TITLE PAGE

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ABSTRACT

As part of our study on “Relationships between seismic properties and rock microstructure”, we have

1. Studied relationships between velocity and permeability.
2. Used independent experimental methods to measure the elastic moduli of clay minerals as functions of pressure and saturation.
3. Applied different statistical methods for characterizing heterogeneity and textures from scanning acoustic microscope (SAM) images of shale microstructures.
4. Analyzed the directional dependence of velocity and attenuation in different reservoir rocks
5. Compared V_p measured under hydrostatic and non-hydrostatic stress conditions in sands.
6. Studied stratification as a source of intrinsic anisotropy in sediments using V_p and statistical methods for characterizing textures in sands.

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INTRODUCTION

In this quarter, we submit results of our research on: impedance mapping on a micrometer scale and its relation to centimeter-scale core measurements of impedance, measurements of clay mineral modulus and the seismic properties of unconsolidated sediments at very low effective pressures. We have also begun work on analyzing the effects of scales on textural, impedance and velocity anisotropy and heterogeneity in consolidated and in unconsolidated materials. Six papers have been published on the work done in this year of the project. This report presents reprints of our publications relevant to this project on impedance mapping, acoustic measurements, and scales analyses.

PAPER 1

Prasad, M., Velocity-permeability relations within hydraulic units.

Paper 1: Total of eleven figures in paper

Velocity-permeability relations within hydraulic units

Manika Prasad*

ABSTRACT

Relationships between seismic velocity and permeability have been difficult to establish. I show that by grouping and sorting rocks into hydraulic units, we can establish relationships between velocity and permeability. The hydraulic units are calculated from measured porosity and permeability values. Correlation between velocity and permeability is significant within each hydraulic unit (the correlation coefficient, R^2 , lies in the range 0.65–0.87). This correlation is an extension of the match between porosity and permeability within a hydraulic unit. I show how the compaction and cementa-

tion history of a sediment can have effects on its physical properties such as porosity and permeability and on its seismic properties. The measured velocity data are further approximated with the Biot model. The velocity-permeability relation and modeling results are demonstrated for a large data set of laboratory measurements. The good match between calculated and measured data demonstrates that this relation can be used to predict permeability from velocity in well logs by zoning the data from training wells into hydraulic units. One possible application is shown where, by using site-specific data, the velocity-permeability relation is vastly improved with a correlation coefficient R^2 of 0.9.

INTRODUCTION

Knowledge about porosity and permeability is essential to evaluate fluid content, fluid flow, and recovery rates in a formation. Information about subsurface formations is generally gathered at three different scales, which vary in resolution, spatial coverage, and number of parameters measured. There is a need to scale up or down to increase reliability of prediction. For example, seismic mapping yields information about subsurface structures and seismic velocities in the formation. It has least resolution (several meters) but maximum coverage (tens of kilometers). This is by far the largest data set available to us, where compressional wave velocity (V_p) is the main measured property. Well logging yields, among other things, information on subsurface formation velocity and porosity. It has medium resolution (several centimeters) and coverage (several meters). The number of measured properties is increased to include, for example, porosity, shale volume, and electrical resistivity, in addition to V_p . In few cases, nuclear magnetic resonance logs and Stoneley wave analyses are used to yield permeability values. Laboratory measurements on cores yield, for example, velocity, porosity, and permeability at discrete depths in the formation. These investigations deliver maximum

resolution (millimeter-micrometer) but minimum coverage (centimeters). A large number of properties can be measured, for example, V_p , shear wave velocity (V_s), P- and S-wave attenuation (Q_p^{-1} and Q_s^{-1}), porosity, permeability, density, mineralogy, and microstructure.

Thus, velocity and porosity are well studied; however, information about permeability, particularly in-situ permeability, is sparse. In absence of measured values of permeability and porosity, proxies are used to infer these parameters from indirect measurements. Attempts to predict permeability from porosity have not been very successful. Similarly, although empirical relations exist for velocity-porosity transforms (for example, Raymer et al., 1980; Nur et al., 1995), establishing a relation between permeability and velocity has met with very little success. It has been shown theoretically that both porosity and permeability control wave propagation in porous rocks (for example, Biot, 1956a, b). Practical application of these theories is precluded by a lack of “correct” choice of the required pore space parameters. A complete description of the pore space parameters will help us to characterize reservoir formations and to understand and model wave propagation mechanisms in porous media. Such a relation will also increase reliability of permeability predictions from velocity.

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Since both velocity and permeability are governed by volumetric *and* geometric considerations, whereas porosity is only a volumetric description, a common link between the three will require some description and quantification of pore space geometry. Amaefule et al. (1993) have shown that a relationship between permeability and porosity exists once the data are separated in hydraulic units. Since velocity depends on pore space volume as well as geometry, I examine the nature of a relation between permeability and velocity. Such a relation is invaluable as a prediction tool. For example, using laboratory data as training data, hydraulic units can be identified in the area of interest. Within these hydraulic units, permeability can be predicted using porosity and velocity from the sonic logs and seismic data. In this paper, I extend the relation between porosity and permeability to the seismic properties of velocity and quality factor using laboratory data. I further show a theoretical basis for this empirical relation between seismic properties, permeability, and porosity.

METHODOLOGY

Many attempts have been made in the past to predict permeability from porosity. Lack of correlation between predicted and measured permeability underscores the difficulties involved in such predictions. The uncertainty is because porosity, by definition, is a volumetric parameter; it is the ratio of pore volume to bulk volume. Permeability, on the other hand, is a measure of the flow properties of a fluid through the pores, which depends not only on the volumetric proportion of the pore space, but also on its geometric distribution and connectivity. Thus, a porosity-permeability relation will be a function of the pore geometry. Various researchers have attempted to model pores with geometrical shapes. The most commonly used porosity-permeability relation is the Kozeny-Carman relation derived for parallel capillary tubes:

$$k = \frac{1}{2\tau^2 S_{Vgr}^2} \frac{\phi^3}{(1-\phi)^2}, \quad (1)$$

where k = permeability (in μm^2), ϕ = fractional porosity, τ = tortuosity, and S_{Vgr} = specific surface area per unit grain volume (Kozeny, 1927; Carman, 1937). In equation (1), the factor 2 accounts for the assumption that the pores are cylindrical with circular cross-sections. Some authors (for example, Paterson, 1983; Walsh and Brace, 1984; Amaefule et al., 1993) generalized this equation to include different pore shapes (F_s). In the notation of Amaefule et al. (1993),

$$k = \frac{1}{F_s \tau^2 S_{Vgr}^2} \frac{\phi^3}{(1-\phi)^2}. \quad (2)$$

Equations (1) and (2) are only partially successful in predicting permeability from porosity. Most pores are not circular cylinders and so equation (1) has limited applicability. Information on τ , S_{Vgr} , and F_s is not readily available and so equation (2) is difficult to apply. Amaefule et al. (1993) have shown that after some rearrangement, equation (2) can be simplified to

$$\log RQI = \log FZI + \log \varepsilon, \quad (3)$$

where ε is the void ratio, the ratio of pore volume to solid volume, defined as

$$\varepsilon = \frac{\phi}{(1-\phi)}, \quad (4)$$

RQI is the reservoir quality index. With permeability expressed in mD, it is defined as

$$RQI = 0.0314 \sqrt{\frac{k}{\phi}}. \quad (5)$$

FZI is the flow zone indicator, expressed as

$$FZI = \frac{1}{\sqrt{F_s \tau S_{Vgr}}}. \quad (6)$$

In equation (3), the parameters which define pore space (F_s , τ , and S_{Vgr}) are grouped together as FZI. Clearly, FZI can be calculated from a set of measured laboratory data on porosity and permeability:

$$FZI = \frac{0.0314}{\varepsilon} \sqrt{\frac{k}{\phi}}. \quad (7)$$

This term can be understood as the relation between the volume of void space (ε) and its geometric distribution (RQI). Rocks with FZI values within a narrow range belong to one hydraulic unit; that is, they have similar flow properties. Amaefule et al. (1993) used this relation to show that samples with similar FZI values plot together on a semilog plot of porosity versus permeability and that the porosity-permeability relation can be defined uniquely in each hydraulic unit. They also show how permeability variations in a reservoir can be understood by binning its cored data in hydraulic units that have similar FZI values calculated using equation (7).

In this paper, I first test this concept of grouping porosity and permeability data according to their FZI units. I then show that this relation can be extended to seismic parameters to create a strong correlation between velocity and permeability. Finally, the FZI units are used in theoretical analysis to model velocity and quality factor.

Porosity analyses

The laboratory data used for this study were compiled by Prasad (1998; 1999) and are given in Appendix A. The data were chosen such that they included velocity and attenuation measurements at similar pressures along with porosity and permeability values.

Figure 1 shows a plot of the porosity-permeability variations in this data set. No apparent relation is observed in Figure 1a. In Figure 1b, the data are color coded according to FZI units calculated from equation (7). The different colors denote FZI intervals as shown by the legend. Separation in hydraulic units (intervals with same FZI values) can be made according to the main FZI intervals present in the data. The solid lines in Figure 1b were calculated from Equation (7) using FZI values marked on the lines. Note that in Figure 1b, lines of constant FZI plot close together at low porosity and are widely separated at higher porosities. Porosity and permeability correlate very well within an FZI interval. At the same porosity, samples with higher FZI have higher permeability. Thus, the FZI values can be understood to denote connectivity; given the same

volumetric amount of pore space, a higher connectivity would produce a larger permeability along with higher FZI.

Velocity analyses

Prasad (1998) has shown that the velocity-porosity relation (Figure 2, open symbols) can be improved by using the modified porosity concept of Galmudi et al. (1999) (Figure 2, closed symbols). Modified porosity takes into account that, in clay-rich rocks, the measured porosity is lower than the total porosity, because the unconnected pore space within clay minerals is underestimated. Galmudi et al. (1999) suggest that the clay porosity (porosity of clay, 0.6, factorized by the clay content) should be added to the measured rock porosity. In Figure 2, the measured porosity data are plotted with open symbols, closed symbols mark modified porosity, modified by assuming a clay porosity of 60% as suggested by Galmudi et al. (1999). Although the scatter in the velocity-modified porosity relation is reduced, it is still significant. Note that the scatter will not decrease by changing the clay porosity. This scatter is due to the geometric distribution of the mineral grains and the pore space. Also plotted in Figure 2 are the upper and lower Hashin-Shtrikman (1963) bounds on velocity in a binary quartz-water system. These bounds describe velocity variations between two endpoints, one at zero porosity, where the velocity is that of quartz, and the other at 100% porosity, where rock velocity is that of water. The lower bound describes the softest arrangement with the water being load-bearing, and the upper bound describes the stiffest arrangement with the quartz being load-bearing. The actual arrangement of the two components can lie between these two extremes. In

Figure 2, the remaining scatter in the velocity-porosity relation after accounting for porosity in the clay minerals is mainly due to the geometric distribution of the grains and pore space.

We now consider the velocity-permeability relation (Figure 3). This figure mimics the porosity-permeability plot in Figure 1. No apparent relation between permeability and velocity is observed in the raw data (Figure 3a). Figure 3b, with colorcoding as in Figure 1b, shows a clear relation between velocity and permeability within narrow confines of FZI units. Separation in different hydraulic units can be made by statistical analyses of the data. The lines in Figure 3b are regression fits between velocity and permeability within a hydraulic unit. Coefficient of determination (R^2) for all lines lies between 0.65 and 0.87. Certainly, narrower intervals could be defined for a larger data set that would improve the correlation. The correlation can also be improved by taking additional constraints, such as lithology, facies, etc.

Attenuation analyses

An important rock property that influences wave propagation is the attenuation of seismic energy. It is a measure of the loss of energy per cycle to peak strain energy stored in the sample. Although attenuation is strongly influenced by pore geometry, a correlation between attenuation and porosity (Figure 4) and between attenuation and permeability (Figure 5) is not as obvious as that for velocity. In Figure 4, although attenuation increases with increasing porosity, the large scatter is not reduced, even after accounting for the additional clay porosity.

FIG. 1. Porosity-permeability relation without color codes (a) and with FZI intervals marked by different colors (b). The legend on the right gives the binning criteria for the hydraulic intervals. The raw data in (a) do not show any apparent relation. In (b), on the other hand, clear relations can be observed between porosity and permeability within narrow FZI intervals. The solid lines mark permeability calculated from equation (7) using FZI values as marked. Note that the color FZI code is in log scale: $\log 10 = 1$, $\log 5 = 0.699$, $\log 2.5 = 0.398$, $\log 1 = 0$, $\log 0.5 = -0.301$, $\log 0.1 = -1.0$.

A relationship between attenuation and permeability is not readily apparent in Figure 5. Although there is some separation

of the data according to the hydraulic units, a clear relation cannot be established. An explanation for this behavior is found by comparing the measured data with modeling results from the Biot theory, as discussed later.

FIG. 2. Velocity-porosity relation. Open symbols denote measured porosity, solid symbols represent porosity modified by accounting for the inaccessible clay porosity. The gray solid line is an exponential fit through the solid symbols. The solid black lines are the Hashin-Shtrikman upper and lower bounds calculated for a quartz-water system. Note that some points lie above the upper bound, because they are either pure carbonates or they contain significant amounts of calcite and/or dolomite.

DISCUSSION

To understand the good correlation between velocity and permeability as observed in Figure 3b, we first examine the FZI unit. By definition, it is a product of pore parameters [shape factor (F_s), tortuosity (τ), and specific surface area to grain volume ratio (S_{vgr})] and describes connectivity between pores (larger FZI implies higher connectivity). In simple terms, it can be understood to describe the initial lithology of a sediment and its compaction and lithification history. In an ideal compaction environment, without sediment influx and cementation effects, porosity reduction is only due to overburden or pressure increase. Porosity-permeability relations follow unambiguous trends governed by initial lithological configurations (for example, grain shape, sorting, clay content). Without external influences, this relation is unique during the compaction history of the sediment. Addition of clay, diagenetic alteration and leaching out of constituent minerals, and cementation of grains, for example, will lead to deviations from this norm. Such deviations will affect both porosity and permeability at higher than normal values. Pore connectivity and FZI values will be high in such samples. On the other hand, clays deposited in pore spaces will reduce porosity and permeability. Pore connectivity and FZI values will be low in such samples.

Figure 6 shows an example of the evolution of marine clay-rich sediments from deposition to burial and lithification.

FIG. 3. Velocity-permeability relation without color codes (a) and with FZI intervals marked by different colors (b). Color codes are same as in Figure 1b. There is a clear relation between velocity and permeability. Solid lines are best fits through all data within each hydraulic unit and have a goodness of fit (R^2) greater than 0.65.

As the sediment is deposited and undergoes compaction, it will move along a normal compaction trend. Without external influences, the porosity reduction (along curve A–B) will be determined only by compaction and volume reduction due to expulsion of water/air from the pores. In Figure 6, line A–B

is an empirical fit to the data. The normal compaction curve (A–B in Figure 6) is constructed from log data for marine

FIG. 4. Attenuation as a function of porosity. Open symbols represent measured porosity, and solid symbols represent porosity modified by accounting for the inaccessible clay porosity. There is large scatter in the data, but low porosity samples appear to have lower attenuation.

FIG. 5. Attenuation-permeability relation. Color codes are same as in Figure 1b. Numbers marks FZI intervals. A clear relation between attenuation ($1/Q_p$) and permeability is not apparent.

silty clay sediments from ODP Leg 161, Site 977 (Shipboard Scientific Party, 1996). Similar compaction curves can be constructed for various sedimentary environments. Within a specific environment, pore space evolution will be more or less similar and the sediments will have the same FZI values. Now, if at a given time (= depth) during its history, the sediment experiences cementation, then the porosity trend will deviate from the normal compaction curve. In Figure 6, point D marks the onset of cementation. At this point, the porosity is “locked-in.” The sediment develops a frame that resists compaction. Porosity reduction will be slower and will move along the hypothetical line D–D’. Similarly, an influx of pore-filling materials in the sediment column will reduce porosity below the normal compaction curve, shown by the hypothetical line C–C’. For both scenarios C–C’ and D–D’, the pore space parameters, or FZI units, will differ from those along the line A–B.

An analogous evolution curve for velocity with depth is shown in Figure 7. Velocity increase with depth progresses along line A–B. Regression curves for velocity logs from the same site as in Figure 6 were used to construct this velocity curve. At the onset of cementation, shown schematically at point D, there is a jump in velocity due to the change in

FIG. 6. Porosity change with depth (A–B). The smoothed data are taken from logging results of Site 977, ODP Leg 161 (Shipboard Scientific Party, 1996). Line C–C’ shows hypothetical porosity-depth relation after influx of pore filling materials at point C, and line D–D’ shows hypothetical porosity changes with depth after initial cementation has locked-in porosity at point D.

compressibility. Now the velocity-depth curve moves along D–D'. Similarly, if there is an influx of pore-filling materials, the velocity in the sediment will move along curve C–C'. The pore-space configuration and, with it, the porosity-permeability and the velocity-permeability relations along the different compaction-cementation histories (A–B, A–C–C', and A–D–D') will be unique. They can be grouped together by the FZI units that account for pore space configurations. These cementation and pore-filling variations in velocity and porosity have been observed, for example, in a silty environment (Prasad and Dvorkin, 2001) and in a carbonate deposition basin (Urmos et al., 1993).

Permeability estimates made with equation (7) with porosity values from Figure 6 and hypothetical FZI values are shown in Figure 8. The dark and light gray curves in Figure 8 are calculated from log porosity and FZI values of 0.1 (light gray) and 0.25 (dark gray) to show their effect on permeability. For example, the cementation event will have a high FZI value, whereas a pore-filling event will lead to a reduction in FZI values. As in the previous Figures 6 and 7, the normal compaction curve is marked by A–B. Early cementation event and subsequent compaction is traced by D–D', whereas C–C' traces a pore-filling event.

Theoretical considerations

I used the Biot theory (Biot, 1956a, b; Stoll, 1980) to model velocity and attenuation for data from Prasad (1998) and for a hypothetical material consisting of quartz particles. Biot's model describes wave propagation in a two-phase system: a porous elastic frame and a viscous, incompressible pore fluid. The losses are due to relative motion between the frame and the pore fluid. Thus, velocity and attenuation are functions of the frame properties, the solid and the pore fluid moduli, and of pore space parameters such as porosity, permeability, pore size, structure constant, and mass-coupling effects. These pore space parameters (for example, structure constant and mass coupling between grains and pore fluid) are difficult to measure. Various studies have attempted to isolate the effect of each of these parameters by keeping all others constant (e.g., Hovem and Ingram, 1979; Dunlop, 1988). The applicability of such an approach is limited because of strong interrelations between the various pore space parameters. I used FZI units to describe pore space parameters.

The calculations presented here were made for quartz mineralogy. The frame modulus was estimated from mineral modulus of quartz and by assuming critical porosity = 36% (Nur et al., 1995). Permeability was calculated from equation (7) for FZI

FIG. 7. Velocity change with depth (A–B). The smoothed data are taken from logging results of Site 977, ODP Leg 161 (Shipboard Scientific Party, 1996). Line C–C' shows hypothetical velocity-depth relation after influx of pore-filling materials at point C and line D–D' marks hypothetical velocity changes with depth at the onset of initial cementation (point D).

FIG. 8. Permeability change with depth (A–B) calculated from log porosity using an FZI value of 0.25. C–C' shows hypothetical permeability-depth relation after influx of pore-filling materials at point C where FZI is decreased to 0.1. Line D–D' marks a hypothetical permeability change with depth at the onset of initial cementation at point D.

values of 0.1, 0.25, 0.75, and 2.5 and porosity values varying between 5% and 35%. The pore-filling medium was assumed to be water. Input parameters for the model are given in Table 1.

The velocity and attenuation variations with permeability are shown in Figures 9 and 10, respectively. The color codes in both figures are same as in Figure 1. There is good agreement between the calculated velocity (solid lines) and the data from Figure 3 (symbols). Within each hydraulic unit, the model predicts velocity variations with permeability fairly well. Although the theoretical attenuation predictions do not appear very successful, they reveal important insights. As noted by Winkler (1985) and Prasad and Meissner (1992), the Biot model predicts attenuation values that are much lower than measured values, mainly because additional loss mechanisms such as frame and squirt-flow attenuations are also active. The model predicts higher losses for larger FZI values. In the model, for each FZI value, there is a peak in attenuation at a certain permeability value. With decreasing FZI value, the

attenuation peak decreases and moves towards lower permeability values. Thus with all other parameters kept constant, position and magnitude of this peak is governed by the pore space parameters, represented by FZI values. This is understandable, since the Biot loss mechanism is due to relative flow between solid grains and pore fluid. Having additional permeability and attenuation data available should allow us to better define the nature of the relation between them. Specifically, by measuring samples with same FZI but controlled permeability (for example, in pressure dependent velocity, attenuation, and permeability measurements), we should be able to track such changes. Furthermore, combining Biot losses with frame and squirt-flow loss mechanisms will allow us to make realistic estimates of attenuation in rocks.

APPLICATIONS

In Figure 11, I show the velocity-permeability trend from Figure 3 again, but with more data values added to the plot. These data are taken from laboratory measurements of core samples from three different wells. Although the data are from very different environments, the velocity-permeability variation is governed by same pore space geometry. From Figure 11, we can derive the following information:

- 1) All data fall within narrow confines of predefined hydraulic units.
- 2) The V_p -permeability relation can be improved for each specific site. This is mainly achieved by narrowing down the hydraulic unit boundaries. An exponential fit through a subset of the data around $FZI = 1.0$ had a correlation coefficient (R^2) = 0.92.
- 3) Such data can be used to establish flow properties within a reservoir. Then, using the velocity-permeability relation

Table 1. Biot equations for longitudinal waves as functions of strain of the frame (ϵ) and displacement of the fluid relative to the frame (ξ), and list of input parameters used in these equations.

Biot equations for longitudinal waves		
$\Delta^2(\mathbf{H}\epsilon - \mathbf{C}\xi) = \partial^2/\partial t^2(\rho_e\epsilon - \rho_f\xi)$		
$\Delta^2(\mathbf{C}\epsilon - \mathbf{M}\xi) = (\rho_f\epsilon - m\xi) - \eta\mathbf{F}(\kappa)/k\partial\xi/\partial t$		
H, C, M are elastic moduli, defined by Stoll (1980) as		
$H = \frac{(K_s - K_{dry})^2}{D - K_{dry}} + K_{dry} + \frac{4}{3}\mu, \quad C = \frac{K_s(K_s - K_{dry})}{D - K_{dry}},$		
and $M = \frac{K_s^2}{D - K_{dry}}$, where $D = K_s(1 + \phi \frac{K_s}{K_f} - \phi)$		
Parameter	Description	Value or equation used
ϕ	Fractional porosity	Between 0.1 and 0.35
ϕ_c	Critical porosity	0.36 (Nur et al., 1995)
ρ_e	Bulk density	$(1-\phi)\rho_s + \phi\rho_f$
ρ_f	Fluid density	1.0 g/cm ³
ρ_s	Mineral density	2.65 g/cm ³
ρ_{dry}	Dry frame density	$(1-\phi)\rho_s$
K_s	Mineral bulk modulus	37 MPa
μ_s	Mineral shear modulus	44 MPa
K_f	Fluid bulk modulus	2.22 MPa
η	Fluid viscosity	0.2 cp
f	Frequency	10 Hz, 1 kHz, 1 MHz
ω	Circular frequency	$= 2\pi f$
K_{dry}	Frame bulk modulus	$K_{dry} = K_s(1-\phi/\phi_c)$ (Nur et al., 1995)
μ_{dry}	Frame shear modulus	$\mu_{dry} = \mu_s(1-\phi/\phi_c)$ (Nur et al., 1995)
FZI	Flow zone indicator	Between 0.1 and 2.5
k	Permeability (mD)	Calculated with equation (7)
a	Pore size	$a = 1.8081 \times k^{0.3932}$
m	Mass coupling factor	$m = c\rho_f/\phi$
c	Shape factor	$c = 1/FZI$
$F(\kappa)$	High-frequency correction factor	
κ		$a \text{ sqrt } (\omega\rho_f/\eta)$

FIG. 9. Velocity-permeability relation. Color codes are same as in Figure 1b. Symbols represent measured data, solid lines mark values calculated by the Biot model. The calculations were made for FZI values of 2.5, 0.75, 0.25, and 0.1 (corresponding to log FZI of 0.398, -0.125, -0.602, -1.0). There is a remarkable agreement between measured and predicted values.

(such as the one established in Figure 11), we can predict and control changes either spatially (in the case of a new field) or temporally (in a producing field).

FIG. 10. Attenuation-permeability relation. Color codes are same as in Figure 1b. Symbols represent measured data, solid lines mark values calculated by the Biot model. The different lines represent calculations for FZI values as marked by numbers on the lines. The calculations were made for FZI values of 2.5, 0.75, 0.25, and 0.1 (corresponding to log FZI of 0.398, -0.125, -0.602, -1.0). Although, the model predictions lie within measured range for FZI = 2.5 (blue line), they are much lower for the lower FZI values.

FIG. 11. Velocity-permeability relation for additional data sets from different environments. Color codes are same as in Figure 1b. The new data are from a tight sand reservoir, marine reservoir sand, and from a data collection of reservoir type sandstones. The black line marks an exponential fit through a subset of the data around FZI = 1.0 with log permeability (mD) = $-0.002999 \times V_p \text{ (m/s)} + 15.55$; $R^2 = 0.92$.

CONCLUSIONS

Using a collection of velocity, porosity, and permeability data from various sources, I have shown that, by grouping the data in different hydraulic units based on pore space properties, a positive correlation between velocity and permeability can be established. The pore space parameters, grouped together as FZI values, are calculated from measured values of porosity and permeability. By using FZI values for pore space properties, velocity modeled with the Biot theory agrees well with measured data and attenuation is underpredicted. I further show how, by defining hydraulic units for a specific site, flow properties can be controlled and predicted.

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APPENDIX A

Data used for this study were reported in Prasad (1998) and are listed in Table A-1. The source numbers in Table A-1 correspond to the following authors: 1 = Assefa and Sothcott (1997), 2 = Prasad et al. (1999), 3 = Klimentos

and McCann (1990), 4 = Lucet (1989), 5 = Prasad and Manghnani (1997), 6 = Tao et al. (1995). All values in the table are given for effective pressures around 40 MPa.

Table A-1. Data used for this study.

Source number	Sample lithology	Porosity (%)	Clay (%)	V_p (m/s)	Q_p	Permeability (mD)
1	dolerite	0.88	1.96	5885	53	0.003
1	gneiss	0.95	4.99	5532	74	0.004
1	gneiss	1.59	0	5931	129	0
1	sandstone	2.8	6.48	4601	18	0
1	sandstone	3.41	6.76	5171	97	0.02
1	sandstone	3.46	0	5091	62	0.005
1	sandstone	4.89	0	5048	42	0.02
1	basalt	4.99	3.17	4608	15	0
1	siltstone	5.71	6.99	4354	24	0
1	sandstone	6.36	26.2	4695	17	0.06
1	basalt	7.08	7.26	4436	21	0.39
1	sandstone	7.52	0	4735	37	0.02
1	sandstone	9.79	9.75	3917	22	0.28
1	basalt	11.08	11.4	3832	24	0.03
1	basalt	13.37	7.28	3824	31	0.002
1	sandstone	13.4	0	4126	33	2.51
1	siltstone	14.23	42	3636	30	0.06
1	siltstone	16.93	25.83	3531	52	0.31
1	sandstone	17.35	2.71	3780	31	0.24
2	sandstone	19.1	10	3488	8	1.4
2	sandstone	23.2	10	3580	34	22
2	sandstone	23.4	5	3751	19	295
2	sandstone	23.9	9	3601	17	68
2	sandstone	25.2	9	3466	13	818
2	sandstone	26.9	4	3726	21	1030
3	sandstone	2.43	0	5835	585	0
3	sandstone	2.72	0	5934	4599	0
3	sandstone	5.98	3	5225	275	0
3	sandstone	8.96	6	4947	22	0.13
3	sandstone	9.96	7	4705	32	0.01
3	sandstone	10.22	9	4895	12	0.16
3	sandstone	11.39	6	4666	37	0.46
3	sandstone	13.11	7	4666	28	3.67
3	sandstone	13.47	14	4498	12	0.06
3	sandstone	13.72	0.5	4950	613	87.65
3	sandstone	14.15	1	4960	79	150.7
3	sandstone	14.37	1	5078	185	255.9

(Continued)

Table A-1. (Continued).

Source number	Sample lithology	Porosity (%)	Clay (%)	V_p (m/s)	Q_p	Permeability (mD)
3	sandstone	14.47	0.2	4788	712	220.9
3	sandstone	15.13	4	4794	34	11.06
3	sandstone	15.18	0.7	4942	394	160.4
3	sandstone	15.41	15	4246	19	52.42
3	sandstone	15.46	15	4152	21	0.05
3	sandstone	15.72	5	4564	127	87.55
3	sandstone	16.11	15	4152	20	50.51
3	sandstone	16.5	15	4149	18	41.74
3	sandstone	16.65	12	4010	29	0.37
3	sandstone	16.71	8	4381	40	0.44
3	sandstone	17.13	12	3933	26	2.21
3	sandstone	17.18	15	4362	9	0.13
3	sandstone	21.19	15	3700	31	11.42
3	sandstone	21.73	22	3572	22	7.1
3	sandstone	22.2	25	3500	10	1.13
3	sandstone	25.41	20	3314	20	5.78
3	sandstone	26.32	20	3299	25	10.27
3	sandstone	27.12	25	3000	11	0.45
3	sandstone	27.33	5	3666	27	305.8
3	sandstone	27.39	23	2952	10	9.3
3	sandstone	27.73	17	3090	19	3.5
3	sandstone	27.87	17	2990	18	9.59
3	sandstone	27.96	15	3675	18	33.67
3	sandstone	28.04	20	3181	18	10.05
3	sandstone	28.79	18	3056	20	5.47
3	sandstone	28.87	23	2909	19	7.03
3	sandstone	30.31	12	3627	17	17.77
3	sandstone	33.59	15	3195	16	2.25
3	sandstone	35.09	30	3108	10	73.26
3	sandstone	36.04	16	3121	19	21.16
4	granite	0.5		5851	74	0.1
4	marble	0.5		7007	189	0.1
4	limestone	6.25		5769	352	0.1
4	limestone	11.1		5019	85	0.2
4	limestone	12.2		5285	18	45
4	limestone	12.2		5437	19	45
4	limestone	12.2		5270	16	45
4	limestone	13.2		5313	15	4.5
4	limestone	13.4		4647	71	0.3
4	limestone	17.8		4593	12	380
4	limestone	17.8		4593	12	380
4	sandstone	20.3		3754	33	8
4	sandstone	20.3		3707	31	8
4	sandstone	20.3		3771	41	8
4	sandstone	20.9		3867		2040
4	sandstone	20.9		3976	48	2040
4	sandstone	20.9		3903		2040
4	sandstone	20.9		3921		2040
4	limestone	21.2		4060	57	4
4	limestone	22		3819	55	176
4	limestone	23.8		3872	11	405
4	sandstone	24.6		3572	38	58
4	limestone	26.3		3397	21	208
4	limestone	28.2		4172	18	145
4	limestone	29.2				106
4	limestone	30.8		4218	53	4
4	limestone	31.15				87
5	sandstone	21	4	4400	100	100
5	sandstone	21	4	4400	100	100
6	sandstone	17	<1	4460	65	1170.7
6	sandstone	17.1	1	4300	35	587.8
6	sandstone	22.1	1	3770	59	794.2
6	sandstone	25.9	3	3060	26	977.1

PAPER 2

Vanorio, T., Prasad, M., and Nur, A., Elastic properties of dry clay mineral aggregates, suspensions and sandstones.

Paper 2: Total of nine figures in paper

Elastic properties of dry clay mineral aggregates, suspensions and sandstones

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SUMMARY

The presence of clay minerals can alter the elastic behaviour of rocks significantly. Although clay minerals are common in sedimentary formations and seismic measurements are our main tools for studying subsurface lithologies, measurements of elastic properties of clay minerals have proven difficult. Theoretical values for the bulk modulus of clay are reported between 20 and 50 GPa. The only published experimental measurement of Young's modulus in a clay mineral using atomic force acoustic microscopy (AFAM) gave a much lower value of 6.2 GPa. This study has concentrated on using independent experimental methods to measure the elastic moduli of clay minerals as functions of pressure and saturation. First, ultrasonic *P*- and *S*-wave velocities were measured as functions of hydrostatic pressure in cold-pressed clay aggregates with porosity and grain density ranging from 4 to 43 per cent and 2.13 to 2.83 g cm⁻³, respectively. In the second experiment, *P*- and *S*-wave velocities in clay powders were measured under uniaxial stresses compaction. In the third experiment, *P*-wave velocity and attenuation in a kaolinite–water suspension with clay concentrations between 0 and 60 per cent were measured at ambient conditions. Our elastic moduli measurements of kaolinite, montmorillonite and smectite are consistent for all experiments and with reported AFAM measurements on a nanometre scale. The bulk modulus values of the solid clay phase (K_s) lie between 6 and 12 GPa and shear (μ_s) modulus values vary between 4 and 6 GPa. A comparison is made between the accuracy of velocity prediction in shaley sandstones and clay–water and clay–sand mixtures using the values measured in this study and those from theoretical models. Using $K_s = 12$ GPa and $\mu_s = 6$ GPa from this study, the models give a much better prediction both of experimental velocity reduction due to increase in clay content in sandstones and velocity measurements in a kaolinite–water suspension.

Key words: acoustic velocities, clay minerals, elastic moduli, porosity.

1 INTRODUCTION

Elastic properties of clay minerals are of great interest for exploration and characterization of shaley formations. Knowledge of elastic moduli of clay minerals is required because: (1) clay minerals are found in most sedimentary environments; (2) clay mineral transitions are considered as important factors in seismogenic zones; and (3) seismic interpretation and modelling seismic response requires knowledge on how the presence of clay changes the physical properties of rocks.

Laboratory data from several investigators (Tosaya 1982; Tosaya & Nur 1982; Kowallis *et al.* 1984; Han 1986; Klimentos 1991) report a systematic decrease of *P*- and *S*-wave velocity with increasing clay content in both well and poorly consolidated sediments. How-

ever, the empirical relations derived from these measurements are restricted to specific lithologies and thus resist generalization. On the other hand, the use of deterministic models to establishing porosity–velocity relations in shaley sandstones requires knowledge of input parameters such as elastic moduli of the solid clay phase.

Elastic properties of clay minerals are almost unknown, mainly because of the difficulty presented by their intrinsic properties. Their small grain size makes it impossible to isolate an individual crystal of clay large enough to measure acoustic properties; reactions with organic polar molecules (Theng 1974) affects handling and clay structure; and low permeability makes saturation difficult. So far the effective elastic properties of clays have been derived either by theoretical computation (Alexandrov & Ryzhova 1961; Berge & Berryman 1995; Katahara 1996), by a combination of theoretical and experimental investigations on clay–epoxy mixtures (Wang *et al.* 2001), or by empirical extrapolations from measurements on shales (Tosaya 1982; Castagna *et al.* 1995; Han 1986). These derived

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values of clay moduli show little agreement. The values reported by Katahara (1996) and Wang *et al.* (2001) (*ca.* 50 GPa for bulk modulus, K and 20 GPa for shear modulus, μ) are higher than those extrapolated (*ca.* 20 GPa for K and 7 GPa for μ) from both laboratory (Castagna *et al.* 1995; Han 1986; Tosaya 1982) and well log measurements in shales and laminated shaley-sands (Eastwood & Castagna 1987). Katahara (1996) explained this discrepancy by acoustic anisotropy: in shaley beds, clay particles tend to align parallel to bedding during compaction. Using theoretical models of shaley sandstones, Berge & Berryman (1995) have shown that the bulk modulus of clay should be around 10–12 GPa. Only one experimental measurement exists on pure clay minerals: Prasad *et al.* (2002) have reported measurements of Young's modulus on clay particles using atomic force acoustic microscopy. Their measured value of Young's modulus for dickite is 6.2 GPa.

We have made direct measurements of the elastic properties of clay minerals by designing experiments that took into account factors affecting clay structures. We used three independent experimental set-ups: (1) a method of obtaining dry clay aggregates having different porosities on which P - and S -velocity (V_p , V_s , respectively) measurements were made as functions of hydrostatic pressure; (2) V_p and V_s measurements on clay powder as functions of uniaxial stress; and (3) P velocity and attenuation (V_p and Q_p , respectively) measurements on water–clay suspension as functions of clay content.

The measurements of elastic moduli for the solid clay phase from all experiments are consistent with each other. In order to verify the reliability of results, the measured moduli values were used in effective medium and granular porous media models and compared with pre-existing data on shaley sandstones (Tosaya 1982; Han 1986; Klimentos 1991). We demonstrate that our measured values of the elastic moduli for clay can be used to predict velocity data in saturated kaolinite samples (Yin 1993), kaolinite–water suspensions (this study) and shaley sandstones (Tosaya 1982; Han 1986; Klimentos 1991).

2 EXPERIMENTAL PROCEDURES AND METHODS

2.1 Collection of clay samples

Samples used in this study were prepared from clay powders obtained from the repository of the Clay Minerals Society. Clay samples included: kaolinite (KGa-1b, reported in the paper as Ka1), Ca-montmorillonite (STx-1, reported as Ca-Mt), Na-montmorillonite (SWy-2, reported as Na-Mt), illite (IMt1, reported as Imt), smectite (SWa-1, reported as Sm) and chlorite (CCa-2, reported as Ch).

2.2 Methods

Our methods of measuring elastic moduli of solid clays were chosen taking into account the various factors affecting clay structures. Brace *et al.* (1969) and Chung (1979) have described a hot-pressing method to determine elastic properties of some mineral powders. This method cannot be used for clays because they are sensitive to temperature and pressure. As temperature increases, clays first dehydrate and then dehydroxylate changing, respectively, their adsorbed water content and their mineral structure. The dehydration temperature depends on pressure (Koster Van Groos & Guggenheim 1984; Colten Bradley 1987). Chemical reagents cannot be used for sample preparation because some clay minerals change structure due to

reaction with polar compounds. These reactions can be exothermic and the temperature may rise to such a level as to overheat the clay specimen (Theng 1974; Newman 1987). The effects of reactions on clay minerals depend largely on their structure (Theng 1974). The basal spacing of the $\text{Si}(\text{O},\text{OH})_4$ -tetrahedra (T) and $\text{M}_{2-3}(\text{OH})_6$ -octahedra (O) layers (where M is either a divalent or a trivalent cation) as well as size, the valence and the charge of the embedded exchangeable cations profoundly influence adsorption of polar molecules other than water. For example, kaolinite shows a fully collapsed structure that is wanting either in water molecules or in exchangeable cations or both. In these structures, intercalation of polar organic liquids or gases is quite complex and adsorption occurs predominantly on the external crystal surfaces (Theng 1974). On the other hand, smectite-type minerals, which host both exchangeable cations and water interlayers, adsorb polar organic molecules by replacing their water interlayers. During the internal adsorption, the organic compound competes with water for the same sites around the exchangeable cation. The removal and replacement of the water interlayer can be so invasive as to provoke dissociation of clay layers (Theng 1974).

An understanding of how these factors affect acoustic measurements is lacking. Since all the above-mentioned effects alter the clay structure, we expect an accompanying change in elastic properties as well. To eliminate these secondary effects, we used the cold-pressing method in which measurements are conducted on pure clay powders pressed at ambient room temperatures. Relatively low-pressure conditions (up to 20 MPa) ensured that the applied pressures did not promote dehydration of the clay minerals. Similarly, in the suspension experiment, a KCl solution was used with kaolinite to limit the effects of adsorption of polar water molecules and the cation exchange capacity of the clay.

2.2.1 Porosity and grain density determination

The grain density of the clay powders was measured using a pycnometer. To avoid trapped air produced by clay flocculation, we used sodium hexametaphosphate in distilled water (40 g of solute/litre) as the dispersive agent (ASTM 1992). For the cold-pressed aggregate samples, porosity, bulk and grain densities were measured at room P – T conditions with a helium porosimeter (with a measurement error of about 1 per cent).

2.2.2 Cold-pressing experiment

Sample preparation. This set-up was designed to make clay aggregate samples from the powders described in Section 2.1. These samples were used for measurements of V_p and V_s as functions of hydrostatic pressure. The equipment consists of a split steel cube with a cylindrical cavity (Fig. 1). Two pistons on either side of this cavity are used to apply uniaxial stress on a powder placed in the cavity. Porous stones coupled to filter paper on either end prevented the samples from bonding to the endplates during the cold-pressing procedure. The split steel cube allows sample recovery without applying additional stress. Powders were cold-pressed at 15 MPa for 1, 7 and 10 d to obtain samples of different porosity. Porosity, density and acoustic properties were measured on these samples at ambient conditions and at different confining pressure (up to 20 MPa).

Acoustic measurements. A complete description of the experimental set-up used in this study can be found in Prasad & Manghnani (1997) and Vanorio *et al.* (2002). Measurements were performed by increasing the confining pressure up to 20 MPa with a pressure

Figure 1. Different stages of the cold-pressing procedure.

interval of 1 MPa. During the down cycle, measurements were taken with a pressure interval of ≈ 5 MPa. Three linear potentiometers were used to measure length changes of the samples as a function of stress. The length change was directly related to changes in porosity by assuming that pore collapse was the main cause of strain. This allowed computation of porosity change with pressure using initial porosity, grain density and length change. V_p and V_s were measured at each pressure step. Traveltimes were measured after digitizing waveforms with 1024 points at a time sweep of 5 μ s, thus allowing a time resolution of about 5 ns or about 0.2 per cent error in velocity. The actual error in velocity measurement is estimated to be around 1 per cent due to operator error in picking the first arrival time. The system delay time was measured by taking the head-to-head time at 2 MPa. The traveltimes were confirmed by measuring an aluminium sample at different pressures.

2.2.3 Uniaxial stress experiment

Samples preparation. Clay powder of kaolinite (KGa-1b), Na-montmorillonite (SWy-2) and Ca-montmorillonite (STx-1) were used for this experiment. The cold-pressing set-up was altered to allow the simultaneous measurement of acoustic velocities during the uniaxial stress compaction of clay powders. PZT crystals with a principal frequency of 1 MHz mounted on the piston endplates were used to generate P and S waves. A high-viscosity bonding medium (Panametrics SWC) was used to bond the endplates to the sample. Acoustic isolation from steel was achieved by jacketing the samples with rubber tubing.

Acoustic measurements. The experimental setup for the uniaxial pressure tests consisted of a digital oscilloscope (Tektronix TDS 420A) and a pulse generator (Panametrics 5052 PR). Measurements were performed by increasing pressure up to 20 MPa with a variable pressure interval from 2.5 to 5 MPa. A caliper was used to measure length changes of the samples as a function of stress (precision = ± 1 mm). The length measurements were made at eight positions around the plate in order to detect possible sample tilting. Computations of both porosity change and traveltimes were made as described in Section 2.2.2. In this experiment, porosities might be underestimated due to deformation of the rubber jacket. However, comparison with our other results and the modelling in Section 4 show this deformation to be minimal. This assumption was further tested using a water sample in the uniaxial experiment because of its highest Poisson ratio value. The test demonstrated that the deformation of the rubber jacket led to a diameter expansion of 1.5 mm. Assuming a

similar behaviour for clays, this corresponds to a porosity error of less than 7 per cent.

2.2.4 Suspension experiment

Sample preparation. Clay powder of kaolinite (KGa-1b), a non-swelling clay, was used in the experiments to limit the effects of polar water molecule adsorption (Theng 1974). To further minimize swelling and the cation exchange capacity, the suspension was prepared in a degassed, 20 000 ppm KCl solution. The size of the K-ion radius, of the order of the kaolinite layer spacing, prevented both cation exchange (Theng 1974) and water molecules from bounding to any cation in the interlayer position. Measured amounts of clay were added to the water to allow computations of clay content and water saturation. To avoid clumps in the suspension, the powder was sieved before adding it to the water. Trapped air was minimized by constant stirring, shaking and by using vacuum to de-air the suspension. Acoustic measurements were performed while increasing clay content in the suspension from 0 to 60 per cent using a step of 2 per cent for clay contents less than 40 per cent, and a step of 5 per cent for clay content above 40 per cent. Measurements were discontinued for clay concentrations greater than 60 per cent: the sample was too viscous to permit a complete outgassing and velocity estimates would be biased by gas content.

Acoustic measurements. The experimental setup consisted of a Plexiglas container that held the kaolinite–water suspension (Fig. 2). Sealed holes on two sides of the container positioned the immersion transducers (1 MHz Panametrics Transducers V303-SU) in the kaolinite–water suspension. In order to avoid reflections from the container sides, the distance between immersion transducers (2.54 cm) was kept shorter than the dimensions of the Plexiglas container ($27 \times 10 \times 27$ cm³). The distance between transducers was fixed during the measurements and was used for velocity computation. The transducers were excited using a Panametrics (5052 PR) pulser. The signals through the suspension were received and digitized by a digital Tektronix oscilloscope (TDS 420 A) and stored on a computer. At each clay concentration two signal acquisitions were taken to ensure the repeatability of measurements. The P -wave quality factor (Q_p) (± 5 per cent) was calculated from the stored signal waveforms by the spectral ratio technique (Toksöz *et al.* 1979; Prasad & Manghnani 1997). The signal from the initial KCl water

Figure 2. Suspension measurement equipment for kaolinite sample. Note that the active faces of the transducers are inside the container.

solution was recorded and used as a reference signal. In this way similar conditions were maintained for the reference and the sample.

3 RESULTS

3.1 Grain density, porosity and sample state

Table 1 compares grain and bulk density, and porosity values measured using the methods described in Section 2.2.1.

Differences in grain densities values for most of samples were less than 2 per cent except for Na-montmorillonite samples (≈ 7 per cent). Swelling of this clay mineral, which affects the volume estimate, was noted during the measurement. For Na-montmorillonite samples we will refer to physical properties measured by the He-porosimeter.

Table 1 also shows that clay mineralogy controls both porosity and the state of samples obtained by the cold-pressing procedure described in Section 2.2.2. Upon removal of samples from the cold-pressing apparatus, well-compacted low-porosity ($4 < \Phi < 20$ per cent) samples of Na-, Ca-montmorillonite (Na-Mt, Ca-Mt), and smectite (Sm) were obtained (Fig. 3a). In contrast, illite (Imt) and chlorite (Ch) samples presented higher porosity ($20 < \Phi < 30$ per cent) and slowly deteriorated with microcracks forming perpendicular to the applied stress direction (for example, sample Ch in Fig. 3b). This effect was most pronounced in kaolinite samples, which completely disaggregated on removal from the housing apparatus (sample Ka1 in Fig. 3b). Since deterioration of samples was typical of clays without water interlayers, and it occurred both at room humidity and under vacuum (samples Ka1 and Ch), this behaviour could be related to clay mineralogy. This hypothesis was confirmed by cold-pressing two dehydrated powders from the smectite group which, in their natural hydrated state, produced good samples (Fig. 3a). Fig. 3(b) shows that the dehydrated Sm and Ca-Mt samples disaggregated upon the removal of the cold-pressing pressure. The results from the cold-pressing experiment show that water interlayers in the clay mineral structure play an important role in the compaction and strength of clay aggregates.

3.2 Acoustic measurements on cold-pressed samples under hydrostatic pressure

Because of the lowest porosity of the suite of studied samples, samples of smectite and Na-montmorillonite were used for the hydro-

Figure 3. (a) Aggregated samples of clay minerals obtained by the cold-pressing procedure: kaolinite (Ka), chlorite (Ch), Na-montmorillonite (Na-Mt), Ca-montmorillonite (Ca-Mt), smectite (Sm), illite (Imt), zeol (zeolites); (b) samples completely disaggregated upon removal of cold-pressing pressure (Ka1) and with microcracks formed perpendicular to the applied stress direction (sample Ch). In the box, samples of dehydrated Sm and Ca-Mt are shown disaggregated upon the removal of cold-pressing pressure (b). Arrows connect the same powders in their natural and in dehydrated states.

static pressure experiment. The experiment was repeated on different samples to ensure repeatability of measurements. In Figs 4 and 5, measurements of velocity and elastic moduli are reported for the smectite sample as functions of pressure and porosity, respectively. Because of the initial low porosity of the smectite sample (4.5 per cent), it was possible to measure velocities and elastic moduli at porosities as low as 0.8 per cent at a confining pressure of 20 MPa. Extrapolation using a polynomial fit to the data (Table 2) shows that, at 0 per cent porosity, $V_p = 2.3 \text{ km s}^{-1}$, $V_s = 1.3 \text{ km s}^{-1}$, $K = 6 \text{ GPa}$ and $\mu = 4 \text{ GPa}$. Fig. 4 shows that velocity values measured during unloading are higher than those measured during increasing pressure. Although the sample is allowed to re-equilibrate after pressure release, porosity is not readily recovered in clay and, as consequence, the velocity is higher. As pressure decreases, this effect is less evident since the sample has had enough time to re-equilibrate; below 5 MPa the time-lag in recovering of porosity is less pronounced.

Table 1. Porosity, bulk (ρ_b) and grain density (ρ_g) of cold-pressed clay samples.

Sample	Pycnometer			He-porosimeter			Sample	Pycnometer			He-Porosimeter		
	ρ_g (g cm^{-3})	Φ (per cent)	ρ_b (g cm^{-3})	ρ_g (g cm^{-3})	Φ (per cent)	ρ_b (g cm^{-3})		ρ_g (g cm^{-3})	Φ (per cent)	ρ_b (g cm^{-3})	ρ_g (g cm^{-3})	Φ (per cent)	ρ_b (g cm^{-3})
Ka1	2.59	43	1.48	Nm ^a	Nm	Nm	Sm3	2.29	6	2.15	2.28	6	2.15
Ka2	2.59	43	1.48	Nm	Nm	Nm	Sm4	2.29	4	2.19	2.30	5	2.19
Na-Mt1	2.30	13	2.00	2.47	11	2.19	Sm5	2.29	22	1.80	2.30	22	1.80
Na-Mt2	2.30	9	2.09	2.44	10	2.19	Sm6	2.29	10	2.05	2.29	10	2.05
Na-Mt3	2.30	14	1.99	2.47	20	1.99	Ca-Mt-1	2.13	22	1.68	2.14	22	1.68
Na-Mt4	2.30	12	2.02	2.48	18	2.02	Ca-Mt-2	2.13	21	1.69	2.13	21	1.69
Na-Mt5	2.30	11	2.05	2.47	17	2.05	Ca-Mt-3	2.13	17	1.76	2.13	17	1.76
Na-Mt6	2.30	19	1.85	2.47	25	1.85	Ca-Mt-4	2.13	21	1.67	2.12	21	1.67
Sm1	2.29	8	2.11	2.29	7	2.13	CH3	2.83	25	2.03	Nm	Nm	Nm
Sm2	2.29	13	1.99	2.29	13	1.99	Imt1	2.68	22	2.10	Nm	Nm	Nm

Nm = samples not measurable.

Figure 4. Velocity (a) and elastic moduli (b) as functions of hydrostatic pressure for smectite. Solid symbols mark increasing and open symbols mark decreasing pressures. The error on velocity is 1 per cent; the error bar in the velocity plots is smaller than the size of the symbols.

3.3 Acoustic measurements under the uniaxial stress condition

In this experiment we monitored velocity and elastic moduli variations during uniaxial stress compaction of kaolinite and Na-montmorillonite powders. Fig. 6 shows the variations in velocity and elastic moduli as functions of porosity for kaolinite (Ka1) and montmorillonite (Na-Mt) in dry condition. At the lowest porosity, a linear fit to the data gives elastic moduli for kaolinite mineral as 11 GPa for K and 6 GPa for μ , respectively. Measurements on Na-montmorillonite provided moduli of 11 GPa for K and 5.5 GPa for μ (Table 2). The different trends shown in the hydrostatic and uniaxial stress compaction experiments could be due to the dissimilar aggregation state of the samples used. In the uniaxial experiment, we capture the entire compaction curve from the clay powder, having negligible pore stiffness, to very low porosity aggregate, whereas in the hydrostatic case, since we have pre-compressed the Sm sample, we are looking at the lowermost porosity of the compaction curve. Future measurements will show whether and how the velocity–porosity trend deviates from linearity at very low porosity.

3.4 Acoustic measurements in kaolinite–water suspension

Fig. 7 shows the variation of bulk modulus (Fig. 7a) and quality factor (Fig. 7b) as functions of clay content in the suspension. The bulk modulus was calculated from V_p and density measurements by assuming $\mu = 0$, that is, the shear wave did not propagate through the suspension. In Fig. 7(a), at clay concentrations below 40 per cent, the bulk modulus (K) increases slightly with increasing clay

Figure 5. Velocity (a) and elastic moduli (b) as functions of porosity for the smectite. Lines show polynomial fits to the data: $M = a\Phi^4 + b\Phi^3 + c\Phi^2 + d\Phi + e$, where Φ = porosity.

Table 2. Coefficients of polynomial curves fit to the data.

Smectite	$M = a\Phi^4 + b\Phi^3 + c\Phi^2 + d\Phi + e$					
	a	b	c	d	e	R^2
V_p	0.011	−0.138	0.613	−1.184	2.321	(0.986)
V_s	0.005	−0.054	0.242	−0.471	1.324	(0.979)
K	0.026	−0.310	1.355	−2.603	5.75	(0.984)
μ	0.049	−0.601	2.682	−5.212	4.02	(0.988)
Kaolinite and Na-Montmorillonite						
M	a	b	R^2	a	b	R^2
V_p	−0.03	2.695	(0.971)	−0.05	2.748	(0.974)
V_s	−0.01	1.436	(0.952)	−0.02	1.496	(0.956)
K	−0.21	11	(0.967)	−0.35	11	(0.994)
μ	−0.10	6	(0.969)	−0.15	5.5	(0.990)

content. Beyond 40 per cent there is a change in the slope of the bulk modulus increase. In Fig. 7(b), the P -wave quality factor (Q_p) decreases with increasing clay content and reaches a minimum at 40 per cent clay concentration. After this minimum, Q_p increases with increasing clay concentration.

The behaviour of K as a function of increasing clay content is similar to the results of V_p in clay–sand mixtures reported by Marion *et al.* (1992). Marion *et al.* (1992) show that when clay volume fraction (C) is less than 0.4–0.45 (as in Fig. 7a), clay particles are suspended in the pore space of the load-bearing medium (sand in their case, water in this study) and velocity measurements reflect mainly a density effect. Q_p decreases due to scattering at possible clumping of the clay. At clay volume fractions greater than 0.4–0.45, clay becomes the load-bearing phase and water (or sand) becomes

Figure 6. Velocity (a) and elastic moduli (b) as functions of porosity from uniaxial stress tests for kaolinite (Ka, filled circles) and Na-montmorillonite (Na-Mt, crosses) in dry conditions. Lines show linear fits to the data: $M = a\Phi + b$. The error on velocity is 1 per cent; the error bar in the velocity plots is smaller than the size of symbols used.

Figure 7. Bulk modulus (a) and P -wave quality factor (b) as functions of clay content in a kaolinite–KCl suspension.

dispersed in the pore space of the load-bearing clay. Above this point, the increased stiffness due to the clay outweighs density and scattering effects and both K and Q_p increase with increasing clay content. Note that although degassing under vacuum was problematic above 40 per cent clay content, we do not expect a significant quantity of gas bubbles to be present. The presence of gas in the

suspension would lower V_p and it would lower Q_p significantly. The increase in Q_p at clay concentrations larger than 40 per cent (Fig. 7b) suggests that gas concentrations were negligible.

4 VELOCITY MODELLING

Both of our experiments on cold-pressed samples gave consistent values of bulk modulus in the range 6–11 GPa and of shear modulus in the range 4–6 GPa that agree with the measurements of Prasad *et al.* (2002). We first compare our results with those reported by Katahara (1996), Wang *et al.* (2001) and Prasad *et al.* (2002) by using them as input in the Hashin–Shtrikman (HS) model (Hashin & Shtrikman 1963) and the modified Voigt average (MVA) (Nur *et al.* 1991, 1998) to calculate the bulk modulus of a kaolinite–water suspension. K_{water} was taken as 2.2 GPa (Mavko *et al.* 1998). Note that in a homogeneous fluid suspension, where $\mu = 0$, the Hashin–Shtrikman lower bound is equivalent to the isostress Reuss average (K_{reuss}) of the composite given by

$$\frac{1}{K_{\text{reuss}}} = \sum_{i=1}^N \frac{f_i}{K_i}, \quad (1)$$

where N is the number of components and f_i is the fraction of the component.

In Fig. 8, Hashin–Shtrikman upper and lower bounds (HS⁺ and HS[−], respectively) of the bulk modulus (Hashin *et al.* 1963) and the modified Voigt average (Nur *et al.* 1991, 1998) of a water–clay mixture are compared with experimental measurements in saturated kaolinite samples (Yin 1993) and in kaolinite–water suspensions (this study). In Fig. 8(a), the calculations were made using 12 and 6 GPa for the bulk and shear moduli of the solid phase, K_s and μ_s , respectively, (Prasad *et al.* 2002, this study). Calculations with $K_s = 56$ and $\mu_s = 32$ GPa (Katahara 1996) and $K_s = 47.9$ and $\mu_s = 19.7$ GPa (Wang *et al.* 2001) are shown in Fig. 8(b). In Fig. 8(a), the lower HS[−] bound lies on the values measured in the kaolinite–water suspensions and the modified Voigt average predictions agree with measurements in the saturated kaolinite sample from Yin (1993). These predictions support the conceptual first-principles-based model of Dvorkin *et al.* (1999) and Prasad & Dvorkin (2001), wherein the softer material (water) is load-bearing above critical porosity (Φ_c) (Nur *et al.* 1998) and the system is adequately described by the lower HS bound. Once the stiffer material (clay) becomes load-bearing, as in the saturated kaolinite sample, the measured velocity values move away from the lower HS bound and are better predicted by the modified Voigt average. As described in Nur *et al.* (1998), critical porosity separates the domain in which a rock sample is frame-supported ($\Phi_{\text{sample}} < \Phi_c$) from that in which it is fluid-supported ($\Phi_{\text{sample}} > \Phi_c$).

The higher values of bulk and shear moduli from Katahara (1986) and Wang *et al.* (2001), respectively, overpredict experimental measurements both for kaolinite–water suspensions and for kaolinite samples (Fig. 8b).

As a second test, we calculated V_p in shaley sandstones using the different K_s and μ_s values as above for clay and $K_s = 36.6$ GPa and $\mu_s = 45$ GPa for quartz (Mavko *et al.* 1998). Han *et al.* (1986) have shown that the velocity in sandstones is dependent on the porosity and on the clay content: at the same porosity, clay-rich sandstones have a lower velocity than clean sandstones. Theoretical velocities calculated with the rock physics model reported in Dvorkin *et al.* (1999) and Prasad & Dvorkin (2001) are compared with velocities measured on clean and clay-rich sandstones at 40 MPa confining pressure (Tosaya 1982; Han *et al.* 1986; Klimentos 1991) in Fig. 9. This model uses the Hertz–Mindlin (Mindlin

Figure 8. Hashin–Shtrikman lower and higher bounds (HS^- and HS^+ , respectively) and the modified Voigt average compared with experimental measurements in saturated kaolinite samples (black diamonds; Yin 1993) and kaolinite suspension (open squares; this study). (a) Elastic clay moduli for theoretical curves taken from the uniaxial stress experiment and (b) values taken from literature Katahara (1996) marked by dotted line and Wang *et al.* (2001) marked by solid line.

1949) contact theory to compute first the effective bulk (K_{HM}) and shear (μ_{HM}) moduli of a dry mixture of quartz and clay at critical porosity (Φ_c) (Nur *et al.* 1998) under a given pressure. Next, the upper and lower Hashin–Shtrikman bounds ($HS^{HM\pm}$) (1963) are calculated between the two end-points at zero porosity (K_s, μ_s) and at the critical porosity (K_{HM}, μ_{HM}) for this two-component mixture of quartz and clay. Finally, the elastic moduli of water-saturated clay–quartz mixture are calculated using Gassmann’s fluid substitution equations (see Dvorkin *et al.* 1999; Prasad & Dvorkin 2001, for details). As in Fig. 8, Fig. 9(a) shows theoretical curves for the clay mineral moduli, $K_s = 12$ GPa and $\mu_s = 6$ GPa and Fig. 9(b) for $K_s = 47.9$ GPa and $\mu_s = 19.7$ GPa. Figs 9(a) and (b), show theoretical curves computed for clay contents values of 5, 15, 25 and 50 per cent. The computations show that in a two-phase system, if the individual phases have similar elastic moduli, then increasing the concentration of one phase over the other will not alter the moduli of the composite system. Thus, in Fig. 9(b), since the moduli of clay ($K_s = 47.9$ GPa and $\mu_s = 19.7$ GPa) and quartz ($K_s = 36.6$ GPa and $\mu_s = 45$ GPa) are very similar, the model does not predict a sig-

Figure 9. P-wave velocity as a function of porosity for shaley sandstones at 40 MPa. The experimental data from Tosaya (1982), Han (1986), Klimentos (1991), are shown by symbols coded by clay content. Data are compared with modelling described in Dvorkin *et al.* (1999) and Prasad & Dvorkin (2001): solid lines mark the upper bounds (HS^{HM+}), dotted lines mark lower bounds (HS^{HM-}). Elastic clay moduli used to generate theoretical curves were taken from values measured by the uniaxial stress experiment (a) and literature values (Katahara 1996; Wang *et al.* 2001) (b).

nificant modulus reduction with increasing clay content. Using the moduli obtained from this study and from Prasad *et al.* (2002), we obtain a much better agreement between model prediction and experimental data (Fig. 9a). Fig. 9(a) shows that the lower moduli of clay explains both the reduction of the velocity as a function of clay content and the separation in measured velocities due to addition of clay to the sandstones.

5 CONCLUSIONS

We have measured elastic moduli values between 6 and 12 GPa for bulk (K_s) and between 4 and 6 GPa for shear (μ_s) in kaolinite, montmorillonite and smectite clay minerals. These results, obtained using different sample preparation and measurements techniques are consistent with one other and those obtained with atomic force acoustic microscopy by Prasad *et al.* (2002). They are also comparable to values extrapolated and modelled from laboratory and log measurements on shales (Han 1986; Castagna *et al.* 1995; Berge & Berryman 1995).

Note that our measurements are much lower than the theoretical extrapolations of Katahara (1996) and measurements on clay–epoxy mixtures (Wang *et al.* 2001). We suggest that a possible reaction between clay minerals and the polar organic epoxy fluid might lead to a super-dry clay with a higher mineral modulus.

In addition to the consistency check, we modelled the elastic moduli in binary systems: porous clay–quartz and water–kaolinite.

We used values measured in this study and those from Katahara (1996) and Wang *et al.* (2001) for clay mineral moduli. Using $K_s = 12$ GPa and $\mu_s = 6$ GPa from this study, the model gives a much better prediction both of experimental velocity reduction due to increase in clay content in sandstones and velocity measurements in a kaolinite–water suspension.

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PAPER 3

Prasad, M., and Mukerji, T., Analysis of Microstructural Textures and Wave Propagation Characteristics in Shales: Expanded Abstract SEG International Exposition and Seventy-Third Annual Meeting.

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Analysis of Microstructural Textures and Wave Propagation Characteristics in Shales

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Summary

We apply different statistical methods for characterizing heterogeneity and textures from scanning acoustic microscope (SAM) images of shale microstructures. Characterizing and understanding the microgeometry, their textures, scales, and textural anisotropy is important for better understanding the role of microgeometry on effective elastic properties. We analyzed SAM images from Bakken shale, Bazhenov shale, and Woodford shale. Our analysis, based on spatial autocorrelation functions shows that there is a small positive correlation between the degree of heterogeneity and the mean spatial correlation length of the microstructure. The textural anisotropy ranges from 10% to 80%. There is considerable variability within each image, as well as from one image to another. A singular value decomposition (SVD) spectral analysis showed that the shale microstructures have characteristics in common with fractal stochastic images. We also obtain a correlation between maturity and textural heterogeneity.

Introduction

Microstructural characteristics of organic rich shales can give important insights on the maturation processes and on oil generation from such formations. Since changes in shale texture and in hydrogen content are closely linked with kerogen maturity, a correlation between them would enhance methods for detecting and prospecting of kerogen rich shales. The problem is complicated by the fact that the intrinsic anisotropic texture of the shales is enhanced by kerogen distribution in the shales. Attempts have been made to relate acoustic velocity and velocity anisotropy to the degree of kerogen maturity of the shales (Vernik and Nur, 1994; Vernik and Liu, 1997). These studies have been instrumental in improving our understanding of the ultrasonic properties in kerogen rich shales. We have revisited the ultrasonic data collected by Vernik and co-authors and interpreted them using recent acoustic microscopy analyses of the impedance microstructure of the shales (Prasad and Nur, 2001)

Optical and scanning electron microscopy methods to analyze kerogen shale microstructure have been utilized in the past. However, due to the opaque nature of the kerogen and the associated pyrite, such methods are rather difficult to implement. In addition to optical and scanning electron microscopy, we used a non-destructive technique to map the impedance microstructure of kerogen-rich shales with a scanning acoustic microscope. With this technique we were able to map changes in elastic properties as the shales

undergo maturation. This paper reports results of acoustic characterization in samples belonging to various maturity grades and with different kerogen contents. Image analyses techniques (described below) are used to detect changes in texture and heterogeneity in the acoustic microscopy images and are related to ultrasonic measurements.

Statistical descriptions of heterogeneity and textures

Figure 1: Spatial autocorrelation and spatial spectrum can be used to stochastically describe image textures and scales. Top left: Synthetic images with increasing spatial correlation; center: corresponding 2D autocorrelation function; right: corresponding 2D Fourier power spectra. Bottom: radial profiles of isotropic Gaussian autocorrelation functions showing the relation between textures and correlation length.

We used statistical descriptors to quantify the heterogeneity and textures observed in the images. The heterogeneity was quantified by the coefficient of variation (CV) given by the ratio of the standard deviation to the mean of the image pixel values. Gray scale image intensity values were converted to elastic impedances using a calibration function before computing their statistics. Textures can be quantified using spatial autocorrelation functions (Figure 1). We used Fourier transform based autocorrelation estimation. Radial profiles of the autocorrelation function along azimuths ranging from 0° to 180° were computed, and the correlation length estimated at each azimuth. The correlation length is taken to be the lag value where the correlation function

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falls to $1/e$ of its maximum value at zero lag. The texture anisotropy was quantified by the anisotropy ratio (AR) defined as the ratio between the maximum and minimum correlation lengths obtained over all azimuths. Textures were also analyzed using a singular value decomposition (SVD) of the images. The spectrum of singular values shows different decay behavior for different textures (Figure 2).

Figure 2: Singular value spectrum shows distinctive behavior for images with Gaussian spatial correlation (top left) versus fractal images (top right). In the bottom figure the black lines with a rapid fall off in singular value corresponds to images with Gaussian autocorrelation, while the dotted red lines correspond to SVD spectra of fractal images.

SAM Images

Despite similar chemical compositions, microstructure of the samples differed considerably. Acoustic micrographs of the Bakken shale series samples, a Bazhenov shale and a Woodford shale are shown in Figure 3. The C-scan surface images were made at 1 GHz with a calibrated color scale. The calibrations were made using standard materials with known impedance (Prasad et al., 2002). From the different maturity shale samples examined in this study, we find following major differences in the impedance microstructural images:

- Increase in impedance as maturity progresses
- Increase in grain size and in the number of coarse grains in mature shales
- Kerogen and grain distribution undergo major change as the maturity progresses. In immature shale, kerogen forms a more or less connected matrix and the higher impedance grains are dispersed in this matrix. In more mature shale, there is a significant increase in number

of coarse grains. The grains appear to form a framework with kerogen globules distributed within this frame.

- Bedding parallel kerogen filled cracks appear to be more common in immature shale.

Figure 3: SAM images of Bakken shales (bk), Bazhenov shale (bz, lower left), and Woodford shale (wf, lower right). The images have differing resolutions. The image size in microns is indicated. The top left image has an order of magnitude higher resolution.

Results of Image Analysis

Figure 4: Left: 2-D spatial autocorrelation of the 8 SAM images in shown above in Figure 3. Right: Azimuthal profiles of the corresponding autocorrelation function from 0 to 180 degrees.

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The results of texture analyses are shown in Figures 4, 5, and 6. Figure 4 shows the estimated autocorrelation function and the azimuthal profiles. The impedance variation ranged from about 7% to 12%, while the mean correlation length ranged from 2 to 4.5 microns (Figure 5). In general the textural anisotropy (AR) increased with increasing heterogeneity (CV), as seen in Figure 6.

Figure 5: CV (heterogeneity) versus mean correlation length. .

Figure 6: CV versus textural anisotropy. Textural anisotropy is measured by the ratio of maximum correlation length to the minimum correlation length.

Figure 7: Variability with images. Bars indicate range of values obtained by sampling different subsections of the same image. Left: CV versus mean correlation length. Right: CV versus anisotropy ratio.

We analyzed variability within images by processing different subsections of the same image, making sure to sample subsections much larger than the mean correlation

length to obtain valid statistics. The variability is displayed in Figure 7. SVD spectra of the images are distinctly different from SVD spectra of synthetic images with Gaussian autocorrelation. They show more resemblance to SVD spectra of fractal images (Figure 8).

Figure 8: SVD spectra of all but one SAM image are similar to SVD spectra of fractal images showing a slower decay than SVD spectra of Gaussian images (compare with Figure 2). The curve with two humps, indicating two spatial scales comes from an image with an order of magnitude higher resolution. The dotted line is the SVD spectrum of a random image with no correlation.

Summary of Acoustic Measurements

We reevaluated the ultrasonic and physical property data from Vernik and Landis (1987) to distinguish between effects of kerogen and porosity on the elastic properties. Figure 9 shows that porosity effects dominate velocity variations in high porosity shales. In low porosity shales, velocity correlates better with kerogen content (Figure 10).

Figure 9: High-porosity kerogen shales resemble sandstones: There is a good correlation between Vp and porosity.

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Figure 10: In low porosity kerogen shales, velocity increases with maturation.

Textural and Acoustic Property Comparisons

Figure 11: The coefficient of variation (CV) of impedance heterogeneity increases with maturity (decreasing kerogen content) implying an influx of stiffer materials in agreement with the observed increase in impedance with maturity.

Figure 12: The ultrasonic velocity anisotropy (Thomsen's Gamma) is directly proportional to kerogen content: It decreases with decreasing kerogen content (increasing shale maturity)

Figure 13: Similar to the anisotropy coefficient Gamma (Figure 12), spatial textural anisotropy ratio (AR) is also directly proportional to kerogen content. It decreases with maturity implying removal of aligned kerogen with maturity

Figure 11 shows the relation between maturity (decreasing kerogen content) and textural heterogeneity as quantified

by the CV. Increasing maturity is linked to increasing textural heterogeneity. At the same time, increasing maturity is linked to a general decrease in both ultrasonic velocity anisotropy measured by Thomsen's gamma parameter (Figure 12) as well as textural anisotropy measured by AR (Figure 13).

Conclusions

Our analysis of SAM images shows the following: the coefficient of variation (CV), which is a measure of heterogeneity, ranges from 7% to about 12% for these shales. The spatial correlation length varies with direction, and the mean correlation length ranges from about 2 to 4.5 microns. There is a small positive dependence between the CV and the mean spatial correlation length. The mean correlation length generally tends to increase with increasing CV. The textural anisotropy, defined as the ratio of the maximum to minimum correlation length within an image, ranges from 20% to about 80% and in general tends to increase with increasing heterogeneity (CV). Singular value analysis of the images indicates that the SVD spectra of the shale microstructure tend to be similar to (but not exactly the same as) SVD spectra of fractal images. The textural heterogeneity (CV) increases with increasing maturity (decreasing kerogen content), while there is a general decrease in ultrasonic velocity anisotropy and textural anisotropy with maturity.

Acknowledgements

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PAPER 4

*Prasad, M., and Nur, A., Velocity and Attenuation Anisotropy in Reservoir Rocks:
Expanded Abstract SEG International Exposition and Seventy-Third Annual Meeting.*

Paper 6: Total of seven figures in paper

Velocity and Attenuation Anisotropy in Reservoir Rocks

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Summary

Sedimentary rocks are generally characterized by textural anisotropy that gives rise to a directional dependence of acoustic velocities. This textural anisotropy in sedimentary rocks can lead to attenuation anisotropy. Our analyses of the directional dependence of velocity and attenuation in different reservoir rocks show that attenuation anisotropy can be quite large. Directional dependences of velocity and attenuation can be very different and are not influenced in the same manner by textural variations. Depending on the texture, the attenuation anisotropy can be opposite to the velocity anisotropy.

Introduction

Knowledge about the pore and confining pressure dependencies of the P-wave velocity (V_p) and quality factor (Q_p) in reservoir rocks is an important requisite for interpreting seismic measurements in terms of subsurface petrophysical parameters. Since permeability is a dominant factor in controlling loss mechanisms in porous media, a study of the quality factor (Q , or attenuation, Q^{-1}) is of special interest in seismic exploration for oil. Theoretical (Biot, 1956a, b; Akbar et al., 1993) and experimental studies (McCann and McCann, 1969; Klimentos, 1990; Prasad, 2003) have shown the importance of permeability in understanding wave propagation characteristics in porous media. Experimental studies on attenuation and velocity have shown the effect of pressure on crack closure and layering, however, a systematic study of attenuation anisotropy is lacking. For AVO analyses, Adriansyah and McMechan (1998) have shown that intrinsic and scattering attenuation can affect the AVO response. Thus, attenuation anisotropy becomes an important factor that needs to be considered for such investigations. Three main causes have been recognized for velocity anisotropy:

1. Preferential alignment of grains and cracks
2. Layering of materials of different acoustic impedance
3. Layers with different amounts of porosity

The question that remains to be answered is whether these factors affect attenuation in the same manner as they affect velocity. Specifically, the effect of permeability anisotropy on attenuation has not been studied systematically. We have investigated the effect of layering and permeability on both, velocity and attenuation on a set of sandstones as functions of confining pressures up to 25 MPa. The sandstones were very similar in mineralogy and porosity.

Permeability, however, varied with direction and between samples. The measurements made in two directions, parallel and perpendicular to borehole axis show different effects of anisotropy on velocity and attenuation.

Experimental Procedure

Cylindrical core samples with 25 mm diameter and 20-30 mm length were prepared with their faces parallel to within 100 μm . Two samples were prepared from each core: parallel to the borehole axis (called vertical) and perpendicular to the borehole axis (called horizontal). The pulse transmission technique was used for P- and S-wave velocity (V_p , V_s , respectively) measurements. Quality factor (Q_p , Q_s) was calculated using the spectrum division method (Toksöz et al., 1979; Sears and Bonner, 1981).

Bulk and grain densities and porosity were measured using a Helium porosimeter. Difference in porosity values measured in two cores from same depth gave an indication of small-scale heterogeneities in the samples. The microstructure of the samples was examined with optical microscopy, CT-scanning, and scanning acoustic microscopy. Mineralogy was determined by XRD analyses of the sample powders. Klinkenberg-corrected air permeability was measured in horizontal and vertical directions.

Samples Used

We have analyzed fluvial sandstone samples from a producing oil field. The chalk, limestone, and sandstone data were taken from Prasad and Manghnani (1996). The Prasad and Manghnani data was measured in two directions simultaneously on the same sample. Measurements on the fluvial sandstones were made on oriented samples.

Results

Porosity, Permeability, and Texture: Figure 1 and Tables 1 and 2 shows XRD mineralogy, permeability and texture from CT scans in two directions for the fluvial sandstones. The samples are very similar in mineralogy and in porosity: They are fairly clean sandstones consisting of quartz and feldspars with clay contents between 5 – 10%. Porosity ranged between 19–25%. The main difference lies in permeability and in permeability anisotropy. Permeability values varied between 0.1 and 1000 mD and permeability anisotropy values ranged between 0 – 200%. In most cases, the permeability anisotropy could be correlated to the density variations revealed by CT scans. Under optical

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microscopy, the permeability variations (and corresponding density layers) were found to be due to alternate layers of coarse and fine-grained sands. Horizontal alignment of the more porous, coarser grained layers gave a higher permeability in that direction. In some cases, the cause for permeability anisotropy could not be recognized on the CT scans, which showed a very homogenous texture. CT scans of some samples (10 and 21) revealed large-scale density variations. These structures, seen as bioturbations and shell fragments under an optical microscope, were larger than the dominant ultrasonic wavelengths.

Table 1. Porosity and permeability in the sandstone samples. hor = horizontal, vert = vertical. Differences in the vertical and horizontal porosity are due to small-scale heterogeneity in the core. The samples were cleaned prior to measurements.

Sample Id	Permeability (mD)		Porosity (%)		Grain density (g/cm ³)
	Hor	Vert	Hor	Vert	
10	68	3.9	23.9	25.2	2.6
11	295	41	23.4	26.9	2.62
12	818	426	25.2	29.2	2.61
15	7.6	7.6	19.8	24.3	2.64
19	91	36	22.4	25.6	2.63
21	1.4	0.21	19.1	22.3	2.62
24	1030	677	26.9	26.2	2.62
25	67	0.84	20	18.8	2.53
26	39	45	20.1	26.3	2.64
28	22	1.5	23.2	23.6	2.64

Table 2. Density and porosity variations in the limestone samples from Prasad and Manghnani (1996). Note low grain density in the chalk samples implying that porosity is underestimated.

Sample Name	depth (m)	Density (g/cm3)		Porosity (%)
		Bulk	Grain	
Sandstone				
Berea	quarry	2.3	2.59	21.18
Chalk				
#10185.6	3104.6	2.46	2.504	3.24
#10185.9	3104.7	2.40	2.688	13.7
#10351	3150	2.44	2.509	4.67
#10438	3180	2.37	2.603	14.7
Limestone				
#4782	1450	2.72	2.84	6.75%
#4701	1430	2.62	2.86	12.8%
Dolomite				
#3200	3200	2.21	2.82	33.4%

Based on the microstructure observed in CT scans and the permeability measurements, the samples were separated into the groups specified in Table 3: Sandstones had heterogeneous, layered, or homogenous textures. The carbonate samples consisted of low and high porosity

chalks, limestones with vuggy porosity, and a high porosity dolomite (Table 3).

Table 3: Textural properties. The symbols “V” and “H” denote vertical and horizontal measurements; no symbol implies same behavior is observed in both directions. The textural groups for the carbonate samples are derived from Prasad and Manghnani (1996).

Group	Texture	Samples
Sandstones		
1.	Heterogeneous texture	10, 21
2.	Layers + high permeability anisotropy	11, 25H, 28H
2a.	Low permeability	15, 25V, 28V
3.	Homogeneous	12, 24, 26
Chalks		
1.	Very low micritic porosity	10185.6, 10351
2.	Medium porosity	10185.9, 10438
Limestone		
1.	Vuggy porosity	4701, 4782
Dolomite		
1.	High porosity	3200

Figure 1: (a) Mineralogy (b) microstructural variations from CT scans, and (c) permeability anisotropy in the fluvial samples. The samples groups based on permeability and microstructural variations are marked in Figure 1b. hor = horizontal, vert = vertical.

Velocity Anisotropy: Figure 2 shows the effect of pressure on V_p for representative samples from the three fluvial sandstone groups. The samples show a general increase in V_p with pressure. Groups 1 and 2 show V_p anisotropy at higher pressures also, mainly due to textural differences. In some cases, a large V_p anisotropy at low pressure decreases to near zero at 25 MPa. In Group 2 samples, V_p anisotropy remains high at 25 MPa confining pressure. The chalks (Figure 3) showed high V_p anisotropy. There is very little change in V_p and in V_p anisotropy with pressure for the low

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porosity chalks. The increase in V_p and decrease in V_p anisotropy with pressure in the high porosity chalk is due to closing of a visible crack. In the limestones (Figure 4), V_p increases very rapidly with pressure. An initially high V_p anisotropy decreases to almost zero beyond 25 MPa. Apparently, the microcracks responsible for V_p reduction and anisotropy close at this pressure. In the high porosity sandstone and dolomite (Figure 4), V_p increases with pressure and V_p anisotropy is low, about 4%.

Figure 2: V_p -pressure variations in vertical and horizontal directions for samples in the textural groups marked by numbers. Black symbols show measured values in horizontal direction, gray in vertical direction.

Figure 3: V_p variation with pressure in chalks. Solid symbols mark values measured in the horizontal direction and open symbols in vertical direction.

Attenuation Anisotropy: Figure 5 shows the effect of pressure on Q_p for the fluvial sandstones. The Q_p – pressure behavior is typical within each group. Thus

- Group 1. Q_p is low and remains constant.
- Group 2. Q_p increases with pressure. The increase is larger for the low velocity direction and leads to a negative Q_p anisotropy.
- Group 2a. Q_p increases equally in both directions with pressure.
- Group 3. Q_p anisotropy is low and positive.

Figure 4: V_p variation with pressure in limestones and dolomite. Solid symbols mark values measured in the horizontal direction and open symbols in vertical direction.

Figure 5: Quality factor-pressure variations in vertical and horizontal directions for samples in the textural groups marked by numbers. Black symbols show measured values in horizontal direction, gray in vertical direction.

Figures 6 and 7 show the effect of pressure on quality factor for the carbonates. Similar to velocity, the low porosity chalks (Figure 6) do not show much change in quality factor anisotropy with pressure. In the high porosity chalks, quality factor increases with pressure.

In the limestones (Figure 7), Q_p is very low and stays constant with pressure. In the high porosity sandstone and dolomite, Q_p increases with pressure. The Q_p anisotropy is also increases with pressure, with the increase being larger for the low velocity direction, leading to a negative attenuation anisotropy.

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Figure 6: Quality factor variation with pressure in chalks. Solid symbols mark show values in horizontal and open symbols in vertical directions.

Figure 7: Q_p variation with pressure in limestones, a dolomite, and a sandstone. Solid symbols mark values measured in the horizontal direction and open symbols in vertical direction.

Conclusions

A summary of the textural properties and their relations to the ultrasonic properties is given in Table 4.

Table 4: Summary of the textural and ultrasonic properties of the fluvial sandstones and carbonates.

Group	Texture	Ultrasonic properties
Sandstones		
1.	Heterogeneous texture	Low Q_p , no change with pressure: scattering losses dominate
2.	Layers + high permeability anisotropy	Negative Q_p - anisotropy, appears to increase with pressure. Large, positive V_p anisotropy
2a.	Low permeability	High Q_p values which increase with pressure.
3.	Homogeneous	Positive and low V_p and Q_p anisotropy

Table 4 (contd.): Summary of the textural and ultrasonic properties of the fluvial sandstones and carbonates.

Chalks		
1.	Micritic texture, layers with equal impedance	Constant V_p -anisotropy, low Q_p -anisotropy
2.	Partially connected pores	V_p and Q_p increase with pressure
Limestone		
1.	Vuggy porosity, cracks	Low Q_p , no change with pressure: scattering losses dominate, high, V_p anisotropy decreasing with pressure
Dolomite		
1.	Connected pores, low impedance contrast	Negative Q_p - anisotropy

We find that not only velocity but also attenuation shows a directional dependence dependent on texture. Depending on texture, attenuation anisotropy can be opposite to the velocity anisotropy.

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PAPER 5

Vega, S., Mukerji, T., Mavko, G., and Prasad, M., Stratification in loose sediments and its seismic signature: Expanded Abstract SEG International Exposition and Seventy-Third Annual Meeting.

Paper 7: Total of five figures in paper

Stratification in loose sediments and its seismic signature

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Summary

In this paper we present a study on stratification as a source of intrinsic anisotropy in sediments using V_p . We compare velocity anisotropy and textural anisotropy for different sediment packs. We use the spatial autocorrelation function and its variation with direction to characterize the stratification texture from images of samples. We found that velocity anisotropy reveals internal packing.

Introduction

In nature, there is stratification in many sedimentary structures. Stratification in mixtures of sediments that contain clay minerals is probably affected by the crystal structures. In general, these mixtures can be very complex. In this study, we focus on the stratification anisotropy in packing of poured sediments.

It has been found that grain size and shape, and deposition speed determine packing. Makse et al., 1997 and Cizeau et al, 1999 have shown that natural stratification occurs in poured granular mixtures of large rounded and small rough grains. However, Baxter et al., 1998 reported that natural stratification also occurs in mixtures of same grain shape and various grain sizes for a relatively low deposition speed.

Acoustic velocities, which are widely used in geophysical studies, are sensitive to both stress anisotropy (Nur and Simmons, 1969; Yin, 1993; Vega et al., 2000) as well as intrinsic anisotropy. The purpose of this paper is to explore compressional velocity and its relation with packing in poured sediments. We seek to answer whether different packings created with poured sediments are detectable with V_p measured under isotropic stress.

Methods

In order to study V_p and its relation with packing, we used a polyaxial apparatus to measure V_p and strain (ϵ) at three perpendicular directions in sand and glass bead samples. We also implemented Fourier transform based codes to compute the spatial autocorrelation function and its directional dependence of the sample images.

Experimental procedure

We used a polyaxial apparatus (Yin, 1993), adapted and improved to make V_p and ϵ measurements in unconsolidated materials (Vega et al., 2003). The principal

frequency of the piezoelectric crystals for P-wave generation was 1 MHz. In the experimental tests, we attempted to apply the same stress in all three directions, $\sigma_z \approx \sigma_x \approx \sigma_y$ in order to create an isotropic stress. V_p and ϵ were measured in the Z, X, and Y directions, V_{pz} , V_{px} , and V_{py} , and ϵ_z , ϵ_x , and ϵ_y , respectively.

We used a beach sand with an average grain size of 0.25 mm and grain density 2.060 g/cc and glass beads of 0.25-0.3 mm (MO-SCI Corporation), 0.5-0.6 mm and 2.794-3.327 mm (Cataphote, INC). Table 1 summarizes characteristic of the samples used in this study.

All samples were poured in the vertical direction (Z) into the aluminum cell for the polyaxial apparatus. We prepared two different samples with the sand, one was only poured and the other was initially poured and then rotated 90° around the horizontal direction X, i.e. Z and Y directions were exchanged in the final configuration. In addition, we made replicas of the samples in transparent plastic containers and took photographs of them as we cannot take photography inside the apparatus cell. As illustrated in Figure 1, images for the XZ (back and front), ZY (back and front), and XY (top) planes were taken (Figure 1).

Table 1: Sample summary

Sample	Sample name	Grain size (mm)	ρ (g/cc)	ϕ
Santa Cruz sand	SCS	0.25	2.060	48%
Santa Cruz sand-rotated	SCR			47%
Glass beads 1	GB1	0.25-0.3	2.5	41%
Glass beads 2	GB2	0.5-0.6		39%
Glass beads 3	GB3	2.794-3.327		41%

Method for textural anisotropy interpretation

We implemented a code that calculates the autocorrelation function of images at azimuth angles between 0° to 180°. For each azimuth, the correlation length is estimated as the lag at which the autocorrelation function falls to e^{-1} its value at zero lag. Finally, we calculate the anisotropy ratio (AR), defined as the ratio between the maximum and the minimum correlation length. The median correlation length is also estimated.

We processed the images in the ZX plane (back and front), ZY (back and front), and XY (top) for all the samples. We used the following steps: (1) the best and more

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representative image of the sample on the chosen plane and side was selected, (2) the image in gray scale was used, (3) the image histogram was equalized, and then (4) filtered until a reasonable median correlation length of at least two grains was obtained, this process was made to reduce the very small grain scale features but retain the layer stratification textures, and finally (5) the anisotropy ratio and the angles of the autocorrelation anisotropy were calculated. In addition, we created an image of a hexagonal 2D packing and we followed the same process, to validate it with a known granular pack.

Figure 1: Image sketch example for ZX and XY planes (SCS sample).

Results

Velocity anisotropy

Figure 2 shows compressional velocities as a function of the mean applied stress, σ . As can be seen in the graphs, most of the samples presented velocity anisotropy in. In the SCS sample (Figure 2a), V_{px} and V_{py} were equal and higher than V_{pz} . In the SCR sample (Figure 2b), V_{pz} and V_{py} were similar and lower than V_{px} . This result indeed revealed different intrinsic anisotropy between the two sand samples, which was expected.

In general, V_{pz} was lower than V_{px} and V_{py} for the glass beads samples. In the GB1 (Figure 2c), all three velocities were similar. In the GB2 sample (Figure 2d), the velocity anisotropy behavior was similar to the SCS. Finally, in the GB3 sample (Figure 2e), all three perpendicular velocities were different with a change in V_{px} and V_{py} at 32 bars.

Textural anisotropy interpretation

We first described the sample images and then estimated their autocorrelation function anisotropy (textural anisotropy), summarized as follow: The SCS sample presented natural stratification as can be seen in Figure 1. Roughly horizontal layers (parallel to the XY plane) are spontaneously created every time that this sand is poured. On the contrary, the resulting packing for the SCR sample

was more complex as grains under and after the rotation slipped due to gravity

Figure 2: V_p versus mean stress, σ . V_{pz} , V_{px} , and V_{py} are the V_p velocities in the Z, X, and Y axes. Sand samples are plotted at the same scale, and glass bead samples are plotted at the same corresponding scale: (a) SCS, (b) SCR, (c) GB1, (d) GB2, and (e) GB3.

The GB1 sample showed light segregation on the ZX and ZY planes, and almost homogenous texture on the XY plane. Moreover, the GB2 seemed to have a high level of segregation on the ZX and ZY planes, and random appearance on the XY plane. In contrast, GB3 sample images presented some layers on the bottom and on the vertical edges combined with some diagonal aligned beads on the ZY and ZX planes. On the XY plane, this sample also showed aligned beads on the walls, and a general orientation to the center. On the whole so far the description of texture in images is qualitative. We attempted to quantify the characterization of stratification texture using the spatial autocorrelation function and its variation with direction.

Figure 3 shows a typical output of the image processing used in this paper: (a) image of the GB1 sample in the XY plane, (b) results on the top right, anisotropy ratio, median correlation length in pixels and in mm, angles with the maximum correlation length, and angles with the minimum correlation length, (c) autocorrelation function of the equalized and filtered image, and (d) autocorrelation function as a function of the lag (correlation distance in pixels) for azimuth angles from 0° to 180° . We found that the maximum values of AR corresponded to the SCS

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sample in the ZX and ZY planes, and to the SCR sample in the XY plane. Nevertheless, the different values of AR for sand, glass beads, and image pack suggest that AR must be relative to the type of sample, or sample composition.

Figure 4 shows the image processing for a 2D hexagonal pack (numerically created). This packing is also equivalent to hcp (hexagonal close pack) and the ts (tetragonal-sphenoidal pack) in 3D, because in 2D these two packs are not distinguishable. Moreover, its autocorrelation function is periodic. We found that the maximum correlation was parallel to the Z direction, and the minimum correlation was in the direction of the hexagon sides. This result agrees with the velocity anisotropy of an hcp pack, which is $V_{pz} \neq V_{px} = V_{py}$ with ZY symmetry plane. However, it may not always be the case that textural anisotropy coincides with acoustic anisotropy.

Velocity anisotropy and packing

We define velocity anisotropy value as the relative difference between two perpendicular velocities

$$\Delta V_{ij} = \left| \frac{V_{pj} - V_{pi}}{V_{pi}} \right| * 100$$

where $i, j = z, x, \text{ or } y$, for $i \neq j$. In Figure 5, the velocity anisotropy values at the initial stress are compared with the corresponding AR. We found the same relative tendency of ΔV and AR for each sample, i.e. for a high ΔV a relative high AR and so on. We also found two general trends between ΔV and AR in Figure 5d, excluding GB2 and SCS samples in the ZY plane. One of the trends contains all the data from the XY plane, which suggests that there might be a relation between ΔV and AR in this plane. However, for the other planes trends are not differentially clear. In addition, ΔV and AR vary with sample, i.e. with packing type. On the other hand, we also found that ΔV changes slightly with stress, which might be related with grains rearrangement during loading and unloading. However, it was not possible to calculate AR as function of stress because pictures cannot be taken inside the apparatus cell.

Discussion

Velocity anisotropy

The velocity anisotropy found in the samples shows that textures in poured sediments present intrinsic anisotropy that can be detectable with Vp. For instance, in the SCS sample, with natural stratification, the lower velocity Vpz was in the direction perpendicular to the layers. This result indicates that the velocity responds to the internal structure of the sample. In addition, in the SCR, the highest velocity Vpx was in the direction of the rotation axis that had some remaining layers. Furthermore, the sample with the lowest ΔV , GB1, was the one with the least visual apparent

segregation in the pictures. The sample with the highest ΔV , GB2, was the one with the most apparent segregation. These results also suggest that there is a relation between velocity anisotropy and packing, although most of the packings were complex to describe.

For all the samples, ΔV was higher than the error of velocity measurement (1%) in 11 of the 15 measures. For that reason, we can assume that the samples presented velocity anisotropy.

Figure 3: Image processing: (a) GB1 sample image in the XY plane, (b) corresponding AR, median correlation length and azimuth angles, (c) autocorrelation function of the equalized and filtered image, and (d) autocorrelation function as a function of lag for azimuth angles between 0° to 180°.

Textural anisotropy interpretation

We found that the SCS sample showed a textural anisotropy that had a higher correlation on the XY plane, i.e. 0° and 180° on ZX and ZY, and lower in the Z direction. This result agrees with the velocity anisotropy in this sample, which was $V_{pz} < V_{px} \approx V_{py}$. In the XY plane, AR is relatively lower than the other planes. The SCR sample showed a more complex textural anisotropy: there was not a clear direction of the maximum and minimum correlation on the ZX and ZY planes, while the maximum correlation was in the X direction. This last result coincides with $V_{px} > V_{py}$ for this sample.

The GB1 sample had a textural anisotropy with a maximum correlation perpendicular to the Z direction and a minimum correlation around 30° of Z; on the XY plane there is not a clear orientation. GB2 showed similar textural anisotropy to the GB1 on the ZX and ZY planes but with higher AR.

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This result agrees with the visual observation that the GB2 presented more segregation than GB1. On the XY plane, the maximum correlation was in the X direction but with lower AR to the other planes, which can imply a lower relative anisotropy in XY. GB3 had a maximum correlation perpendicular to the Z direction and the minimum slightly varies from ZX and ZY, around the Z direction. On the XY plane, the maximum and the minimum correlation were close to the Y direction, i.e. no clear textural anisotropy was seen. In general, these results mostly agree with the velocity anisotropy seen in the glass beads samples. That is, the maximum correlation was perpendicular to the Z direction corresponding to a V_{pz} lower than the V_{px} and V_{py} .

Figure 4: Hexagonal packing: (a) 2D image, (b) autocorrelation function of the image, and (c) autocorrelation function as a function of lag (correlation distance in pixels) for azimuth angles between 0° to 180°.

Conclusions

All the results indicate that velocity anisotropy and grain packing textures are related. The anisotropy ratio calculated from spatial autocorrelation function of images gives an estimation of packing structure. The results also suggest that if there is a relation between ΔV and AR, it might also depend on the sample composition. In other words, it seems that velocity anisotropy is closely associated with the intrinsic texture anisotropy. Thus, if it were possible to separate the intrinsic anisotropy effect and stress anisotropy effect from velocity anisotropy, it would be possible to improve seismic processing and interpretation.

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Figure 5: Velocity anisotropy, ΔV , compared with anisotropy ratio, AR, for all the samples in the (a) ZX plane, (b) ZY plane, and (c) XY plane, and (d) in all the planes, red and blue lines indicate general trends.

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PAPER 6

Vega, S., Prasad, M., and Mavko, G., Comparative study of velocities under hydrostatic and non-hydrostatic stress in sands: Expanded Abstract SEG International Exposition and Seventy-Third Annual Meeting.

Paper 8: Total of six figures in paper

Comparative study of velocities under hydrostatic and non-hydrostatic stress in sands

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Summary

In this paper, we compare V_p measured under hydrostatic and non-hydrostatic stress conditions in a sand. We describe how we apply isotropic stress using a polyaxial apparatus. We examine velocities in three perpendicular directions as function of stress and find that V_p under hydrostatic pressure is higher than V_p measured under non-hydrostatic isotropic stress. Furthermore, we observe that velocity anisotropy revealed the intrinsic anisotropy in the sands.

Introduction

Laboratory measurements of acoustic velocity in sands are most often made under hydrostatic pressure (Domenico, 1997; Zimmer et al., 2002; Wang, 2002). Stresses in the lithosphere are generally non-hydrostatic and anisotropic (Sinha and Kostek, 1996; Winkler et al., 1994; Zoback and Zoback, 2000). Although Mavko et al. (1995) suggested a method for rocks to predict velocity anisotropy in an anisotropic stress from hydrostatic lab measurements of V_p and V_s , a similar prediction for sands is not yet well understood.

Domenico (1977) measured acoustic velocities under hydrostatic pressure in a sand and in glass beads of similar grain size and porosity. The velocity, pore volume, porosity, and pore compressibility as functions of pressure found for the dry and brine saturated sample are useful for a better understanding of unconsolidated formations. Wang (2002) measured velocity anisotropy under hydrostatic pressure in the lab on sands, shales, and rocks. A relation to estimate V_p anisotropy from V_s anisotropy and vice versa was found. However, as all these correlations have been measured under hydrostatic pressure, they have to be carefully extrapolated to in situ stress.

In this paper, we present an experimental study of velocity measurements in sands at isotropic stress condition in a hydrostatic oil-loading system and a polyaxial piston-loading system. We compare velocities measured under these two conditions and show that they are not equal. This is the first step toward a better understanding of how to relate velocities measured under hydrostatic pressure and in situ stress velocities in sands.

Experimental setup and procedure

We used a hydrostatic and a polyaxial apparatus to compare compressional velocity (V_p) and strain (ϵ) under

hydrostatic and non-hydrostatic stress in a sand. We implemented a test called quasi-hydrostatic which consists of creating isotropic stress ($\sigma_z \approx \sigma_x \approx \sigma_y$) in the polyaxial apparatus.

Quasi-hydrostatic test

For the quasi-hydrostatic stress test, we adapted a polyaxial apparatus (Yin, 1993) to make V_p and ϵ measurements in unconsolidated materials (Vega et al., 2003). In this apparatus the sample is contained in an aluminum cell that can be loaded in three orthogonal directions with pistons.

In the quasi-hydrostatic stress test, the same stress was applied in all three directions, $\sigma_z \approx \sigma_x \approx \sigma_y$. We loaded (and unloaded) the samples by successively incrementing σ_z , then σ_x , and following with σ_y . To check the possible influence of loading sequence on the results, we also changed the order of loading increments to σ_x , σ_y , and then σ_z . We waited after the first stress increment eight to ten hours before starting the V_p and ϵ measurements. At each step, we followed the same order of loading (and unloading) and the stresses were allowed to stabilize (average about half an hour) before making the acoustic measurements.

Hydrostatic test

For the hydrostatic stress tests, we used an oil-loading pressure vessel (Vanorio et al., 2002). The sample is contained in a cylindrical tygon jacket and is subjected to a confining pressure generated with oil.

In order to be consistent with the polyaxial test, the confining pressure steps were the same as in the quasi-hydrostatic test and the sample was allowed to stabilize (average about half an hour) before making the acoustic measurements. The principal frequency of the piezoelectric crystals for P-wave generation was 1 MHz in both apparatus.

Samples and sample preparation

We used a beach sand with an average grain size of 0.25 mm and grain density 2.060 g/cc. The grain size analysis was made by sieving, and the grain density was measured using a pycnometer. This sand shows natural stratification when it is poured (Figure 1a). Natural stratification has also been observed in different mixtures of granular materials (Makse et al., 1997; Baxter et al., 1998; Cizeau et al., 1999).

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We made four sand samples, one for the hydrostatic test, HNS, and three for the quasi-hydrostatic test QNS1, QNS2, and QRS. All samples were poured in the vertical direction (Z) creating the natural stratification characteristic of this sand. The HNS sample was poured in a cylindrical rubber jacket for the hydrostatic apparatus and the QNS samples were poured into the aluminum cell for the polyaxial apparatus. QRS was first poured and then rotated 90° around the X-axis, i.e. Z and Y directions were exchanged in the final configuration (Figure 1b). For the hydrostatic apparatus, a sample similar to QRS was not possible due to limitations of the setup. The samples were flattened by tamping for a better transducer-sample coupling.

Figure 1: (a) Natural stratification shown in the poured sand: the black line shows one of the layers naturally formed. (b) Rotated sand around X direction, equivalent to the QRS sample: the black lines show mainly interpreted features. (Pictures taken in a transparent container outside the aluminum cell).

Initial porosity (ϕ) was calculated from the volume and grain density of the sand samples. Average porosity was 47% with a sample-to-sample error lower than 2% (Table 1). All samples were measured dry during the loading and unloading of three stress cycles up to 40 bars. In this paper, we show the loading path of the first cycle as velocity and strain results did not reveal a significant hysteresis. QNS1 was used for the Z→X→Y loading order, and QNS2 for the X→Y→Z loading order. Table 1 summarizes the general characteristics of the samples.

Table 1: Sample summary

Sample name	Test*	Fabric**	Loading order	ϕ
HNS	H	NS		46.1
QNS1	Q	NS	Z→X→Y	47.9
QNS2	Q	NS	X→Y→Z	46.8
QRS	Q	RS	Z→X→Y	47.3

* H: hydrostatic, Q: quasi-hydrostatic

** NS: natural stratification, RS: rotated stratification

Results

Fabric anisotropy

Figure 2 and Figure 3 show V_p as a function of the mean stress, $\sigma = (\sigma_z + \sigma_x + \sigma_y) / 3$, for QNS1 and QRS, respectively. Figure 2 displays the velocity anisotropy in QNS1 due to its depositional anisotropy. V_{pz} , which was in the perpendicular direction to the layering, was lower than V_{px} and V_{py} , as in a typical TI medium. A change in the loading direction (QNS2) produced similar results. To test whether the sample had textural anisotropy, we measured velocity in a rotated cell, QSR (Figure 3). Velocity in the unchanged horizontal direction V_{px} remained high. However, velocities in the exchanged vertical (V_{pz}) and horizontal (V_{py}) directions were now equal and lower than V_{px} .

Figure 2: Velocity as a function of stress for sample QNS1. Open circles, squares, and triangles denote QNS1 V_{pz} , V_{px} , and V_{py} , respectively. Close circles represent HNS velocity in the Z direction, V_h .

Figure 3: Velocity as a function of stress for sample QRS. Open circles, squares, and triangles denote QRS V_{pz} , V_{px} , and V_{py} , respectively. Close circles represent HNS velocity in the Z direction, V_h .

Figure 4 shows strain in the Z direction (ϵ_z) as a function of the stress for all samples. We found the same trend ϵ_z - σ for the samples measured in the polyaxial apparatus, and a slightly different trend for the HNS sample. The strain for QRS was higher than for QNS samples, which must be influenced by the fabric anisotropy as the sample-to-sample porosity difference was relatively small (0.5-0.6%). In

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addition, there was a difference between the strain of QNS1 and QNS2 that is most likely a consequence of the porosity, which varied by 1.1% between these samples.

Figure 4: Strain measured in the quasi-hydrostatic test as a function of the strain measured in the hydrostatic test. Symbols denote measured values, solid lines represent second order polynomial fits.

Vp under hydrostatic and non-hydrostatic stress

Figure 2 shows V_p in the Z direction for HSN (V_h) as a function of pressure, and it is compared with the QNS1 velocities. As revealed by the graph, V_h was higher than the QNS1 velocities. We can also see that the V_h -pressure curve increases faster than the V_{pz} -stress curve for QNS1 at stresses lower than 22 bars, and increases slower at stresses higher than 22 bars.

Figure 3 shows V_h as a function of the pressure, and it is compared with the QRS velocities. As for QNS1 in Figure 2, V_h was also higher than the velocities in the Z direction. The V_h -pressure curve and V_{pz} -stress curve for QRS increase equally at stresses lower than 22 bars, and the V_h -pressure curve increases slower than this V_{pz} -stress curve at stresses higher than 22 bars.

Table 2 summarizes the velocity comparison between the samples under hydrostatic and non-hydrostatic stress. It is clear that velocities measured under hydrostatic pressure are not equal to velocities measured under non-hydrostatic stress fields, even for an approximately isotropic stress field with the same fabric anisotropy (HNS, QNS1, and QNS2).

Table 2: Velocity comparison summary

Sample	Velocities		$\frac{V_h - V_z}{V_h}$	$\frac{V_h - V_x}{V_h}$
	Same direction	Different direction		
HNS QNS1 QNS2	$V_h > V_{pz}$	$V_h > V_{px} \approx V_{py}$	27%	7% - 12%
HNS QRS	$V_h > V_{pz}$	$V_{px} \approx V_h > V_{py}$	8% - 12%	4%

Discussion

Fabric anisotropy

The velocity anisotropy detected in QNS1 (Figure 2) and QNS2 is consistent with the natural stratification shown in Figure 1a. The velocity anisotropy detected in QRS (Figure 3) deviates from a TI anisotropy. For instance, Figure 1b shows that the packing of QRS is more complex than QNS. For QRS, it seems that some of the original layers formed during pouring became more curved after the rotation due to gravity. In a simple way, this can explain why V_{pz} and V_{py} are lower than V_{px} . V_{py} is measured in the perpendicular direction to the layers, V_{pz} is perpendicular to the layer edges, and V_{px} is in the direction of the still parallel layers.

V_h was higher than V_{pz} for the two packings, QNS and QRS. It was 27% higher for QNS, and 8-12% for QRS (Table 2). This difference between QNS and QRS indicates that the divergence between velocities measured under hydrostatic and non-hydrostatic conditions can be also affected by the fabric anisotropy of the sands. Therefore, it must be necessary to know the direction of the velocity measurement with respect to the geological formation anisotropy to extrapolate hydrostatic lab V_p to non-hydrostatic in situ V_p .

Vp under hydrostatic and non-hydrostatic stress

In order to see if there is a correlation between velocities measured under hydrostatic pressure and quasi-hydrostatic stress, we plotted V_{pz} as a function of V_h for all samples with same fabric anisotropy (Figure 6). As revealed by the graph, the samples have the same trend.

Figure 6: Compressional velocity measured in the quasi-hydrostatic test as a function compressional velocity measured in the hydrostatic test.

In order to check edges effects in the polyaxial apparatus, we evaluated the stress in the borders of the cell using a surface line load on a semi-infinite solid region approximation (Jaeger and Cook, 1979). We found that stress at the edges were approximately 2% lower than the

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applied stress. However, the relative difference between V_h and V_{pz} changes less than 1% when this edge stress effect is added.

Conclusions

The velocity anisotropy under isotropic stress revealed intrinsic anisotropy in the sand samples. Moreover, V_p measured under hydrostatic pressure was higher than V_p measured under non-hydrostatic stress in the sand, for the same fabric anisotropy and similar isotropic stress. The difference between hydrostatic V_p and non-hydrostatic V_p was not due to border effects in the semi-infinite solid approximation. Nevertheless, line forces in granular material are non-uniform and might affect the velocity-stress results. Further work is required to fully explain the results found in this paper.

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