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ABSTRACT

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

1. Kerogen-rich shales using Scanning Acoustic Microscopy and ultrasonic wave propagation. We find that an increase in elastic modulus with increasing kerogen maturity can be directly related to the microstructural acoustic impedance changes. A positive relation is established between microstructural changes and velocity variations as functions of kerogen maturity.
2. Elastic properties of clay minerals using Atomic Force Acoustic Microscopy and Scanning Acoustic Microscopy. We show the effect of clay minerals in contact zones as load-bearing constituents of rocks
3. Elastic properties of unconsolidated sediments in an effort to quantify attributes for detection of overpressures from seismic and for effects of stress-induced velocity anisotropy in sediments
4. We have initiated efforts for velocity upscaling to quantify long-wavelength and short-wavelength velocity behavior and the scale-dependent dispersion caused by sediment variability in different depositional environments.

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LIST OF GRAPHICAL MATERIALS

Paper 1: Total of eight figures in paper	7
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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 modulus using Atomic Force Acoustic Microscopy (AFAM) and measurements of seismic properties of unconsolidated sediments at very low effective pressures. Five papers have been published on the work done in the first year of this project. This report presents reprints of our publications relevant to this project on impedance mapping, AFAM measurements, and overpressure analyses.

PAPER 1

Prasad, M., Reinstaedtler, M., Nur, A., and Walter A.: Quantitative Acoustic Microscopy: Application to Petrophysical Studies of Reservoir Rocks: Acoustical Imaging, Vol. 26, 493 – 502.

We demonstrate here applications of scanning acoustic microscopy (SAM) to quantify microstructure as variations in acoustic impedance. The acoustic maps can then be used to understand the seismic response of the rock. For example, ultrasonic measurements on kerogen shales have shown that the elastic modulus increases with increasing kerogen maturity. SAM of these shales shows a stiffening of the shale framework with increasing kerogen maturity. The load-bearing material changes from low impedance kerogen to high impedance shale. The impedance values determined from SAM are comparable with those obtained from ultrasonic time-of-flight data. In contrast, the cement is load bearing in sandstones where high impedance quartz grains are cemented together by low impedance clay. The cement determines not only the total impedance of the rock but also its pressure dependence.

Paper 1: Total of eight figures in paper

PAPER 2

Zimmer, M., Prasad, M., and Mavko, G.: Empirical Velocity-Pressure and Porosity-Pressure Trends in Unconsolidated Sands: 72nd. Ann. Internat. Mtg., Soc. Expl. Geophys., Expanded Abstracts, Paper RP2.5.

Presence of clay minerals, hydrous aluminosilicates with grain size smaller than 2 mm, We developed empirical velocity-pressure and porosity pressure trends from compressional and shear wave velocity and porosity measurements on unconsolidated sand and glass bead samples over a pressure range from 100 kPa to 20 MPa. We found the S-wave velocity to vary with approximately the fourth root of the effective pressure over the entire range of pressures, while the P-wave velocity demonstrates a slightly lower pressure dependence. Preconsolidation reduces the pressure dependence of both velocities, with the effect being larger for the P-wave velocities, and results in largely unrecoverable porosity loss. Samples prepared with varying sorting qualities produced a range of initial porosities, but the effect on the P and S-wave velocities and on their pressure trends was minimal. Water-saturated P and Swave velocities were modeled with Gassmann fluid substitution. The water-saturated P-wave velocities show a large porosity dependence, which results in a significant porosity dependence in the VP-VS ratio, especially at low pressures.

Paper 2: Total of five figures in paper

PAPER 3

Prasad, M., Kopycinska, M., Rabe, U., and Arnold, W.: Measurement of Young's modulus of clay minerals using atomic force acoustic microscopy: GRL, Vol. 29, #8.

Presence of clay minerals, hydrous aluminosilicates with grain size smaller than 2 mm, alters the elastic and plastic behavior of materials significantly. Clay in contact zones has significantly lower impedance than the quartz it cements and so ultrasonic velocities are reduced in clay-rich sandstones when the clay is load-bearing. Despite their ubiquitous presence, measurements of the elastic properties of clay minerals remain a challenge. Since clay minerals are an integral part of many formations and seismic measurements are the main tools for identification of subsurface lithologies, knowledge about the elastic properties of clays is essential for interpretation and for modeling the seismic response of clay-bearing formations. However, elastic properties of clay minerals are almost unknown. Until now, estimates of single crystal elastic properties have been either theoretical or based on extrapolations from measurements on clay-epoxy mixtures. We have used Atomic Force Acoustic Microscopy (AFAM) to measure elastic properties of clay minerals. We demonstrate the AFAM technique for measuring elastic properties of soft materials. Using this technique, we present first-ever quantitative measurements of Young's modulus in clay. The Young's modulus of dickite was measured as 6.2 GPa.

Paper 3: Total of three figures in paper

PAPER 4

Zimmer, M., Prasad, M., and Mavko, G., Pressure and porosity influences on V_p - V_s ratio in unconsolidated sands: The Leading Edge, vol. 21, no. 2.

We have investigated the sensitivity of seismic methods for detection of shallow water flows and over-pressured zones. Using P-wave information alone can be ambiguous, because a drop in P-wave velocity (V_p) can be caused both by overpressure and by presence of gas. The ratio of P-wave velocity to S-wave velocity (V_p/V_s), which increases with overpressure and decreases with gas saturation, can help differentiate between the two cases. Since P-wave velocity in a suspension is slightly below that of the suspending fluid and $V_s = 0$, V_p/V_s and Poisson's ratio must increase exponentially as a load-bearing sediment approaches a state of suspension. On the other hand, presence of gas will also decrease V_p but V_s will remain unaffected and V_p/V_s will decrease. Our analyses of ultrasonic P- and S-wave velocities in sands show that the V_p/V_s ratio, especially at low effective pressures, decreases rapidly with pressure.

Paper 4: Total of four figures in paper

PAPER 5

Zimmer, M., Prasad, M., and Mavko, G., Velocity – pressure trends in sands from Ultrasonic P- and S-wave velocity measurements: SAGEEP expanded abstract.

The pressure dependence of compressional (P) and shear (S) wave velocities in unconsolidated sediments is an important consideration in a number of engineering applications. This pressure dependence is often used to correct or project velocities to depths or locations where measurements have not been made. It also allows velocity changes to be used to monitor pressure changes in weak reservoirs or aquifers. The hazards posed to offshore drilling by unknown overpressures at shallow depths have also prompted attempts to use the seismic velocities or V_p - V_s ratio to quantitatively predict in situ effective pressures.

Our measurements of compressional (P) and shear (S) wave velocity in unconsolidated sand and glass bead samples at pressures from 100 kPa to 20 MPa show that the S-wave velocity varies with approximately the fourth root of the effective pressure over this entire range of pressures, while the P-wave velocity shows a slightly lower pressure dependence.

Paper 5: Total of seven figures in paper

BIBLIOGRAPHY

- Prasad, M., Reinstaedtler, M., Nur, A., and Walter A.: Quantitative Acoustic Microscopy: Application to Petrophysical Studies of Reservoir Rocks: Acoustical Imaging, Vol. 26, 493 – 502.
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Seismic and Rock-Physics Diagnostics
of Multiscale Reservoir Textures
Program/Project Title

Identification Number	Description	Planned Completion Date	Actual Completion Date	Comments
1	Impedance microtexture imaging	09/30/2002		
2.	Laboratory measurements in sediments	09/30/2002		
3.	Define impedance lithofacies	09/30/2003		Ongoing effort
4.	Assess fluid and heterogeneity effects	09/30/2003		Ongoing effort
5.	Seismic attributes of overpressured zones	09/30/2004		Ongoing effort
6.	Scale relations for velocity transforms	09/30/2004		Ongoing effort

* SAME DATES AS PLANNED.

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QUANTITATIVE ACOUSTIC MICROSCOPY: Application to Petrophysical Studies of Reservoir Rocks

Manika Prasad, Michael Reinstaedtler, Amos Nur, Walter Arnold*

1. INTRODUCTION

Rocks are composites made of grains that are cemented together by different materials. The strength and seismic properties of the rock depend on the stiffness of the cement. Traditional techniques, such as scanning and transmission electron and optical microscopy can help identify variations in type of cement. However, they provide qualitative descriptions only. We demonstrate here applications of scanning acoustic microscopy (SAM) to quantify microstructure as variations in acoustic impedance. The acoustic maps can then be used to understand the seismic response of the rock. For example, ultrasonic measurements on kerogen shales have shown that the elastic modulus increases with increasing kerogen maturity. SAM of these shales shows a stiffening of the shale framework with increasing kerogen maturity. The load-bearing material changes from low impedance kerogen to high impedance shale. The impedance values determined from SAM are comparable with those obtained from ultrasonic time-of-flight data. In contrast, the cement is load bearing in sandstones where high impedance quartz grains are cemented together by low impedance clay. The cement determines not only the total impedance of the rock but also its pressure dependence.

Such a quantitative study can help in understanding and correlating geophysical data that are gathered at various scales to lithology and microstructure. Analyses of measurements at a common scale will increase the reliability of prediction about subsurface lithology. For example, the seismic scale provides least resolution (tens of meters) and maximum spatial coverage (km). P-wave velocity and impedance are the measured properties. The well-log scale has a medium resolution and coverage (m). The measurements include properties such as P-wave velocity, porosity, shale volume, and density. The maximum resolution is achieved at the laboratory scale, by sacrificing the coverage. In the laboratory, the lithology and physical properties can be fully characterized on cm sized core plugs. The challenge lies in extending relationships gathered from laboratory measurements to field seismic studies.

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2. EXPERIMENTAL METHODS

We have used SAM at 1 GHz to measure impedance in various rocks, for example sandstones and shales. The instrument (ELSAM) operated at 1 GHz. SAM¹ can be best understood in geophysical terms as microscopic analogs to side-scan sonar and reflection seismic mapping. The main principle is to study surface and sub-surface features with the help of high-frequency acoustic waves. Images of microstructure are prepared on the basis of the reflected acoustic waves, that is, on the impedance changes in the sample. Acoustic waves that are incident on a sample are mode converted, partly transmitted into the sample, and partly reflected. The signal intensity received back at the transducer, is governed by changes in acoustic impedance and hence by changes in the elastic constants of the material.

C-scans of surface and subsurface features can be used to study impedance changes in the sample quantitatively^{2,3}. By making a gray scale calibration using materials of known impedance, the acoustic images of the rocks can be evaluated for impedance variations as functions of cementation strength, kerogen maturity, and reflectivity. To make a calibrated image, a set of standard materials is selected. The expected impedance in the unknown sample should lie within the impedance values of the standard materials. The standard samples are imaged at a fixed setting of signal and video gains, such that all samples can be imaged at that setting. The gray scale of each sample image is associated with its impedance and reflection coefficient with respect to the coupling medium, water. This gives rise to a calibration of the gray scale from 0 (= black) to 255 (= white), the depth of digitization in the ELSAM. Next, the unknown sample is first imaged at an optimum gain setting. Then, the gain setting is reverted to the calibration values and an image is recorded. The gray scale variations in the unknown image can then be determined from the calibration plot of reflection coefficient versus impedance based on a least-square fit. Approximately 10-15% error is to be expected due to instrumental drift, especially if calibrations are not performed often or if the sample has large variations in surface topography.

Table 1 lists velocity, density, and impedance values of the standard samples used in this study along with values for common rock-forming minerals. Expected variations in the impedance of reservoir rocks lie between 1.5 Mrayls (water) and 37 Mrayls (pyrite). Two sets of calibration standards were used to map the impedance variations in the rock samples, a low (8 – 1.5 Mrayls) and a high (8.5 – 43 Mrayls) impedance group.

3. SAMPLES AND PETROPHYSICAL CONSIDERATIONS

We present here results from samples of various lithologies, such as sandstone and shale, to demonstrate the applicability of quantitative scanning acoustic microscopy to petrophysical studies. Table 2 gives a list of the samples along with their main petrophysical characteristics and composition. The samples used were either thin sections or rock billets with the final polish made with 0.1 μm polishing powder.

The petrophysical properties of sedimentary rocks at depth depend largely on their sedimentation history and their evolution from deposition to compaction and cementation during burial. Typical sedimentation and burial scenarios for marine sediments are presented in Fig. 1 along with the velocity and porosity variations to be associated with them⁶.

Table 1. Impedance values and reflection coefficients for water-sample interface of materials used for calibrating the gray scale of ELSAM images and of common rock-forming minerals. Group I consists mainly of plastics with low impedance; Group II has higher impedance metals.

Sample	P-wave velocity (Km/s)	Density (g/ccm)	Impedance (Mrayls)	Reflection coefficient
GROUP I: Low Impedance				
Arsenic sulfide	2.58	3.27	13.74	0.80
Acrylic	1.18	2.68	3.16	0.36
Polycarbonate	1.20	2.22	2.67	0.28
HDPE	0.95	2.56	2.43	0.24
Polystyrene	1.05	2.11	2.21	0.19
LDPE	0.92	2.06	1.90	0.12
GROUP II: High Impedance				
Copper	4.76	8.93	43.09	0.93
Titanium	3.32	7.3	24.24	0.88
Silicon	8.68	2.2	19.10	0.86
Aluminum	6.42	2.7	17.33	0.84
Glass	5.42	2.54	13.76	0.80
Magnesium	5.77	1.7	9.81	0.74
Arsenic sulfide	2.58	3.27	8.44	0.70
Major rock-forming minerals				
Quartz	6.04	2.65	16.0	0.83
Feldspars	5.59 – 7.05	2.56 – 2.73	14.3 – 19.3	0.81 – 0.86
Calcite	6.26	2.7	16.9	0.84
Dolomite	6.93	2.87	19.9	0.86
Pyrite	7.7	4.81	37.0	0.92
Clay	1.44	1.58	2.3	0.21
Carbon	2.25	1.47	6.3	0.62
Kerogen	4.76	1.3	2.9	0.32

As the sediment is deposited and undergoes compaction, it will move along a normal compaction trend. Without external influences, porosity is governed by compaction and volume reduction due to expulsion of pore fluids. Porosity will decrease and velocity will increase with depth. In Figure 1, the solid black line marks this normal compaction curve constructed from the best fit to measured log-data. Both, cementation and pore-filling events cause the porosity and velocity trends to deviate from the normal. At the onset of cementation, the porosity is "locked-in". The sediment develops a frame that resists compaction. Porosity reduction will be slower as shown schematically by the dark gray line in Figure 1. The increase in sediment-frame stiffness due to cementation also causes a jump in velocity. On the other hand, an influx of pore filling materials will reduce porosity below the normal compaction curve, shown schematically by the light gray line in Figure 1. Velocity will be slightly lower than normal. Such variations in velocity and porosity have been observed, for example in a silty environment⁷ and in a carbonate deposition basin⁸. The correlation between microstructure and velocity was made by qualitative visual core observation. The compliance of the cement in sandstones also affects their velocity – pressure behavior⁵. Figure 2 illustrates this effect on velocity – pressure relations. In Berea sandstone, the contact regions have lower impedance than the quartz grains they cement. Under pressure, the compliant contacts deform and velocity changes with pressure. In Boise sandstone the

quartz grains are cemented together by quartz. These contacts are stiffer, the material offers resistance to pressure and so the velocity shows a smaller pressure-dependence.

Table 2. Samples used in this study. P- and S-wave velocities (V_p and V_s) were measured on cm-sized core plugs at 1 MHz frequency and 25 MPa pressure. The fast-direction kerogen shale velocities are taken from ref. 4, sandstone data is from ref. 5.

Sample	Lithology	Maturity	Porosity	V_p (km/s)	V_s (km/s)	Hydrogen Index
Berea	Sandstone	n/a	20	4.12	2.57	n/a
Boise	Sandstone	n/a	22	4.08	2.23	n/a
Bakken	Kerogen shale	II	0.70	3.93	2.31	584
Bakken	Kerogen shale	III	0.68	4.19	2.57	319
Bakken	Kerogen shale	IVa	0.75	4.48	2.55	294
Bakken	Kerogen shale	IVb	0.97	4.67	2.74	175
Bakken	Kerogen shale	V	1.08	4.36	2.59	122

Clay minerals can affect wave velocities considerably. The velocity is lower in sandstones containing clay than in those without clay⁹. The effect of clay minerals on wave propagation in glass is shown in Figure 3. Addition of a clay layer to glass discs dampens waves significantly. Using ultrasonic (1 MHz) transducers, we propagated P- and S-waves through two glass discs with a thickness of 19.6 mm. The signals displayed show signal damping after applying a 0.05 mm thick clay layer in the contact region between the glass discs. The effect is more pronounced in the S-waves than in the P-waves. The amplitude of both P- and S-signals decreases significantly, in particular amplitude of the S-signal.

Figure 1. Velocity and fractional porosity variations with depth (in mbsf: meters below sea-floor) in silty marine sediment. The dotted black line shows log P-wave velocity, V_p (left), and fractional porosity (right). The black line is a best-fit line to the data. Also shown schematically are the effects of cementation (dark gray line) and pore filling (light gray line). On cementation, a stiffening of the frame increases V_p significantly and porosity remains at a higher value due to compaction resistance offered by the frame. A pore-filling event, on the other hand, does not change V_p much, but reduces porosity considerably⁵.

Figure 2. Top: P- and S-wave velocity (V_p and V_s , respectively) variations as functions of pressure (P_c) in Berea (black dots) and Boise sandstones (gray dots) measured in cm-sized core-plugs. Bottom: Microstructure and impedance variations imaged with acoustic microscopy in Berea (left) and Boise (right). Impedance contrast is low between the grains and the cement region in Boise sandstone whereas it is quite significant in the Berea sandstone.

4. RESULTS

We now present how quantifying microstructure as impedance variations can help understand the effects of maturity changes in kerogen shales. Figure 4 shows porosity- velocity relations in shales as compared with sandstones^{4,10}. Plus signs mark low porosity kerogen shales and open circles mark high porosity kerogen shales. The sandstone data⁹, from samples with varying amounts of clay content, is marked by filled circles with varying gray shade: black denotes clean sandstones, progressively lighter gray shades mark increasing clay contents. High porosity kerogen shales resemble sandstones. Velocity increases with decreasing

Figure 3. P- and S-waves through glass discs as shown on the right. Gray lines depict signals traveling through glass only and black lines represent signals through two glass discs separated by a 0.05mm layer of clay. Signals deteriorate on addition of clay to the contact area between the glass discs. The shear wave marked by a thin black line was made with the transducers at cross-polar position to ascertain shear arrival (thick black line).

porosity and clay content. This trend does not apply to the low porosity kerogen shales. Thus, a different approach is required to understand their velocity variations.

Plotting the velocity as a function of maturity in Figure 5 shows that the velocity increases with kerogen maturity stage. In stage II maturity, the shales appear uncorrelated with maturity stage. These shales have higher porosity than the higher maturity kerogen shales. As maturity increases from stage II up to stage VI, the velocity also increases.

An examination of the microstructure shows variations in acoustic impedance that can explain this increase in velocity. Figure 6 shows a comparison of the impedance microstructure in four Bakken shales with increasing maturity marked by Hydrogen Index (in yellow numbers). To emphasize the change in position of the low-impedance kerogen material, the areas with impedance between 0 – 2% of the calibrated values are marked by white. Thus, white denotes impedances from 1.5 – 4 Mrayls. In the lowest maturity shale with highest Hydrogen Index value, the texture is dominated by the low impedance material, which also appears to be load bearing. As maturity increases and Hydrogen Index decreases, the low impedance material also reduces until it is relegated to isolated pockets and the higher impedance shale becomes load bearing.

5. COMPARISON WITH CORE PROPERTIES

The quantitative acoustic microscopy results showed significant changes in microstructural impedance of the kerogen shales with maturation. We now compare the microstructural impedance changes (Figure 6) with velocity increase with maturity (Figure 5). Figure 7 shows a comparison between micro-impedance measurements made with acoustic microscopy and impedance values measured for core samples in ultrasonic pulse transmission experiments. Despite the difference in scale and frequency of measurement, there is a remarkable agreement

Figure 4. Porosity-velocity relations in shales compared with sandstones. Shale data is taken from references 4 and 10 and sandstone data is from ref. 9. Plus signs mark low porosity kerogen shales and open circles mark high porosity kerogen shales. The sandstone data were measured on samples with varying amounts of clay content marked by variations in gray shade: Black denotes clean sandstones, progressively lighter gray shades mark decreasing clay contents.

Figure 5. Velocity variations in shales as function of their maturity. Above maturity stage II, V_p is directly related to the maturity stage of the shales: increasing maturity leads to an increase in velocity.

agreement between the two values. This gives us a valuable tool to diagnose and interpret remote measurements to detect presence and maturity of kerogen shales.

Our results show a correlation between the impedance of shale matrix, the total organic content and the hydrogen index of the shale. Presence of pyrite, a common accessory mineral in such rocks, increases the impedance of the altered areas as compared to the unaltered kerogen material. This change in impedance should be taken into consideration when analyzing the causes of velocity anisotropy measured in organic shales. Figures 8a and b show examples of Acoustic Microscopy and Acoustic Sounding results of kerogen-rich shale, respectively. The very high reflectivity in the B-scans from Acoustic Sounding is due to high impedance contrast observed in the SAM images (Figure 8a) between grains (white colored) and matrix (black colored). The line drawing in Figure 8b represents an A-scan made at that position.

6. CONCLUSIONS

We have shown how quantitative acoustic microscopy can be used for petrophysical studies of reservoir rocks. The gray scale of acoustic microscopy images is calibrated with materials of known impedance. Variations of micro-impedance values in unknown samples can be determined by comparing the gray scale of their images with the calibration values. Our studies show that the microstructural changes in sediment texture are controlled also by its sedimentation history. In our case study of kerogen shales with varying maturity, quantitative acoustic microscopy helps to interpret and understand the impedance variations in kerogen shales as they undergo maturation. The change in core velocity as a function of maturity is understood in terms of change in load-bearing properties of the shales.

Figures 6. (a) Effect of maturity on impedance microstructure of the shales showing the transition from the load-bearing kerogen in the immature shales (HI=319) to grain-supported matrix in the mature shales (HI=122). White marks lowest impedance materials in the shales. In immature shales (HI=319 and 294), low impedance materials dominate the image. In the mature shales (HI=175 and 122), they are isolated in small pockets.

Figure 7. Impedance variations in shales as function of their maturity given by Hydrogen Index measured on in a 1 mm area and on a core plug of 2.5 cm length. There is a remarkable agreement between the impedance values measured at two different scales. The micro-impedance values lie between the fast and slow directions of wave propagation in the core plugs.

7. ACKNOWLEDGEMENTS

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Figure 8. Comparison between impedance microstructure (a) and reflectivity (b) of a kerogen-rich shale. 3D reconstructions are shown in (c) and (d). Lowest impedance (less than 7.5) is masked by gray color in Fig. 8a. The 15 MHz images (c, d) are made of the same sample from opposite directions: Top image has the coarse grained, high impedance layer from Fig. 8a on top (contact is at 5.5 mm = 1.4 ms). Bottom image has the coarse grained, high impedance layer at bottom (contact is at 22.7 mm = 6 ms). A prominent reflection from this contact is only seen in the bottom image. Scattering from the coarse-grained layer in the top image masks it. The white line in the single B-scan at left is the line trace of a single A-scan.

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Empirical Velocity-Pressure and Porosity-Pressure Trends in Unconsolidated Sands

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Summary

We developed empirical velocity-pressure and porosity-pressure trends from compressional and shear wave velocity and porosity measurements on unconsolidated sand and glass bead samples over a pressure range from 100 kPa to 20 MPa. We found the S-wave velocity to vary with approximately the fourth root of the effective pressure over the entire range of pressures, while the P-wave velocity demonstrates a slightly lower pressure dependence. Preconsolidation reduces the pressure dependence of both velocities, with the effect being larger for the P-wave velocities, and results in largely unrecoverable porosity loss. Samples prepared with varying sorting qualities produced a range of initial porosities, but the effect on the P and S-wave velocities and on their pressure trends was minimal. Water-saturated P and S-wave velocities were modeled with Gassmann fluid substitution. The water-saturated P-wave velocities show a large porosity dependence, which results in a significant porosity dependence in the V_P - V_S ratio, especially at low pressures.

Introduction

The hazards posed to offshore drilling by unknown overpressures at shallow depths have prompted the use of P-wave velocities or V_P - V_S ratios to detect overpressures prior to or during drilling. In general, pressures are assessed by developing an empirical, site-specific, normally pressured P-wave interval-travel-time vs. depth trend. This trend is usually assumed to be linear when depth is plotted against the log of the interval travel-time, and deviations above this trend are assumed to be the result of high pore pressures (Pennebaker, 1970; Pilkington, 1988). Recently Huffman and Castagna (2001) and Prasad (2002) have demonstrated the potential of the V_P - V_S ratio as an indicator of pressure, which may be extracted from multi-component or large-offset reflection data. The V_P - V_S ratio is especially sensitive to pressure for water-saturated sediments. At low effective pressures the shear velocity drops to near zero while the P-wave velocity is limited to values above 1500 m/s, and so the V_P - V_S ratio rises dramatically at low pressures.

Because of the frequent use of the dynamic shear modulus in geotechnical applications, a great deal of research has been conducted on the pressure dependence of the shear modulus and of the shear wave velocity in soils at very low pressures. A large body of experimental work at pressures generally below 500 kPa has demonstrated that the shear velocity of sands demonstrates a pressure dependence of

approximately the fourth root of the pressure: $p^{1/4}$ (Hardin, 1980; Hryciw and Thomann, 1993).

Hardin and Blandford (1989) developed semi-empirical expressions for the shear modulus and P-wave modulus as functions of the effective pressure and porosity. A simplified form for an isotropic stress state is given by:

$$M_{ij} = \frac{OCR^k}{F(e)} S_{ij} p_a^{1-n} \sigma'_{av} \quad (1)$$

where M_{ij} is the modulus in the plane of propagation, σ'_{av} is the mean effective stress, and p_a is the atmospheric pressure. Equation (1) includes two free parameters: S_{ij} , a multiplier to account for textural factors and structural anisotropy, and n , which dictates the pressure dependence of the modulus. The void ratio function, $F(e) = 0.3 + 0.7e^2$, is meant to account for the effect of porosity differences, whether from textural differences between samples or from the compaction of a given sample. The OCR^k term corrects the pressure dependence for the effects of preconsolidation, where OCR is the overconsolidation ratio, and k is a function of the plasticity index. The overconsolidation ratio is defined as the preconsolidation pressure divided by the current pressure, and so the pressure dependence of the modulus for unloading or reloading paths is simply the effective pressure, p , to the $n-k$. Little experimental data has been collected on unconsolidated sediments to constrain the free parameters for the P-wave modulus.

We have run a series of experiments on reconstituted sand and glass bead samples to measure the pressure dependences of the velocities in unconsolidated sands and to investigate the effects of sorting and preconsolidation on the velocities and porosity. We measured the P and S-wave velocities at pressures from below 100 kPa up to 20 MPa in order to test the pressure dependence over this entire pressure range. In this paper we report our results and discuss the porosity and velocity trends observed.

Experimental Method

The experimental apparatus consists of a sample holder that is inserted into a hydrostatic pressure vessel. The two end caps of the sample holder contain both P and S-wave transducers made with 200 kHz piezoelectric (PZT) crystals. The samples, 3.8 cm (1.5 in.) in diameter, were generally prepared to be about 3 cm long. Velocities were calculated by picking first arrivals from pulse-transmission signals. With this arrangement we were able to get interpretable shear signals at pressures as low as 50 kPa, with errors of approximately 2% for the P-wave velocities and 4% for the S-wave velocities.

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The data presented here are from samples of a fine grained, well sorted, quartz sand, called the Santa Cruz Aggregate, as well as from synthetic samples made from sieved fractions of this sand and of glass beads. Four samples of the sand were run, two dry and two water-saturated. Two other samples were made of sieved fractions of this sand, both of which were run dry: one entirely of a large grain size fraction, and a second sample of 65% by mass of the large size fraction, and the other 35% of grains about $\frac{1}{4}$ the size (sample Sa-35%-Small). A total of seven glass bead samples were run, all dry. Three samples consisted of different narrow size ranges of beads. Three samples were made with a “bimodal” mixture of grain sizes, with 35% of the mass made up of smaller grains and 65% of larger grains. Finally, one sample was made up of a broad range of particle sizes. More detailed descriptions of the samples and of the sample preparation procedures can be found in Zimmer et al. (2002a,b).

The initial porosity of the samples was calculated from the grain density, dry sample mass, and sample volume. The changes in the sample volume and porosity with pressure were then monitored by measuring changes in the length and circumference of the samples. An error analysis of our porosity measurements estimates the error at 0.012 to 0.016, or 3 to 4% of the total porosity.

The pressure paths followed generally included a number of pressure cycles of increasing peak pressure for each cycle. The velocities and porosity were measured at the same set of pressures during each cycle (e.g. 0.1, 0.2, 0.5... MPa). This allowed us to compare the velocities and porosities measured at the same pressure for a sample that had been preconsolidated to a range of higher pressures. Five of the samples were cycled through 8 or 9 cycles, while the rest were cycled through between 1 and 5 cycles.

Pressure Trends and Preconsolidation Effects

We treated n , k , and S as free parameters and fit Equation (1) to the moduli calculated from our velocity measurements. We show the empirical fit for the two moduli from sample Sa-35%-Small in Figure 1. These figures demonstrate that there is a slight preconsolidation effect to the samples, where the trends for the unloading and reloading paths can be seen to diverge slightly. The coefficients for this sample and for the dataset as a whole are given in Table 1. The coefficients of all samples are given in Zimmer et al. (2002b). For the shear modulus we found the value of n to vary from between 0.352 to 0.612, so the pressure dependence of the S-wave velocity then varies between p to the 0.176 and the 0.306. The value of k for the shear modulus varied between -0.073 and 0.101 and averaged -0.003 (effectively zero). These values of k are relatively small compared to the values predicted for clays,

which can be as large as the value of n for high plasticity clays (Hardin and Drnevich, 1972). The pressure dependence of the P-wave modulus is consistently lower than that of the shear modulus, with n ranging from 0.304 to 0.478 and averaging 0.06 less than the n for the shear modulus for the same sample. The value of k , on the contrary, was almost always higher for the P-wave modulus than for the shear modulus, averaging 0.026 and ranging from -0.033 to 0.104 . We did not observe a significant systematic initial porosity effect on the pressure dependence (n) or on the preconsolidation dependence (k), thus the variation in initial porosity that comes from the differences in sorting does not significantly affect these dependences over the porosity range tested.

Porosity reduction of the samples was fit with an empirical expression of the form:

$$\phi = \phi_0 (1 - A p^m OCR^l) \quad (2)$$

where ϕ is the porosity, ϕ_0 is the initial porosity of the samples, p is the effective pressure, OCR is the overconsolidation ratio, and A , m , and l are free parameters. The fit of this expression is shown in Figure 2, where $1 - \phi/\phi_0$ is plotted against the pressure, and the coefficients for sample Sa-35%-Small are given in Table 1. Here the unloading path can be seen to be very different from the loading path, and most of the porosity is not recovered during unloading.

Figure 1: The velocity results from sample Sa 35% Small: A) V_P vs. pressure, B) V_S vs. pressure, The black lines represent fits to the Equation (1).

Figure 2: Normalized porosity-pressure fit for sample Sa-35%-Small, after Equation (2).

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Table 1

Sample	S	n	k	A	m	l
Sa 35% Small P-wave Mod.	1754	0.47	0.059	0.0039	0.65	0.59
Shear Mod.	539	0.51	0.029			
Entire Dataset P-wave Mod.	2336	0.41	0.032			
Shear Mod.	604	0.47	0.001

Figure 3: The velocity data from all of the samples plotted against porosity and color-coded according to the pressure at the time of measurement: A) V_P B) V_S . The lines represent the Reuss average between the moduli of the high-porosity dry-frame end-member and solid quartz.

Sorting Effect

Figure 3 shows the velocity data from all of the samples plotted together against porosity and color coded according to the pressure. These plots illustrate the relative effects of the pressure and sorting on the velocity and porosity. Increased pressures result in large increases in the velocity for these coarse sediments, but the porosity is only slightly reduced. On the contrary, the porosity reductions that result from a decrease in the sorting quality produce only minor increases in the velocity.

The lines shown in Figure 3 demonstrate the velocity effect predicted for the sorting-induced porosity changes by the Reuss average between the dry frame moduli and the solid quartz moduli. The Reuss (isostress) average, the weighted harmonic average between two end member moduli, simulates the weakest possible way to combine two materials. Here the Reuss average is used to represent the theoretically minimal possible change in the velocities produced by adding solid grain material (quartz) to the high porosity sample (assuming constant packing) (Dvorkin and Nur, 1996). The high porosity end-member moduli come from the highest porosity data for each pressure. The fact that the trend is so well described by the Reuss average implies that the small grains added tend to sit more or less passively in the pores, reducing the porosity but not adding significantly to the stiffness of the sediment.

Figure 4 shows the results of Gassmann fluid substitution with water of the P-wave velocities of all the dry samples. As seen in this figure, the water saturated P-wave velocities

have a significant porosity dependence, with lower porosities correlating to higher P-wave velocities. The presence of water in the pores stiffens a low porosity sediment more than a high porosity one, though the trend is still described reasonably well by the Reuss average between the moduli from the water-saturated, high porosity end-member and solid quartz. The shear wave velocities, which only require a density substitution, show a slight porosity dependence at the higher pressures, but no porosity dependence at low pressures.

Implications for Pressure Prediction and Monitoring

The V_P - V_S ratios for all of the Gassmann fluid-substituted data are shown in Figure 5, plotted against pressure and color-coded by porosity. The V_P - V_S ratio increases from below 3 at 10 MPa to a mean value of about 7 at 0.5 MPa. Unfortunately there is a considerable amount of scatter in the V_P - V_S ratio, especially at low pressures, that would generate a significant amount of uncertainty in the in situ effective pressure determined from V_P - V_S ratio measurements. For example, if the measured V_P - V_S ratio is 5, the in situ pressure (based on our fluid-substituted data) could vary from between 0.2 and 2 MPa.

Figure 5 demonstrates that there is a porosity dependence to much of this scatter in the V_P - V_S ratio at low pressures. Since these non-cohesive sediments demonstrate only relatively small changes in both the velocity and porosity with compaction, preconsolidation has only minor influences on the V_P - V_S ratio. This implies that for water-saturated sands there should be only very little difference between the V_P - V_S ratio signatures of overpressures produced through undercompaction or through repressurization mechanisms such as fluid expansion. As preconsolidation can cause a much larger reduction in the porosity of clay-rich sediments (Bowers, 1995), this effect could potentially be more important for cohesive sediments.

Figure 4: The P-wave velocity results after Gassmann fluid substitution with water, plotted against porosity, again showing the Reuss averages between the high porosity end-member and solid quartz.

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Figure 5: The V_p - V_s ratio plotted in a log-log plot against pressure and color coded according to the porosity.

The porosity variation between the samples comes primarily from the differences in the initial porosities of the samples due to differences in their sorting. The scatter in the V_p - V_s ratio is therefore a product of the different effects that the porosity has on the water-saturated P- and S-wave velocities. The small porosity-dependence of the S-wave velocities at lower pressures allows the porosity dependence of the water-saturated P-wave velocity to show through. The implication is that for robust pressure estimates to be made from measured V_p - V_s ratios it will be necessary to correct for this porosity effect at pressures below about 10 MPa. Such a correction could potentially be developed through a suite of laboratory measurements on natural sands, and would need to be developed in order to permit the use of the V_p - V_s ratio as a quantitative pressure indicator.

Conclusions

We have measured the pressure dependence of the P and S-wave velocities through sand and glass bead samples with a variety of initial porosities produced by varying their sorting. We also ran the samples through a series of pressure cycles of increasing peak pressures in order to observe the effects of preconsolidation on the samples. We have observed the following results:

- 1) The S-wave velocities vary with the fourth root of the effective pressure over the entire pressure range measured (100 kPa to 20 MPa). For the P-wave velocities the pressure dependence is slightly lower – about $p^{0.22}$.
- 2) These non-cohesive sediments do exhibit a small preconsolidation effect on the velocities, though the effect is generally smaller for the S-wave velocities than for the P-wave velocities.
- 3) While poorer sorting can result in large decreases in the porosity, it has a relatively small effect on the S-wave velocities and dry P-wave velocities. The porosity reduction does produce a larger effect on the water-saturated P-wave velocities, as calculated from Gassmann fluid-substitution of our dry velocity results. This porosity

dependence of the water-saturated P-wave velocities, and the limited porosity dependence of the S-wave velocities, results in a porosity dependence in the V_p - V_s ratio. Higher porosities correspond to a lower V_p - V_s ratio.

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Measurement of Young's modulus of clay minerals using atomic force acoustic microscopy

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[1] The presence of clay minerals, hydrous aluminosilicates that are smaller than 2 μm can alter the elastic and plastic behavior of materials significantly. We have used Atomic Force Acoustic Microscopy (AFAM) to measure elastic properties of clay minerals. We demonstrate the AFAM technique for measuring elastic properties of soft materials. Using this technique, we present first-ever quantitative measurements of Young's modulus in clay. The Young's modulus of dickite was measured as 6.2 GPa. **INDEX TERMS:** 5102 Physical Properties of Rocks: Acoustic properties; 3909 Mineral Physics: Elasticity and anelasticity; 5112 Physical Properties of Rocks: Microstructure; 3994 Mineral Physics: Instruments and techniques

1. Introduction

[2] Presence of clay minerals, hydrous aluminosilicates with grain size smaller than 2 μm , alters the elastic and plastic behavior of materials significantly. Using acoustic microscopy images of sandstones, Prasad [2001] has shown that the clay in contact zones has significantly lower impedance than the quartz cements. Ultrasonic velocities are reduced in clay-rich sandstones when the clay is load-bearing [Tosaya, 1982; Han *et al.*, 1986]. As pore filling materials, they block hydraulic pathways and decrease permeability. In the presence of water, clay minerals can swell and cause considerable formation damage. Despite their ubiquitous presence, measurements of the elastic properties of clay minerals remain a challenge.

[3] Since clay minerals are an integral part of many formations and seismic measurements are the main tools for identification of subsurface lithologies, knowledge about the elastic properties of clays is essential for interpretation and for modeling the seismic response of clay-bearing formations. However, elastic properties of clay minerals are almost unknown [Alexandrov and Ryzhova, 1961]. Until now, estimates of single crystal elastic properties have been either theoretical [Katahara, 1996], or based on extrapolations from measurements on clay-epoxy mixtures [Wang *et al.*, 2001].

[4] Recent studies have shown that it is possible to make measurements of dynamic Young's modulus at nanometer resolution on various materials using Atomic Force Acoustic Microscopy (AFAM) [Kester *et al.*, 2000; Rabe *et al.*, 2001; Amelio *et al.*, 2001]. We have investigated the feasibility of such measurements on clay minerals. We describe the methodology for making

quantitative measurements of stiffness and elastic moduli using AFAM and show results of such measurements on clay samples.

2. Atomic Force Acoustic Microscopy

[5] In scanning Atomic Force Microscopy (AFM), the topography of the sample surface is measured from deflections of the cantilever tip from its equilibrium position. AFAM is an enhancement of the AFM technique in which ultrasonic transducers are used to insonify the contact zone between the cantilever tip and the sample surface. To this end, either the sample or the cantilever holder might be excited by an ultrasonic transducer [Rabe *et al.*, 1996; Yamanaka *et al.*, 1999]. The dynamic Young's modulus is determined by measuring the difference of the cantilever contact-resonance frequencies relative to its free resonances. The resonance frequencies of the flexural modes of the cantilever will depend, amongst other parameters, on the stiffness of the tip-sample contact, which in turn is a function of the Young's modulus of the sample and the tip, their Poisson's ratios, the cantilever tip radius and shape, the load exerted on the tip, and the geometry of the sample surface. In an isotropic material, if the elastic modulus of the tip and its radius are known, the Young's modulus of the sample can be determined if the Poisson's ratio is either known or assumed [Kester *et al.*, 2000; Rabe *et al.*, 2001]. In the anisotropic case, the so-called indentation modulus [Vlassak and Nix, 1993] is determined. For reliable measurements, the sample surface must be of nanometer smoothness over an area of tens of nanometers where the measurements are taken. The Young's modulus is measured with a spatial resolution of a few tens of nanometers.

[6] In an actual measurement, an area with low topographic variations is first selected with an AFM. Then, the free and contact resonances of the cantilever are measured. The excitation frequency is varied and the cantilever vibration amplitude is measured as a function of frequency. The frequency range used (10 kHz to about 3 MHz) covers typically the first three flexural modes of the cantilever in contact. The procedure is repeated with different normal forces on the tip in order to obtain information on the tip shape and to check whether a Hertzian contact prevails. For a complete set of measurements on an unknown sample, a standard sample is first measured with AFAM to determine the tip radius. The unknown sample is measured next. Finally, the standard sample is measured again to ascertain that the tip remained intact during the measurement. An averaged value between both standard measurements is used as tip radius for the calculation of the Young's modulus in the unknown sample [Kester *et al.*, 2000; Rabe *et al.*, 2001].

[7] Kester *et al.* [2000] give an accuracy of the Young's modulus from the average of the scatter and considerations of the simplifications mentioned above to be about 40%. By comparing indentation moduli of silicon, Rabe *et al.* [2001] have shown that it is possible to differentiate between indentation moduli of Si (100) and Si (111) crystals, that differ only by 4%. However, our

Figure 1. Contact mode AFM image of kaolinite (a) and dickite (b) powders. The topography is coded with gray colors: white = highs, black = lows. Although topography on each stack of grains is not large, the cantilever position on the sample is not stable, most probably due to the small grain size. Kaolinite grains show a typical rosette texture. Typical clay booklets with topography variations below 10 nm can be observed in the dickite sample.

measurements on a softer material, polystyrene, showed that an accuracy of at least 40% can be expected for such materials. We present here initial results of AFM and AFAM measurements for clay minerals and the Young's modulus derived from them.

3. Samples and Sample Preparation

[8] Clay mineral powders were used in this study. Fused silica, mica and polystyrene served as the standard samples with known Young's modulus [Amelio, 2000]. The Young's modulus of polystyrene measured with ultrasonic pulse-echo was 3.5 GPa.

[9] Kaolinite, dickite, and mica powders were immersed in distilled water and insonified for 30 minutes. After allowing the larger particles (mainly silts) to settle, a dilute water-clay mixture was extracted from the top of the container. This suspension was diluted further, mixed, and allowed to settle for 5 minutes. The water-clay mixture taken from the top of this solution was put on a glass slide heated by a glass lamp. In this manner thin layers of kaolinite and dickite samples were obtained. Mica particles were co-mingled with the clay. With this method, the clay minerals were aligned with their C-axis perpendicular to the glass slide.

4. AFM Imaging

[10] AFM images of the kaolinite and dickite samples are shown in Figures 1a and 1b, respectively. In these images, as in all AFM images, the topographic differences are originally color coded and then converted to a gray scale (black = dips and white = highs). The kaolinite image shows grain stacks with a typical

(a) Topography image, scale 500 nm (b) AFAM image at 987 kHz

Figure 2. Contact mode AFM of image of dickite powder showing mineral texture (left image). Topography variations, color coded as in Figure 1 are below 500 nm. The right-side image is an acoustic image with AFAM: It is made with contact mode AFM after insonifying the sample at a resonance frequency of the sample-tip system. Although, topographic effects are small, the AFAM image has some noise and disturbances marked by arrows probably due to sample movements on the glass slide.

rosette structure of the kaolinite. Each mineral stack is less than 1 μm wide. Although the topography varies only by a few tens of nanometers within each stack, the image shows numerous disturbances caused mainly by movement of the sample with respect to the cantilever. The dickite sample with slightly larger grain size shows the typical clay booklets observed in SEM studies. The topography of this stack of about 7 layers is less than 10 nm. The measurements were made in contact mode with a stiff triangular cantilever. This mode of operation was found to create disturbances in the imaging mode due to sample movement and grain deformation. The so-called tapping mode [Zong *et al.*, 1993] gave better resolution and imaging results. Using this mode, we ascertained that the texture shown in Figure 1b is repeated at random locations throughout the dickite sample.

5. AFAM Quantitative Analyses

[11] Despite the good quality AFM images, quantitative AFAM of clay measurements were more difficult to obtain. The AFAM image (Figure 2) often had disturbances in the scanning mode. However, good repeatability was observed in the single-location contact-resonance spectra obtained at random clay locations with a rectangular cantilever (stiffness = 1.5 N/m). Quantitative analyses were made at 10s different clay locations. A typical spectra sequence is shown in Figure 3 for different materials. The contact-resonance spectra for polystyrene are at a lower frequency than mica and fused silica. The second contact-resonance spectra

Figure 3. Contact-resonance spectra in polystyrene, dickite, mica, and fused silica. While there is a small shift in first contact-resonance towards lower frequencies (a) in the low impedance polystyrene and clay, a clear impedance based separation in frequency is observed in the second contact-resonance (b). The resonance spectra are broader in the low impedance materials, possibly due to attenuation in the sample.

Table 1. Calculated Effective Stiffness (k^*) and Young's Modulus (E) for Various Positions on the Dickite Sample

Sample	k^*		E1 (GPa)		E2 (GPa)	
	30	45	30	45	30	45
			(load in nN)			
Polystyrene	56	58	3.2	3.7		
Mica	301	300	56.9	60.2		
Position 1	81	76	5.99	5.69	6.7	6.13
Position 2	72	100	4.99	8.75	5.58	9.45
Position 3	81	83	5.99	6.52	6.7	7.04
Position 4	80	83	5.88	6.52	6.57	7.04
Position 5	78	75	5.65	5.57	6.32	6.0
Position 6	76	77	6.08	6.26	5.16	5.78
Position 7	77	78	6.19	6.38	5.26	5.9
Position 8	80	81	6.57	6.77	5.56	6.25
Position 9	78	59	6.32	4.14	5.37	3.83
Position 10	77	85	6.19	7.3	5.26	6.74

E1 corresponds to the tip radius obtained from measurements on the standard sample **before** measurements on clay and E2 corresponds to the tip radius obtained from measurements on the standard sample **after** the measurements on clay. The mean value of E is 5.9 GPa at 30 nN and 6.4 GPa at 45 nN static loads and the standard deviation for each load step is 0.54 GPa and 1.27 GPa, respectively.

are as much as 500 kHz apart. Note that the relative frequency difference of the contact resonances on different materials depends on the mode of the cantilever vibrations [Turner *et al.*, 1997]. The contact-resonance spectra in polystyrene and clay also appear to be broader, possibly due to internal friction losses in the sample. Young's modulus for dickite was calculated from the spectra shown in Figure 3 by assuming a Poisson's ratio of 0.3. The values obtained for all 10 locations are given in Table 1 along with the standard deviation for the calculations at different locations (0.54 GPa and 1.27 GPa for 30 nN and 45 nN static load, respectively). The mean value for Young's modulus from our measurements is 5.9 GPa at 30 nN and 6.4 GPa at 45 nN static load.

[12] The Young's modulus was calculated by assuming a spherical tip and Poisson ratio = 0.3 for the dickite. The average value according to Table 1 is 6.2 ± 1.0 GPa. Although the values show a high reproducibility, a larger error of about 40% (± 2.5 GPa) needs to be assessed for the absolute values at present. This error arises from uncertainties in the theories required to model the cantilever vibration modes and the contact area between tip and sample. The input parameters for these calculations, such as beam geometry and tip shape, are only known to a certain approximation. The Young's modulus is calculated by comparing the vibration modes of the AFM cantilever on a known reference sample with those on the unknown sample. The error can be relatively small if the elastic constants of the calibration sample are close to those of the unknown sample. A future set of measurements will improve the precision of our measurement by using new reference samples. Furthermore, the complex clay-tip interactions will be modeled to account for deviations from the assumed model. For example, surface tension forces of clay might lead to higher lateral forces than assumed. Elastic moduli in the c_{11} and c_{33} directions in sheet silicates can differ by a factor of 3 to 3.5 [Alexandrov and Ryzhova, 1961]. Future work will improve accuracy and resolve the issue of anisotropy in these samples.

6. Conclusions

[13] Our analyses of the AFAM measurements and comparisons with measurements on similar materials give Young's modulus values of 6.2 GPa (± 2.5 GPa) in dickite in the c_{11} direction. This value is in agreement with the theoretically derived bulk modulus

value of 10 – 12 GPa by Berge and Berryman [1995]. Our study resolves the controversy surrounding elastic modulus values of clay minerals. Various values have been suggested for bulk modulus of clay minerals. For example, extrapolation from shale measurements [Tosaya, 1982; Castagna *et al.*, 1985; Han *et al.*, 1986], theoretical models using measured values on other sheet silicates [Katahara, 1996], and extrapolation from measurements on clay-epoxy mixtures [Wang *et al.*, 2001] gives bulk modulus values of kaolinite, which has a similar structure and mineralogy as dickite, between 21 and 55 GPa - much higher than our measured value of 6.2 GPa. We also demonstrate for the first time how the AFAM technique can be used for soft materials. Implications of this study go beyond the applications for earth sciences.

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Pressure and porosity influences on V_P - V_S ratio in unconsolidated sands

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Elevated pore pressures, commonly encountered in the shallow, unconsolidated section of the sedimentary column, present a significant hazard during the drilling and completion of offshore wells. The porosity of the ocean-floor sediments is high and a cover of low-permeability clay can prevent the underlying sediments from draining, even at very shallow depths below the seafloor. As further deposition loads the sediment, the entrapped fluids impede normal compaction by becoming pressurized. The lowered effective stress that results from the higher pore fluid pressure produces a proportional drop in the strength of non-cohesive sediments, resulting in very weak shallow sands. This weakness can result in washouts or fracturing in the shallow strata, potentially leading to the loss of the well and of neighboring wells.

These problems are exacerbated in deepwater wells where the long column of drill mud makes it difficult to control the pressure placed on the formation. In addition, slight overpressures result in a very narrow mud-weight window. A mud-weight below the pore pressure can result in shallow water flow, analogous to large scale washouts of unconsolidated sand layers. On the contrary, high mud weights can fracture these weak, shallow strata, with fractures potentially propagating all the way up to the surface and out to other wells in the same template (Ostermeier et al., 2000). The detection of high pore-pressure regions prior to drilling and the remote measurement of in-situ pressures has the potential to prevent the loss of many deepwater wells and of the millions of dollars that would be spent on them.

For a number of years, pore pressure evaluation has mainly relied on the use of empirical relationships between pressure and seismic velocities or interval transit times. Recently, Huffman and Castagna (2001) and Prasad (2002) have demonstrated the use of the V_P/V_S ratio as an indicator of pressure, as extracted from AVO analysis or multicomponent reflection data. One weak link in these analyses is the paucity of laboratory data available to calibrate the measured velocities or V_P/V_S to the in-situ pressure, especially for unconsolidated samples at the low effective pressures appropriate to this environment. This is in a large part due to the difficulties in propagating an ultrasonic signal through unconsolidated sediments at such low pressures. In an effort to constrain the velocity-pressure trends and to investigate the influence of porosity reduction on these trends, we have undertaken a systematic set of experiments to measure the ultrasonic velocities through loose sediments at effective pressures from below 100 kPa up to 20 MPa. Here we present measurements of P - and S -wave velocities through sand and glass bead samples at a range of porosities. We also discuss the effects of pressure, sorting, and compaction on the velocities and porosities, and implications for the evaluation of pore pressures using V_P/V_S in unconsolidated sands. This study did not include a number of other factors that could also influence the velocity-pressure relationship, including the age of the sediment, clay content, packing (or depositional environment), or stress-induced or inherent anisotropy, though they should be considered when

Figure 1. P - and S -wave velocity plotted against effective pressure for all of our sand and glass bead samples, and for similar measurements on clean sands from Prasad and Meissner (1992), Domenico (1977), Yin (1992), Estes et al. (1994), and Robertson et al. (1995).

attempting to predict in-situ pressure from velocity or V_P/V_S measurements.

Experimental methods and samples. We adapted a hydrostatic pressure apparatus to be able to make P - and S -wave velocity measurements on unconsolidated samples at low effective pressures. One adaptation was the use of lower frequency (200 kHz) transducers to produce and receive the ultrasonic signals. We also made the transducer face plates out of a glass-filled polycarbonate (shatterproof glass) in order to improve the acoustic impedance matching between the samples and the transducers. With this arrangement, we were able to get interpretable shear signals at pressures as low as 0.05 MPa, with errors of approximately 2% for P -wave velocities and 4% for S -wave velocities.

The data presented are from samples of a fine grained,

Table 1.

Sample Name:		Initial Porosity					
Sands:	Dry 1	0.414	clean, quartz sand; well sorted				
	Dry 2	0.432	D ₁₀ =0.16 mm ¹				
	Wet 3	0.400	D ₆₀ =0.3 mm				
	Wet 4	0.411					
			Fraction	Size, μm (φ scale) 2	Fraction	Size, μm (φ scale)	
Glass Beads:	Big Sand	0.409	1	295-350 (1.5-1.75)	—	—	
	GB Big	0.381	1	295-350 (1.5-1.75)	—	—	
	GB Small	0.411	1	74-88 (3.5-3.75)	—	—	
	GB 35% Small	0.315	0.65	295-350 (1.5-1.75)	0.35	74-88 (3.5-3.75)	
	GB Tiny	0.421	1	37-44 (4.5-4.75)	—	—	
	GB 35% Tiny 1	0.296	0.65	295-350 (1.5-1.75)	0.35	37-44 (4.5-4.75)	
	GB 35% Tiny 2	0.258	0.65	295-350 (1.5-1.75)	0.35	37-44 (4.5-4.75)	
	GB Broad	0.338		37-710 (0.5-4.75)	—	—	

¹ D_{10} and D_{60} are the grain diameters below which 10%, or 60% respectively, of the mass of the sample is found.

² $\phi = -\log_2 d$, where d is the grain diameter in mm.

well sorted, quartz sand called the Santa Cruz Aggregate, and from synthetic glass bead samples. Four samples of the Santa Cruz Aggregate were run, two dry (Dry 1, Dry 2) and two water-saturated (Wet 3, Wet 4). A fifth sample (Big Sand) was also prepared by sieving the 0.295-0.350 mm grain sizes out of this sand, and running this fraction dry. A total of seven glass bead samples were run, all dry. Three samples (GB Big, GB Small, and GB Tiny) consisted of different narrow size ranges of beads. Three samples (GB 35% Small, GB 35% Tiny 1 and 2) were made with a "bimodal" mixture of grain sizes, with 35% of the mass made up of smaller grains and 65% of larger grains. Finally, one sample (GB Broad) was made up of a broad range of particle sizes. Table 1 summarizes the grain sizes and initial porosities of all samples.

Special attention was given to preparing the samples in such a way as to ensure complete mixing of the different grain sizes, and to maintain full saturation of the water-saturated samples. However, the sample preparation varied based on whether the sample was a single grain size and dry (air pluviated) or water-saturated (water pluviated), or was a mixture of grain sizes (some samples mixed dry, some mixed while moist and allowed to dry once in the sample holder). These differences in sample preparation produced variations in the packing of the grains and some scatter in the data.

The porosity was calculated from the grain density, dry sample mass, and sample volume. The changes in the sample volume and porosity with pressure were then monitored by measuring changes in the length and circumference of the samples. The error in the porosity is estimated to be about 0.02, or about a 5% relative error.

The pressure path followed generally included a number of pressure cycles of increasing peak pressure for each cycle, up to 20 MPa. The velocities and porosity were measured at the same set of pressures during each cycle (e.g., 0.1, 0.2, 0.5... MPa). This allowed us to compare the velocities and porosities measured at the same pressure for a sample that had been preconsolidated to a range of higher pressures. Four samples were cycled through 8 cycles, while the rest were cycled through between 1 and 5 cycles.

Ultrasonic results. Figure 1 shows the P - and S -wave velocities plotted against pressure for all our samples, and

for measurements on clean sands from Prasad and Meissner (1992), Domenico (1977), Yin (1992), Estes et al. (1994), and Robertson et al. (1995). This figure demonstrates similar trends between the data from all sources. The scatter is presumably a result of textural differences between the samples such as their sorting, angularity, packing (sample preparation), and mineralogy.

Figure 2a shows all dry velocity results plotted against porosity, and color-coded according to the effective pressure. The lower porosity data (below 0.35) come from the glass bead samples where two size fractions of beads were mixed, with the smaller grains partially filling the pore space between the larger grains. The lines in Figure 2a represent velocities calculated with the Reuss (isostress) average between the pure mineral moduli (quartz) and the high-porosity dry frame moduli for a given pressure. The high porosity end-member moduli come from the highest porosity data for each pressure, taken from the virgin compaction curve of sample Dry 2, which are highlighted with black circles in Figure 2a. The Reuss average, the weighted harmonic average between the two end member moduli, simulates the weakest possible way to combine two materials. Here the Reuss average is used to represent the minimum possible effect on the velocities of adding solid grain material (quartz) to the high porosity sample. The Reuss average was calculated as follows:

$$\frac{1}{M} = \frac{f_D}{M_D} + \frac{f_Q}{M_Q}$$

M is the resulting average modulus, M_D is the modulus of the dry frame at the pressure of interest, and M_Q is the modulus of pure quartz. The fraction of dry frame, f_D , is simply given by $f_D = \phi / \phi_0$, where ϕ is the porosity, and ϕ_0 is the initial porosity at that pressure from sample Dry 2. The fraction of pure quartz, f_Q is then just $1 - f_D$, or $(1 - \phi / \phi_0)$.

The data from a single sample (the streaks of points that rise more or less vertically in Figure 2a—e.g., the black or blue highlighted points) demonstrate the effect of the pressure on the velocity-porosity trend: a significant velocity increase with relatively little change in the porosity. Data collected at the same pressure, shown in the same color, illustrate the effects of sorting. Poorer sorting produces large decreases in the porosity, but relatively small increases

Figure 2. Velocity data plotted against porosity and color coded with effective pressure, along with lines showing the Reuss average between the quartz mineral modulus and the highest porosity dry-frame moduli for different pressures. The high porosity dry-frame moduli are taken from the virgin compaction curve of sample Dry 2, highlighted with black circles around the points. (a) All of the data. (b) Only data from the large size fraction sand (Big Sand, highlighted with blue circles in part (a), run dry through eight cycles of increasing peak pressure. The arrows indicate the effects on the velocity-porosity trend of pressure (for virgin loading and reloading of a preconsolidated sample), and of porosity reduction from changes in sorting or compaction.

in the velocities. While there is some scatter in the data above and below the Reuss average, likely because these data come from glass bead and sand data prepared in slightly different ways, all in all the data collected at similar pressures do tend to lie along this trend. This result confirms the supposition of Dvorkin and Nur (1996) that the porosity variation produced by variations in sorting has a minimal (Reuss average) effect on the velocities. The implication is that the poorer sorting results in small grains lying passively in the pores between larger grains. Most small grains reduce the porosity but do not add significantly to the stiffness of the sediment, and so the velocity is only slightly increased.

Figure 2b illustrates the effect on the velocities of preconsolidation or compaction. This figure shows a velocity-porosity plot of the data from only the large size fraction sand, Big Sand, again color-coded by the pressure of the measurement. The data from this sample are highlighted with blue circles in Figure 2a. Also shown in Figure 2b are

the same Reuss average lines as in Figure 2a, based on the virgin compaction curve of sample Dry 2. Sample Big Sand was run through eight pressure cycles, during which measurements were made at the same eight pressures for each cycle, once the peak pressure had exceeded that measurement pressure. Velocity measurements taken at the same pressure but compacted under higher preconsolidation pressures are seen to rise slightly as the porosity is reduced by compaction. This effect is more pronounced and more consistent for the higher measurement pressures. The rise is also slightly more pronounced for the P -wave velocities than for the S -wave velocities. For the higher measurement pressures, this effect is noticeably greater than the basically flat trend predicted by the Reuss averages based on the Dry 2 data. At the lower measurement pressures, the data follow the Reuss average trend fairly well, although the S -wave velocities rise slightly, then fall again with greater consolidation pressures. Nevertheless, as for the case of sorting, the total effect on

Figure 3. Velocity data from the glass bead samples plotted against effective pressure, and color coded according to porosity. (a) Gassmann fluid-substituted P -wave velocities. (b) Fluid(density)-substituted S -wave velocities. At low pressures the P -wave data still show a consistent porosity dependence, while the shear-wave velocities show less of a spread in the velocities at low pressure.

Figure 4. V_P/V_S ratio data for the water-saturated (Gassmann fluid-substituted) glass bead samples plotted against the effective pressure and color-coded for porosity. (a) Linear scale pressure-axis. (b) Log scale pressure-axis. The different porosity dependences of the water-saturated P - and S -wave velocities can be seen to produce a porosity dependence in the V_P/V_S ratio at low pressures.

the velocities of this compaction-driven porosity reduction is relatively minor for these cohesionless sediments.

Figure 3 shows the results of Gassmann fluid substitution (with water) of the velocities of all dry glass bead samples, plotted against pressure and color-coded for porosity. As seen in this figure, higher porosities produce lower P -wave velocities for the water-saturated samples. The shear-wave velocities, which require only a density substitution, show similar porosity dependence at the higher pressures, with the low-porosity data resulting in higher velocities. At lower pressure, however, the porosity effect on the S -wave velocities is relatively small, as shown in the reduced scatter in Figure 3b and the rather flat Reuss average lines for the lower pressures in Figure 2a.

Pressure prediction. V_P/V_S ratios for all the Gassmann fluid-substituted glass bead data are shown in Figure 4, plotted against pressure and color-coded by porosity. As discussed by Prasad (2002) and Huffman and Castagna (2001), the water-saturated V_P/V_S shows a dramatic rise as the effective pressure decreases. The V_P/V_S ratio increases from below 3 at 10 MPa to a mean value of about 7 at 0.5 MPa. Unfortunately, at the lower pressures there is a considerable amount of scatter in the V_P/V_S ratio that would generate a significant uncertainty in the in-situ effective pressure determined from V_P/V_S measurements. For example, if the measured V_P/V_S is 5, the in-situ pressure (based on our fluid-substituted glass bead data) could vary between 0.15 and 2 MPa.

In this study, we considered the effect of two factors

on the relationship between V_p/V_s and pressure: preconsolidation and sorting. Both factors influence the porosity of the sediment: preconsolidation by compaction of the grain matrix primarily through rearrangement of grains, and sorting by smaller grains filling pore spaces between larger grains.

Preconsolidation: Because these noncohesive sediments demonstrate only relatively small changes in both the velocity and porosity with compaction, we do not expect preconsolidation to produce a significant V_p/V_s signature. Preconsolidation does have a slightly larger effect on the dry P -wave velocities than on the S -wave velocities, as demonstrated in Figure 2b. At a given pressure, this would result in a slightly higher V_p/V_s in overconsolidated sediments than in normally consolidated ones. This effect does not persist once the data is fluid-substituted, which, except for some scatter at the very lowest pressures, shows no significant V_p/V_s signature from preconsolidation. So for water-saturated sands our results demonstrate no difference in V_p/V_s signature of overpressures produced through undercompaction or through repressurization mechanisms such as fluid expansion. As preconsolidation can cause a much larger reduction in the porosity of clay-rich sediments (Bowers, 1995; Sayers et al., 2001), this effect could potentially be important for cohesive sediments.

Sorting: At low pressures, the fluid-substituted V_p/V_s ratio is a function of both the pressure and the porosity, with low porosities corresponding to high V_p/V_s ratios (see Figure 4b). The porosity variation comes primarily from the differences in the initial porosities of the samples due to differences in their sorting. The spread in the V_p/V_s ratio comes from the different effects that the porosity has on the velocities, as demonstrated in Figure 3 and discussed above. As the relative porosity effect at high pressures is

similar for the P - and S -wave velocities, there is relatively little scatter in V_p/V_s above about 10 MPa. The diminishing porosity-dependence of the S -wave velocities at lower pressures allows the porosity dependence of the P -wave velocity to show through. The implication is that for robust pressure estimates to be made from measured V_p/V_s , it will be necessary to correct for this porosity effect at pressures below about 10 MPa. Laboratory measurements on natural sands could potentially produce a relatively general porosity correction, but at present we would recommend cautious use of V_p/V_s as a quantitative pressure predictor.

Suggested reading. "Elastic properties of unconsolidated porous sand reservoirs" by Domenico (GEOPHYSICS, 1977). "Elasticity of high porosity sandstones: Theory for two North Sea data sets" by Dvorkin and Nur (GEOPHYSICS, 1996). "Overpressure detection from compressional- and shear-wave data" by Dvorkin et al. (*Geophysical Research Letters*, 1999). "The petrophysical basis for shallow-water flow prediction using multicomponent seismic data" by Huffman and Castagna (TLE, 2001). "Dealing with Shallow-Water Flow in the Deepwater Gulf of Mexico" by Ostermeier et al. (Proceedings - 32nd Offshore Technology Conference, 2000). "Acoustic measurements in sands at low effective pressure: Overpressure detection in sands" by Prasad, (GEOPHYSICS, 2002).

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VELOCITY-PRESSURE TRENDS IN SANDS FROM ULTRASONIC P AND S-WAVE VELOCITY MEASUREMENTS

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Abstract

We present both compressional (P) and shear (S) wave velocity measurements from unconsolidated sand and glass bead samples at pressures from 100 kPa to 20 MPa. We found the S-wave velocity to vary with approximately the fourth root of the effective pressure over this entire range of pressures, while the P-wave velocity shows a slightly lower pressure dependence. The pressure dependence of both velocities was influenced by preconsolidation, though the effect was greater for the P-wave velocities. We prepared samples with a range of initial porosities by varying the sorting of the samples, generally by mixing two different sizes of sand grains or glass beads. The samples with lower sorting qualities had significantly lower porosities, but the effect on the P and S-wave velocities was minimal. The Reuss average between the moduli of the highest porosity sample and of solid quartz described the effect of sorting on the velocities and porosity relatively well. The variation in initial porosities due to the sorting did not produce a systematic variation in the pressure dependences. Water-saturated P and S-wave velocities were modeled with Gassmann fluid substitution. The water-saturated P-wave velocities show a large porosity dependence, which results in a significant porosity dependence in the V_P - V_S ratio, especially at low pressures.

Introduction

The pressure dependence of compressional (P) and shear (S) wave velocities in unconsolidated sediments is an important consideration in a number of engineering applications. This pressure dependence is often used to correct or project velocities to depths or locations where measurements have not been made, especially for site amplification predictions or liquefaction susceptibility analyses (Youd and Idriss, 1996). It also allows velocity changes to be used to monitor pressure changes in weak reservoirs or aquifers. The hazards posed to offshore drilling by unknown overpressures at shallow depths have also prompted attempts to use the seismic velocities or V_P - V_S ratio to quantitatively predict in situ effective pressures.

Because of the frequent use of the dynamic shear modulus in geotechnical applications, a great deal of research has been conducted on the pressure dependence of the shear modulus and of the shear wave velocity in soils. Hardin and Blandford (1989) developed a semi-empirical expression for the small-strain shear modulus. The form for an isotropic stress state is as follows:

$$G_{ij}^e = \frac{OCR^k}{F(e)} \frac{S_{ij}}{2(1+\nu)} p_a^{1-n} \sigma_{av}'^n \quad (1)$$

where G_{ij}^e is the shear modulus in the plane of propagation, which experiences the mean stress σ_{av}' in all directions, p_a is the atmospheric pressure, and ν is the Poisson's ratio of the grain material. Equation (1) includes two free parameters: S_{ij} , a multiplier to account for textural factors and structural anisotropy, and n , which dictates the pressure dependence of the modulus. The void ratio function, $F(e)=0.3 + 0.7e^2$, is meant to account for the effect of porosities differences, whether the result of textural differences between samples or of the compaction of a given sample. The OCR^k term corrects the pressure dependence for the effects of compaction or preconsolidation of the sample, where OCR is the

overconsolidation ratio, and k is a function of the plasticity index usually assumed to be zero for sands. Since the overconsolidation ratio is defined as the preconsolidation pressure the current pressure, the pressure dependence of the modulus for unloading or reloading paths is simply the effective pressure, p , to the $n-k$. A large body of work has demonstrated that the value of n for sands is generally near 0.5 (Hardin, 1980; Yu and Richart, 1984; Hryciw and Thomann, 1993). Most of this work has been conducted at pressures below a few hundred kPa. Hryciw and Thomann (1993) measured the pressure dependence of a number of texturally different sands at pressures up to 300 kPa and found n to vary between values of 0.39 and 0.72, and to correlate to the compressibility of the sand. Their measurements showed $S_{ij}/[2(1+\nu)]$ to vary from 478 to 734, and to be inversely related to n . They also recognized that k can be greater than zero for loose, compressible sands.

For offshore applications compressional waves are more commonly measured than shear waves. In these applications pressures are generally assessed by developing a normal P-wave interval travel-time vs. depth trend. This trend is usually assumed to be linear when depth is plotted against the log of the interval travel-time. Deviations above this trend are assumed to be the result of high pore pressures and low effective pressures (Hottman and Johnson, 1965; Pennebaker, 1970; Pilkington, 1988).

Hardin and Blandford (1989) also propose a semi-empirical relationship for the constrained (P-wave) modulus, M_i , similar in form to that of the shear modulus:

$$M_i = \frac{OCR^k}{F(e)} \frac{S_i(1-\nu)}{(1+\nu)(1-2\nu)} p_a^{1-n} \sigma_i^n \quad (2)$$

where σ_i^n is the stress in the direction of wave propagation, and all the other symbols are the same as for equation (1). Little experimental data has been collected on unconsolidated sediments to constrain the free parameters for the P-wave modulus.

Recently Huffman and Castagna (2001) and Prasad (2002) have demonstrated the potential of the V_P - V_S ratio as an indicator of pressure, as extracted from AVO analysis or multi-component reflection data. The V_P - V_S ratio is especially sensitive to pressure for water saturated sediments. At low pressures the shear velocity drops to near zero. The water-saturated compressional wave velocity is limited by the compressional resistance of water to values above 1500 m/s, and so the V_P - V_S ratio rises dramatically at low pressures. The appeal to using the V_P - V_S ratio as a pressure indicator is that it can be easily extracted from AVO data, whereas pure shear wave velocities are difficult to measure in situ in offshore environments. It is also thought that this parameter might reduce the ambiguity in the compressional wave velocity-pressure trend that results from the presence of gas in the sediment, and at the same time might reduce the effect of textural effects inherent in both of the velocities.

A number of theoretical models (Digby, 1981; Walton, 1987; Goddard, 1990) have been proposed to describe the elastic moduli of granular materials. These models generally assume that the material is made up of an assemblage of perfect spheres, with the stiffness of the contacts described by Hertz and Mindlin solutions (Mindlin, 1949) to the displacements of two spheres subject to normal and shear forces. The behavior at the contacts is then used to predict the shear or compressional moduli either by assuming a regular packing (Santamarina and Cascante, 1996) or by assuming a random arrangement of contacts (Walton, 1987). In either case these models predict both the bulk and shear moduli to have a pressure dependence of $p^{1/3}$. If the contacts are assumed to be a cone in contact with a plane, or if the number of inter-particle contacts are allowed to vary by buckling of the particle chains, the pressure dependences are predicted to be $p^{1/2}$ at low pressures, transitioning to $p^{1/3}$ at high pressures (Goddard, 1990).

We have run a series of experiments on sand and glass bead samples designed to measure the effects of porosity changes due to sorting and compaction on the velocities and on the velocity-pressure relationship. We measured the P and S-wave velocities at pressures from below 100 kPa up to 20 MPa in order to test the pressure dependence over this entire pressure range. In this paper we report our results

and discuss the porosity-velocity and pressure-velocity trends observed. We also discuss the implications of our results for quantitative pressure prediction over the range of pressures measured.

Figure 1: The experimental apparatus. A) The pressure vessel, about 80 cm long and 10 cm in diameter, with the sample holder. B) The sample holder.

Experimental Apparatus and Samples

The experimental apparatus consists of a sample holder that is inserted into a hydrostatic pressure vessel (see Fig. 1). The two end caps of the sample holder contain both P and S-wave transducers made with 200 kHz piezoelectric (PZT) crystals. In order to improve the impedance matching between the transducers and the sample, the faces of the transducers are made of 30% glass filled polycarbonate (shatter proof glass). Velocities were calculated by picking first arrivals from pulse-transmission signals. With this arrangement we were able to get interpretable shear signals at pressures as low as 50 kPa, with errors of approximately 2% for the P-wave velocities and 4% for the S-wave velocities.

The data presented here are from samples of a fine grained, well sorted, quartz sand, called the Santa Cruz Aggregate, as well as from synthetic samples made from sieved fractions of this sand and of glass beads. Four samples of the Santa Cruz Aggregate were run, two dry (Sa Dry 1, Sa Dry 2) and two water-saturated (Sa Wet 3, Sa Wet 4). Two other samples were made of sieved fractions of this sand, both of which were run dry. One sample (Sa Big) was made of 0.295-0.350 mm grain sizes. A second sample (Sa 35% Small) was made up of 65% by mass of the 0.295-0.350 mm size fraction, and the other 35% of 0.053-0.088 mm size grains. A total of seven glass bead samples were run, all dry. Three samples (GB Big, GB Small, and GB Tiny) consisted of different narrow size ranges of beads. Three samples (GB 35% Small, GB 35% Tiny 1 and 2) were made with a “bimodal” mixture of grain sizes, with 35% of the mass made up of smaller grains and 65% of larger grains. Finally, one sample (GB Broad) was made up of a broad range of particle sizes. Table 1 summarizes the grain sizes used and the initial porosities of all the samples.

Special attention was given to preparing the samples in such a way as to insure complete mixing of the different grain sizes, and to maintain full saturation of the water-saturated samples. However, the sample preparation varied based on whether the sample was a single grain size and dry (air pluviated) or water-saturated (water pluviated), or was a mixture of grain sizes (some samples mixed dry, some mixed

while moist and allowed to dry once in the sample holder). These differences in sample preparation produced variations in the packing of the grains and led to some scatter in the data.

The initial porosity of the samples was calculated from the grain density, dry sample mass, and sample volume. The sample dimensions were determined initially by measuring the diameter of the jacket around the sample and the distance between the two end caps once the sample had been prepared in the holder. The samples, 3.8 cm (1.5 in.) in diameter, were generally prepared to be about 3 cm long. The changes in the sample volume and porosity with pressure were then monitored by measuring changes in the length and circumference of the samples. An error analysis of our porosity measurements estimates the error at 0.012 to 0.016, or 3 to 4% of the total porosity.

The pressure paths followed for the different samples generally included a number of pressure cycles of increasing peak pressure for each cycle. The velocities and porosity were measured at the same set of pressures during each cycle (e.g. 0.1, 0.2, 0.5... MPa). This allowed us to compare the velocities and porosities measured at the same pressure for a sample that had been preconsolidated to a range of higher pressures. Five of the samples were cycled through 8 or 9 cycles, while the rest were cycled through between 1 and 5 cycles.

Pressure Trends

The shear and compressional velocity results from all of our samples are shown plotted against effective pressure in a log-log plot in Figure 2. Also shown are other measurements made on clean sands over similar pressure ranges from Domenico (1977), Prasad and Meissner (1992), Yin (1992), Estes et al. (1994), and Robertson et al. (1995). Figure 2 also includes lines showing pressure dependences of $p^{1/6}$, $p^{1/4}$, and $p^{1/3}$. This figure demonstrates that the measurements are fairly consistent, and are comparable to other data collected on loose sands over the same pressure range. It also illustrates that in general the pressure dependence of these samples tends to remain close to $p^{1/4}$ for the shear wave velocities, but with a slightly lower pressure dependence for the P-wave velocities.

Figure 3 shows the data from sample Sa 35% Small: plotted against effective pressure in (A) and (B), and plotted against porosity and color coded according to the measurement pressure in (C) and (D). These figures demonstrate that there is a slight preconsolidation effect to the samples, especially visible in (C) and (D) where the velocities measured at the same pressure rise as the porosity decreases with compaction from higher preconsolidation pressures.

We treated n , k , and $S/F(\nu)$ (lumping the Poisson's ratio terms, $F(\nu)=2(1+\nu)$ for the S-waves and $F(\nu)=(1+\nu)(1-2\nu)/(1-\nu)$ for the P-waves, in with S) as free parameters and fit equations (1) and (2) to the moduli calculated from our velocity measurements. We show the empirical fit for these two equations to the data from sample Sa 35% Small in Figure 3. The coefficients for the fits for each of the samples are given in Table 1, while the coefficients for the fits to the dataset as a whole are given in Table 2. The coefficients are plotted against the initial porosity of the samples in Figure 4. For the shear modulus we found the value of n to vary from between 0.352 to 0.612, with most of the values in the range from 0.45 to 0.5. This implies that the pressure dependence of the S-wave velocity would range from p to the 0.176 and the 0.306. These values are within the range of the results of Hryciw and Thomann (1993). The value of k for the shear modulus varied between -0.073 and 0.101 and averaged -0.003 (effectively zero), generally slightly lower than most of the values measured by Hryciw and Thomann (1993). These values of k are still relatively small compared to the values predicted for clays, which can be as large as 0.5 for high plasticity clays (PI=100) and would be about 0.1 for a plasticity index of 10 (Hardin and Drnevich, 1972). Some of the samples, especially a couple of the glass bead samples, actually demonstrates a velocity drop with consolidation, indicated by a negative k value.

Figure 2: P and S-wave velocities plotted against effective pressure for all of our sand and glass bead samples, and for similar measurements on clean sands from Prasad and Meissner (1992) (water sat.-▼,▲), Domenico (1977) (dry-●, water sat.-◆), Yin (1992) (dry-◄, water sat.-►), Estes et al. (1994) (dry-■, water sat.-★), and Robertson et al. (1995) (water sat.-+,x).

Figure 3: The velocity results from sample Sa 35% Small: A) V_P vs. pressure, B) V_S vs. pressure, C) V_P vs. porosity and color-coded by the pressure, and D) V_S vs. porosity. The black lines in parts (A) and (B) and the black dots in parts (C) and (D) represent fits to the Hardin and Blandford (1989) equations. The lines, the same as shown in Figure 5, are the Reuss average and Hardin and Blandford predictions for the velocity-porosity trend based on the moduli of the high porosity end-member (Sa Dry 2).

Figure 4: The values of: A) n , k and B) $S/F(v)$ plotted against the initial porosity for all of the dry samples.

The pressure dependence of the P-wave modulus is consistently lower than that of the shear modulus, with n ranging from 0.304 to 0.478 and averaging 0.07 less than the n for the shear modulus for the same sample. While the value of n was lower for the P-wave modulus of each sample, the value of k was almost always higher for the P-wave modulus than for the shear modulus, averaging 0.026 and ranging from -0.033 to 0.104 .

Figure 4 demonstrates that there is not a significant systematic initial porosity effect on the pressure dependence (n) or on the preconsolidation dependence (k). Thus the variation in initial porosity that comes from the differences in sorting does not significantly affect these dependences over the porosity range tested.

For one of the samples, GB Big, the pressure dependence of the P-wave modulus is actually below $p^{1/3}$, the dependence predicted by Hertz-Mindlin contact models. For all of the rest of the samples the pressure dependences of both moduli are higher. This might indicate that the pressure dependence is influenced by the grain angularity, as suggested by Goddard (1990), and grain size, though there are a number of other factors that we also did not systematically investigate that might influence the pressure dependence as well, including the packing, age, and mineralogy.

Sorting Effects

Figure 5 shows the velocity data from all of the samples plotted together against the porosity and color coded according to the pressure. The lower porosity data (below 0.35) come from the glass bead samples where two size fractions of beads were mixed, with the smaller grains partially filling the pore space between the larger grains. These plots illustrate the relative effects of the pressure and the sorting on the velocity and porosity. While for these coarse sediments, increased pressures result in large increases in the velocity, the porosity is only slightly reduced. On the contrary, the porosity reductions that result from a decrease in the sorting quality produce only minor increases in the velocity.

The lines shown in Figure 5 demonstrate the velocity effect predicted for the sorting-induced porosity changes in two ways: 1) by the Reuss average between the dry frame moduli and the solid quartz moduli (solid lines), and 2) by the Hardin and Blandford void ratio equation, $F(e)$ (dashed lines). The Reuss (isostress) average, the weighted harmonic average between two end member moduli, simulates the weakest possible way to combine two materials. Here the Reuss average is used to represent the theoretically minimal possible change in the velocities produced by adding solid grain material (quartz) to the high porosity sample (assuming constant packing) (Dvorkin and Nur, 1996). The high porosity end-member moduli come from the highest porosity data for each pressure, taken from the virgin compaction curve of sample Sa Dry 2. The Reuss average was calculated as follows:

$$M = \frac{1 - \frac{\phi}{\phi_{df}}}{M_{qtz}} + \frac{\frac{\phi}{\phi_{df}}}{M_{df}}$$

where M is the resulting average modulus, M_{df} is the modulus of the dry frame at the pressure of interest, and M_{qtz} is the modulus of pure quartz. The fraction of dry frame is given by $\frac{\phi}{\phi_{df}}$, where ϕ is the current porosity, and ϕ_{df} is the initial dry frame porosity at that pressure from sample Sa Dry 2. The fraction of quartz is then just $1 - \frac{\phi}{\phi_{df}}$. The Hardin and Blandford porosity effect on the moduli was calculated as follows:

$$M = \frac{F(e_{df})}{F(e)} M_{df} = \frac{0.3 + 0.7e_{df}^2}{0.3 + 0.7e^2} M_{df}$$

where M is the resulting porosity-corrected modulus, M_{df} is the modulus of the high porosity dry frame at the pressure of interest (calculated from the velocities and density of sample Sa Dry 2 at that pressure), e_{df} is the void ratio of the dry frame, and e is the current void ratio. Either relationship predicts the sorting effect relatively well, given the scatter in the data, though the Reuss average is likely to fit lower porosity data better than Hardin and Blandford's expression. The fact that the trend is so well described by the Reuss average implies that the small grains added tend to sit more or less passively in the pores, reducing the porosity but not adding significantly to the stiffness of the sediment.

Figure 6 shows the results of Gassmann fluid substitution with water of the velocities of all the dry samples, plotted against pressure and color-coded for porosity. As seen in this figure, the water saturated P-wave velocities have a significant porosity dependence, with lower porosities correlating to higher P-wave velocities. The presence of water in the pores stiffens a low porosity sediment more than a high porosity one, though the trend is still described reasonably well by the Reuss average between the moduli from the water-saturated, high porosity end-member and solid quartz. The shear wave velocities, which only require a density substitution, show a slight porosity dependence at the higher pressures, with the low porosity data resulting in higher velocities. At lower pressures, however, the porosity effect on the S-wave velocities is smaller than the scatter in the data.

Figure 5: The velocity data from all of the samples plotted against porosity and color-coded according to the pressure at the time of measurement: A) V_p B) V_s . The solid lines represent the Reuss average between the moduli of the high-porosity dry frame end-member and solid quartz. The dashed lines represent the porosity correction to the high porosity moduli proposed by Hardin and Blandford (1989).

Figure 6: The velocity results after Gassmann fluid substitution with water: A) the water saturated V_P results plotted against porosity, again showing the Reuss averages between the high porosity end member and solid quartz. B) Both velocities plotted against pressure and color coded for porosity.

Implications for Overpressure Prediction

The V_P - V_S ratios for all of the Gassmann fluid-substituted data are shown in Figure 7, plotted against pressure and color-coded by porosity. As discussed by Prasad (2002) and Huffman and Castagna (2001), the water saturated V_P - V_S ratio shows a dramatic rise as the effective pressure decreases. The V_P - V_S ratio increases from below 3 at 10 MPa to a mean value of about 7 at 0.5 MPa. Unfortunately there is a considerable amount of scatter in the V_P - V_S ratio, especially at low pressures, that would generate a significant amount of uncertainty in the in situ effective pressure determined from V_P - V_S ratio measurements. For example, if the measured V_P - V_S ratio is 5, the in situ pressure (based on our fluid-substituted data) could vary from between 0.2 and 2 MPa.

Figure 7(B) demonstrates that there is a porosity dependence to much of this scatter in the V_P - V_S ratio at low pressures. Since these non-cohesive sediments demonstrate only relatively small changes in both the velocity and porosity with compaction, preconsolidation has only minor influences on the V_P - V_S ratio. This implies that for water-saturated sands there should be only very little difference between the V_P - V_S ratio signatures of overpressures produced through undercompaction or through repressurization mechanisms such as fluid expansion. As preconsolidation can cause a much larger reduction in the porosity of clay-rich sediments (Bowers, 1995; Sayers et al., 2001), this effect could potentially be more important for cohesive sediments.

The porosity variation between the samples comes primarily from the differences in the initial porosities of the samples due to differences in their sorting. The scatter in the V_P - V_S ratio is therefore a product of the different effects that the porosity has on the water-saturated P- and S-wave velocities, as demonstrated in Figures 5 and 6 and discussed above. As the relative porosity effect at high pressures is similar for the P and S-wave velocities, there is relatively little scatter in the V_P - V_S ratio above about 10 MPa. The small porosity-dependence of the S-wave velocities at lower pressures allows the porosity dependence of the water-saturated P-wave velocity to show through. The implication is that for robust pressure estimates to be made from measured V_P - V_S ratios it will be necessary to correct for this porosity effect at pressures below about 10 MPa. Such a correction could potentially be developed through a suite of laboratory measurements on natural sands, and would need to be developed in order to permit the use of the V_P - V_S ratio as a quantitative pressure indicator.

Figure 7: The V_p - V_s ratio plotted against pressure and color coded according to the porosity: A) linear plot, and B) log-log plot.

Conclusions

We have measured the pressure dependence of the P and S-wave velocities through sand and glass bead samples with a variety of initial porosities produced by varying their sorting. We also ran the samples through a series of pressure cycles of increasing peak pressures in order to observe the effects of preconsolidation on the samples. We have observed the following results:

- 1) The pressure dependence of the S-wave velocities is around the fourth root of the effective pressure (p). The pressure dependence of the shear modulus ranges from p to the 0.352 to p to the 0.612. For the P-wave velocities the pressure dependence is slightly lower. The P-modulus dependence varies between p to the 0.304 and p to the 0.478, though generally still higher than the $p^{1/3}$ dependence predicted by Hertz-Mindlin contact models.
- 2) These non-cohesive sediments do exhibit a small preconsolidation effect on the velocities, though the effect is generally smaller for the S-wave velocities than for the P-wave velocities.
- 3) While poorer sorting can result in large decreases in the porosity, it has a relatively small effect on the S-wave velocities and dry P-wave velocities. The porosity reduction does produce a larger effect on the water-saturated P-wave velocities, as calculated from Gassmann fluid-substitution of our dry velocity results. This porosity dependence of the water-saturated P-wave velocities, and the limited porosity dependence of the S-wave velocities, results in a porosity dependence in the V_p - V_s ratio. Higher porosities correspond to a lower V_p - V_s ratio.
- 4) The sorting appears to have no significant effect on the pressure dependence of the velocities or on the effect of preconsolidation. The variation in the values of n and k for these non-cohesive sediments could be largely a result of differences in sample preparation that lead to inconsistency in the packing from sample to sample, or of other textural parameters that were not systematically investigated here.

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Table 1: Sample summary and fit parameters for individual samples

Sample Name:				Vs: $S/F(\nu)$				Vp: $S/F(\nu)$							
Initial Porosity	# of cycles			$S/F(\nu)$	n	k	R^2	$S/F(\nu)$	n	k	R^2				
Sands:	Sa Dry 1	0.414	5	clean, quartz sand, well sorted $D_{10}=0.16$ mm ¹ $D_{60}=0.3$ mm	610	0.459	0.101	0.9918	1897	0.439	0.104	0.9937			
	Sa Dry 2	0.432	1		601	0.490	-0.0202	0.9972	2352	0.430	0.0300	0.9918			
	Sa Wet 3	0.400	2												
	Sa Wet 4	0.411	4												
	Sa Big	0.409	9												
	Sa 35% Small	0.378	9												
	Glass Beads: GB Big	0.381	8		1	295-350 (1.5-1.75)	1075	0.352	-0.0413	0.9952	4088	0.304	-0.0045	0.9872	
		GB Small	0.411		8	1	74-88 (3.5-3.75)	671	0.448	-0.0037	0.9951	2634	0.378	0.0363	0.9841
		GB 35% Small	0.315		8	0.65	295-350 (1.5-1.75)	526	0.490	-0.0186	0.9973	2400	0.400	0.0284	0.9926
		GB Tiny	0.421		3	1	37-44 (4.5-4.75)	704	0.448	0.0098	0.9926	2798	0.375	0.0437	0.9825
GB 35% Tiny 1		0.296	3	0.65	295-350 (1.5-1.75)	564	0.503	-0.0346	0.9974	2655	0.414	-0.0091	0.9948		
GB 35% Tiny 2	0.258	3	0.65	295-350 (1.5-1.75)	287	0.612	0.0050	0.9969	1802	0.468	-0.0333	0.9934			
GB Broad	0.338	3		37-710 (0.5-4.75)	393	0.494	-0.0729	0.9940	1719	0.420	-0.0273	0.9940			

Table 2: Fit parameters to whole dataset

	$S/F(\nu)$	n	k	R^2
V_S	604	0.474	0.0008	0.9431
V_P (dry)	2336	0.411	0.0318	0.9541