismic guided wave in a poroviscoelastic medium. Source center frequency = 400 Hz. Quality factor $Q_v = 50$, and well separation = 100 m. The source is at 4.2 m and the detector is at 4 m.

Figure 5. Dispersion of the seismic guided wave given in Fig. 4.

Figure 6. U_x component full waveform seismic wave in a poroviscoelastic medium. Source center frequency = 400 Hz. Quality factor $Q_v = 50$, and well separation = 100 m. The source is at 4.2 m and the detector is at 4 m.

Figure 7. Time-frequency analysis of the full waveform seismic wave given in Fig. 6.

Figure 8. Dispersion of the full waveform.

Figure 9. Fluid-pressure synthetic seismogram for Rayleigh-type modes in a water saturated poroviscoelastic medium. The source center frequency is 1000 Hz and the layer thickness is 8 m. $Q_v = 50$.

Figure 10. U_x component synthetic seismogram for Rayleigh-type modes in a water saturated poroviscoelastic medium. The source center frequency is 1000 Hz and the layer thickness is 8 m. Well separation of 100 m. $Q_v = 50$.

Figure 11. U_x component synthetic seismogram for Rayleigh-type modes in a water saturated poroviscoelastic medium. The source center frequency is 400 Hz and the layer thickness is 8 m. Well separation of 200 m. $Q_v = 50$.

FIGURE 12.

Figure 12. Ux component synthetic seismogram for Rayleigh-type modes in a water saturated poroviscoelastic media. The source center frequency is 400 Hz and the layer thickness is 8 m. Well separation of 400 m. $Q_v = 50$.

Figure 13. Spectra of zero-vertical offset seismic waveforms from seismograms given in Figs. 5 and 7.

CHAPTER 5

INTERWELL SEISMIC LOGGING FOR RESERVOIR CONTINUITY AT THE GYPSY TEST SITE, OKLAHOMA

The gas and oil industry is continuing to emphasize research and development efforts on advanced technology to improve reservoir description through the producing life and development history of heterogeneous hydrocarbon reservoirs. Existing methods for reducing uncertainty in heterogeneous reservoir compartments, such as VSP and multiple offsets crosswell techniques may enhance resolution, but these methods are not economically justifiable in oil and gas fields. Continuity logging using guided waves is an alternative approach to analyzing interwell seismic data to confirm the continuity of heterogeneous hydrocarbon reservoir compartments; in particular, the continuity of sand and shale stratigraphy in reservoirs. The present work includes a study of rock physical properties and seismic data for hydrocarbon reservoir continuity using interwell logging techniques. Processing methods are used to delineate the reservoir structure and to relate seismic wave dispersion with rock physical parameters. The principal application of this study is to identify and analyze the propagation characteristics of guided waves in the Gypsy-fluvial sandstone formation at the Gypsy test site in Oklahoma. The integration of well logs and interwell seismic data using guided waves indicated the presence of two low-velocity clean sands to be continuous in the Gypsy sandstone interval. In addition, the results of this work confirmed the presence of a continuous shale above the sandstone. The correlation of sand and shale stratigraphy with well logs and leaky modes is consistent with crosswell and pulse test data acquired by BP Exploration. These pulse tests indicated that the sands are in pressure communication between the wells as well as the existence of an impermeable barrier between the top and the bottom of the sand units in the Gypsy sandstone reservoir.

In this chapter we integrate rock physical properties and interwell seismic data for hydrocarbon reservoir continuity using interwell seismic logging techniques (Krohn, 1992). In particular, analyze methods of processing to delineate the reservoir architecture and to relate dispersion to rock physical properties of the formation. The architecture includes structure of flow units, sealing surfaces and boundary surfaces that result from deposition and diagenesis. Several of these surfaces may have elastic properties that will support the propagation of guided waves. In fact, it is recognized that guided waves contain valuable information related to the continuity and rock physical properties of thin layers and surfaces which serve either as conduits or barriers to hydrocarbon flow (Turpening et al, 1992; Krohn, 1993; and Parra and Xu, 1994).

The principal application of this study will be to identify and analyze the propagation characteristics of guided waves in the Gypsy-fluvial sandstone formation at the Gypsy test site in Oklahoma where BP has drilled six wells. This test site is formed by highly heterogeneous water-saturated formations, and is thoroughly documented by well log records, detailed lithological descriptions, vertical seismic profiles, 3D seismic surveys, crosswell seismic surveys, etc.

The original purpose of Gypsy test site study was to relate sedimentary features to reservoir flow units as well as to provide a data base for assessing reservoir characterization technology and test reservoir properties in controlled field conditions. In addition, since the site is a water saturated aquifer it removes complications of multi-phase fluid systems from the goal of establishing a methodology for guided wave propagation. The Gypsy site is a fluvial depositional environment of Pennsylvania age in the Vamoosa formation located in northeastern Oklahoma, and is formed by channel fill deposits. The Gypsy sandstone interval is a series of stacked channel-fill, point bar deposits that are surrounded and subdivided by flood plain deposits. An overall fining upward trend is present within each individual sandbody (Doyle et al., 1992).

Regionally, the Gypsy sandstone dips to the northwest. In the vicinity of the Gypsy test site the dip is to the northwest at 1° . No faulting has been detected at the site, but a regional fracture system has been identified. The fracture trends are north-south and east-west. The geological cross-sections suggest that the most continuous zones for guided wave propagation are the shales bounding the Gypsy interval. Alternatively, the shales that separate the channel sandbodies appear not to be continuous between two wells. To properly apply the guided wave technology we have conducted petrophysical characterization of the Gypsy interval and integration of the core and log data, and further effort to delineate the geometries of the shales and the geometry of some channel sandbodies. The concepts and applications of continuity logging using guided waves are given in Krohn (1990), Krohn (1992), and Buchanan (1983).

The geological and petrophysical data was analyzed to determine the following rock properties: lithology, clay content, porosity, permeability, water saturation, layer geometry, etc. In addition, compressional and shear wave velocities and density associated with selected geological units of interest were derived. In addition, geological cross-section between selected wells were compiled using the well logs and core information. These geological cross-sections were integrated with existing interwell seismic data recorded by BP Exploration. Select full waveforms were correlated with geological units of interest, processed and interpreted to identify dispersion characteristics which may be associated with seismic guided waves. Guided wave events were extracted from the interwell seismic data using time-frequency representation techniques for reservoir continuity applications (Parra and Zook, 1992; Cohen, 1989).

Overview and Geological Setting

The Gypsy test site is located in Pawnee County, Oklahoma, about 35 miles northwest of Tulsa. The site lies within the Central Oklahoma Platform (Landes, 1970). The test site consists of six test holes (1300 to 1400 ft deep) arranged in a five-acre, inverted five spot pattern with a sixth observation well midway between two of the others. The wells penetrate Pennsylvanian sandstones and shales. Figure 1 illustrates the location of the six wells at Gypsy.

The focus of this study is the interval from 800 to 1100 ft, which is the lower part of the Virgilian age Vamoosa Formation and the upper half of the Missourian age Tallant

Formation. The lower-most sandstones in the Vamoosa Formation (informally known as the Gypsy sandstone) and the upper-most portion of the Tallant Formation were cored in each well. To investigate the seismic wave propagation between wells we selected the cross-section through the Dallas 7-9, 7-5, and 7-7 which contains the calculated lithology and the compressional wave velocity logs as shown in Figure 2. In this cross-section there are six selected intervals having high-degree of probability of being continuous. Five high-velocity (HV) zones and four low-velocity (LV) zones are correlated across the section. The Gypsy sandstone interval (HV4) is composed by stacked channel sands, mudstones, and siltstones which indicates an uncertain and very difficult correlation of individual sandstone units between wells.

The Gypsy interval (HV Zone 4) was deposited as a series of stacked channel-fill, point bar sandstones that incised into and are now surrounded and subdivided by floodplain and crevasse splay deposits (Doyle, et al., 1992). The number of Gypsy channel sandbodies in each well varies from 2 to 5, and with sand thickness ranging from 10 to 50 ft.

The Gypsy interval was deposited on a regional unconformity, represented by terrestrial shales, that developed on a marine sandstone in the Tallant formation (Doyle and Sweet, 1992). This sandstone is a lower to middle shoreface sequence with root traces on its erosional upper surface (Doyle and Sweet, 1992). The Gypsy interval is overlain by a marine flooding surface consisting of shales (e.g. LV Zone 4) and low porosity, calcareous siltstone/sandstone (e.g. HV Zone 5). These marine units occur in all six wells and are inferred to be continuous between wells.

It is very difficult to correlate individual Gypsy channel sandbodies between wells. Only the base of the Gypsy interval, which as the base of a paleovalley is a sixth-order bounding surface, is correlateable between all the wells. The units that can be correlated with a high degree of confidence are those surrounding the Gypsy interval: the marine flooding surface (HV Zone 5 and LV Zone 4) overlying the Gypsy interval and several of the marine sandstones and shales in the Tallant Formation (HV Zones 1, 2, and 3; LV Zones 1 and 2).

Doyle and Sweet (1992) identified seven lithofacies in the Gypsy interval: mudclast-rich sandstone, cross-bedded sandstone, planar-laminated sandstone, ripple-laminated fine sandstone, sandstone with soft sediment deformation structure, red mottled silty mudstone, and bioturbated, shaly sandstone. They interpreted the red mottled silty mudstone facies to be floodplain deposits because of the presence of interbedded paleosol horizons. Some mudstones and siltstones occur within channel sandbodies as abandoned channel fill deposits. The bioturbated, shaly sandstones occur within the floodplain deposits and those that are ripple-laminated are probably splay deposits. The other five facies occur in the channel-fill sandstones.

Channel architectural elements are difficult to distinguish in either the cores or the logs. Although channel bases (fifth-order bounding surfaces) are identifiable, lower order bounding surfaces are difficult to define with certainty in the cores (Doyle and Sweet, 1992).

Thomas (1990) conducted a petrographic analysis of some of the Gypsy sandstones in two wells. The Gypsy sandstones are moderately sorted, very fine to medium grained sublitharenites. Quartz grains are the dominant framework component, with igneous and metamorphic rock fragments common. Clay laminations and mud clasts are common in most of the sandstones and mud clasts. The clays are illite and kaolinite, with lesser amounts of chlorite. Quartz overgrowths, calcite, iron-rich calcite, and iron-rich dolomite are the cements. Most of the porosity is intergranular, but some is moldic, created by the partial to complete dissolution of framework grains (Thomas, 1990).

The core porosity of the Gypsy sandstones averages 19.7% and permeability averages 558 md (Doyle, 1992). Porosity values reach as high as 26% and permeability as high as 2000 md.

INTERPRETATION OF WELL LOGS AND INTERWELL SEISMIC DATA

This part of the paper is a well log and seismic analysis of the geological cross section between wells 7-5 and 7-7 at the Gypsy test site (see Figure 2). The integration of the geology and well logs with the crosswell data indicates the presence of two low-velocity sands to be continuous in the Gypsy sandstone interval at the depths of 966 to 974 ft (see Figure 3). These two sand units exhibit high porosity (26%) and high permeability (2 darcy). Also, Figure 3 illustrates gamma ray and density logs as well as a comparison between the vertical compressional wave velocity log and the horizontal compressional wave velocity log determined from zero-vertical offset seismic waveforms. This comparison shows strong vertical velocity anisotropy in the shale, and weak anisotropy in the clean sands.

In order to better characterize the interwell seismic response of the target zones of interest the horizontal compressional wave velocity log is plotted separate in Figure 4. In general, the horizontal velocity log (see Figure 4), the density log, and the gamma ray log have similar responses. In particular, a bottom low-velocity zone can be easily identified in the horizontal velocity log at the depth interval of 970-976 ft, and when the source and detectors are located within this zone the measured interwell seismic waveforms exhibits dispersion characteristics associated with guided waves trapped in low-velocity layers. In addition, the logs in Figures 5 and 6 show the Gypsy sandstone interval which is composed of a series of stack channel of sandstones, mudstones and siltstones. In Figure 5, there are two sand units at depths of 966 ft and 974 ft in wells 7-5, both having the same petrophysical characteristics. The porosity and permeability of both sands are about 25% and 2000 md, respectively. In addition, these units are low-density clean sands within a low-velocity zone which has the appropriate velocity contrast that may support propagation of guided waves. Below the second clean sand at a depth of about 978 ft there is a sand which has porosity and permeability values less than those associated with the clean sand units. This type of sand appears to be contaminated with mudstones and siltstones and correlates with a compressional wave velocity high in the horizontal velocity log. Below this sand there is a low-velocity shale which separates the Gypsy sandstone interval and the Tallant sandstone formation.

In Figure 6, the Gypsy interval is homogeneous between the depths of 950 ft and 990 ft in well 7-7. In this zone the sand thicknesses and the well log signatures are different from

those intercepted by well 7-5. These heterogeneous conditions makes the interpretation very difficult for establishing continuity within the units of the Gypsy interval between wells 7-7 and 7-5 using only well logs and the petrophysical information. However, the connectivity of the clean sand beds between wells 7-7 and 7-5 have been already demonstrated by crosswell and pulse test data acquired by BP Exploration at Gypsy. The results of these tests indicated that many of the sands are in pressure communication with many of the sands in the receiver well 7-5. In addition, the pulse test data have confirmed the existence of an impermeable barrier between the top and bottom sand units (Turpening *et al.*, 1992).

Continuity of Shales Using Normal Modes

A different method to confirm continuity of the sand and shale units is the interwell seismic method using guided waves. In fact, the most continuous low-velocity zones for seismic guided wave propagation appears to be the shales bounding the Gypsy interval. To confirm continuity of these low-velocity zones or any other zone we have used time-frequency analysis (Cohen, 1989). This processing technique can be applied to extract guided waves (i.e., normal modes and leaky modes) from full waveforms and determine how such signals may be useful to interpret rock physical properties from waveguides and the surrounding medium. The applicability of the time-frequency analysis for guided wave characterization is illustrated for a zone of interest which is studied in detail using interwell seismic data: a shale layer at a depth of 874 ft. Figure 7 shows the time-domain signal at a depth of 874 ft in the shale layer. The group velocity versus frequency plot is illustrated in Figure 8. Only data below the approximate shear wave velocity is given.

A dispersive event having a signature similar to that of a normal mode is clearly evident in the range of 400-800 Hz. The group velocity starts at a shear wave velocity of about 2300 m/s and decreases as a function of frequency toward a group velocity of 1750 m/s (Airy phase). The values of 2300 m/s and 1750 m/s are greater than the vertical polarized shear wave velocities observed in the shear velocity log given in Figure 9 (i.e., 1540 m/s and 1210 m/s). That is, the group velocity of the dispersive event is associated with waves traveling between wells 7-7 and 7-5 in the horizontal direction. We believe that compressional waves radiated by the source are reflected and converted to shear waves at a sandstone shale boundary above the source (which is located at 874 ft) to return to the waveguide and excite dispersive normal modes [i.e., incident P and SV waves generate qSH waves in the presence of anisotropy (Keith and Crampin, 1977)]. To prove this observation we estimate approximately the appropriate rock physics parameters to calculate theoretical dispersion and attenuation curves (Parra and Xu, 1994) by assuming that the modes were excited by a plane shear wave, as shown in Figure 10a.

In addition, energy coupled to the low-velocity shale as tube wave excited Rayleigh-type modes which group velocity starts at the vertical polarized shear wave velocity of the sandstone (about 1540 m/s) and decreases as a function of frequency to the vertical polarized shear wave velocity of the shale (about 1210 m/s). To analyze this concept, theoretical normal modes were calculated using the compressional and vertical polarized shear wave velocities obtained from well logs and zero-vertical offset interwell seismic data. The Rayleigh-type and attenuation curves are illustrated in Figure 10b.

To examine the mode content and to verify the group identification we superimposed on the upper part of contour plots of Figure 8 the theoretical wave dispersion curve of the third mode given in Figure 10a. Also, superimposed on the lower part of the contours is a Rayleigh-type wave dispersive curve of the second normal mode given in Figure 10b. The dispersion characteristics of this second event in the range of 400-700 Hz is not evident from the contours of Figure 8. The main reason is that the energy radiated by the cylindrical bender source near the low-frequency part of the spectrum below 400 Hz is practically zero. Therefore, seismic events are not completely excited and their dispersions are not fully characterized at frequencies near 400 Hz. However, signal enhancement can be obtained by plotting the group velocity contours in dB as shown in Figure 11. Superimposed on the contours is the theoretical group-velocity curve given in Figure 10b.

Continuity of Sands and Shales Using Leaky Modes

In addition, we have searched for leaky modes since they are more detectable than normal modes when hydrophones are used (Krohn, 1992). For example, in Figure 12, the time-frequency analysis shows the normal modes are visible with an Airy phase near a group velocity of 1750 m/s, although its power is about 10 dB below the energy of the faster wave. The faster event is broadband, having some interesting structure, which we believe to be a leaky mode (Laster et al, 1965; and Stalmach and Bremaecker, 1973). In many of the plots, it is difficult to readily observe both the leaky and normal modes at the same time; this particular trace shows them more clearly than most. Furthermore, leaky modes are also indications of propagation through a continuous layer (Krohn, 1993).

The continuity of thin shales and sands in the Gypsy sandstone reservoir between wells 7-7 and 7-5 can be confirmed using leaky modes. This can be done first selecting the appropriate low-velocity intervals from the horizontal compressional wave velocity log illustrated in Figure 4. Low-velocity zones are observed between depths of 960 ft and 980 ft, and between depths of 990 ft through 1060 ft. Second, we selected a common source fan containing waveforms produced when the seismic source was placed at the depth of 970 ft in the low-velocity zone of interest in well 7-7. We extracted several waveforms recorded by detectors placed at depths of 960 ft through 990 ft and 2 ft intervals. These waveforms were extracted from the common source fan shown in Figure 13. The waveforms were processed using time-frequency analysis to determine group-velocity contours which are given in Figures 14 through 17.

Figure 14 illustrates group velocity contours associated with detectors located at depths of 960 ft through 968 ft. In particular, the contours produced from the waveform recorded at a depth of 960 ft exhibit a dispersion signature similar to that of a leaky mode which starts about a compressional wave velocity of 3100 m/s. This velocity is less than the group velocity of the surrounding medium and it decreases as a function of frequency to a minimum at 600 Hz, then increases towards the compressional wave velocity of the target zone. This weak signature in the range of 450-900 Hz indicates a loss of energy towards the surrounding medium as the waves propagate between wells 7-7 and 7-5. Most of the energy appear to be concentrated in the frequency range of 900-1500 Hz. Although, the leaky mode signature is weak its dispersion suggests continuity of the thin shale zone at a depth of 960 ft.

Alternatively, the detectors placed at depths of 962 ft and 964 ft have recorded waveforms containing no dispersion characteristics associated with a transition zone formed by a combination of shales and sands. However, group velocity contours of waveforms recorded at depth of 966 ft exhibits a dispersion signature which correlates with a low-velocity clean sand.

Figures 14 and 15 show group velocity contours of waveforms recorded by detectors at depths between 968-978 ft. No dispersion characteristics are observed between 968-972 ft. The 4 ft interval of shale intercepted by well 7-5 corresponds to a shale of less than a foot of thickness within the sand interval of well 7-7. As a consequence, we may conclude that these shales are not connected between wells 7-5 and 7-7.

However, dispersive group velocity signatures generated from waveforms recorded at depths between 974 ft and 978 ft are observed in Figure 15. These signatures correspond to a thin continuous low-density and low-velocity clean sand which is intercepted by both wells 7-7 and 7-7. Below this continuous sand layer between depths of 980 through 990 ft the gamma ray and density logs show a transition zone of shale and sand associated with a horizontal high compressional wave velocity. This zone is a discontinuous heterogeneous shale/sand sequence associated with seismic responses having no dispersions, as shown in Figure 16. Finally, Figure 17 shows dispersive group velocity signatures associated with a low-velocity shale zone in the 992-998 ft interval. In this case, the group velocity contours show dispersion characteristics similar to those of continuous low-velocity zones observed in previous examples.

In summary, the observation of group velocity contour plots and well logs suggest that the energy radiated by the source is coupled to the low-velocity sand at the depth of 970 ft in well 7-7. This coupled energy excited leaky modes in the thin sands and shales within the lower Gypsy sandstone. In particular, energy leakage from the low-velocity sand excited modes in the bottom low-velocity shale zone. As a consequence, the shales at the depth of 960 ft and 996 ft are permeability barriers that may prevent fluid-flow between the upper and the lower Gypsy sandstone reservoir.

CONCLUSIONS

Since guided waves are observed in the interwell seismic logging data at Gypsy, seismic guided waves can propagate in heterogeneous sandstone reservoirs. The correlation of the lithology with well logs and interwell seismic data indicates the origin of the waveforms associated with continuous and discontinuous geological units. Interwell seismic logging techniques using guided waves provides useful information for continuity of sandstone and shale stratigraphy in heterogeneous fluvial sandstone reservoirs. In particular, the continuity of low-velocity clean sands in the Gypsy sandstone interval using guided waves was confirmed and it was consistent with the previous results of crosswell and pulse test data. Leaky mode dispersion characteristics were associated with thin continuous mudstones and sands within the Gypsy interval. Alternatively, normal modes were associated with continuous shale boundary surfaces above the upper Gypsy sandstone interval.

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Figure 1. A plane view of the six wells at the Gypsy borehole test facility in Cleveland, Oklahoma.

Figure 2. A geological cross-section through the Dallas 7-9, 7-5, and 7-7.

ZERO VERTICAL OFFSET WAVEFORMS IN WELLS SPACED 105 m

GYPSY WELL 7-5

Figure 3. Zero-vertical offset seismic waveforms produced between wells 7-5 and 7-7 showing gamma ray, density and compressional wave velocity logs.

GYPSY WELL 7-7

ZERO VERTICAL OFFSET WAVEFORMS IN WELLS SPACED 105 m

Figure 4. Zero-vertical offset seismic waveforms produced between wells 7-5 and 7-7 showing gamma ray, density, and horizontal compressional wave velocity logs.

Figure 5. Well log display for well 7-7 including: core permeability (PERM), gamma ray (GR CORR), compressional wave velocity (SONIC VE), density-neutron crossplot porosity corrected for shale volume (DNPISHC) and core porosity (CORE PHI).

Figure 6. Well log display for well 7-5 including: core permeability (PERM), gamma ray (GR CORR), compressional wave velocity (SONIC VE), density-neutron crossplot porosity corrected for shale volume (DNPHISHC) and core porosity (CORE PHI).

Figure 7. Zero-vertical offset waveform in the shale low-velocity layer recorded with a cylindrical bender source. The detector and source depth is 874 ft.

Figure 8. Time-frequency analysis of a zero-vertical offset waveform in the shale low-velocity layer recorded at a depth of 874 ft.

Figure 9. Well log display for well 7-5 showing compressional (VC) and shear wave (VS) velocity logs calculated by Schlumberger and New England Research.

Figure 10. Dispersion and attenuation curves for normal modes associated with (a) quasi SH waves, (b) quasi SV waves in an anisotropic low-velocity layer.

Figure 11. Group velocity of a Rayleigh-type mode in the shale low-velocity layer at a depth of 874 ft.

Figure 12. Time-frequency analysis of the zero-vertical offset full waveform in the shale low-velocity layer recorded at a depth of 874 ft.

Figure 13. Common source seismic waveforms with source at 970 ft. Waveforms produced between wells 7-5 and 7-7 showing gamma ray, density and horizontal compressional wave velocity logs.

Figure 14. Group velocity plots for the waveforms recorded at detector depths of 960 through 968 ft. The source depth was 970 ft.

Figure 15. Group velocity plots for the waveforms recorded at detector depths of 970 through 978 ft. The source depth was 970 ft.

Figure 16. Group velocity plots for the waveforms recorded at detector depths of 980 through 988 ft.

Figure 17. Group velocity plots for the waveforms recorded at detector depths of 990 through 998 ft.

CHAPTER 6

INTERWELL SEISMIC LOGGING EXPERIMENTS FOR SOURCE-INDEPENDENT SEISMIC ATTENUATION STUDIES AT THE GYPSY TEST SITE, OKLAHOMA

INTERWELL SEISMIC EXPERIMENTS AND DATA ACQUISITION

Interwell seismic logging experiments were conducted at the Gypsy Test Site in Oklahoma (Figure 1), where BP Exploration drilled six 1,310 ft boreholes to relate sedimentary features to reservoir flow units, and test reservoir properties in controlled field conditions. A plane view of the location of the wells is shown in Figure 2. The aim of the experiments was to acquire interwell seismic data for reservoir characterization using high-resolution seismic instrumentation capabilities to efficiently study and interpret reservoir rock physical parameters in full scale.

To date, instrumentation for high-resolution delineation of oil and gas reservoirs has emphasized the development and field applications of specialized borehole seismic sources and single-channel hydrophone detectors. These devices have proven to be sufficient for demonstrating high-resolution interwell seismic measurements but have imposed limitations on the amount of field data that could be efficiently acquired because of single-channel operation. Furthermore, the errors in collecting repeat measurements at a given downhole depth station using a single-channel detector reduce the effective accuracy of high-resolution data analysis, particularly in deriving the seismic attenuation parameters of the formation being proved. As a consequence, a broadband 1.2 kJ arc discharge seismic source (sparker) developed by Southwest Research Institute was used, together with two multi-channel downhole hydrophone arrays developed by Innovative Transducers, Inc. The specifics of the arc discharge technology as a seismic source can be found in Owen and Schroeder (1987).

Recent development in new piezoelectric polymer film hydrophones for ocean acoustic applications are now adaptable in an inexpensive way for use in boreholes. This downhole sensor technology was recently tested at the Buckhorn test site, a shallow oil field reservoir located in Illinois (Parra et al., 1992). In these tests, a 24-channel piezoelectric film hydrophone array manufactured by Innovative Transducer, Inc. (ITI), Fort Worth, Texas was used in combination with the Southwest Research Institute 1.2 kJ impulsive arc-discharge source to obtain excellent high-resolution hole-to-hole seismic measurements to depths of about 200 m in a limestone oil-producing formation (see Chapter I of this report).

The design characteristics of the borehole hydrophone array tested in Illinois were discussed with technical representatives of ITI, leading to a revised array design suitable for use in deeper borehole and providing improved detection sensitivity. The basic hydrophone element utilized in this new array is a type DH-5 piezoelectric film sensor with a self-contained low-noise preamplifier and cable driver molded directly in a downhole rubber jacketed cable. The complete technical specifications of this hydrophone array are presented in Table 1.

TABLE 1
TECHNICAL SPECIFICATIONS OF THE 12-CHANNEL BOREHOLE
SEISMIC HYDROPHONE ARRAY

HYDROPHONE SENSOR ELEMENT

Multilayer PVDF piezoelectric polymer film sensor mounted on a grooved aluminum cylinder 1 inch in outside diameter and molded in a urethane rubber jacket 1.25 inch in outside diameter. (Innovative Transducers, Inc. Type DH-5 transducer)

Overall Dimensions:	Length 6.25 inch (16.9 cm) Diameter 1.25 inch (3.2 cm)
Operating Temperature:	-5° to +55°C
Maximum Operating Depth:	2,500 ft. (760 m)
Sound Pressure Sensitivity:	-190 dB re 1V/ μ Pa $\pm$ 1 dB
Sensitivity vs. Depth:	-1 dB at maximum operating depth
Frequency Response:	$\pm$ 0.5 dB over the frequency range of 7-10,000 Hz.

SELF-CONTAINED PREAMPLIFIER

Noise Spectral Density:	50 nV/Hz ^{1/2}
Noise Floor:	2.2 μ V rms in frequency band 100-2,000 Hz (6,960 μ Pa rms)
Dynamic Range:	127.2 dB 5 V peak maxim
Operating Power:	12 VDC 100 ma (From surface supply)

CABLE AND HYDROPHONE ARRAY

Number of Hydrophones:	12
Hydrophone Spacing:	13.12 ft (4.0 m)
Total Cable Length:	1,650 ft (503 m); Extendable to 2,500 ft (760 m)
Cable Construction:	Hydrophones molded into cable Cable outside diameter 0.5 inch (1.27 cm) Insulation: Neoprene rubber Internal Tensile Strength Member: 0.125 inch Kevlar Number of Conductors: 27 Molded Depth Markers: 10 m spacing; Color Coded

CABLE REEL

Electric Power Operation
 Mechanical Brake
 Type SS-1-NK-27 Terminal connector on Reel

Two complete borehole hydrophone arrays having the specifications listed in Table 1 were fabricated by Innovative Transducers, Inc. This array design is capable of future expansion to 24 hydrophone elements and a total cable length of 2,500 ft (760 m) if desired.

The interwell seismic transmission measurements between pairs of borehole were performed by placing the arc discharge source transducer in well 7-1 and two separate multi-sensor arrays of seismic detectors in wells 7-11 and 7-7, respectively, to permit simultaneous measurements between the source borehole and the two detector boreholes. Full waveform seismic traces were recorded in digital form and stored on magnetic tapes, using the Bison 9024 data acquisition system. The boreholes used and the measurement depth intervals were selected based upon the inspection of available well logs and other site data provided by BP Exploration. As a result of the seismic experiments, two multi-offset data sets were produced at 75 m and 150 m well separations, respectively. The data were recorded with 50 microsecond sample rate, 5,000 samples per trace, and a bandpass filter of 256-2000 Hz. The source was fired to record 4-stack data sets. The number of stacks was determined by conducting several repeatability tests and sparker signal quality.

DATA ANALYSIS AND PROCESSING

Integration of Interwell Seismic Data and Well Logs

The focus of this attenuation and dispersion study is the interval from 800 to 1100 ft of the Pennsylvanian sandstone and shales. The permeable and porous Gypsy sandstone interval in the Vamoosa formation is selected to investigate seismic wave propagation characteristics between wells, and to extract attenuation and dispersion parameters. Specifically, the geological cross-section through the wells 7-1, 7-11, and 7-7 shown in Figure 3 with the lithology and compressional wave velocity logs is selected. The Gypsy sandstone interval (HV4) is composed of stacked channel sands, mudstone, and siltstones which indicates a very difficult correlation of individual sandstone units between wells.

The Gypsy sandstones are moderately sorted, very fine to medium grained sublitharenites. Quartz grains are the dominant framework components, with igneous and metamorphic rock fragments. Clay lamination are common in most of the sandstone and mud clasts. Most of the porosity is intergranular, but some is moldic, created by the partial to complete dissolution of framework grains. The core porosity of the Gypsy sandstones averages 19.7% and permeability averages 558 md (Doyle, 1992). Porosity values reach as high as 26% and permeability as high as 2000 md.

In addition, a Gypsy pilot-site hydraulic conductivity study performed by BP Exploration have demonstrated that in the Gypsy sandstone interval between wells 7-11 and 7-7, the average permeability of a sand layer of about 20 ft of thickness is 1640 md. This permeable sand is a low-velocity zone that is observed at a depth of about 980 ft in the Gypsy sandstone interval as shown in Figures 4 and 5. Here the density and gamma logs are illustrated together with the horizontal compressional wave velocity profiles, and the zero-vertical offset seismograms recorded at the well separations of 75 m and 150 m, respectively.

The horizontal compressional wave velocity profiles were determined from the zero-vertical offset waveforms.

The low-density clean sandstone correlates with the velocity low observed in the horizontal compressional wave velocity profile between the depth of 960-990 ft. To analyze the full-waveforms recorded in this zone, we selected common source seismograms for a source placed in well 7-1 at the depth of 300 m (984 ft) in the clean sandstone zone (see Figures 6 and 7). The six waveforms shown in Figure 8a and 8b were selected from the common source seismograms to investigate in more detail the propagation characteristics of the zone of interest. In fact, time-frequency analysis was applied to these six waveforms to produce the group velocity contour plots illustrated in Figures 9a and 9b. These plots show the dispersion characteristics and reduction in frequency content of the seismic events transmitted between wells.

We have observed that the waveforms recorded at the detector depth of 296 m (970 ft) between wells 7-1 and 7-11 is similar in character with the waveform recorded at the same depth between wells 7-1 and 7-7. That is, the direct P-wave recorded in well 7-11 is the same direct P-wave recorded in well 7-7. The direct P-wave was not modified by the heterogeneous conditions associated with the presence of the shale. In fact, the density and gamma ray logs in the receiver wells 7-11 and 7-7 exhibits similar signatures associated with a low-density clean sand. Alternatively, the compressional waves recorded at the detector depths of 292 m (958 ft) and 300 m (984 ft) have been modified by the presence of reflections and transmission waves associated with the heterogeneous conditions of the Gypsy sandstone interval. As a consequence, in this case the signal observed at the wells 7-11 and 7-7 are quite different. This will produce an attenuation effect caused by the combination of sand and shale units. It will be more difficult to relate permeability with attenuation.

Attenuation and Dispersion Parameters From Interwell Seismic Data

Measurements of seismic wave attenuation is a valuable source of information on the physical properties of a rock. This is a field where interwell seismic logging is very appealing. In particular, observations at points within the medium make possible study of changes in pulse waveform where it travels through a lithologic zone of interest.

Determination of attenuation from field data requires careful analysis and assumptions concerning the absorption model. For the determination of attenuation from interwell seismic data (recorded between two detector wells) we used the spectral ratio method given by Garley and Kanasewick (1980). This is a relatively simple method that does not depend on the nature of the seismic source. The method is outlined next.

At a distance R from the source, the seismic field amplitude in the frequency domain is given by

$$A(\omega) = A_{\text{source}}(\omega) \frac{I(\theta)}{R} e^{i\tilde{k}(\omega)R}, \quad (1)$$

where

$A_{\text{source}}(\omega)$ = source spectrum,

$I(\theta)$ = source radiation pattern (at angle θ),

and $\tilde{k}(\omega)$ = complex propagation constant.

The complex propagation constant can be written as

$$\tilde{k}(\omega) = \frac{\omega}{c(\omega)} + ia(\omega), \quad (2)$$

where

$c(\omega)$ = phase velocity,

and $a(\omega)$ = attenuation.

If we measure the field at two locations (R_1, θ_1) and (R_2, θ_2) , we can take the ratio of the detected fields (use $R_2 > R_1$):

$$\frac{A_2(\omega)}{A_1(\omega)} = \frac{I(\theta_2)}{I(\theta_1)} \frac{R_1}{R_2} e^{i\tilde{k}(\omega)(R_2 - R_1)} \quad (3)$$

Separating Equation (3) into its amplitude and phase components, we obtain

$$\ln \left| \frac{A_2(\omega)}{A_1(\omega)} \right| = \ln \left[\frac{I(\theta_2)}{I(\theta_1)} \frac{R_1}{R_2} \right] - a(\omega)(R_2 - R_1), \quad (4)$$

and

$$\text{Phase} \left(\frac{A_2(\omega)}{A_1(\omega)} \right) = \frac{\omega}{c(\omega)} (R_2 - R_1). \quad (5)$$

Notice that the amplitude information yields the attenuation as a function of frequency while the phase information yields the velocity dispersion. Equation (5) indicates the spectral

method can be used to determine velocity dispersion. While true, the phases of the spectra will be very sensitive to the exact determination of arrival times.

A common assumption is that attenuation increases linearly with increasing frequency. This is often related to a frequency-independent "Q"-factor by

$$a(\omega) = \frac{\omega}{2Qc(\omega)} \quad (6)$$

Dispersion is often neglected in Equation (6), by making $c(\omega)$ a constant over the frequency range of interest.

The application of the spectral method is as follows. First, find a linear fit to the logarithmic amplitude versus frequency, over a frequency range. That is, find slope m such that

$$\ln \left| \frac{A_2(\omega)}{A_1(\omega)} \right| = m \cdot f + b \quad (7)$$

Note that m will be negative for attenuating media. Second, find Q by using Equations (4) and (6):

$$Q = -\frac{\pi(R_2 - R_1)}{mc} \quad (8)$$

According to Equation (3), the determination of Q does not depend on geometric factors. Figures 10 show an example use of this method on simulated data. The simulation was done to be compared with measured data. The calculated Q of 58.6 compares well with the actual Q of 60. The linear fit was done over the frequency range of 1000–2500 Hz.

The spectral ratio method was done with synthetic data, to allow us to find the best ways of using this method to accurately determine the attenuation. We found that measuring attenuation with noisy data can be difficult, so that refinement of the technique can be very important in getting accurate results.

There are a few observations to make from Figures 10. First, choice of the frequency range over which to do the linear fit is critical. Comparison of Figures 10b and 10c show that the proper range essentially begins after the spectral peak and ends before the amplitude falls to near the noise floor. Note that we want to wait until sufficiently after the spectral peak that the sloping/decaying nature of the spectrum has become well-established. Second, in addition to the reduction in amplitude, the attenuation also causes the effective spectral peak to shift to lower frequency. This makes sense, since attenuation is approximately proportional to frequency.

The spectral method, as derived above, implicitly uses a single arrival. That is, in actual seismic data, there are multiple arrivals of the transmitted waveform, from reflections, refractions, etc. In order to use the spectral method with actual data, we use only a portion of the time-domain data, selected manually to be a single arrival. This cannot be done very well, however, as arrivals often overlap in time.

Next, we calculate attenuation and dispersion curves from real seismic data using Equations (4) and (5). That is, the effective quality factor Q is determined as a function of depth using the spectral ratio method from the zero-vertical offset waveforms recorded in wells 7-11- and 7-7 at well separations of 75 m and 150 m, respectively. The effective quality factors are plotted together in Figure 11 with the gamma ray log to illustrate the variability of the effective quality factor with the lithology (i.e., the sand-shale sequence). Significant differences in the Q value in the various lithologies are observed. In particular, the absorption of the compressional waves in the clean sandstone at the depth of about 980 ft (in the low-velocity zone) was twice the absorption associated with the seismic waves propagating in the shales, above and below the Gypsy sandstone interval. In general, the absorption of seismic waves propagating in clean sands is greater than the absorption of those seismic waves traveling in the shales and transition zones.

Since we are interested to relate attenuation with permeability, we will focus on the Gypsy sandstone interval where a low-velocity zone is observed. The common source waveforms given in Figures 8a and 8b, were selected to calculate attenuation and phase velocity curves using the spectral ratio method. The waveforms recorded at the detector depths of 292 m (958 ft), 296 m (970 ft) and 300 m (984 ft), at 75 m and 150 m well separations, respectively, are analyzed. Figures 12-18 illustrate the waveforms, the spectra, and the corresponding attenuation and phase velocity curves.

The direct P waves recorded at the detector depth of 296 m (970 ft) in wells 7-11 and 7-7 appear to be the only waves that travel within the low-velocity zone (i.e., a clean permeable sand). The spectra of these waveforms detected at 75 m and 150 m have similar responses within the frequency range of 50-1000 Hz, as was previously indicated using time-frequency analyses. The attenuation increases to a peak value of 0.5 dB/m at 1000 Hz, and then decreases as a function of frequency. That is, a loss of energy of 37.5 dB occurred for the P-wave traveling between wells 7-11 and 7-7. The absorption and dispersion characteristics observed in the attenuation and phase velocity curves are consistent as shown in Figure 13b. On the other hand, the spectra signatures and attenuation curves associated with the seismic waves recorded at the detector depths of 292 m (958 ft) and 300 m (984 ft) are affected by transmission and reflections produced at the boundaries corresponding to lithologic changes within the Gypsy sandstone interval. Although the seismic signatures observed at the detector depths of 292 m (958 ft) and 300 m (984 ft) are different, their attenuation losses are the same, about 0.25 dB/m.

The attenuation of the compressional wave detected at the receiver depth of 296 m (970 ft) is practically twice the attenuation of those waveforms recorded at 292 m (958 ft) and 300 m (984 ft) in the same frequency range. The attenuation peak of 0.5 dB/m at 1000 Hz is caused by presence of the notch in the spectrum of the waveform traveling between

wells 7-1 and 7-7. If the notch is suppressed the resulting attenuation peak is about 0.25 dB/m at 1000 Hz. Also, attenuation peaks are observed at 750 Hz for the waveforms recorded at 292 m (958 ft) and 300 m (984 ft). There is a shift of the attenuation peak towards the high-frequency associated with the low-velocity clean sandstone, which is more permeable than its surrounded medium (as indicated by the air permeability data). The shift of the attenuation peak towards the high-frequency is consistent with the squirt-flow phenomena described by Dvorkin and Nur (1993). To analyze this phenomena in more detail using the present seismic data it will require to correct the interwell seismic waveforms for the effect of heterogeneities and transmission and reflections effects caused by the presence of boundaries. Thus, a more appropriate interwell seismic signature can be analyzed to relate more directly the sand attenuation with permeability.

CONCLUSIONS

The high-resolution interwell seismic field experiments and data analysis have yielded important results which demonstrates that high-sensitivity hydrophone detector arrays respond to seismic waves produced by a 1.2 kJ arc-discharge source in the sedimentary formations at the Gypsy test site in Oklahoma. The integration of petrophysical and geological information with well logs and interwell seismic waveforms have determined the presence of a low-velocity permeable sandstone zone at a depth of about 980 ft in the Gypsy sandstone interval between wells 7-1, 7-11, and 7-7. The presence of this sandstone zone has been already demonstrated by pulse test data. The results of those pulse tests indicated that many of the sands are in pressure communication between the wells at Gypsy. The attenuation and dispersion results suggested that the loss of energy of seismic waves propagating within the low-velocity zone are caused by the porous and permeable conditions of the sandstone. Specifically, a shift of the attenuation peak towards the high-frequencies (associated with the presence of a more permeable sandstone) was observed in the attenuation curves.

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Figure 1. Site map indicating location of the Gypsy test site.

Figure 2. A plane view of the six wells at the Gypsy borehole test facility in Cleveland, Oklahoma.

Figure 3. A geological cross-section through the wells 7-1, 7-11, and 7-7.

Figure 4. Zero-vertical offset seismic waveforms produced between wells 7-1 and 7-11 showing gamma ray, density and compressional wave velocity logs.

Figure 5. Zero-vertical offset seismic waveforms produce between wells 7-1 and 7-7 showing gamma ray, density, and compressional wave velocity.

Figure 6. Common-source seismic waveforms for a source located at 300 m in the Gypsy sandstone formation.
Well separation = 75 m

Figure 7. Common-source seismic waveforms for a source located at 300 m in the Gypsy sandstone formation.
Well separation = 150 m.

Figure 8. Common-source waveforms selected from the seismograms given in (a) Figure 6, and (b) Figure 7 for attenuation studies.

Figure 9. Group velocity contour plots determined from waveforms given in (a) Figure 8a and (b) Figure 8b.

(a)

(b)

(c)

Figure 10. (a) Zero-vertical offset synthetic waveforms calculated for detectors located at the horizontal distance of 75 m and 150 m from the source, (b) Spectra; (c) Spectral ratio.

Figure 11. Effective quality factor as a function of depth with gamma ray log in well 7-7.

Figure 12. Common-source waveforms produced between wells (a) 7-1 and 7-11; (b) 7-1 and 7-7. Source at a depth of 300 m and detector at 292 m.

Figure 13. Spectra of the common source waveforms given in Figure 12.

(a)

(b)

Figure 14. (a) Dispersion and (b) Attenuation curves determined from the spectra given in Figure 13.

Figure 15. Common source waveform produced between wells (a) 7-1 and 7-11; (b) 7-1 and 7-7. Source at 300 m and detector at 296 m.

Figure 16. Spectra of the common source waveform given in Figure 15.

(a)

(b)

Figure 17. (a) Dispersion and (b) attenuation curve determined from the spectra given in Figure 16.

Figure 18. Common-source waveform produced between wells (a) 7-1 and 7-11 (b) 7-1 and 7-7. Source at a depth of 300 m and detector at a depth of 300 m.

Figure 19. Spectra of the common-source waveform given in Figure 18.

(a)

(b)

Figure 20. (a) Dispersion and (b) attenuation curve determined from the spectra given in Figure 19.

CHAPTER 7

SOURCE-INDEPENDENT SEISMIC ATTENUATION MODEL STUDIES

SEISMIC WAVE PROPAGATION IN POROELASTIC MEDIA INCLUDING BIOT AND SQUIRT-FLOW MECHANISMS

The dynamic poroelasticity given by Biot (1941 and 1956) have served as rigorous and formal foundations to several practical solutions concerning the acoustic wave propagation in fluid-filled saturated sediments (e.g., Stoll, 1977). In the Biot solution, the pore fluid is forced to participate in a solid oscillatory motion by viscous friction and inertial coupling. A different mechanism of fluid flow during seismic and acoustic wave propagation is associated with the squirting of the pore fluid out of the cracks as they are deformed by passing seismic waves. Mavko and Nur (1979) have shown that squirt-flow mechanism results in much higher and realistic attenuation in partially saturated rocks than those predicted by Biot mechanism.

A consistent theory dealing simultaneously with Biot and squirt-flow mechanisms has been presented by Dvorkin and Nur (1993). The theory takes those mechanisms by considering the fluid motion parallel (the Biot mechanism) and transverse (the squirt-flow mechanism) to the direction of a planar P-wave propagation. That is, the fluid is free to move not only parallel but also perpendicular to this direction. The fluid flow is induced by the compression of the pore volume due to wave excitation, which results in increasing poropressure.

The Biot-squirt model developed by Dvorkin and Nur (1993) relates the dynamic poroelastic behavior of a saturated rock to the poroelastic parameters, porosity, permeability, fluid compressibility and viscosity, and the characteristic squirt-flow length. The model can be used to relate compressional phase velocity and attenuation to rock and fluid properties such as permeability, porosity, viscosity, and compressibility.

In this section we recall the dynamic poroelastic coupled Biot equations for a force or a volume source. We modify the wave propagation coefficients to include the squirt-flow mechanism in the poroelastic Green's function. Thus, the solution for the pressure and the displacement components are obtained for a point source in a poroelastic unbounded media modeled with the Biot and squirt-flow mechanisms.

Formulation

The theory of seismic wave propagation for point source in porous and permeable media requires the solution of the coupled system of differential equations (Parra, 1991).

$$\mu \nabla^2 u + (\lambda + \mu) \nabla \nabla \cdot u + \omega^2 \rho u - \alpha \nabla p = -F \delta(r - r_0) \quad (1a)$$

$$\theta \nabla^2 p - \beta p - \alpha \nabla \cdot u = -V \delta(r - r_0), \quad (1b)$$

where

$$\begin{aligned} u &= \text{displacement of the bulk rock matrix} \\ p &= \text{the fluid pressure} \\ \lambda &= K_b - 2\mu/3 \\ \mu &= (1 - \phi) \bar{\mu} \\ \rho &= \rho + \rho_f \omega^2 \theta \\ \alpha &= \alpha + \rho_f \omega^2 \theta \\ \alpha &= 1 - K_b/K_s \\ \rho &= (1 - \phi) \rho_s + \phi \rho_f \\ \beta &= (\alpha - \phi)/K_s + \phi/K_f \\ K_f &= \text{pore fluid modulus} \\ \text{and } \theta &= -K(\omega)/j\omega. \end{aligned}$$

In this equation $K(\omega)$ is the frequency-dependent generalized Darcy coefficient (Biot, 1962):

$$K(\omega) = \frac{j\phi}{\omega \rho_f} \left[\frac{\rho_a/\rho_f + \phi}{\phi} + \frac{j\omega_c}{\omega} \right]^{-1}, \quad (2a)$$

which leads θ to be

$$\theta = -\frac{\phi}{\omega^2 \rho_f} \left[\frac{\rho_a/\rho_f + \phi}{\phi} + \frac{j\omega_c}{\omega} \right]^{-1}. \quad (2b)$$

where ρ_a is the additional coupling density. The critical angular frequency is given by $\omega_c = \eta\phi/\bar{k}\rho_f$, where $\bar{k}$ is the intrinsic permeability of the material and η is the viscosity of the saturant fluid.

The fluid pressure in terms of the mean displacement in the solid and in the pore fluid can be obtained directly from Equation (9) given in Parra (1991):

$$p = -\frac{\phi}{\beta} \left[\nabla \cdot U + \frac{(\alpha - \phi)}{\phi} \nabla \cdot u \right], \quad (3)$$

where U is the mean displacement in the pore fluid, and

$$\frac{\phi}{\beta} = \frac{1}{\frac{1}{K_f} + \frac{1}{\phi Q}} \equiv F ,$$

in which

$$\frac{1}{Q} = \frac{1}{K_s} \left(1 - \phi - \frac{K_b}{K_s} \right) .$$

Following the work of Dvorkin and Nur, we introduce the squirt-flow mechanism in the Biot Equations (3) to produce the average local fluid pressure:

$$P_{av}(\omega) = -\frac{\phi}{\beta} \left[1 - \frac{2 J_1(\xi \tilde{R})}{\xi R J_0(\xi \tilde{R})} \right] \left[\nabla \cdot U + \frac{(\alpha - \phi) \nabla \cdot u}{\phi} \right] , \quad (4)$$

where $\tilde{R}$ is the squirt-flow length, and ξ is obtained from

$$\xi^2 = \frac{\omega^2 \rho_f}{(\phi/\beta)} \left(\frac{\rho_a/\rho_f + \phi}{\phi} + \frac{j\omega_c}{\omega} \right) . \quad (5)$$

The parameter ξ can be expressed in terms of the parameter θ defined by Equation (2b) as

$$\xi^2 = -\beta/\theta . \quad (6)$$

Since the effect of squirt flow is taken into account in Equation (4), this equation has to be used to derive the second couple Equation (1b). Thus, the new β parameter containing the squirt-flow mechanism is given by,

$$\beta = \phi / F_{sq} ,$$

where

$$F_{sq} = F \left[1 - \frac{2 J_1(\xi \tilde{R})}{\xi R J_0(\xi \tilde{R})} \right] ,$$

$$F = \phi / \left((\alpha - \phi) / K_s + \phi / K_f \right) ,$$

$$\text{and } \xi^2 = -\frac{\phi}{\theta F}.$$

For values $\frac{\omega_c}{\omega} \gg 1$ which may be of practical importance even for high frequencies, the approximate expression for θ given by Equation (2b) is reduced to

$$\theta = jk/\eta \omega,$$

$$\text{and } (\xi R) = \sqrt{j} \sqrt{\frac{\bar{R}^2 \omega}{\kappa}}, \quad \kappa = \frac{\bar{k} F}{\eta \phi}.$$

Thus, the only parameter that has been modified in Equation (1) by considering the squirt-flow mechanism is the parameter β which becomes frequency dependent in the Biot-squirt flow model. To determine the wavefield solution for a point source in a poroelastic layered medium associated with the Biot-squirt-flow mechanism, we use the Green's tensor in the wavenumber domain and express the displacement stress matrix, the fluid pressure, and the volumetric flux component, w , in terms of upgoing and downgoing waves in each layer (Parra, 1991).

The propagation wavenumbers λ_1 and λ_2 associated with fast and slow compressional waves, and the wavenumber, k_s , for shear waves are required to be determined before the layered media boundary value problem is addressed. These wave parameters given by Parra (1991) are summarized as follows

$$\lambda_1 = [b + \Delta]/2]^{1/2} \quad (9a)$$

$$\lambda_2 = [(b - \Delta)/2]^{1/2} \quad (9b)$$

$$k_s = [\omega^2 \rho / \mu]^{1/2}, \quad (9c)$$

where

$$\begin{aligned} b &= k_p^2 - \alpha^2/(\lambda + 2\mu)\theta - \beta/\theta \\ \Delta &= [b^2 + 4k_p^2\beta/\theta]^{1/2} \\ \text{and } \beta &= \phi/F_{sq}. \end{aligned}$$

The phase velocities and attenuations of the compressional waves (fast and slow) and the shear waves are given by

$$v_{\text{fast}} = \omega / \text{real}(\lambda_1)$$

$$Q_{\text{fast}}^{-1} = 2 \text{imag}(\lambda_1) / \text{real}(\lambda_1) ,$$

$$v_{\text{slow}} = \omega / \text{real}(\lambda_2)$$

(10)

$$Q_{\text{slow}}^{-1} = 2 \text{imag}(\lambda_2) / \text{real}(\lambda_2)$$

and

$$v_{\text{shear}} = \omega / \text{real}(k_s)$$

$$Q_{\text{shear}}^{-1} = 2 \text{imag}(k_s) / \text{real}(k_s) .$$

Analytic solution for a point source in an unbounded poroelastic medium

The Green's tensor in the wavenumber domain leads directly to closed-form expressions for the vector wave displacement and the pressure produced by a point source in a poroelastic unbounded medium. Thus, the displacement components and the pressure at the detector position z , for a force at the source depth z_s are obtained in terms of the derivatives of Sommerfeld integrals.

For the horizontal force (F_r)

$$u_r^{(h)} = \frac{F_r(\omega)}{4\pi \rho \omega^2} \left(\frac{\partial^2 G}{\partial r^2} + J \right)$$

$$u_z^{(h)} = \frac{F_r(\omega)}{4\pi \rho \omega^2} \frac{\partial^2 G}{\partial r \partial z} \quad (11a)$$

and
$$p^{(h)} = \frac{F_r(\omega)}{4\pi \rho \omega^2} \frac{\partial T}{\partial r} .$$

For the vertical force (F_z)

$$\begin{aligned} u_r^{(v)} &= \frac{F_z(\omega)}{4\pi\beta\omega^2} \left(\frac{\partial G}{\partial z^2} + J \right) \\ u_z^{(v)} &= \frac{F_z(\omega)}{4\pi\beta\omega^2} \frac{\partial^2 G}{\partial r \partial z} \end{aligned} \quad (11b)$$

and $p^{(v)} = \frac{F_z(\omega)}{4\pi\beta\omega^2} \frac{\partial T}{\partial z}$.

For the volume source (V_o)

$$\begin{aligned} u_r &= \frac{V_o(\omega)}{4\pi\beta\omega^2} \frac{\partial T}{\partial r} \\ u_z &= \frac{V_o(\omega)}{4\pi\beta\omega^2} \frac{\partial T}{\partial z} \end{aligned} \quad (11c)$$

and $p = \frac{V_o(\omega)}{4\pi\omega^2\beta} \left[\frac{\delta_1}{\alpha_3} \frac{e^{j\lambda_1 R}}{R} - \frac{\delta_2}{\alpha_3} \frac{e^{j\lambda_2 R}}{R} \right]$

where

$$\begin{aligned} G &= \frac{\alpha_1}{\alpha_3} \frac{e^{j\lambda_1 R}}{R} + \frac{\alpha_2}{\alpha_3} \frac{e^{j\lambda_2 R}}{R} + \frac{e^{jk_s R}}{R} \\ J &= k_s^2 \frac{e^{jk_s R}}{R} \\ T &= \frac{\beta_1}{\alpha_3} \left(\frac{e^{j\lambda_2 R}}{R} - \frac{e^{j\lambda_1 R}}{R} \right) \\ \text{and } R &= [r^2 + (z - z_s)^2]^{1/2} \end{aligned}$$

The coefficients α_1/α_3 , $\frac{\alpha_2}{\alpha_3}$, β_1/α_3 and δ_1/α_3 are given in Parra (1991). The propagation wavenumbers λ_1 , λ_2 and k_s are given in Equation (9).

The phase velocities and attenuations of the compressional waves (fast and slow) and the shear waves are given by

$$\begin{aligned} v_{\text{fast}} &= \omega / \text{real}(\lambda_1) \\ Q_{\text{fast}}^{-1} &= 2 \text{imag}(\lambda_1) / \text{real}(\lambda_1) , \\ v_{\text{slow}} &= \omega / \text{real}(\lambda_2) \end{aligned} \tag{10}$$

and

$$\begin{aligned} v_{\text{shear}} &= \omega / \text{real}(k_s) \\ Q_{\text{shear}}^{-1} &= 2 \text{imag}(k_s) / \text{real}(k_s) . \end{aligned}$$

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$$\begin{aligned} u_r &= \frac{V_o(\omega)}{4\pi \rho \omega^2} \frac{\partial T}{\partial r} \\ u_z &= \frac{V_o(\omega)}{4\pi \rho \omega^2} \frac{\partial T}{\partial z} \end{aligned} \quad (11c)$$

and $p = \frac{V_o(\omega)}{4\pi \omega^2 \rho} \left[\frac{\delta_1}{\alpha_3} \frac{e^{j\lambda_1 R}}{R} - \frac{\delta_2}{\alpha_3} \frac{e^{j\lambda_2 R}}{R} \right]$

where

$$\begin{aligned} G &= \frac{\alpha_1}{\alpha_3} \frac{e^{j\lambda_1 R}}{R} + \frac{\alpha_2}{\alpha_3} \frac{e^{j\lambda_2 R}}{R} + \frac{e^{jk_s R}}{R} \\ J &= k_s^2 \frac{e^{jk_s R}}{R} \\ T &= \frac{\beta_1}{\alpha_3} \left(\frac{e^{j\lambda_2 R}}{R} - \frac{e^{j\lambda_1 R}}{R} \right) \\ \text{and } R &= [r^2 + (z - z_s)^2]^{1/2} \end{aligned}$$

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$$\begin{aligned} \text{and} \quad v_{\text{shear}} &= \omega / \text{real}(k_s) \\ Q_{\text{shear}}^{-1} &= 2 \text{imag}(k_s) / \text{real}(k_s) . \end{aligned}$$

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$$\text{and} \quad p^{(h)} = \frac{F_r(\omega)}{4\pi\beta\omega^2} \frac{\partial T}{\partial r} .$$

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and
$$p^{(v)} = \frac{F_z(\omega)}{4\pi\beta\omega^2} \frac{\partial T}{\partial z} .$$

For the volume source (V_0)

$$\begin{aligned} u_r &= \frac{V_0(\omega)}{4\pi\beta\omega^2} \frac{\partial T}{\partial r} \\ u_z &= \frac{V_0(\omega)}{4\pi\beta\omega^2} \frac{\partial T}{\partial z} \end{aligned} \quad (11c)$$

and
$$p = \frac{V_0(\omega)}{4\pi\omega^2\beta} \left[\frac{\delta_1}{\alpha_3} \frac{e^{j\lambda_1 R}}{R} - \frac{\delta_2}{\alpha_3} \frac{e^{j\lambda_2 R}}{R} \right]$$

where

$$\begin{aligned} G &= \frac{\alpha_1}{\alpha_3} \frac{e^{j\lambda_1 R}}{R} + \frac{\alpha_2}{\alpha_3} \frac{e^{j\lambda_2 R}}{R} + \frac{e^{jk_s R}}{R} \\ J &= k_s^2 \frac{e^{jk_s R}}{R} \\ T &= \frac{\beta_1}{\alpha_3} \left(\frac{e^{j\lambda_2 R}}{R} - \frac{e^{j\lambda_1 R}}{R} \right) \\ \text{and } R &= [r^2 + (z - z_s)^2]^{1/2} \end{aligned}$$

The coefficients α_1/α_3 , $\frac{\alpha_2}{\alpha_3}$, β_1/α_3 and δ_1/α_3 are given in Parra (1991). The propagation wavenumbers λ_1 , λ_2 and k_s are given in Equation (9).

NUMERICAL RESULTS

The objective of the present analytical studies is to relate permeability with attenuation and dispersion of acoustic waves propagating in the Gypsy sandstone formation. Part of this study is to select representative rock physical parameters from logs and interwell seismic data at the Gypsy site. The selected rock physical parameters are associated with a low-velocity sandstone which has been identified by integrating geological and petrophysical data with interwell seismic waveforms. The petrophysical characteristics of this low-velocity sandstone were described early in pages 6-7 and 6-8. This zone was characterized as a thin permeable and porous fully-saturated sandstone surrounded by a low-permeability and higher velocity host medium. Since quartz grains are the dominant framework components in the sandstone, we selected the grain bulk modulus, K_s , to be 38 GPa and the grain density, ρ_s , to be 2.59 gr/cm³. The average porosity, ϕ , of the sand is about 24 percent. The compressional wave velocity and shear wave velocity of the sandstone for drained conditions (i.e., $\bar{v}_p$ and $\bar{v}_s$) were determined in terms of the rock physical parameters for saturated conditions (i.e., v_p and v_s). The horizontal compressional wave velocity, v_p , was determined from zero-vertical offset waveforms, and the shear wave velocity, v_s , was approximately estimated from group velocity contour plots and shear wave velocity logs. The Gassmann equations were used to calculate the bulk modulus, K_b , and the velocities of the sandstone:

$$\bar{v}_p = \left[\frac{K_b + 4/3 \mu}{(1 - \phi) \rho_s} \right]^{1/2} ,$$

$$\bar{v}_s = \left[\frac{\mu}{(1 - \phi) \rho_s} \right]^{1/2} ,$$

$$\mu = v_s^2 \rho_e ,$$

$$\rho_e = (1 - \phi) \rho_s + \phi \rho_f ,$$

$$K_b = \frac{K_f(K_s - K) - \phi K(K_s - K_f)}{K_f(K_s - K) - \phi K_s(K_s - K_f)} K_s$$

$$K_f = \rho_f v_f^2$$

and

$$K = \rho_e \left(v_s^2 - \frac{4}{3} v_s^2 \right) .$$

The physical properties of the sandstone and saturant fluid characteristics are given in Tables 1 and 2.

TABLE 1. Formation parameters

Region	K_s (GPa)	ρ_s (gr/cm ³)	$\bar{v}_p$ (Km/s)	v_s (Km/s)	$\bar{k}$ (darcy)	ϕ (%)
Sandstone	38	2.59	3.48	1.97	1.5	24

TABLE 2. Saturant fluid parameters

Fluid	v_f (Km/s)	ρ_f (gr/cm ³)	η (poise)
Water	1.5	1.0	0.01

Attenuation and Dispersion of Acoustic Waves Propagating in a Thick Sandstone

Figures 1a and 1b show the attenuation and phase velocity curves in the range of 10-4000 Hz for the model parameters given in Tables 1 and 2 and permeabilities of 500 md and 1000 md, and a squirt-flow length of 0.4 mm. The attenuation curves show practically the same attenuation peak values as the permeability varies, however, the peaks shift towards the lower frequencies when the permeability decreases. This influence of the permeability on the attenuation was also observed by Dvorkin and Nur (1993).

Next, we calculate synthetic seismic waveforms using the model parameters given in Tables 1 and 2. For this purpose a Ricker wavelet source pulse signal was assumed. Pressure and displacement seismograms (horizontal and vertical) were calculated for a volume source signal centered at 300 Hz and located at a horizontal distance of 75 m from the detectors. The traces in the seismograms shown in Figures 2-4 correspond to outputs of detectors spaced 4 m apart. Also, a pressure seismogram was produced for the same volume source and a new set of detectors located at a horizontal distance of 150 m from the source as shown in Figure 5. In order to illustrate the effect of attenuation and dispersion on seismic waves propagating in permeable media we selected the zero-vertical offset waveforms arriving at 20 ms and 40 ms for detectors placed at 75 m and 150 m, respectively.

The spectral-ratio method was used to calculate attenuation and phase velocity curves. Time-windows of about 10 ms were used to analyze the P-wave pulses calculated at 75 m and 150 m. These direct P-wave pulses extracted from the zero-vertical offset traces are shown in Figure 6. These waveforms indicate a reduction of amplitude of about 5 times for waves traveling in the permeable sandstone. The spectra of these waveforms are illustrated in Figure 7. Here most of the energy appears to be distributed within the frequency range of 1000 Hz, and sharply decreases at higher-frequencies. The estimate pulse center frequency of both P-waves is about 300 Hz, which is equal to the peak frequency of the Ricker wavelet.

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Next, we calculate synthetic seismic waveforms using the model parameters given in Tables 1 and 2. For this purpose a Ricker wavelet source pulse signal was assumed. Pressure and displacement seismograms (horizontal and vertical) were calculated for a volume source signal centered at 300 Hz and located at a horizontal distance of 75 m from the detectors. The traces in the seismograms shown in Figures 2-4 correspond to outputs of detectors spaced 4 m apart. Also, a pressure seismogram was produced for the same volume source and a new set of detectors located at a horizontal distance of 150 m from the source as shown in Figure 5. In order to illustrate the effect of attenuation and dispersion on seismic waves propagating in permeable media we selected the zero-vertical offset waveforms arriving at 20 ms and 40 ms for detectors placed at 75 m and 150 m, respectively.

The spectral-ratio method was used to calculate attenuation and phase velocity curves. Time-windows of about 10 ms were used to analyze the P-wave pulses calculated at 75 m and 150 m. These direct P-wave pulses extracted from the zero-vertical offset traces are shown in Figure 6. These waveforms indicate a reduction of amplitude of about 5 times for waves traveling in the permeable sandstone. The spectra of these waveforms are illustrated in Figure 7. Here most of the energy appears to be distributed within the frequency range of 1000 Hz, and sharply decreases at higher-frequencies. The estimate pulse center frequency of both P-waves is about 300 Hz, which is equal to the peak frequency of the Ricker wavelet.

The attenuation and dispersion curves obtained using the spectral ratio method are given in Figure 8. The dispersion characteristics are clearly observed in the phase-velocity curve by showing an increase in velocity within the 500 Hz interval. The attenuation increases linearly as a function of frequency in the range of frequency of about 100 Hz to 1200 Hz by reaching a value of 0.2 dB/m (i.e., an average Q of about 19). The energy losses of seismic waves propagating between detector wells at 75 m and 150 m is about 15 dB within the frequency range of about 1000 Hz. The 15 dB corresponds to a wave amplitude reduction of 5 to 6 times.

In order to compare the above results with those determined using a higher pulse center frequency, we calculated pressure synthetic seismograms using the same parameters and source detector geometries for a volume source signal centered at 750 Hz (see Figures 9 and 10). The waveforms and their spectra as well as attenuation and phase velocity curves are given in Figure 11-13. In this example, most of the energy is distributed between 100 Hz to 2000 Hz, and slowly decreases towards the higher frequencies. The dispersion characteristics shown in the phase velocity curve are practically free of distortions between 10 Hz to 3000 Hz. That is, at the source center frequency of 750 Hz, we are moving further away in frequency from the relaxation frequency of the permeable sandstone (1500 md) where the dispersion effect is dominant. In addition, the attenuation curve exhibits a linear trend which started at 100 Hz to reach a value of 0.3 dB/m at 1000 Hz and a maximum value of 0.4 dB/m at 2500 Hz. The attenuations of 0.4 dB/m and 0.3 dB/m are equivalent to losses of 30 dB and 22.5 dB, respectively, when waves propagate between the 75 m and 150 m well separation. That is, a greater loss of energy occurs at high-frequencies when the seismic waves propagate in a permeable medium of 1500 md.

Attenuation and Dispersion of Acoustic Waves Propagating in a Thin Sandstone Layer

The next example was also constructed using rock physical parameters estimated from well logs and interwell seismic data. The model consists of a low-velocity permeable sandstone of thickness 8 m in a high-velocity host medium. The rock physical parameters of this model are given in Table 3.

TABLE 3. Formation parameters

Region	K_s (GPa)	ρ_s (gr/cm ³)	$\bar{v}_p$ (Km/s)	$\bar{v}_s$ (Km/s)	$\bar{k}$ (darcy)	ϕ (%)
Upper and lower medium	35	2.6	3.6	2.01	0.1	15
Layer	38	2.59	3.48	1.97	1.5	25

As was done in the previous example, a Ricker wave source pulse signal was assumed. A pressure and displacement seismograms (horizontal and vertical) were produced

for a volume source signal entered at 750 Hz and located at a horizontal distance of 75 m from a vertical line of detectors. The traces in the seismograms shown in Figure 14-16 correspond to outputs of detectors spaced 1 m apart. The volume source was placed in the low-velocity sandstone at a depth of 297.5 m (975.8 ft). The ninth detector is in the top layer interface at $z = 292$ m (957.8 ft), and the seventeenth detector is in the bottom layer interface at $z = 300$ m (984 ft). The pressure seismograms shown full waveforms including compressional waves, reflections and leaky modes. The leaky modes are produced because the compressional wave velocity of the low-velocity layer is greater than the shear wave velocity of the surrounding medium. In particular, stronger wave amplitude wavetrains are observed at detectors near and at the layer interfaces. These events are associated with the permeability contrast between the sandstone and the surrounding medium. In this case, the fluid-flow at the layer interface induces pressure gradients which modify the global flow. The fluid-pressure in the layered medium is due to the Biot and squirt-flow mechanisms. Since the wave amplitudes observed in the fluid-pressure seismogram increase at the layer interface, where fluid flow interaction occurs, the reflections and leaky modes may contain information associated with the permeability of the low-velocity sandstone.

To analyze the effect of dispersion and absorption of acoustic waves propagating in the layered medium, the pressure seismogram for a vertical line of detectors placed at 150 m from the volume source was produced in Figure 17. The waveforms recorded at detector depths of 296 m (970 ft) were selected from Figures 14 and 17 to calculate attenuation and phase velocity curves as well as the spectra (see Figures 18-20). In this example most of the energy is distributed between 100 Hz and 2000 Hz. The estimate pulse center frequency of both events is about 750 Hz, which corresponds to the peak frequency of the volume source wavelet. The attenuation increases linearly with frequency to reach a relaxation peak of 0.15 dB/m at 1200 Hz, equivalent to an energy loss of 11.25 dB for seismic waves propagating between the 75 m well separation. Comparison of this attenuation with that produced by a volume source in a thick permeable sandstone (i.e., about 22.5 dB) suggests that the presence of the low-permeability host medium has reduced (to about 10 dB) the attenuation of acoustic waves propagating in the thin sandstone layer.

CONCLUSIONS

The formulation of seismic wave propagation in poroelastic media including Biot and squirt-flow mechanisms has been presented that allows to produce synthetic seismic waveforms in layered media and relate attenuation and dispersion characteristics of a fluid-filled saturated medium to porosity, permeability, fluid compressibility, and viscosity, and the squirt-flow length. The absorption and dispersion of seismic waveforms modeled with Gypsy sandstone reservoir parameters concurred with the attenuation and dispersion characteristics determined from real interwell seismic data at Gypsy. These results suggested that most of the seismic energy losses and velocity dispersion in the Gypsy sandstone interval are due to fluid flowing between saturated large irregular-shaped porous of different orientation.

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Figure 1. (a) Compressional wave velocity curves, and (b) inverse attenuation quality factor Q^{-1} for Gypsy sandstone having permeabilities of 0.5, 1 and 1.5 md as predicted by the Biot-squirt theory.

Figure 2. Common-source pressure synthetic seismogram for a volume source located at a depth of 300 m in a sandstone formation. The borehole separation is 75 m. Source center frequency = 300 Hz.

Figure 3. Common-source horizontal displacement seismogram for a volume source located at a depth of 300 m in a sandstone formation. The borehole separation is 75 m. Source center frequency = 300 Hz.

Figure 4. Common-source vertical displacement for a volume source located at a depth of 300 m in a sandstone formation. The borehole separation is 75 m. Source center frequency = 300 Hz.

Figure 5. Common-source pressure synthetic seismogram for a volume source located at a depth of 300 m in a sandstone formation. The borehole separation is 150 m and the source center frequency is 300 Hz.

(a)

(b)

Figure 6. Zero-vertical offset P-waves for the sandstone formation calculated for detectors at (a) 75 m and (b) 150 m from the source. Source and detector are at a depth of 300 m. Source center frequency = 300 Hz.

Figure 7. Spectra for the P-waves illustrated in Figure 6.

(a)

(b)

Figure 8. (a) Attenuation and (b) dispersion curves determined from the spectra of Figure 7.

Figure 9. Common-source pressure synthetic seismogram for a volume source located at a depth of 300 m in a sandstone formation. The borehole separation is 75 m and source center frequency is 750 Hz.

Figure 10. Common-source pressure synthetic seismogram for a volume source located at a depth of 300 m in a sandstone formation. The borehole separation is 150 m and source center frequency is 750 Hz.

(a)

(b)

Figure 11. Zero-vertical offset P-waves for the sandstone formation calculated for detectors at (a) 75 m and (b) 150 m from the source. Source and detector are at a depth of 300 m. Source center frequency = 750 Hz.

Figure 12. Spectra for the P-waves given in Figure 11.

(a)

(b)

Figure 13. (a) Attenuation and (b) dispersion curves determined from the spectra of Figure 12.

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Figure 14. Common-source pressure seismogram for a volume source located at a depth of 297.5 m in a thin sandstone layer. The borehole separation is 75 m and source center frequency is 750 Hz.

Figure 15. Horizontal displacement seismogram for a volume source located at a depth of 297.5 m in a thin sandstone layer. The borehole separation is 75 m and source center frequency is 750 Hz.

Figure 16. Vertical displacement seismogram for a volume source located at a depth of 297.5 m in a thin sandstone layer. The borehole separation is 75 m and source center frequency is 750 Hz.

Figure 17. Common-source pressure seismogram for a volume source located at a depth of 297.5 m in a thin sandstone layer. The borehole separation is 150 m and source center frequency is 750 Hz.

(a)

(b)

Figure 18. Zero-vertical offset waveforms for the layered sandstone formations calculated at (a) 75 m and (b) 150 m from the source. Source center frequency is 750 Hz.

Figure 19. Spectra for the waveforms given in Figure 18.

(a)

(b)

Figure 20. (a) Attenuation and (b) dispersion curves determined from the spectra of Figure 19.

APPENDIX A
SOFTWARE DESCRIPTIONS

APPENDIX A

SOFTWARE DESCRIPTIONS

Although the purpose of this project was to investigate interwell seismic logging techniques for indirectly interpreting hydrocarbon reservoir geology and rock physical properties, instrumentation and computer programs were developed to assist on the field operations and data interpretation, respectively. Processing and modeling computer programs were written in FORTRAN IV and C for the HP 9000 series 400 computer. Some of the programs require external libraries such as IMSL and intrinsic subroutines in order to run. Processing and plotting programs contain the necessary documentation for analyzing high-resolution seismic data produced using a cylindrical bender source. A brief description of each of the computer programs is given below.

- PICKGEO.** This program was developed to pick arrival times from a data base. The time-picking algorithm is a modification of the method suggested by Coppens (1985) which consists of comparing a short-term running average of the trace energy to a large-term running average. Each common source seismogram is done at once. The program displays the information on the screen, after all the picks have been made and the output file has been generated.
- TOMO.** This program was developed to calculate the compressional wave velocity distribution from travel time data extracted from interwell seismic waveforms using either straight or curve ray algorithms.
- CORRGEO.** This program was developed to correlate a common source sweep seismogram with a given pilot sweep when a cylindrical bender source is used.
- ZTPLOT.** This program is used to display seismic crosswell data on a separation versus time scale. The program reads an ASCII input file and then displays the data graphically on either a terminal or the 7550A pen plotter.
- TFR4.** This program implements time-frequency representation analysis. The program reads interwell seismic waveforms and displays spectrograms using short-time Fourier transform.
- SPECRAT.** This program is an implementation of the spectral ratio-method given by Ganley and Kanaswich (1980). The program displays attenuation and phase velocity curves and uses IMSL Exponent Graphics package for plotting.
- INTPRS.** This program calculated the vector displacement and pressure associated with a point source in one borehole and a series of detectors in a second borehole in a layered poroelastic medium. The program uses IMSL subroutines and SwRI numerical integration subroutines. The method of solution in the program is based on the theoretical development given by Parra (1991).

INTSOLID. This program calculates the vector displacement associated with a point source in one borehole and a series of detectors in a second borehole in a layered viscoelastic medium. The program uses IMSL subroutines and SwRI numerical integration subroutines. The method used in the program is based on the theoretical solution given by Parra et al. (1993).

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