ELASTIC AND VISCO-ELASTIC LABORATORY PROPERTIES IN CARBONATES

by Ludmila Adam A thesis submitted to the Faculty and the Board of Trustees of the Colorado School of Mines in partial fulfillment of the requirements for the degree of Doctor of Philosophy (Geophysics).

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ABSTRACT

This thesis is a study of the elastic and visco-elastic properties of carbonate samples at seismic and ultrasonic frequencies for varying fluid saturation and pressure. We explore in detail some of the assumptions of Gassmann's fluid substitution relation, especially rock frame sensitivity (shear modulus change) to fluid type and bulk modulus frequency dependence. We find that for our carbonate rock samples, at high differential pressures and seismic frequencies, the saturated bulk modulus of rocks with high aspect ratio pores is predicted by Gassmann's relation. Ultrasonic velocities should be used with caution when predicting seismic time-lapse data, because laboratory data show that velocities are dispersive from 10 to 10^6 Hz. Nevertheless, the measured dispersion seems proportional such that there is good agreement between laboratory seismic and ultrasonic elastic properties, for example, in the V_P/V_S ratio.

Sample heterogeneity is observed in carbonates, which could bias the measured velocity. Therefore, we developed and tested a fiber-optic strainmeter to estimate bulk velocity and attenuation at seismic frequencies with greater sensitive than the conventional strain gages. However, measuring samples is challenging because the new strainmeter's strength (high sensitivity) makes it also vulnerable to ambient noise.

Intrinsic attenuation at low and high frequencies is also measured in our carbonate suite. Contrary to most observations in sandstones, we observe that the bulk compressibility losses dominate over the shear-wave losses for dry samples and samples fully-saturated with liquid-butane or brine. Attenuation modeled from the measured modulus corroborates this conclusion. On average, P-wave attenuation can increase by almost a factor of three when brine substitutes for a light hydrocarbon in these carbonate rocks.

Finally, the scenario of brine substituting a light hydrocarbon in a reservoir is analyzed with synthetic seismograms to study the feasibility of estimating time-lapse attenuation. For when fluid substitution occurs, we propose that time-lapse apparent attenuation is dominated by the change in intrinsic attenuation rather than scattering attenuation. Although apparent attenuation at a specific time is sensitive to receiver separation and random noise, we are still able to predict time-lapse intrinsic attenuation for these synthetic traces.

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CHAPTER 1

INTRODUCTION

Carbonates are important targets for rock property research, because they currently present over half of the major oil and gas reservoirs in the world. However, rock physics measurements and theories related to changing reservoir conditions, mostly relate to sandstones.

Generally, when the initial oil production of a reservoir starts to decline there is a program in place to enhance the productivity by injecting a fluid in the reservoir (e.g. water, CO_2). This recovery process means the original fluid in the rock is replaced by another with different physical properties. In exploration geophysics we study the seismic reservoir response to infer these fluid or gas properties in the pore space. In rock physics, several theories and mechanisms describe the rock-fluid interactions that influence the seismic signature. However, compared to sandstones, carbonate rocks have few laboratory measurements and valid models to describe them. The complexity and variety of carbonate rock textures is the main reason that these rocks have not been intensely studied from the rock physics perspective. Figure 1.1 shows the classification of carbonate porosity by Choquette and Pray (1970). Primary fabric in the rock is created at different energy environments in the sedimentary carbonate platform; while the secondary fabric results from following dissolution and recrystallization processes (i.e. diagenesis) of the original rocks. Therefore, measuring, interpreting and generalizing elastic carbonate rock properties is challenging and not well understood in some areas of rock physics. To put this in perspective, for example, in a reservoir we can have a carbonate rock with a frame composed of deposited particles (interparticle porosity in Figure 1.1, pink is pore space) and the same rock but that has been altered during burial for which the original particles have been dissolved: what used to be the particle frame is now the pore space (moldic porosity in Figure 1.1, black is pore space), while the original pores recrystallize into the rock frame. The elastic properties of these two rocks, originally the same, can be very different and thus the need to study carbonate geophysical properties in controlled laboratory settings.

The linear response of the strain (i.e. deformation) to an applied stress (i.e. pressure) with no chronological memory is called linear elasticity. Rock properties are linearly elastic when the strain response to an applied stress is small ($< 10^{-6}$). A propagating seismic wave produces a stress variation small enough that rocks behave linearly elastic. Nonetheless, rocks can have a strain memory or time dependent relaxation. This irreversible behaviour, known as linear visco-elasticity, means there is energy dissipation when the rock is deformed. A rock behaves as a visco-elastic material if the modulus or velocity are complex. The real



Figure 1.1. Carbonate porosity classification by Choquette and Pray (1970) showing the complexity of carbonate textures. Primary fabric is created from deposition, while secondary fabric results from diagenetic processes. (Modified from Moore (2001))

part of this modulus describes the elastic behaviour, while the imaginary component controls visco-elasticity. Rock physics models are used to predict changes in the moduli under varying fluid and pressure conditions, among other reservoir parameters. The increase or decrease of the elastic bulk modulus due to fluid change in the pore space is routinely studied by applying Gassmann's equation; while the commonly applied theories for saturated rocks behaving as visco-elastic materials are Global or Squirt fluid flow mechanisms. Rock physics models also exist to describe anisotropic velocity in rocks, but in this thesis we assume that the studied rocks are isotropic.

Controlled laboratory experiments on rock samples are used to test rock-physics mechanisms and theories. Most rock properties measured in the laboratory are acquired at KHz or MHz frequencies, significantly higher than seismic frequencies. The Rock Physics Lab at Colorado School of Mines has a unique system capable of measuring elastic and viscoelastic rock moduli in the seismic frequency range (3-1000 Hz), as well as the more common ultrasonic frequency. A significant part of this thesis focuses on studying the carbonate rock moduli at different frequencies, and the impact of this frequency dependence when experimental data is included in the large picture of reservoir characterization. Gassmann's relation is customarily used in industry and academia to model the bulk modulus change with fluid from seismic, log (KHz) or laboratory data (KHz and MHz). However, this relation is derived for low frequency; zero frequency to be specific. chapter 2 analyzes the applicability of Gassmann's equation to this particular set of carbonate samples. This research is performed with an emphasis in understanding the role of frequency, and the validity of the assumptions in Gassmann in the context of fluid and pressure. Chapter 3 studies the experimental visco-elastic moduli for fully-saturated carbonates. If the rock modulus is frequency dependent, then there are intrinsic amplitude losses (attenuation) for that particular modulus. This chapter analyzes the different types of intrinsic losses in the rock which can be used to interpret the saturation state of a rock.

As previously mentioned, in the laboratory we could record a different frequency response of the modulus (frequency scaling). However, there is also the issue of heterogeneity and sample size scale. In chapter 4 we briefly analyze how these scaling factors influence the experimentally estimated elastic parameters for this carbonate set. The measured seismic rock moduli can be biased by sample heterogeneity, such as large vugs (> 5 mm) in carbonate samples. In chapter 5 we develop and test a fiber-optic strainmeter system that would potentially average such heterogeneity at the core scale. The strainmeter is more sensitive that the current seismic frequency sensors when estimating the elastic and visco-elastic properties of rocks.

Routinely, laboratory data is used to forward model logs or the seismic response to study the feasibility of monitoring a dynamic reservoir with geophysical methods (e.g. production, enhanced oil recovery, CO_2 sequestration). Over a period of time (time-lapse) fluids in the reservoir change, varying the elastic and visco-elastic moduli in the rock. An unexplored topic in reservoir characterization is time-lapse attenuation. In seismic data or full-waveform sonic logs the preferential loss of amplitude with frequencies is referred to as apparent or effective attenuation. This attenuation is the combination of intrinsic and scattering loss. In chapter 6 we explore the feasibility of obtaining information about the fluid change in the reservoir by studying the time-lapse change in apparent attenuation from synthetic zero-offset VSP and surface seismic. The main idea is to analyze ways to cancel the contribution of scattering losses on the apparent attenuation by studying timelapse data. By doing so, time-lapse changes in apparent attenuation are only related to the intrinsic losses of the rock: the fluid signature. Here the thesis comes full circle, because the experimental elastic and visco-elastic moduli are used as the model parameters to generate synthetic seismic data. For that we assume that the core-scale properties measured in the laboratory represent the elastic and visco-elastic modulus and velocity in the in-situ reservoir rocks.

I refer to work presented in this thesis in first person plural (*we*) because each chapter is the result of discussions and collaborations by several people involved in this research. The outcome of the chapters are papers or abstracts with the following co-authors: chapter 2 is published in Geophysics, co-authored by Michael Batzle (Colorado School of Mines) and Ivar Brevik (StatoilHydro), chapter 3 is accepted for publication in the Journal of Geophysical Research, Solid Earth, co-authored by Michael Batzle, Kyle Lewallen (ExxonMobil URC) and Kasper van Wijk (Boise State University), chapter 4 is published in The Leading Edge, co-authored by Michael Batzle (Colorado School of Mines), chapter 5 is published as an SEG Expanded Abstract, co-authored by John Scales and Michael Batzle (Colorado School of Mines), and Tim Niebauer (Micro-g LaCoste). And chapter 6 will be submitted to 4

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CHAPTER 2

GASSMANN FLUID SUBSTITUTION AND SHEAR MODULUS VARIABILITY IN CARBONATES AT LABORATORY SEISMIC AND ULTRASONIC FREQUENCIES

2.1 Introduction

An important area of research for carbonate rocks is the fluid substitution effect on elastic moduli and velocities. One of the widely used relations to estimate the effect of fluids on bulk modulus is Gassmann's fluid substitution theory (Gassmann, 1951), which we will examine in the following section. Laboratory measurements on carbonates have been performed at ultrasonic frequencies (~ 0.8 MHz) to estimate the validity of Gassmann's equations for limestones and dolomites (Wang, 2000; Wang et al., 1991; Marion and Jizba, 1997; Baechle et al., 2005; Røgen et al., 2005; Capello and Batzle, 1997; Capello, 1995). In most cases Gassmann's predictions under-estimate the observed ultrasonic velocities for either oil or brine saturated samples, although for some samples Gassmann theory overestimates the measured velocities (Wang, 2000; Baechle et al., 2005; Røgen et al., 2005).

Presently, the applicability of Gassmann's equation to carbonate rocks is unresolved. With our work we hope to make inferences about the uncertainties and interpretation on the applicability of Gassmann's equation. Our work focuses on understanding Gassmann's fluid substitution theory at seismic and ultrasonic frequencies. We also analyze the validity of some of the assumptions for Gassmann's theory, especially rock frame sensitivity to fluids. Our carbonate samples consist of different fabrics, mineralogies, porosities, and permeabilities, still we must be careful in generalizing our results to all carbonate reservoirs.

First, we present Gassmann's theory and its assumptions. Second, we describe the laboratory acquisition, processing, and data uncertainty analysis at seismic and ultrasonic frequency. Then, we introduce shear modulus variability with fluid substitution and the possible mechanisms that could explain these changes. And finally we compare our measured bulk modulus to Gassmann's predictions for these carbonate rocks.

2.2 Gassmann's equation

Gassmann's fluid substitution relation is commonly applied to predict the bulk modulus for the saturated rock with different fluids:

$$K_{sat} = K_{dry} + \frac{\left(1 - \frac{K_{dry}}{K_{min}}\right)^2}{\frac{\phi}{K_{fl}} + \frac{1 - \phi}{K_{min}} - \frac{K_{dry}}{K_{min}^2}}.$$
(2.1)

Gassmann's equation (2.1) estimates the saturated bulk modulus (K_{sat}) through the bulk modulus of the forming minerals (K_{min}) , the bulk modulus of the frame or dry rock (K_{dry}) , the bulk modulus of the fluid (K_{fl}) and the rock porosity (ϕ) (Gassmann, 1951). Note that in Gassmann's relation the considered property of the fluid in the rock is only the fluid bulk modulus.

Gassmann's derivation is based on the following assumptions for a porous system: 1) Pore pressure is in equilibrium between pores. This can be achieved at very low frequencies, usually at seismic frequencies or lower, where the fluid has enough time to reach relaxation or equilibrium. However, the relaxation time depends also on fluid viscosity and density, and rock permeability. 2) The porous frame consists of a single solid material (monomineralic). 3) Pores are in flow communication and homogeneously fully-filled with a non-viscous fluid. 4) The system is closed (undrained). 5) The pore fluid does not chemically influence the solid frame. Although implied, a constant rock shear modulus from dry to any fluid type saturation is not an assumption but an outcome of Gassmann's theory (Berryman, 1999).

The beauty of Equation (2.1) is its simplicity as well as the fact that variables have physical significance and are usually well constrained or can be directly measured. Other fluid substitution theories require the knowledge of such factors as symmetry of the rock, the geometry of the inclusions and the crack density among others. For example, in the low frequency limit, where no pore pressure gradients exist, Brown and Korringa (1975) relate the anisotropic rock effective elastic compliance tensor to the same rock filled with fluid, and for an isotropic and monomineralic rock their relations reduce to Gassmann's equation. For this fluid substitution theory knowledge of the anisotropic symmetry and pore space compressibility are required. Other fluid substitution theories mostly assume isolated inclusions and their geometries in the derivation of the equations. Isolated cavities should then also be isolated with respect to fluid flow (presence of pore pressure gradients). Therefore, theories that assume isolated inclusions (Kuster and Toksoz, 1974; O'Connell and Budiansky, 1974; Hudson, 1981) may be more applicable to the high frequency range and require knowledge of parameters related to pore space.

2.3 Carbonate samples

Our carbonates are from two wells in a single reservoir with depths between 2915 and 3180 m below sea level. The reservoir has lagoon, ramp and shoal depositional environments. These different depositional systems create different textures, porosities and permeabilities (Figure 2.1). Some reservoir regions have been dolomitized. Dolomitization is evident from high porosity and high permeability, because dissolved grains or fossils become pore space,



Figure 2.1. Thin sections for some of our carbonate samples. Pink represents calcite, gray dolomite, white anhydrite and blue pore space.

increasing the connectivity between pores, thus increasing permeability. The reservoir is not fractured and has little clay minerals, but does have minor anhydrite (Brevik, 2004). The available samples are nine carbonates with varying porosity (5-35%), permeability (0.001-800 mD), mineralogy (dolomite and limestone), and texture. The samples are either almost pure calcite or dolomite (95% total volume) with less than 3% clays and 5% anhydrite of total volume. Samples with large anisotropy or vuggy pores are avoided, because any heterogeneity of the size of the strain-gage (~ 0.5 cm) can bias the elastic modulus of the rock. Table 2.1 summarizes the petrological data for our samples. Porosity and permeability are measured using standard helium porosimetry and air permeability equipment at atmospheric pressure. Permeability values are corrected for Klinkenberg gas slippage. The samples are cylindrical, 3.75 cm in diameter and 3.75 to 5 cm in length.

Velocity and elastic modulus data are acquired at nine pressure points. Confining pressure varies from 3.5 to 34.5 MPa while pore pressure is held constant at 3.5 MPa, thus reaching a maximum differential pressure of 31 MPa. The low frequency system in the laboratory is pressurized with nitrogen gas, but for safety reasons the system is not able to reach the reservoir differential pressure (34.5 MPa). Samples are measured dry, under butane (C₄H₁₀), and brine (200,000 ppm NaCl) saturations. Butane at 3.5 MPa is in liquid state. Samples are measured with some amount of moisture because even less than 1% of water can reduce the bulk and shear moduli significantly (Clark et al., 1980). Because samples show sensitivity to water, several are kept in a high-humidity chamber to provide an initial brine saturation (less than 1%). Samples A, C, E, F and G are humidified previous

SAMPLES	А	В	С	D	Е	F	G	Н	Ι
ϕ (%)	1.6	4.6	21.0	24.9	28.5	34	23.6	29.6	34.7
$\kappa (mD)$	0.03	0.03	5.50	1.20	0.43	0.31	25.00	103.00	432.00
$D_g (gm/cm^3)$	2.73	2.84	2.70	2.71	2.70	2.69	2.84	2.80	2.86
Calcite(%)	83.0	0.7	76.0	99.6	-	97.0	-	-	0.4
Dolomite(%)	11.0	97.0	21.0	0.0	-	0.0	-	-	93.0
Anhydrite (%)	0.5	0.5	0.0	0.0	-	0.7	-	-	4.9
Phy (%)	3.4	0.8	2.4	0.0	-	2.3	-	-	1.1
Quartz(%)	0.6	0.6	1.2	0.4	-	0.2	-	-	0.8
Feldspar (%)	2.0	0.0	0.0	0.0	-	0.0	-	-	0.0
K_{min} (GPa)	70.70	78.96	71.59	71.26	71.59	70.35	85.00	78.96	77.67
Texture	wacke	mud	grain	grain	grain	bound	pack	wacke	mud

Table 2.1. Petrological data for the carbonate set. Mineralogy was obtained from XRD analysis and are reported in percent per volume (samples E, G and H had no XRD analysis). ϕ is porosity and κ permeability. D_g stands for grain density and K is bulk modulus. Mineral bulk modulus is computed using Voigt-Reuss-Hill average. Texture follows modified Dunham's carbonate classification (Moore, 2001): mud = mudstone, wacke = wackestone, pack = packstone, grain = grainstone and bound = boundstone. *Phy* is phyllosilicate.

to measurements, thus dry for these samples means humidified. Samples *B*, *D*, *H* and *I* are measured at room conditions (30 % humidity). Samples are coated with a thin impermeable polyimide film (kapton), over which strain gages are glued to measure rock deformations at seismic frequencies. This film keeps the moisture inside the rock and prevents nitrogen diffusion.

2.4 Data example: acquisition and processing

Samples are measured at low (seismic: 3-3000 Hz) and ultrasonic frequencies (~0.8 MHz), although sample G is measured at ultrasonic frequencies only. Seismic frequency moduli and velocities are derived from the strain-stress method (Spencer, 1981; Batzle et al., 2001). Appendix A describes the methodology to estimate modulus and velocity. As an example of the estimated bulk modulus over the entire frequency range, we show results for sample H in Figure 2.2. The computation of the error bars and the linear fit are also discussed in Appendix A.

Observe that the rock bulk modulus increases with saturating fluid. However, the change in rock bulk modulus from dry to butane saturated is small compared to when the rock is saturated with brine. This is because butane has a lower fluid bulk modulus than brine. Figure 2.2 also shows bulk modulus dispersion (higher frequencies have larger modulus). Several theories exist to explain the nature of this dispersion. A primary cause for dispersion can be pore pressure disequilibrium caused by non-zero pore pressure gradients.



Figure 2.2. Seismic and ultrasonic frequency bulk modulus least-squares estimates (solid lines) and measured data for sample H at 31 MPa. Observe the modulus dispersion for different fluid. Error bars are two standard deviations of the estimated bulk modulus.

This unrelaxed pressure is described by several mechanisms: grain-fluid inertial and viscous coupling (Biot, 1956a), patchy saturation (White, 1975; Dutta and Ode, 1979) and squirt or local fluid flow (Mavko and Jizba, 1991), among others. Our goal here is not to decide which of frequency dependent modulus or velocity theories are causing the dispersion. We do want to point out differences in modulus estimates as a result of the dispersion from seismic to ultrasonic frequencies. As previously mentioned, Gassmann's theory is the low frequency limit, meaning that this theory may not be suitable to predict ultrasonic data because of possible dispersion in the elastic moduli and velocities. Wang (1997), Marion and Jizba (1997), Baechle et al. (2005) and Røgen et al. (2005) have shown how in most cases Gassmann's theory under-predicts ultrasonic frequency measurements. Pore pressure can equilibrate if there is enough time for the fluids to relax. This means there is a characteristic frequency f_c of the rock perturbation. For measurements acquired at a frequency less than f_c the pore pressure has reached equilibrium, while for higher frequencies than the f_c , pore fluids are not equilibrated, producing higher values for modulus and velocity.

Differential pressure also controls the modulus dispersion of a rock. At low differential pressures where compliant pores or cracks are open, pore pressure disequilibrium is more likely to occur. Wang (2000) shows in a compilation of ultrasonic laboratory data of carbonate samples, that Gassmann's theory substantially (up to 30%) under-predicts the measured velocities at low differential pressures. At high differential pressures, compliant pores close, and Gassmann's theory predicts the measured data within 10%.

Carbonates are heterogeneous and vugs or moldic structures can have comparable length to the ultrasonic wavelength (0.5 cm for a wave at 0.8 MHz and with a velocity of 4500 m/s). Some of our samples showed inclusions of different densities or voids with dimensions on the order of ultrasonic wavelengths. Therefore scattering of ultrasonic waves is possible in carbonate samples, especially in dry rocks where the density contrast between voids and matrix is large. When scattered, the wave loses energy to multiple reflections from grains mostly resulting in lower moduli and velocities at higher frequencies. The larger modulus contrast will be for air-grain and butane-grain interfaces.

2.4.1 Poisson's ratio: a correction

Samples B, F and I show higher values of Poisson's ratio at low frequency than expected in carbonates. Rock heterogeneity is probably not the cause, since placing the strain gages on large heterogeneities (visible to the eye) on measured core plugs are avoided. The observed larger deformations of the sample in the horizontal direction probably result from end effects in our stress-strain system. This large deformation or bulging can result from the combination of intrinsically large Poisson's ratios in carbonates (> 0.25) and short samples (our sample length is close to its diameter). This bulging has been confirmed with preliminary finite element modeling at our laboratory. Poisson's ratio depends on the V_P/V_S , but because the dispersion in V_P and V_S are similar for our samples the resulting dispersion in Poisson's ratio is negligible, making it possible to correct the low frequency data with the estimates we obtain from ultrasonic data. Domenico (1984), Anselmetti and Eberli (1993), Mavko et al. (1998), Assefa et al. (2003), and Han (2004, Fluids and DHI Consortia Meeting Report) measured carbonate samples ultrasonically and derived empirical relations for V_P and V_S . We use their relations to compute Poisson's ratio for water/brine saturated carbonates and compare their values to our samples Poisson's ratio measured at ultrasonic frequencies (Figure 2.3). Agreement between modeled Poisson's ratio and our measurements lets us use the ultrasonic values to correct the Poisson's low frequency data. The correction consists of multiplying the seismic frequency Poisson's ratio by a factor less than one. This factor is obtained from the ratio of the ultrasonic and the biased seismic frequency Poisson's ratios.

2.5 Variations in shear modulus

Fluids have a shear modulus of zero, so we expect that the dry or fluid saturated rock shear modulus to be constant (true for many rocks that are isotropic and homogeneous). Together with the assumption in Gassmann's theory that pore fluids do not chemically alter the mechanical properties of a rock, Gassmann's theory predicts that the shear modulus remains constant under different saturations. Thus, a measure of the shear modulus is one way to validate Gassmann's theory.

However, our carbonate samples show rock shear modulus changes from dry to brine saturation of up to 20%. Several laboratory studies have also reported shear modulus changes between 5 % and 20 % from dry to water or brine saturation in carbonates (Vo-Thanh, 1995; Assefa et al., 2003; Baechle et al., 2005; Røgen et al., 2005; Sharma et al., 2006). The shear modulus of the rock is also sensitive to small amounts of moisture (Clark



Figure 2.3. Modeled Poisson's ratio from empirical relations from ultrasonic data for carbonate rocks saturated with water/brine. Squares and circles are the Poisson's ratio obtained from our measurements at ultrasonic frequencies. This plot shows that our values of Poisson's ratio are in agreement with the empirical equations. Therefore, we use the ultrasonic Poisson's value to correct the low frequency data for three of our samples (B, F and I)represented by circles.

et al., 1980). They show that this weakening of the matrix is related to the amount of surface area in rocks.

Rock weakening resulting from fluids has also been observed in field data. Water weakening the rock frame in carbonates is invoked as a primary factor controlling subsidence of the Ekofisk field. Sylte et al. (1999) show that compaction of Ekofisk chalks occurs only in chalks that are being water flooded. High porosity chalks that have original water content (pre-water flooding) are not compacting and behave elastically throughout the lifetime of the field. They conclude that the injected water weakens invaded chalks resulting in compaction and porosity loss. In their study they compare observation to geo-mechanical models, but do not give the physical-chemical mechanisms that could be producing this weakening.

Khazanehdari and Sothcott (2003) compile rock-fluid interactions that explain the rock shear modulus (μ) variability with fluids. They define rock weakening when $\mu_{saturated} < \mu_{dry}$, and strengthening for $\mu_{saturated} > \mu_{dry}$. Cardona et al. (2001); Cardona (2002), based on work from Brown and Korringa (1975), show that for transversely isotropic with a horizontal symmetry axis rocks, the vertically propagating shear waves are sensitive to the compressibility of the saturating fluid. However, our rocks are largely isotropic at the core scale, although they might be anisotropic at field scale. Therefore, in our work we will focus on the rock-fluid interactions that are responsible for rock shear modulus changes.



Figure 2.4. Sample C showing shear modulus weakening and strengthening at seismic and ultrasonic frequencies respectively. Measurements are performed from high to low differential pressures. Circles represent repeated differential pressures going from low to high differential pressures after the initial unloading cycle was finalized. Note that as we decrease the differential pressure more compliant pores and cracks open. Error bars are one standard deviation (one σ for seismic frequency data are contained in the size of the symbol).

2.5.1 Data examples of shear modulus sensitivity to fluids and possible explanations

Figure 2.4 shows the rock shear modulus for sample C at seismic and ultrasonic frequencies when dry and brine saturated. Error bars represent one standard deviation of the shear modulus. Two main observations are to be drawn from Figure 2.4. First, the rock shear modulus can either weaken or strengthen upon brine fluid saturation compared to the dry rock. At 100 Hz we observe shear modulus weakening from dry to wet, while for 0.8 MHz data the shear modulus strengthens when brine fills the pore space. This implies that more than one rock-fluid mechanism is active.

Second, for the 100 Hz frequency measurements, the shear modulus weakens more for low than for high differential pressures. Our measurements are performed going from high to low differential pressures (unloading cycle). After the experiment with brine saturation reached 3.5 MPa, we increased differential pressure again for three pressure stages (circles in Figure 2.4). Observe that the rock shear modulus sensitivity to brine saturation for both 100 Hz and 0.8 MHz is repeatable; thus, the shear modulus weakening is not affected by hysteresis. This reversible weakening or strengthening of the frame is likely associated with opening and closing of compliant pores or micro-cracks. Some of these cracks are intrinsic to the rock, while others might have been induced while drilling or coring. Other samples with significant shear modulus weakening show similar pressure dependence as sample C.

Figure 2.5 compares the dry and brine saturated rock shear modulus for all samples for

100 Hz at 3.5 and 31 MPa differential pressure. The solid line indicates equal dry and brine saturated shear modulus. Most samples have a rock shear modulus around 10 MPa. This cluster of data correspond to samples with high porosity (24-35%), while the low porosity samples have shear modulus larger than 15 MPa. The error bars of the shear modulus (one standard deviation) are within the size of the marker. Observe that at low differential pressures (3.5 MPa) all samples show shear modulus weakening, while at higher pressures (31 MPa) shear weakening is still present but less significantly than for low pressure (see also Figure 2.4).

Most samples at ultrasonic frequency and at both 3.5 and 31 MPa differential pressure show neither weakening or strengthening of the rock shear modulus within the data uncertainty (Figure 2.6). Weakening is observed in samples B and D, but less than for seismic frequency (Figure 2.5).

When we compare Figures 2.5 and 2.6, the shear modulus for brine saturated rock at ultrasonic frequency is greater than for seismic frequency. This strengthening can be due to modulus dispersion as a result, for example, of global and squirt fluid flow in the pore space. However, for samples B and D the chemical softening of the rock can be dominating over modulus dispersion. Alternatively, our ultrasonic wave velocity represent the fastest path (stiffest area in the rock). If the chemical weakening is occurring in an isolated area of the sample, the stress-strain experiment measures the effective rock deformation (frame softening), while the ultrasonic wave will avoid this area and propagate in the unperturbed rock.

We also saturated the carbonate rocks with butane, a highly compressible light hydrocarbon (in liquid phase at our elevated pore pressures). The sensitivity of the rock shear modulus to this fluid is much less than for brine (Figure 2.7).

We can now examine what are the possible weakening and strengthening mechanisms acting on our carbonate rocks based on work of Khazanehdari and Sothcott (2003). They compile several mechanisms that can cause the shear modulus to either weaken or strengthen when a fluid contacts the solid matrix.

Pores and micro-fractures create surface area in a rock. Surface energy reduction (Murphy et al., 1986; Tutuncu and Sharma, 1992) and sub-critical crack growth (Atkinson, 1984) mechanisms relate to the amount of surface area in a porous rock. Compliant pores and micro-fractures are observed in our samples from thin sections. We also know from modulus as a function of differential pressure that compliant pores and micro-fractures open, increasing the surface area, as the differential pressure decreases (Figure 2.4). For our samples open low aspect ratio pores might exhibit growth as well as breakage of solid bounds due to interaction with brine. These two mechanisms acting on our carbonate samples are consistent with the fact that a non-polar fluid, such as butane, saturating the rock does not show significant shear modulus variation (Figure 2.7). Another rock-fluid mechanism such as viscous-coupling (Bourbié et al., 1987) is probably not the cause of shear modulus variability in carbonates because the sensitivity to brine is large while it is not



Figure 2.5. Shear modulus weakening in carbonates samples resulting from dry to brine saturation at seismic frequency (100 Hz) for differential pressures of 3.5 and 31 MPa. Error bars, representing one standard deviation, are within the size of the marker for most samples.



Figure 2.6. Carbonate samples showing that the shear modulus remains almost constant from dry to brine saturation at ultrasonic frequency for differential pressures of 3.5 and 31 MPa. Error bars, representing one standard deviation, are within the size of the marker for most samples.



Figure 2.7. Carbonate samples showing little shear modulus weakening and strengthening resulting from dry to butane saturation compared to the dry-brine saturation case. Both plots are at a differential pressure of 3.5 MPa for seismic and ultrasonic frequencies. Error bars, representing one standard deviation, are within the marker size for most samples.

significant for liquid butane, with both fluids having similar and low viscosities (0.2 cP for liquid butane and 1 cP for brine). Dissolution of carbonate minerals could also be occurring. Dissolution of calcite and dolomite minerals depend on the pH of the fluid, temperature, and the reaction order of the cations (Ca, Mg, Ba) which control the dissolution rate of carbonate minerals (Chou et al., 1989).

By acquiring data at seismic and ultrasonic frequencies we observe evidence of at least three mechanisms for which the shear modulus weakens (surface energy reduction and crack growth) or strengthens (modulus dispersion). Changes in shear modulus could be observed from seismic time-lapse data, especially in the presence of compliant pores and polar fluids such as water. When injecting water into an oil reservoir, the nature of this polar fluid, its viscosity, pressure, temperature, etc. will likely interact with the rock solid phases creating weakening or strengthening on the shear modulus (and maybe in some cases the bulk modulus) compared to the original fluid saturation.

Also, when logging data are available in a field, the analysis has to consider that modulus dispersion can be significant and should be taken with care if compared to seismic data. Log data will fall in between our measured frequency ranges (~ 10 KHz). Having knowledge of the characteristic frequency (f_c) might help the interpretation of log data. The f_c separates the behavior for relaxed and unrelaxed fluids. If $f_{log} < f_c$ and we have compliant pores, we could observe weakening of the shear modulus upon water saturation. On the other hand, if the $f_{log} > f_c$, strengthening of the shear modulus might be observed. Sharma et al. (2006) compile results for the shear modulus change from dry to water saturation from several authors. In this study, the shear modulus strengthens at ultrasonic frequencies and weakens for sonic frequencies(~ 10 KHz) for data by Lucet (1989). This observation is in agreement with our observations on shear modulus change from seismic to ultrasonic frequencies.

2.6 Gassmann's fluid substitution

We introduced Gassmann's theory with its assumptions, and in this section we compare and analyze the computed saturated bulk modulus using Gassmann's theory to the measured rock bulk modulus. Our experimental setting for seismic frequency data acquisition let us acquire data when the fluid is at equilibrium. The pore pressure is held constant thus the fluid modulus is 0.5 GPa for butane, and 3.4 GPa for brine.

Figure 2.8 compares bulk modulus calculated using Gassmann theory to the measured modulus for butane saturated carbonates at frequencies: 100 Hz and 0.8 MHz, and at differential pressure of 31 MPa. The solid line represents the case where the butane-substituted modulus predicted by Gassmann's theory and measured bulk modulus are equal. Error bars represent one standard deviation for the bulk modulus. Gassmann's theory is correctly predicting the observed butane saturated modulus for our carbonate samples, partly because the influence of butane on the rock bulk modulus is not large. Butane is a highly compress-

ible fluid, thus, the fluid influence on rock compressibility is not significantly different from the dry rock (see Figure 2.2).

For brine saturation, Gassmann-calculated and measured bulk modulus at 100 Hz and 0.8 MHz, and at differential pressures of 3.5 and 35 MPa are compared in Figures 2.9 and 2.10 respectively. The solid line represents the case where the fluid-substituted and measured moduli are equal. Error bars represent one standard deviation for the bulk modulus. Observe that some samples match the predictions well, while others do not.

In Figure 2.9 at a frequency of 100 Hz, none of the predictions fit the observed bulk modulus within the associate uncertainty, while at 0.8 MHz, for the same differential pressure of 3.5 MPa, the fit to the predicted bulk modulus is better. At low differential pressure and at 100 Hz, the bulk modulus for all of the samples but F is over-predicted by Gassmann's theory. We observe shear modulus weakening for all samples (and the least for sample F, Figure 2.5), therefore if the rock frame has weakened in the presence of brine so could the bulk modulus, a factor not accounted for Gassmann's theory. Therefore the over-prediction of the bulk modulus by Gassmann's theory at low differential pressure is probably because the rock frame has been altered (softened).

The bulk modulus is under-predicted for 100 Hz, yet it is well predicted at 0.8 MHz (Figure 2.9). This is largely a result of modulus dispersion. Gassmann's theory estimates the saturated modulus for low frequencies. Gassmann's theory uses the bulk modulus of the dry rock, which is not dispersive, to predict the saturated rock modulus. However, modulus dispersion exists in most of our brine saturated carbonates (see Figure 2.2). This bulk modulus dispersion is evidenced in the shifting of data points in Figure 2.9 as the frequency increases from 100 Hz to 0.8 MHz. The bulk modulus dispersion at ultrasonic frequency can lead to errors when comparing ultrasonic to seismic data. Thus, a better fit at ultrasonic frequency might be somewhat of a paradox on Gassmann's theory applicability on carbonates.

At a differential pressure of 31 MPa (Figure 2.10), the 100 Hz data show that the bulk modulus of four brine saturated carbonates (B, E, I and H) is predicted well by Gassmann's theory. The bulk modulus for samples A and C are largely over-predicted by Gassmann's theory. Samples A and C have the highest content of non-calcareous minerals, especially clay. We ignore that softening of clays is a possible mechanism for elastic moduli weakening for most of our samples. However, that $K_{Measured}$ is significantly less than $K_{Gassmann}$ for samples A and C. These two samples have some clay content in the pore space, and thus the decrease in bulk modulus can result from water softening the pore space and clays of these carbonates.

We focus now on data at 100 Hz, where frequencies are low enough that we expect the fluid pressures gradients are zero as Gassmann's theory requires. Still, at high differential pressure, we observe that some samples are well predicted by Gassmann's theory while others not. So where can this difference come from? On one hand, we have observed rock



Figure 2.8. Butane saturated bulk modulus measured and estimated with Gassmann's theory for 100 Hz and 0.8 MHz at 31 MPa differential pressure. Solid line represents equal measured and estimated bulk modulus. Error bars are one standard deviation of the bulk modulus.



Figure 2.9. Brine saturated bulk modulus measured and estimated with Gassmann's theory for 100 Hz and 0.8 MHz at 3.5 MPa differential pressure. Solid line represents equal measured and estimated bulk modulus. Error bars are one standard deviation of the bulk modulus.



Figure 2.10. Brine saturated bulk modulus measured and estimated with Gassmann's theory for 100 Hz and 0.8 MHz at 31 MPa differential pressure. Solid line represents equal measured and estimated bulk modulus. Error bars are one standard deviation of the bulk modulus.


Figure 2.11. Bulk modulus for carbonates with brine saturation as a function of differential pressure $(P^{1/3})$ for 100 Hz.

shear modulus sensitivity to brine saturation. On the other hand, for low differential pressures we expect to have open compliant pores or cracks. Gassmann's equations are derived without assuming any specific pore geometry, and can be applied to any pore type as long as the assumptions for Gassmann's theory are satisfied, i.e. pore pressure is in equilibrium. The mismatch between observed and Gassmann predicted bulk modulus could relate to differences in pore type creating pressure gradients or chemical reactions which violate Gassmann's assumptions. Therefore, samples yielding better predictions by Gassmann's theory might be explained through the dependence of bulk modulus with differential pressure. Figure 2.11 plots the bulk modulus of brine saturated carbonates as a function of differential pressure. The anomalous behavior of sample D at 20.7 MPa is due to a small gas leak into the rock when the sample was saturated with brine. This dramatically lowered the bulk modulus of samples D at low frequencies for pressures lower than 20.7 MPa. In Figure 2.11 we observe a consistent linear behavior of bulk modulus with differential pressure from Hertz-Mindlin model: $K = mP^{1/3}$ (Mavko et al., 1998), still the slopes (m) of the linear trends are different for different rocks. Higher slopes mean larger dependence on differential pressure, indicating the existence of compliant pores or micro-cracks. Table 2.2 compares Gassmann's predictability, shear modulus weakening, mineralogy, and pressure effect on all samples at 100Hz. Gassmann's predictability and shear modulus weakening are reported for the highest differential pressure reached at 31 MPa. The pressure effect is measured by the slope of the linear dependence (m) of the bulk modulus (Figure 2.11).

SAMPLES	А	В	С	D	Е	F	Η	Ι
GASSMANN'S THEORY FITS		Х			Х		Х	Х
SHEAR MODULUS WEAKENING		Х		Х			Х	
BULK MODULUS VS. PRESSURE (m)	11.8	5.6	7.9	11.2	5.2	5.4	4.5	3.2
MINERALOGY	С	D	С	С	С	С	D	D

Table 2.2. Gassmann's theory applicability correlated with shear modulus weakening and bulk modulus dependence with pressure (slopes of Figure 2.11). Gassmann and shear modulus analysis corresponds to 100 Hz at a differential pressure of 31 MPa. X means the statement is true. Dominant mineralogy: C = calcite, D = dolomite. Note correspondence of good Gassmann's theory fit with low pressure dependence (m).

There seems to be no correlation between the shear modulus weakening and the observed match between measured and computed bulk modulus for brine saturated carbonates at high differential pressure (Table 2.2). For example, both B and D samples show significant shear modulus weakening at 31 MPa differential pressure, still sample B is well predicted by Gassmann's theory while sample D is not. It might seem confusing that although Gassmann's assumption that the rock frame stays unaltered by the fluid is violated for some samples, the measured brine saturated bulk modulus is well predicted by Gassmann's theory for these samples. A likely reason for this is because the increase in bulk modulus in absolute percent from dry to brine saturation (35% in average) is more significant than the shear modulus weakening in absolute percent (6% in average).

Examining the pressure dependence, the saturated bulk modulus for samples with lower slopes (B, E, H and I) is well predicted by Gassmann's theory. Low slopes mean the sample has less compliant pores or cracks. Samples A, C and D have high slopes, and Gassmann's theory is not predicting the observed saturated bulk modulus. Sample F has an intermediate slope, but the saturated bulk modulus is not well predicted by Gassmann's theory. For sample F the bulk modulus as a function of pressure is less smooth than for other samples, leading to a higher variance in the slope calculation. As previously mentioned, our experimental setup could not quite reach the differential pressure of the reservoir at 34.5 MPa. This could result in some compliant pores still being open at these pressures. From this we conclude that open compliant pores are a possible factor affecting the mismatch between observed and predicted bulk modulus. Samples B, H and I are dolomites, but we do not have enough statistical data to make correlations with rock grain density. Nevertheless, these dolomite samples have high porosity and permeability probably satisfying Gassmann's assumption on pore connectivity and fluid distribution in the porous space.

From our observations, carbonates with round pores, small vugs or micritic textures are well predicted by Gassmann's theory for low frequencies. Even at reservoir pressures, open compliant pores or cracks might be present at reservoir in-situ conditions. In this case an anisotropic fluid substitution theory, such as that of Brown and Korringa (1975) is perhaps more appropriate. However, knowledge of the anisotropic symmetry with all of the stiffness coefficients of the rock and the pore space compressibility are required for this theory. Using additional parameters might allow one to fit the data better, but the estimated parameter could not be realistic or representative of the rock.

2.7 Conclusions

We present data over a large range of frequency and under varying saturation and pressure conditions to investigate the applicability of Gassmann's theory for our carbonate data set. We observe that the rock shear modulus is sensitive to brine saturation, especially at seismic frequencies. Weakening of the solid matrix occurs possibly due to surface energy loss and/or sub-critical crack growth in compliant pores, mostly at low differential pressures. These mechanisms violate an assumption of Gassmann's theory that the fluid does not influence the the solid matrix of the rock. However, we find no positive correlation between the rock shear modulus weakening and the failure of Gassmann's theory to predict the saturated bulk modulus at seismic frequencies. We do find that the brine saturated bulk modulus for carbonates with small differential pressure dependence (round pores or vugs) is well predicted by Gassmann at seismic frequencies, while for carbonates strongly influenced by pressure (compliant pores or micro-cracks), Gassmann's theory does not match the observations. Therefore, knowledge of the reservoir pore space geometry can aid in the understanding and applicability of Gassmann's theory.

Predicting the saturated bulk modulus at ultrasonic frequencies violates Gassmann's low-frequency assumption. Nevertheless, we test our carbonate samples at ultrasonic frequencies to show the role of modulus dispersion. For some of our samples, the measured and Gassmann-calculated bulk modulus at ultrasonic frequencies show better agreement compared to seismic frequency. This match is apparent, resulting from bulk modulus dispersion which we observe in our carbonates when saturated with brine. We also observe shear modulus dispersion. Little change from dry to brine saturation is present in the rock shear modulus at ultrasonic frequencies, but this modulus is always higher than the shear modulus obtained at seismic frequency. This increase could be a result of dispersion or preferential propagation path, which avoids altered (weakened) sections in the saturated rocks. Although our conclusions are based on samples with different texture and mineralogy we must be careful to generalize these results to all carbonate rocks.

Our observations are applicable particularly to the analysis of time-lapse data. Ultrasonic laboratory data are used in some cases to calibrate time-lapse seismic reflection data. We should be aware that bulk modulus in carbonate rocks can have significant dispersion affecting the applicability of Gassmann's fluid substitution theory at ultrasonic frequencies (and maybe at log frequencies). Also, when water or brine replaces a non-polar fluid such as oil, shear modulus weakening could be observed in the field.

CHAPTER 3

SEISMIC WAVE ATTENUATION IN CARBONATES

3.1 Introduction

Velocity and amplitude analysis of elastic waves is common practice in reservoir rock physics. Variations in reservoir seismic properties can be related to fluid changes within them. Intrinsic wave attenuation, or the conversion of mechanical energy into heat, has been a topic of research for many years. Although numerous studies examine mostly clastic rocks (Winkler and Nur, 1982; Best et al., 1994; Murphy, 1982; Wyllie et al., 1962; Spencer, 1979; Best et al., 2007), there are few laboratory measurements of attenuation in carbonates rocks, and even fewer experimental data at exploration seismic frequencies and reservoir pressures (Lienert and Manghnani, 1990; Paffenholz and Burkhardt, 1989; Spencer, 1981; Batzle et al., 2005; Gautam, 2003). Because more than half of the current major oil and gas reservoirs in the world are in carbonates, these rocks have become important topics of rock property research. Our study analyzes wave attenuation in five carbonate samples at seismic frequencies ($10^1 - 10^3$ Hz), and at an ultrasonic frequency (10^6 Hz) for three of the samples. The rocks are measured dry and fully-saturated with a light hydrocarbon and with a brine, at reservoir pressures.

There are several ways to describe intrinsic seismic losses, and in this paper we use the definition of attenuation as the inverse of the quality factor (Q). Intrinsic attenuation can be defined as $Q^{-1} = \frac{Im[M^*]}{Re[M^*]}$, where M^* is the complex modulus or velocity; and the imaginary part (Im) of the complex modulus is small. Different waves and flexural modes allow us to study the shear-wave (Q_S^{-1}), compressional-wave (Q_P^{-1}), extensional (Q_E^{-1}) and bulk compressibility (Q_K^{-1}) attenuation. Winkler and Nur (1979) and Winkler (1979) analytically show that one of the following relations between the attenuation modes will always be true:

$$Q_S^{-1} > Q_E^{-1} > Q_P^{-1} > Q_K^{-1}, (3.1)$$

$$Q_K^{-1} > Q_P^{-1} > Q_E^{-1} > Q_S^{-1}, (3.2)$$

$$Q_K^{-1} = Q_P^{-1} = Q_E^{-1} = Q_S^{-1}.$$
(3.3)

For a porous medium, and based on the conceptual model of two identical cracks perpendicular to each other and touching at their tips, Winkler (1979) interprets relation 3.1 as the expected behavior when the rock is fully-saturated, while equation 3.2 is for partially saturated rocks. This is generally true in sandstones for either logging or ultrasonic frequencies: the shear losses at full-saturation are mostly observed to be greater than the bulk or compressional-wave losses (Winkler and Nur, 1982; Best et al., 1994; Murphy, 1982; Wyllie et al., 1962; Spencer, 1979; Toksoz et al., 1979). Pore shape and distribution in sandstone could dictate that either P-wave or S-wave attenuation can dominate at logging frequencies (Assefa et al., 1999; Prasad and Meissner, 1992).

However, published data on carbonates are mixed regarding which attenuation mode dominates. Lucet (1989) and Lucet et al. (1991) measured Q_E^{-1} and Q_S^{-1} on 15 saturated limestones at sonic and ultrasonic frequencies. $Q_E^{-1} > Q_S^{-1}$ for virtually all the fully-water saturated samples. Paffenholz and Burkhardt (1989) and Cadoret et al. (1998) observe that $Q_E^{-1} > Q_S^{-1}$ at seismic and sonic frequencies respectively for water-saturated samples. Recently, Best et al. (2007) and Agersborg et al. (2008) measured carbonate samples at ultrasonic frequencies and high differential pressures. Differential pressure is the difference between hydrostatic confining pressure and pore pressure. The three carbonate samples from Best et al. (2007) show mixed relations between Q_P^{-1} and Q_S^{-1} , because the samples have velocity and attenuation anisotropy resulting from the complex pore structure. Four out of the six samples from Agersborg et al. (2008) have estimates of Q_P^{-1} greater than Q_S^{-1} under dry and fully-brine saturated conditions at 30 MPa differential pressure. Most of the previous work, except for Lucet (1989) and Assefa et al. (1999), has been performed at room conditions and with the samples either dry or saturated with water and measured at high frequencies. Here we study our samples near the reservoir differential pressures over a broad frequency range, dry and saturated with brine, as well as with a light hydrocarbon for the purpose of reservoir characterization. For our fully-saturated carbonate samples, the correlation among the different attenuation modes is described by relation 3.2.

For our carbonate sample set, we address five important relationships. Specifically, we examine the relation between the modes of attenuation, compare Q at seismic and ultrasonic frequencies, measure the sensitivity of Q to fluids (either a light hydrocarbon or a brine), investigate the frequency dependence of Q in the exploration seismic bandwidth (10-100 Hz), and determine if there is a correlation between attenuation and permeability. Both the sensitivity of attenuation to fluids and its correlation to permeability are important for studying the evolution of reservoir systems, because one possible application for utilizing attenuation as a monitoring tool is during enhanced oil recovery where water, steam or CO_2 is injected into the reservoir to stimulate oil production. Although quantitative intrinsic attenuation data from surface seismic have been an interpretive challenge historically, characterizing attenuation becomes more robust as acquisition and processing technologies continue to improve.

3.2 Samples

The core samples in this study come from two Middle Eastern carbonate reservoirs undergoing enhanced oil recovery. The core plugs from the first reservoir are samples 100, 200 and 300. For this field, light hydrocarbon production is stimulated by injecting brine

SAMPLES	100	200	300	В	С
Porosity(%)	32.79	30.46	20.39	4.60	21.00
Permeability(mD)	6.75	5.19	58.10	0.03	5.50
Grain density (gm/cm^3)	2.71	2.71	2.71	2.84	2.70
$\operatorname{Calcite}(\%)$	90.6	89.4	98.6	0.7	75.8
Dolomite(%)	8.5	9.6	TR	97.0	20.6
Quartz (%)	0.6	0.7	1.2	0.6	1.2
Phyllosilicates $(\%)$	0.0	TR	TR	0.8	2.4
Texture	wacke	wacke	pack	mud	grain
Depth (m)	2479	2452	2456	3181	3138
Reservoir Pd (MPa)	34.5	34.5	34.5	35.8	35.8
Heterogeneity	none	none	vertical bands	horizontal bands	none

Table 3.1. Petrological data for the carbonate set. Mineralogy was obtained from XRD analysis and are reported in percent per volume. TR means less than 0.3%. Texture follows modified Dunham's carbonate classification (Moore, 2001): mud = mudstone, wacke = wackestone, pack = packstone and grain = grainstone. Porosity and permeability are measured at 18.3 MPa confining pressure for samples 100, 200 and 300; and at reservoir pressures for samples B and C. By heterogeneity we refer to visual or CT-scan heterogeous features.

into the reservoir (Soroka et al., 2005). Samples from the second reservoir are named B and C; corresponding to the same samples as in chapter 2. The samples are either almost pure calcite or dolomite (>95% total volume), with less than 3% clays. Table 3.1 summarizes the petrographical analysis. Porosity and permeability are measured using standard helium porosimetry and air permeability equipment. Permeability values are corrected for Klinkenberg gas slippage. The samples are cylindrical, and either 2.5 or 3.75 cm in diameter, with lengths varying from 4.3 to 5.5 cm. We now describe the petrographical characteristics of the samples, and thin sections for the samples are shown in Figure 3.1.

Sample 100: Mud-supported wackestone texture. Blue haze in the thin sections implies abundant micro-porosity (Figure 3.1). The presence of micro-porosity suggests the sample has a large content of irreducible water in-situ. Micro-pores are 1-5 microns. The main fossil observed is Orbitolina. The sample is mostly calcite with calcite cement. Dolomite is deposited in small vugs after the dissolution of calcite (burial dolomite). There is autogenic quartz and linear clam fragments (Scholle and Ulmer-Scholle, 2003).

Sample 200: Mud-supported wackestone texture. This sample is similar to sample 100, except that micro-pores are larger (20-50 microns). The sample is partially burrowed and dolomite crystals are less coarse. This sample has larger amounts of fossils fragments than sample 100 (e.g. rudist fragments (Scholle and Ulmer-Scholle, 2003)).

Sample 300: Packstone borderline grainstone texture. Abundant skeletal debris with inter-particle porosity. This type of porosity results in better permeability. Echinoderm fossils are surrounded by a synaxial overgrowth of calcite cement. Different textures between



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Figure 3.1. Thin section of samples a) 100, b) 200, c) 300, d) B and e) C. Shown in plane polarized light. Dyed epoxy is blue and represents pore space.

stylolytes are probably the result of high and low energy environments.

Sample B: Mudstone texture. Composed of relatively pure dolomite with a non-planar fabric and a unimodal, aphanocrystalline to very-finely crystalline texture. The sample has horizontal solution seams (or microstylolites) and small vugs (<1mm), possibly infilled by an iron oxide.

Sample C: Skeletal grainstone texture with finely-crystalline spar matrix. Skeletal grains have been largely dissolved, but those remaining are foraminifera (mostly miliolids and other rotalids), ranging from 0.25 to 1.5 mm in diameter, moderately sorted. Contains moldic porosity, partially filled with euhedral dolomite rhombs (30-60 microns in diameter), and interparticle porosity, partially filled with subhedral finely-crystalline calcite.

Based on the fossil types, samples 100, 200 and 300 were deposited in the photolytic zone. In all three samples the cement is calcite and there are no clays.

Samples 100 and 200 are similar in most petrographic aspects, with the main difference being the pore size estimated from the digital thin section images. Sample 100 has 1-5 micron pores, compared to 20-50 microns for sample 200. Small radius pores (\sim 10 microns) generally have small pore throats, while larger radius pores (>10 microns) could either have large or small pore throats (Lindquist et al., 2000). Thus, sample 200 may have larger pore throats than sample 100. Pore throats are important, as they control the time fluid pressure takes to reach equilibrium after being disturbed by a passing wave. This fluid pressure disequilibrium controls attenuation. Figure 3.2 shows the CT scans of the samples. Gray scales represent densities contrasts. Sample 300 is the most heterogeneous,



Figure 3.2. CT scans of samples: a) 200, b) 300 and c) B. Gray color variations represent density contrasts. Observe how sample 300 is heterogeneous with bands of different textures/porosity.



Figure 3.3. Scanning electron microscope (SEM). Sample 100 shows dolomite crystals. There are no SEM images for Samples B and C.

with vertical textural bands. Although we do not have the CT scan for sample 100, the scanning electron microscope (SEM), thin section and visual inspection show similarities to sample 200. Sample C, in spite of abundant fossils, is overall homogeneous. Finally, Figure 3.3 contains the SEM images of samples 100, 200 and 300. Crystals of dolomite are observed for samples 100 and 200, but are mostly absent for sample 300, verifying the XRD data (Table 3.1).

Velocity and attenuation anisotropy is not expected to be significant on these samples based on CT scans and visual inspection. Still, we do not quantitatively show the presence or absence of anisotropy.

3.3 Experimental Procedure

The core samples are measured dry (humidified, as described below), 100% saturated with liquid-butane (C_4H_{10} , a light hydrocarbon) or with high salinity brine (180,000 ppm NaCl). Samples are first measured dry, then butane is injected into the sample until there is no more flow from the fluid pump to the sample, indicating we have reached a full saturation state. Butane at room conditions is a gas, but becomes liquid at pressures greater than 0.3 MPa. To remove the liquid butane from the sample pore space, we slowly decrease the pore pressure until we reach the room pressure condition. We then open the fluid lines to let the butane gas escape. Brine is injected into the sample the same way as for butane. Samples are coated with an impermeable polyimide film (kapton), over which semi-conductor strain gages are glued to measure rock deformations at seismic frequencies. This film keeps moisture inside the rock, prevents nitrogen diffusion and eliminates the Biot-Gardner effect observed in unjacketed rocks (Gardner, 1962; White, 1986; Mörig and Burkhardt, 1989).

All dry samples, except sample B, are initially humidified by storing the samples in a humidifying chamber at 98% humidity for 2 weeks, to create matrix softening that results from moisture in the pore space. Clark et al. (1980) showed that introducing water vapor ($\sim 1\%$ water saturation in pores) to oven dry samples can significantly reduce the rock bulk and shear moduli, and increase attenuation. The water content is not expected to be less than 1% in reservoirs, as even gas filled reservoirs have irreducible water in the pore space. The rock sensitivity to water vapor is directly related to the pores surface area. Nonetheless, we will show here that as the brine saturation of the rock increases, the rock frame softens. Such weakening could well be observed in reservoirs where brine displaces the original gas or fluid.

Figure A.1 is a drawing of the sample setup and a sketch of the recorded low frequency strains for the different gages. At seismic frequencies, moduli and attenuation are measured by applying a sinusoidal stress to the rock and measuring the resulting strain in different directions on the rock sample and on the reference material (aluminum). The measured strain amplitudes are low ($\sim 10^{-7}$), the same amplitude as the deformation imposed on the reservoirs in the field by seismic waves. The samples and reference are cylindrical and



Figure 3.4. Schematic drawing the sample setup for low frequency measurements with strain gages and ultrasonic frequencies with transducers. Confining pressure is applied to the whole system, and pore pressure is controlled through the pore fluid lines.

are attached in series with the cylinder axis coinciding with the axis of applied stress. The moduli are estimated from the amplitude of the strain signatures (for details see Batzle et al., 2006). To measure ultrasonic frequencies we use transducers embedded in the aluminum reference (Figure A.1). We record the waveform transmitted through the rock core (Birch, 1960). From the measured time of flight and sample length, we estimate the P- and S-wave velocities and from these, the rock moduli. Samples with visible heterogeneity or large vugs the size of the strain gages (~ 0.5 cm) are avoided because the gages will measure the properties of the heterogeneity, biasing the estimate of attenuation. For the ultrasonic pulse, these heterogeneities can create scattering losses.

The whole system as shown in Figure A.1 is lowered into a high pressure vessel. Confining (hydrostatic) pressure (P_c) is applied with nitrogen gas; while the pore pressure (P_p) is controlled through the pore fluid lines with different types of fluids. In this paper we present attenuation data recorded under differential pressure (P_d) : $P_d = P_c - P_p$. Velocity and attenuation data are measured at 31 MPa differential pressure for samples 100 and 200, $P_d=24$ MPa for sample 300, and for samples B and C, $P_d=21$ MPa. For all rocks, P_p was held constant at 3.5 MPa. Because the sample setup is pressurized with nitrogen gas, for safety reasons the system is not able to quite reach the reservoir's confining pressure. However, the pressures in the experimental setup are close to the reservoirs differential pressure (see Table 2.1).

In our experiments, we estimate Q_E^{-1} and Q_S^{-1} at seismic frequencies, and Q_P^{-1} and Q_S^{-1} at ultrasonic frequencies. In a visco-elastic material, the strain lags in phase with respect to the stress. This phase lag (θ) can be described in terms of the complex modulus: $Q^{-1} = \frac{Im[M^*]}{Re[M^*]} = \tan \theta \approx \theta$. Because the phase lag is small, we can approximate the tangent by the angle itself. In our low frequency experiments, attenuation is estimated from the phase lags between the different strains. The phase is obtained by using a lock-in amplifier. We assume that the phase of the strain on the aluminum represents the phase of the applied stress. This is a reasonable assumption because the attenuation for aluminum is low ($Q^{-1} \sim 10^{-5}$). Thus, extensional attenuation (Q_E^{-1}) is estimated from the phase lag of the vertical strain on the rock and the reference aluminum. The vertical strain on the rock has the largest strain waveform amplitudes, yielding robustness in the estimate of Q_E^{-1} . To obtain the shear-wave attenuation we use the following relation (White, 1965):

$$\frac{1}{Q_S} \simeq \frac{1}{Q_E} - \frac{\nu tan\theta_\nu}{1+\nu} \tag{3.4}$$

where ν is Poisson's ratio and θ_{ν} is the phase lag between the vertical and horizontal strain on the rock sample. Poisson's ratio at seismic frequencies is obtained from the ratio of the horizontal and vertical strain amplitude on the rock. The estimate of Q_S^{-1} is not overly sensitive to an error in ν . A ±10% bias in ν translates into a ±2% error Q_S^{-1} . Winkler and Nur (1982) derive equations that enable the estimation of a specific mode of attenuation for isotropic materials from two other attenuation modes and the Poisson's ratio. We estimate the P-wave and the bulk compressibility attenuation from (Winkler and Nur, 1982):

$$\frac{(1-\nu)(1-2\nu)}{Q_P} = \frac{1+\nu}{Q_F} - \frac{2\nu(2-\nu)}{Q_S}$$
(3.5)

$$\frac{1-2\nu}{Q_K} = \frac{3}{Q_E} - \frac{2(\nu+1)}{Q_S}$$
(3.6)

For the ultrasonic data, attenuation is estimated by the spectral ratio method (Toksoz et al., 1979; Hauge, 1981). An aluminum sample of the same shape and dimensions as the rock sample is used as a reference material to establish the base spectra. Aluminum has a high bulk modulus (K=76 GPa) and a high Q. The waveforms on the aluminum sample were measured at room conditions and under 31 MPa confining pressure, but there is no significant difference in the ultrasonic waveform or on the amplitude spectrum at these conditions. The direct arrivals for the sample and aluminum are windowed by a Hanning-type function, and the amplitude spectra is obtained from a discrete Fourier transform to these isolated events. Q is estimated from the logarithm of the ratio of amplitude spectra as a function of frequency. P- and S-wave attenuation are directly estimated from the waveforms, and we use the relations in equation 3.6 to estimate the remaining two attenuation modes.

3.4 Parameter uncertainty

Estimates of the error in our data are represented in terms of the standard deviation (σ) between our noisy observations and data predicted by an assumed true model (for details see appendix A). For the low frequency data, this is the standard deviation of our phase measurements and the Poisson's ratio. The repeatability or standard deviation of the phase measurements with our lock-in amplifier is on average 0.03 degrees (Gautam, 2003). One standard deviation of the Poisson's ratio (ν) is on average 0.002. These phase and ν standard deviations are propagated to estimate the error in Q⁻¹ at seismic frequencies. On average at low frequencies, one standard deviation in $1/Q_K$, $1/Q_P$, $1/Q_E$, $1/Q_S$ are 0.0042, 0.0090, 0.0023 and 0.0031, respectively. The average σ in P-wave and S-wave velocity are 121 m/s and 42 m/s, respectively.

Attenuation at ultrasonic frequencies is estimated by the spectral ratio method on the recorded waveforms. On average, the estimated standard deviation for $1/Q_K$, $1/Q_P$, $1/Q_E$, $1/Q_S$ at ultrasonic frequencies are 0.0019, 0.0007, 0.0019 and 0.0012, respectively. The average σ for P-wave and S-wave ultrasonic velocities are 22.4 m/s and 7.8 m/s, respectively.

We cannot quantify the bias on our Q estimate because we do not know the true model, but we can estimate part of a bias on Q based on the difference in attenuation estimates between two methodologies. The frequency shift (Quan and Harris, 1997) and the spectral ratio methods assume Q to be frequency independent, but Q is estimated differently by integrating or taking the ratio of amplitude spectra, respectively. On average, the ultrasonic attenuation estimates between these two methods differ by 18%. Other sources of bias on Q are possible if we have errors in two parameters. 1) If there is a $\pm 10\%$ variation in this window size, the error in Q is $\pm 20\%$ because we either include late arrivals or the window is not long enough to contain the whole amplitude information of the direct arrival. 2) If there is a $\pm 5\%$ error in the picked first arrival time, Q has an error of $\pm 6\%$.

3.5 Dispersion

Figure 3.5 shows the P-wave velocity as a function of frequency for samples 100, 200 and 300, estimated at seismic frequencies between 10 and 1000 Hz, and accompanied by a single ultrasonic data point at 0.8 MHz. We plot the low frequency velocity at three representative points (10, 100 and 1000 Hz) obtained from the previously described parameter estimation and error analysis. The data points near 10^4 Hz are obtained from sonic log data in the wells from which the rock samples were cored. This velocity is an average over a 0.6 meter depth range centered at the sample depth. From the resistivity and bulk density logs, we conclude that the samples were extracted from brine-saturated reservoir intervals. Overall, the P-wave velocity for the three samples consistently increases with frequency. At low frequencies, sample 300 has large error bars in the velocity, so the velocity in this sample could also be interpreted as non-dispersive in the seismic frequency range.

There are two reasons why the ultrasonic velocity is higher than seismic and log velocities. High velocities at ultrasonic frequencies can result from the bulk stiffening due to pore fluid pressures being unable to reach equilibrium within one wave period (Biot, 1956b). At low frequencies there is enough time for the pore fluid pressure to relax, the system is therefore more compliant, which results in lower velocities. Our observed velocity dispersion is in agreement with those from the broad frequency range experiment on sandstones by Sams et al. (1997) and Best and Sams (1997). This velocity dispersion could be modeled by existing dispersive media theories (Biot, 1956a,b; O'Connell and Budiansky, 1977; Dvorkin et al., 1995; Müller and Gurevich, 2005). However, choosing the wave dispersion mechanisms that best describe our data is beyond the scope of this paper.

An alternative explanation why the ultrasonic velocity is higher than for seismic and log frequencies is that the ultrasonic wave propagation is path dependent. This phenomenon is particularly relevant given our transducer size. Each ultrasonic crystal (P- and S-wave) in our transducer package has a diameter of 0.75 cm compared to the aluminum casing diameter of 3.75 cm. Small transducers are used because the aluminum casing is also our reference material for the strain measurements, and a large crystal embedded into the casing would change the elastic properties of the aluminum. Because the transducer face is small with respect to the sample diameter, the generated high-frequency wave can propagate through the fastest path in the sample. For a heterogeneous sample with high and low compressibility regions, a portion of the ultrasonic pulse preferentially propagates along incompressible, high-speed wave regions according to Fermat's principle. For sample 300, the log data agree better with the low frequency data than with the ultrasonic data (Figure 3.5). From the CT-scan for sample 300 (Figure 3.2), zones of high and low compressibility are aligned with the pulse propagation direction, accentuating the fast path phenomenon. The ultrasonic energy then can traverse this heterogeneous sample more quickly, leading to an overestimate of the average wave speed of sample 300. This analysis is qualitative at this point, but it shows how a relatively high ultrasonic velocity can be due to dispersion as well as path dependence.

3.6 Attenuation Modes

Figures 3.6 to 3.10 show the different attenuation modes for all the samples as a function of frequency for three pore-filling fluids. For low frequency data we first estimate extensional and shear-wave attenuation, and from these and the Poisson's ratio we obtain bulk and P-wave attenuation. For ultrasonic data, we use the P- and S-wave attenuation estimates to calculate the extensional and bulk attenuation. For most samples and frequencies, bulk compressibility losses dominate the shear-wave attenuation for dry (humidified) and 100% saturated samples. These carbonate samples show that relation 3.2, previously interpreted as a condition of partial saturation (Winkler, 1979), governs our fully-saturated carbonate samples. For sample 300, the P-wave transducer failed, therefore there is no ultrasonic frequency P-wave attenuation data. We also did not acquire the ultrasonic attenuation for samples B and C because we did not have the ultrasonic aluminum standard measurements to go with it. Sample 300 has low Q, and overall the attenuations are equal for all modes. High attenuation is probably responsible for the inability to distinguish between attenuation modes.

There are two ways we try to ensure full-saturation of the connected pore space. First, we calculate the connected pore volume from the porosity and dimensions of the sample. We then monitor the fluid flow into the sample until there is no more flow from the pump to the sample. The volume of injected fluid is within 2% of the computed pore volume. Second, we observe the bulk modulus as a function of brine saturation. Figure 3.11 shows the bulk modulus as a function of saturation for sample 300. The addition of pore fluid stiffens the sample, which we observe as an increase in bulk modulus. The significant increase of the bulk modulus at 100% saturation supports the assumption that sample 300 is fully-saturated. Observe that a small amount of gas (~ 8%) drops the bulk modulus close to its dry-sample value for seismic frequencies. For ultrasonic frequencies, the bulk modulus starts increasing at a saturation lower (~ 23%) than for the seismic frequencies (~ 62%). As mentioned previously, the velocity or modulus increase with frequency can result from rock-fluid dispersion mechanisms or path effects. Samples partially saturated by imbibition can have regions that are saturated while others remain dry. These saturated regions can form preferential high speed paths with high bulk moduli.

The dominance of the bulk over shear attenuation for our saturated samples can be



Figure 3.5. P-wave velocity as a function of frequency for samples 100, 200 and 300, measured dry and fully saturated with liquid-butane and brine. A linear fit is applied to the low frequency data (3, 100 and 1000 Hz). The data at 10^4 Hz is obtained from the sonic logs. One σ for the ultrasonic data is the size of the markers.



Figure 3.6. P-wave, S-wave, extensional and bulk compressibility attenuation for sample 100 at 31 MPa differential pressure. The sample is measured a) dry (humidified), b) 100% saturated with liquid butane and c) 100% saturated with brine.



Figure 3.7. P-wave, S-wave, extensional and bulk compressibility attenuation for sample 200 at 31 MPa differential pressure. The sample is measured a) dry (humidified), b) 100% saturated with liquid butane and c) 100% saturated with brine.



Figure 3.8. P-wave, S-wave, extensional and bulk compressibility attenuation for sample 300 at 24 MPa differential pressure. The sample is measured a) dry (humidified), b) 100% saturated with liquid butane and c) 100% saturated with brine.



Figure 3.9. P-wave, S-wave, extensional and bulk compressibility attenuation for sample B at 21 MPa differential pressure. The sample is measured a) dry (humidified), b) 100% saturated with liquid butane and c) 100% saturated with brine.



Figure 3.10. P-wave, S-wave, extensional and bulk compressibility attenuation for sample C at 21 MPa differential pressure. The sample is measured a) dry (humidified), b) 100% saturated with liquid butane and c) 100% saturated with brine.



Figure 3.11. Bulk modulus as a function of brine saturation for sample 300 at seismic and ultrasonic frequencies. At 100 Hz, the bulk modulus increases 25% from 97% to 100% brine saturation.

explained two ways. The first is related to the fact that relations 3.1 and 3.2 are associated with full or partial saturation respectively. This interpretation is based on a conceptual model of two identical low aspect ratio pores perpendicular to each other at their tips (see Figure 15 in Winkler and Nur (1982)). Because we are dealing with carbonate rocks of complex texture, the variety of pore shapes may not be well described by this two crack model. Thus, the saturation state of carbonate rocks might not be possible to determine from the relation among the different attenuation modes. Second, Dvorkin et al. (1995) shows that squirt flow from softer pores or cracks to stiffer pores in fully-saturated rocks can predict that the compressional-wave attenuation is greater than the shear-wave losses. This model is hard to implement with our data, because attenuation for low pressure is calculated from data at high pressure, where cracks or compliant pores are assumed to be closed. Also, pore scale squirt models are probably not dominant at seismic frequencies (Pride et al., 2004), but the same idea regarding losses has been extended to seismic frequencies by substituting the notion of pores by mesoscopic regions (Pride et al., 2004; Masson and Pride, 2007).

In our carbonate samples, no micro-cracks are observed from the thin sections, but the existence of compliant pores is interpreted from the pressure dependence of the shear modulus in Figure 2.4. At lower pressures, compliant pores open, softening the rock frame and lowering the modulus and velocity. All the samples show similar pressure dependence for the modulus and velocity as for Figure 2.4. Due to the textural complexity of carbonate rocks and because hydrostatic stress is applied, compliant pores can remain open even at high differential pressure, creating soft rock regions. Therefore, losses resulting from fluid movement from softer to stiffer regions is possible for our carbonate samples at high differential pressures, and can result in compressional losses dominating over shear losses.

There are several experimental challenges when measuring attenuation that must be

accounted for when interpreting the results plotted in Figures 3.6- 3.10. First, the fluid lines in Figure A.1 have a manual valve which we close when acquiring low frequency data. The fluid lines are designed to allow high fluid flow into and out of the sample. Unless the valve is completelly closed, the applied stress at low frequencies (< 50 Hz) creates significant fluid movement from the sample and into the fluid lines. This fluid flow effect is illustrated in Figure 3.6c for frequencies less than 50 Hz, for which high attenuation is not related to the visco-elastic properties of the rock, but to a valve malfunction.

A second challenge is that the low frequency apparatus has system resonances at frequencies that depend on the vessel as well as the geometry and properties of the rock and aluminum samples. Resonance peaks at 150 and 257 Hz affect the attenuation measurements in Figure 3.8c. This resonance is evidenced by the attenuation peaks at those two frequencies, biasing the attenuation estimates in the 150 to 1000 Hz range. Therefore, to avoid the attenuation instability generated by the system resonances for sample 300, we only analyze frequencies between 10 and 100 Hz.

A third challenge is to interpret how the heterogeneous rock texture observed on sample 300 can affect the low frequency attenuation and modulus estimates. Twelve strain gages are glued on the sample and reference in groups of three every 90 degrees. One of the four sample sides is sketched in Figure A.1. Gages on the 0° and 180°, and 90° and 270° faces are averaged into one strain during the acquisition recording. We later average the resulting 0°-180° and 90°-270° strains into one final strain estimate. Therefore, significant heterogeneity on one or more of the sample faces can affect the estimate of modulus and attenuation through this averaging process. It is possible that for sample 300, the textural heterogeneity evidenced in Figure 3.2 causes the high attenuation values plotted in Figure 3.8. Another possibility, however, is that because this sample has the largest permeability of the five measured samples, the high fluid mobility is responsible for the increased attenuation. The attenuation for sample 300 between 10-100 Hz is frequency dependent, increasing with increasing frequency.

In general, directly correlating attenuation between seismic and ultrasonic frequencies is not straightforward. At exploration seismic frequencies (10 to 100 Hz), all samples but sample 300 show that attenuation is frequency independent (Figures 3.6 to 3.10). However, for frequencies between 10 and 1000 Hz, attenuation estimates increase with increasing frequency for samples 100, 200 and 300. For a single relaxation mechanism, attenuation as a function of frequency is described by a bell-shaped curve with a peak which can depend on the rock permeability and fluid viscosity (Biot, 1956a,b; Mavko and Nur, 1979). Multiple attenuation mechanisms can result in multiple maxima. Our data (with a gap in observations between 10^3 and 10^6 Hz) does not require a more complex model than the most conservative model of a single relaxation mechanism. For the measured samples, the attenuation maximum may be located in the data gap between 10^3 and 10^6 in Figures 3.7c and 3.8, or at a frequency greater than 10^6 Hz in Figures 3.6, 3.7b and 3.7c.

3.7 Bulk modulus attenuation and fluids

Figure 3.12 shows the estimated bulk modulus attenuation as a function of frequency for all measured fluids and samples, for different fluids. Attenuation is non-zero when samples are dry (humidified) because a fluid layer on carbonate grains can increase attenuation compared to an oven-dry sample when the stress generated by the passing wave creates movement of the small amount of water (Clark et al., 1980; Murphy et al., 1986; Winkler and Nur, 1982). Winkler et al. (1979) analyze the effect of frictional loss with strain for sandstone samples, and show that strain becomes a loss mechanism for strain amplitudes greater than 10^{-6} . In their study, for dry or slightly saturated samples measured at strains between 10^{-7} and 10^{-6} , Young's attenuation is non-zero ($Q_E \approx 100$), but the loss mechanism is not attributed to friction. We do not attribute our observed attenuation to friction between grains because our measurements are at a single strain amplitude of 10^{-7} .

Attenuation estimates when the rock is dry or fully-saturated with butane are comparable for all samples, but we observe an increase in attenuation when brine replaces liquid butane. The changes from butane to brine resemble a reservoir fluid substitution process. We observe that the P-wave attenuation can increase by 150% to 400%, depending on the sample, when a brine replaces a light hydrocarbon in the pore space. By comparison, the P-wave velocity changes only 4% to 10% when brine substitutes liquid-butane in these samples. For a producing field and if time-lapse data is available, this attenuation sensitivity to fluids may be useful as an additional seismic attribute for dynamic reservoir monitoring.

Samples 100 and 200 have significantly different attenuation when saturated with brine, although petrographically they are similar. Sample 200 could have larger pore throats than sample 100, so that the induced pore pressure gradients responsible for attenuation will reach equilibrium differently for these two samples. It is important to keep in mind that our samples remain undisturbed in the apparatus during the exchange of fluids. There is no handling of the sample between attenuation measurements at varying saturating conditions. Therefore, even if the absolute values of the attenuation might have some bias, especially for sample 300, we expect this bias to be consistent between fluids.

We now describe possible reasons for the attenuation to increase when brine substitutes liquid butane. Fluid viscosity controls the frequency at which the maximum peak of attenuation occurs. At 3.5 MPa and 25°C, the viscosity of liquid butane and brine are similar, 0.2 cP and 1 cP, respectively. Nonetheless, the small viscosity difference between water and butane could still control the attenuation of the rock (Gautam, 2003; Best et al., 1994). Another possible reason for high attenuation in brine-saturated carbonates is related to the weakening of the rock frame upon water saturation.

Figure 2.4 shows the shear modulus as a function of differential pressure for sample C at 100 Hz and ultrasonic frequencies (0.8 MHz). We observe that as differential pressure decreases and compliant pores open, the rock shear modulus decreases (weakens) from dry (humidified) to fully-brine saturated. Therefore, there is a correlation between increasing



Figure 3.12. Bulk compressibility attenuation for all samples: a) 100, b) 200, c) 300, d) B and e) C. Dry measurements were performed with the samples humidified, not oven dry. Therefore, some water exists in the pore space. Butane and brine are for full-saturation. The differential pressures are as reported in Figures 3.6 to 3.10.



Figure 3.13. Shear modulus weakening with brine saturation for sample 300 at 100 Hz and 21 MPa differential pressure.

numbers of compliant pores and an increase in the shear modulus weakening. If the solid frame of the rock does not change, the rock shear modulus is not sensitive to the saturating fluid, because the shear modulus of fluids is zero. However, the modulus decrease due to brine can be explained by the weakening of the rock frame resulting from the interaction of a polar fluid such as water with the solid frame of the rock (Khazanehdari and Sothcott, 2003; Adam et al., 2006; Risnes et al., 2005). Rock surface energies or solid bounds between grains can break in the presence of water, weakening the rock. This process increases the pore space and can increase attenuation. Spencer (1981) measured Young's modulus attenuation in a fully-water saturated limestone at low frequencies (10-100 Hz), and showed that attenuation is caused by the reduction in surface energy in the grains (matrix softening). Vo-Thanh (1995) shows that the shear-wave attenuation depends on the amount of water saturation in limestone and sandstone. We observe that frame weakening is also dependent on the amount of water in the pore space. This rock frame softening is shown in Figure 3.13, where the shear modulus weakens by increasing brine saturation for sample 300.

An alternative attenuation mechanism can result from local fluid flow between the gap in the grain contact and the adjacent pores (Murphy et al., 1986). This wave attenuation is dependent on the saturation state of the rock, and the frequency at which the peak of maximum attenuation occurs. The latter is largely a function of the fluid and the size of the grain contact gap. As the aspect ratio of this gap decreases, the attenuation maximum moves towards lower frequencies. After brine saturation, grain contacts could have lower aspect ratios (solid bounds break), which under the grain contact loss mechanism means that the maximum attenuation peak shifts to lower frequencies. For our limited low frequency interval, a peak is hard to define. Nonetheless, the attenuation maximum shift to lower frequencies can be observed as an increase in attenuation (see for example Figure 3.12 b). Finally, for our samples, brine is not believed to be dissolving the carbonate grains, because Figure 2.4 shows that the frame weakening process is reversible (black circles). In conclusion, the increase of soft regions in the rock due to brine (newly opened grain contacts compared to dry or butane-saturated carbonates), can increase the differential fluid-rock movements, and therefore attenuation.

3.8 Numerical modeling

In previous sections we have suggested reasons why $Q_K^{-1} > Q_S^{-1}$ in our carbonate samples. Here we support these interpretations of measured attenuation with numerical modeling constrained by our estimated values of elastic modulus and velocities. The amplitude and phase of our measured waveforms provide estimates of the elastic moduli and attenuation, respectively. These two attributes can be assumed independent. For an elastic sample, the strain amplitude yields moduli and the strain phase shifts are zero. That same sample under different saturation conditions can behave visco-elastically, with the same strain amplitude as for the elastic sample, but with a non-zero phase shift. Based on that strain amplitude is independent of phase, we model attenuation from the measured modulus and show that $Q_K^{-1} > Q_S^{-1}$.

The modulus' frequency dependence in a material is directly related to the attenuation. In the extreme case of no modulus dispersion, there is no attenuation; which is a fundamental relation described by the Kramers-Kronig equations. In practice, Cole and Cole (1941) obtain an equation that describes the complex modulus (M^*) dispersion as a function of frequency (w):

$$M^*(w) = \frac{(M_0 - M_\infty)}{1 + (iw\tau_0)^{1-\alpha}} + M_\infty,$$
(3.7)

where M_0 and M_{∞} are the moduli at zero and infinite frequency, respectively. τ_0 is the relaxation time and α is a parameter that controls the curvature of the S-shaped modulus dispersion. α is bounded between 0 and 1, and if $\alpha = 0$, equation 3.7 reduces to the modulus dispersion relation for a standard linear solid or Debye model. The α value controls the amplitude of the attenuation peak.

We use the Cole and Cole (1941) equations to model attenuation under the assumption that there is only one relaxation mechanism (*i.e* one attenuation peak) between 10 and 10^6 Hz. We model the imaginary part of the complex modulus by applying a least-squares fit to the real part of the measured modulus. We use the 3 Hz data as M_0 , the ultrasonic data point as M_{∞} , and τ_0 and α follow from the best fit to the modulus vs. frequency data. We model attenuation by taking the ratio of the imaginary and real parts of the modeled complex modulus. The solid lines in Figures 3.14a and Figures 3.14b show the best fit to the real bulk modulus and the modelled attenuation for sample 100 saturated with brine.

Our measurements cover a large range of frequencies, but within three orders of mag-



Figure 3.14. Bulk modulus and attenuation for sample 100 fully-brine saturated. Dots are the measured data and the lines are the modeled moduli and attenuation using the Cole-Cole relation.

nitude $(10^3 - 10^6)$ we only have one datum. Modeling attenuation as a function of frequency is highly sensitive to this ultrasonic datum. To show this sensitivity we vary the ultrasonic data point. 20 ultrasonic bulk moduli are generated from a Gaussian distribution of the ultrasonic errors. The dashed lines in Figure 3.14a are the best fits for each of the 20 ultrasonic points. The corresponding modeled attenuation are plotted as dashed lines in Figure 3.14b. For the large range of possible ultrasonic bulk modulus, the modeled attenuation peak (1000/Q) can vary from 25 to 75. Therefore, quantitatively estimating Q from measured bulk modulus is sensitive to the uncertainty in the ultrasonic data.

Figure 3.15 shows that the modeled bulk attenuation is greater than the shear-wave attenuation for the modeled brine-saturated samples. Sample B is not modeled because the ultrasonic bulk modulus is slightly lower than the low frequency data; probably because the ultrasonic pulse is slowed down due to scattering in the horizontal layers in this sample. The same observation holds for when the samples are dry or fully-saturated with liquid-butane. The fact that bulk losses dominate over shear losses is in agreement with the experimental attenuation data shown in Figures 3.6 to 3.10.



Figure 3.15. Modeled bulk and shear-wave attenuation for fully-brine saturated samples: a) 100, b) 200, c) 300 and d) C.

3.9 Conclusions

In contrast to clastics, our measurements in five carbonate rocks show that bulk modulus attenuation and compressional-wave attenuation are greater than the shear-wave attenuation when samples are fully-saturated with either liquid butane or brine at reservoir conditions. This observation is consistent over a large range in frequencies (10 Hz-1000 Hz, and 0.8 MHz) and is repeatable using two different experimental methodologies. Attenuation is observed to be frequency independent for the exploration seismic frequencies (10-100 Hz) for four of five samples. From published laboratory data, the saturation state of sandstones could be interpreted based on whether the bulk or shear losses dominate; however, from our observations, this correlation will probably not be straight-forward for carbonates. The sensitivity of attenuation, particularly the bulk modulus losses, to the type of fluid in the pore space is much greater than for velocities. Bulk modulus attenuation could change from 150% to 400% (depending on the sample) when brine replaces a light hydrocarbon. By comparison, the average compressional-wave velocity changes by 7% for our samples.

The laboratory measurements described here could aid in the interpretation of surface seismic data. Even thought estimating intrinsic attenuation from surface seismic still faces some challenges, the recent improvements in data quality should make this estimation more reliable.

CHAPTER 4

COMPARISON OF SEISMIC AND ULTRASONIC FREQUENCIES

Rocks saturated with a fluid can be described as visco-elastic materials. The velocity and elastic moduli of visco-elastic materials increase with frequency. Therefore, the elastic rock properties that we measure at high frequency might not resemble the observations at lower frequencies. Laboratory measurements of velocity and elastic moduli are mostly performed at frequencies higher than the exploration surface seismic data, but with the stress-strain experimental procedure the moduli and velocity of laboratory samples can be measured at seismic frequencies. In this study we compare measurements at 10 Hz and 0.8 MHz on eleven carbonate samples from the Middle East at reservoir differential pressures. We compare how the measurements at these two frequencies probe the sample. We make observations on the dispersion of the different rock properties and its effect when performing fluid substitution with Gassmann's relation. And finally, we will show that for these samples the variation of elastic properties from low to high frequencies is significant; but if the ratio or difference between dispersive parameters is analyzed this difference between measurements at different frequencies can be reduced.

The carbonate samples are measured dry and fully-saturated with either a light hydrocarbon (liquid butane) or brine (180,000 ppm NaCl) at reservoir differential pressures. Differential pressure is the difference between confining and pore pressure. The samples have a range of porosity, permeability and textures (Table 2.1 and 3.1). The moduli and velocities at seismic frequencies (10 Hz) are measured by applying a sinusoidal stress to the rock and measuring the resulting strain in different directions on the rock sample and on the reference material (aluminum). The measured strain amplitudes are at the same scale as for seismic waves ($\sim 10^{-7}$). At ultrasonic frequencies (0.8 MHz), we measure the time of flight of a wave transmitted through the rock sample. From this time we estimate the Pand S-wave velocities and from these the rock moduli.

The core samples belong to two different reservoirs with a differential pressure of about 35 MPa. Velocity and elastic modulus are measured at a differential pressure (Pd) of 31 MPa for all samples except for sample L, which is measured at a Pd of 24 MPa. For all rocks the pore pressure was constant at 3.5 MPa. The stress-strain system in the laboratory is pressurized with nitrogen gas, but for safety reasons the system is not able to reach the reservoirs confining pressure. However, the pressures in the experimental setup are close to the reservoirs differential pressure. Most samples are initially humidified to avoid the softening matrix effect of initial introduction of moisture in the pore space. Small amounts of water (less than 1% of the pore volume) can reduce the bulk and shear moduli of the

rock significantly. The data in this paper is the estimated mean (markers in the plots) and one standard deviation of the random error (error bars on the markers). These values are obtained from a statistical analysis between the measured experimental data and an assumed *true* model. Throughout the paper, the comparison in modulus and velocity are between the mean values.

Figure 3.5 shows the P-wave velocity for two of the samples over a broad frequency band. Three representative points are presented for the low frequency data (3, 100 and 1000 Hz). The ultrasonic data is collected simultaneously with the low frequency data at 800 KHz. The data at 10^4 Hz is obtained from the P-wave sonic log at the same wells and depths from where the rock samples were cored. This velocity is an average of the velocities from the sonic log over a 0.6 meter window centered at the sample depth. Observe in Figure 3.5 that the velocity estimated at different frequencies is dispersive: the P-wave velocity consistently increases with frequency.

The observation that the velocity at ultrasonic frequency is higher than at seismic frequency can be explained by visco-elasticity of the material. Alternatively, the ultrasonic wave could be path dependent propagating through the fastest path. This preferential higher velocity path would be a lower compressibility region of the rock. For example, for partial saturation, the fluid-saturated region is less compressible than the dry region in the rock. However, the ultrasonic data could also be influenced by the combination of both visco-elasticity and path dependence.

There are several rock-fluid mechanisms that describe the visco-elasticity of porous media. Two of the most common visco-elastic theories are the Biot and squirt flow mechanisms. For both models the modulus increases with frequency. Briefly, the passage of a wave stresses the rock creating a fluid movement relative to the rock frame. The fluid is then in an unrelaxed state and would need time to return to its original condition (pre-stress condition). Probing a rock at different frequencies translates into different times that the fluid has to relax. Therefore, the rock properties vary with respect to whether the fluid has had enough time to relax (low frequencies), is completely unrelaxed (high frequencies) or somewhere in between. The specific values of low and high frequencies are dependent on the fluid and rock properties such as viscosity and permeability among others.

One of the goals in reservoir characterization is to quantify the fluid content of the rock. The bulk modulus is an intuitive choice to predict fluid content as it describes how resistant a rock is to being compressed, which depends on the amount of fluid in the pore space. Figure 3.11 shows a partial saturation experiment on sample L and compares the bulk modulus estimate at low and high frequency. The bulk modulus increases with the amount of fluid in the pore space. First, observe how the ultrasonic data becomes sensitive to the brine in the pore space after 20% saturation, while the 10 Hz data does not show stiffening of the rock until 80% saturation. The fluid sensitivity at low saturation can be explained by the ultrasonic wave following a preferential path through the saturated regions in the rock. At 10 Hz, the bulk modulus remains equal to the dry bulk modulus

until most of the pore space is filled with fluid. The stress-strain experiment measures the deformation of the rock sample as a whole. Stress is evenly applied on the rock and the measured strain is the average of the different locations of the strain sensors. Usually, this type of setup averages the dry and saturated regions in the rock. Therefore, measurements at seismic frequencies are not significantly affected by patchy saturation. The observations at 10 Hz are in agreement with producing reservoirs where a small amount of gas coming out of solution ($\sim 5\%$) drops the bulk modulus of the rock close to the dry condition. Dry samples do not exhibit significant dispersion because there is practically no fluid to produce modulus dispersion. The pink arrow at zero saturation (dry) shows the difference between 10 Hz and ultrasonic bulk moduli. The higher bulk modulus at ultrasonic frequency is probably caused by a preferential path in the sample rather than modulus dispersion. This sample is significantly heterogeneous with vertical bands of different textures and pore space observed in the x-ray tomographic scan (CT-scan). Most of the measured samples were overall homogeneous. The same size arrow is overlaid at 100% brine saturation and observe that in this case the difference between 10 Hz and ultrasonic is greater. The remaining difference (green arrow) is possibly caused by rock-fluid mechanisms that cause bulk modulus dispersion.

Figure 4.1 shows the bulk and shear modulus for all the samples 100% brine saturated. The dashed line represents that seismic (10 Hz) and ultrasonic moduli are equal. Observe that for most samples the ultrasonic moduli are greater than the seismic moduli. On average, the moduli increase from 10 Hz to ultrasonic by 23% and 12% for bulk and shear moduli respectively. The isotropic P-wave velocity is the combination of the bulk and shear moduli and the rock density. When performing fluid substitution, there will be a trade-off among these parameters affecting the velocity. Figure 5.4 is the P- and S- wave velocity for all samples fully-saturated with brine at high differential pressure. As expected, the velocity at ultrasonic frequency is greater than the seismic frequency. However, the velocity increase from low to high frequency data is on average 9.5% and 5.8% for the P- and Swave, respectively. These lower dispersion values with respect to the moduli results from the trade-offs among rock properties (i.e. weakening and strengthening of the shear modulus, see chapter 2) and the definition of the isotropic P-wave velocity (mostly because of the square root in its equation). Nevertheless, the observations that the moduli as well as the velocities are dispersive for all samples fully-saturated with brine hold for the case of full-saturation with liquid butane.

If both the P- and S-wave velocities are dispersive, what about the Vp/Vs ratio? Figure 4.3 shows the Vp/Vs ratio for all samples 100% saturated with butane and brine. For brine, the average increase in the Vp/Vs ratio from 10 Hz to ultrasonic is 2.6%; and for butane is 1.9%. Therefore, the velocity dependence on frequency is significantly reduced by taking the ratio of two dispersive parameters.

Laboratory data can be used to perform time-lapse feasibility studies. For example, from the measured velocities we can compute the impedance and the reflection coefficient to



Figure 4.1. Carbonate samples bulk and shear moduli at 31 MPa differential pressure and fully brine saturated. The plot compares measurements at 10 Hz and 0.8 MHz. The dashed line represents where the moduli at both frequencies are equal.



Figure 4.2. Carbonate samples P- and S-wave velocities at 31 MPa differential pressure and fully brine saturated. The plot compares measurements at 10 Hz and 0.8 MHz. The dashed line represents where the moduli at both frequencies are equal.

analyze the possible changes in amplitudes resulting from enhanced oil recovery processes in the reservoir. We compute how much would the zero-offset P-wave reflection coefficient (RC) change from 10 Hz to 0.8 MHz data based on the laboratory measurements. We assume that the whole reservoir is represented by the rock properties of one sample and is overlaid by a sandstone layer with a P-wave velocity of 4 km/s and bulk density equal to 2.4 g/cm³ (Figure 4.4). For a specific fluid, the absolute change in the reflection coefficient amplitude (e.g. $|RC_{brine,10Hz} - RC_{brine,0.8MHz}|$) from seismic to ultrasonic frequency is significant and is in average 0.05 for brine and 0.04 for butane (see for example the arrows in Figure 4.4, where the arrow size represents this difference for sample L). Let us call the variation in the reflection coefficient from but ne to brine ΔRC - think of a time-lapse experiment (e.g. $\Delta RC_{10Hz} = RC_{brine,10Hz} - RC_{butane,10Hz}$). If we now look at the absolute values of ΔRC at 10 Hz and at 0.8 MHz we observe that the difference between seismic and ultrasonic frequencies is smaller than when studying the RC for one fluid (see for example the brackets in Figure 4.4, where the bracket size represents this difference for sample L). The difference in ΔRC between 10 Hz and 0.8 MHz is on average 0.01. We have to take the absolute value of ΔRC as the RC sign is controlled by the choice of the properties of the top layer. What we can say so far is that for these samples the ultrasonic data agrees with the seismic frequency data if the geophysical property is an outcome of the ratio (Vp/Vs) or the difference (ΔRC) of dispersive rock properties. However, when performing quantitative analysis in time-lapse reservoir characterization, our interest focuses on the fluids and its contribution into the rock modulus or velocity. The most well known equation to study the sensitivity of the rock to fluids is Gassmann's equation (Equation 2.1)

One of the applications for rock physics laboratory data is to verify the validity of Gassmann's relation under different conditions. In the laboratory we usually measure the modulus for the sample dry (K_{dry}) , and saturated with a fluid (K_{sat}) . In chapter 2 we discuss how especially at low differential pressures the ultrasonic bulk modulus seems better predicted by Gassmann's relation than 100 Hz data. However, Gassmann's theory was derived for zero frequency. Therefore, interpreting whether Gassmann is appropriate for carbonates rocks at ultrasonic frequencies could be biased if the measured data in the MHz range is dispersive with respect to the low frequencies or even more appropriate, to zero frequency.

The trade-off between bulk and shear moduli and density in the isotropic P-wave velocity equation is also important to consider when studying Gassmann's relation. For example, at reservoir differential pressures and 10 Hz, the bulk modulus deviates in average by 14.1% from perfect prediction of Gassmann to the measured data. The absolute minimum and maximum deviations for all samples are 0.7% and 35.7%, respectively. If we now analyze the measured and Gassmann estimated P-wave velocity, the average deviation from perfect prediction is 4.5%, while the absolute minimum and maximum deviations are 0.02% and 10.3%, respectively. Typically, a 5% deviation in velocity may be considered low, and therefore most interpretations would be that Gassmann's theory is valid for these velocity



Figure 4.3. Vp/Vs ratio for the measured samples at 31 MPa differential pressure and fully butane and brine saturated. The plot compares measurements at 10 Hz (squares) and 0.8 MHz (circles).



Figure 4.4. Zero offset P-wave reflection coefficient for a two layer reservoir, for which the bottom layer is represented by the measured samples at 31 MPa differential pressure and either fully saturated with butane (green) or brine (blue). The plot compares measurements at 10 Hz (squares) and 0.8 MHz (circles).

measurements. However, Gassmann's theory was developed for the bulk modulus and zero frequency, and in this domain the deviation is close to 15% for our measured carbonates. Therefore, the variation in the modulus from low to high frequencies is more significant than when studying velocities alone. Although in geophysics our main data are velocity and amplitude changes, Gassmann's applicability should probably be analyzed through the bulk modulus rather than the velocity.

In reality our prime goal in interpreting seismic velocity changes is to identify pore fluid type so that Gassmann's equation (2.1) can be used to invert for the fluid bulk modulus K_{fl} . A 5% systematic error in velocity or 15% in rock bulk modulus would correspond to a 50% and 200% over-prediction in K_{fl} for brine and a light hydrocarbon respectively. Thus if we are seeking information on fluid types, dispersion is a serious issue to consider.
CHAPTER 5

FIBER-OPTIC STRAINMETER

5.1 Introduction

Laboratory calibration of the effects of fluid saturation on the seismic signature should be done at seismic frequencies and amplitudes. Stress-strain techniques can be used to measure moduli at low frequencies and amplitudes. The experimental data can also be used to estimate seismic wave intrinsic attenuation (inverse of the quality factor - Q). The preferential loss of wave amplitude with frequency can be an indicator of fluids in the pore space and, rock permeability and saturation. One of the major challenges when measuring core samples is rock heterogeneity.

Our experimental approach measures strains in the sample and a calibrating material (aluminum) while applying a sinusoidal stress for a range of frequencies - from less than 1Hz to a few KHz - as described by Batzle et al. (2001, 2006). Deformation (strain) is used to estimate moduli and, combined with the density yields velocity. The rock is assumed isotropic and linearly elastic meaning that the applied stress on the sample and aluminum standard must be equal. Because the intrinsic attenuation of aluminum is low, the phase of the strain on the aluminum represents the phase of the applied stress. The phase lag between the strain on the aluminum and the measured strain on the rock is directly related to seismic attenuation. Avoiding heterogeneities (grains, vugs, mineralizations, etc.) is a challenge when using semi-conductor strain gages (dimension $\sim 1x5$ mm) to estimate rock properties in a stress-strain experiment. This research is based on using fiber-optic cable as a strain sensor rather than the small strain gages.

A joint research program was begun with the Colorado School of Mines and Micro-G Solutions (now Micro-g LaCoste) to develop a fiber-optic strainmeter (FOS) to provide spatially averaged properties as well as higher accuracy. Using the FOS as the acquisition system for the stress-strain experiment setup would sample bulk rather than point properties as the fiber can be wrapped around the sample to average heterogeneities. Increased accuracy and sensitivity when measuring rock deformations undergoing an applied sinusoidal stress will permit better phase discrimination between stress and strain providing better measures of attenuation.

A few previous studies describe systems to measure strain in geo-mechanical applications for rock or structure failure. Butter and Hocker (1978) estimate strains on a cantilever bar by counting diffraction fringes, while Schmidt-Hattenberger et al. (2003) approach geomechanical problems by measuring strains via a Bragg grating fiber. These two systems 60

have been successful for their applications. However, the experimental strain has to be in the order of 10^{-7} to be able to compare laboratory data to field seismic observations. The strain magnitudes of the two systems previously described are large and their sensitivity are not as high as the FOS be used for exploration rock physics.

The research on the fiber-optic strainmeter has gone through several stages of experimental setup as well as trial-and-error. Attenuation has been measured with limited success on a Plexiglas sample. Further tests where focused on improving the system. The largest challenge in measuring moduli and attenuation using fiber-optic is at the same time its largest advantage: high sensitivity to deformations. In the next sections we summarize the different setups and a preliminary test on a Plexiglas sample. The last section focuses on ideas in the road ahead for the development of this technology.

5.2 Setup 1

Figure 5.1 shows the initial setup for the fiber-optic strainmeter in a Michelson interferometer geometry. The light of an HeNe laser (red light, $\lambda = 633 \eta m$) is split into two fiber-optical cables. The unjacketed sections of the fibers (cladding) are wrapped and glued around a sample and a piezo electric device, both of cylindrical shape. The end of each fiber is terminated with a high reflectivity mirror so that, ideally, all of the light is reflected back to the fiber splitter. From there, half of the light intensity is sent back towards the laser source but is blocked by an optical isolator (not shown). The other half of the recombined light is directed to a detector inside the compensator (See Appendix B for a description of the compensator). In theory, when both fiber arms are the same length and unstrained, the light detector will record the maximum intensity I_{max} . In practice the fiber arms will not be identical, producing a lower combined light intensity I_o (Figure 5.3). By changing the fiber length in one arm, the light intensity resulting from combining the light from the two fiber arms interferes constructively and destructively. In Figure 5.3, the horizontal axis is the light path disturbance, meaning how much differential distance the light travels through the strained fiber. The vertical axis is the light interference from the combined arms. When stress is applied to the sample the shape changes, straining (extending) the fiber. For identical length arms and when the path increases by $n\lambda/2$ n is an integer) the light from the two arms interferes destructively giving a minimum intensity. But when the path increases by $n\lambda$ the interference is fully constructive producing a maximum intensity (I_{max}) . This light intensity pattern are the fringes sketched in Figure 5.3. For this setup with mirrors, if the fiber length changes by $n\lambda/4$ means a path disturbance of $n\lambda/2$ because the mirrors reflect the light, thus traveling twice through the stretched fiber. For a different setup not involving mirrors (discussed latter), the path disturbance and fiber length change are the same.

Lets assume our non-identical and undisturbed fiber arms interfere to create I_o . For example, consider the case where one fiber is wrapped around a cylindrical test sample



Figure 5.1. First setup of the fiber-optic strainmeter. The solid lines are the fiber cables jacketed and unjacketed - at the sections glued to the sample and piezo. The dashed line is the voltage input to the piezo crystal produced by the compensator.



Figure 5.2. Fiber-optic cable cross-section. The glass cladding is coated with acrylate and the light propagates through the core. The fiber can have several buffers and jackets. (Source: www.wikipedia.org)

(e.g. Plexiglas sample). Applying a stress to the sample creates a small strain and the fiber stretches decreasing the light intensity. When the driving force (stress) is sinusoidal, the intensity will decrease and increase from I_o (represented by the arrow in Figure 5.3).

If we drive the sample such that the response is less than one fringe, the expected signal is phase locked to the driving signal and can therefore be extracted with a phase sensitive detector (e.g. lock-in amplifier). We can lock the interferometer by providing a compensating strain on one of the arms while applying stress to the other fiber arm. In this case, the bandwidth of this compensation servo (the compensator) can be large and encompass the signal frequency so that there is no change observable on the interferometer output. The signal is then given by the deformation required to lock the interferometer at a certain light intensity. In our experiments the compensating deformation is given by applying a voltage to a piezo electric crystal, expanding this device to match the sample strain. The applied voltage can be either multiplied by different factors to provide deformation units or measured on a reference material (aluminum) in mechanical series with the sample and estimate strain ratios. The second approach is how moduli and attenuation are estimated



Figure 5.3. Fringe sketch describing the constructive and destructive interference of light resulting from varying the distance the light travels between two fiber arms (path disturbance).

using strain gages. Modulus and velocity (strain amplitude) depends on the amount of fiber glued to the sample (i.e. light intensity). While, attenuation is not dependent on the amount of fiber in contact with the sample or light intensity, but is related to the phase shift. This means there is a greater advantage of measuring seismic wave attenuation over velocity.

5.2.1 Velocity and attenuation on a Plexiglas sample

As a preliminary test to the system, three fibers are glued to a Plexiglas sample and an aluminum standard. Two fibers are wrapped around the Plexiglas and the aluminum standard, and one is set vertically on the Plexiglas. Only one of the fibers in the sample or the aluminum can be measured at a time, as there is only one free arm to be compensated by the piezo (Figure 5.1). The Plexiglas sample is 2.5 cm in diameter and 8 cm in length.

The Plexiglas (sample) was set up in the same shaker (apparatus that applies axial sinusoidal stress) used for low frequency strain gage measurements Batzle et al. (2001). The sample was measured for a range of seismic frequencies on a laboratory bench at room pressure and temperature. Both, strain gage and fiber deformation where recorded for this setup. The measured deformations are used to calculate the moduli of the material. In turn velocities are calculated from the density and moduli. The comparison of P- and S- wave velocities using strain gages, fiber-optic, and a propagating ultrasonic wave are shown in Figure 5.4. The velocity dependence on frequency is opposite when comparing measurements from strain gages and fiber-optic, but for this first test the overall bulk values are encouraging.

Extensional attenuation (Young, E) is computed from the phase difference between the sinusoidal vertical strain on the Plexiglas and the aluminum. The strain phase angle is estimated with a lock-in amplifier. Figure 5.5 shows the extensional mode attenuation estimated from fiber-optic and semi-conductor stain gages. The fiber-optic measurements are performed at room temperature ($\sim 30^{\circ}$ C). Different Plexiglas samples can have differences in their composition and curing history giving different attenuation values. For the attenuation data, the strain gage measurements correspond to a different Plexiglas sample and were measured at 25°C and 40°C. A better correlation between attenuation estimated with fiber-optic and strain gages is observed than that for velocity. This is probably because the phase estimation is not dependent on the fiber length in contact with the sample. Also the quality of the light intensity affects the magnitude of the strain amplitude and thus the velocity.

We list a couple of possible sources of error in the strain measurements: 1) Not enough light was received at the detector (mirrors did not reflect 100%). 2) The fiber length glued to the sample is used to normalize deformations, and the estimated modulus is sensitive to this contact length. 3) These measurements were performed on a regular bench that is not isolated from vibrations that can be generated by walking, wind, etc, as an optics bench would be.

A dry Berea sandstone sample was prepared with fiber and strain gages to perform measurements similar to the Plexiglas, but the efforts to estimate strain and attenuation where unsuccessful. The fringes were unstable and the compensator would not lock to an intensity, probably because there was not enough light. We did record the phase shift for the few times the compensator did lock, estimating a phase shift at about 6 degrees ($Q_E=10$), compared to 0.4 degrees ($Q_E \sim 120$) as predicted by strain gages (on the same sample) and the literature Toksoz et al. (1979). Varying the input voltage to generate the stress from 0.05 V, 0.1 V and 0.2 V did not change the phase estimate significantly, always obtaining the phase shift in the order of several degrees.

5.3 Setup 2

Our first experiment shows that attenuation estimates broadly correspond to those obtained with strain-gages, and that velocities are probably affected by light intensity or fiber contact length. To increase the light intensity we modified the interferometer setup to a Mach-Zehnder geometry. Figure 5.6 shows the first of three approaches for this type of interferometer. The light intensity increased, however the fiber splitters are very sensitive to motion. We then isolate the system on an optical bench and, greatly reduced the noise generated by the splitter. Because of logistics from this point on we use a small shaker compared to the one in the previous section. We first try the setup as in Figure 5.6 with a Plexiglas sample. Since the compensator was not functioning, therefore we switch to a light detector that converts the combined light intensity into voltage.



Figure 5.4. Compressional and shear wave velocity as a function of frequency for a Plexiglas. The sample is measured with semi-conductor strain gages and fiber-optic cable following a stress-strain experimental procedure. The data at 0.8 MHz is obtained from an ultrasonic pulse propagation technique. The three measurements are performed on the same Plexiglas sample.



Figure 5.5. Extensional mode attenuation as a function of frequency on two Plexiglas samples. One was measured with semi-conductor strain gages, while the other sample was measured with fiber-optic cable. Observe that the attenuation peak is sensitive to temperature. The absolute values of attenuation can dependent on the Plexiglas composition.



Figure 5.6. Modified setup for the FOS to a Mach-Zehnder geometry.



Figure 5.7. Input signal (stress) and recorded fringes in the vertical direction of a strained Plexiglas sample. The driving frequency is 300 Hz. The two plots are recorded at different times.

The main challenge in measuring strain with this system is fringe instability. Over a short time (few periods of the imposed strain) the fringes are sinusoidal with low distortion at the frequency of the applied stress (Figure 5.7 left). However, for the same acquisition just a short period later (Figure 5.7 right), the overall fringe intensity decreases and the phase flip 180°. To estimate amplitude and phase we need to average many cycles. This light instability over short period of time is the major challenge for the interferometer. The light intensity coming out of the (warmed) laser is stable. Ambient factors (e.g. temperature, wind) are probably responsible for these fringe instability. The fiber is isolated on an optical table, but the sensitivity to external factors is still significant. As we will see later, we can isolate the fiber into an aluminum chamber to decrease these effects.

The unjacketed fiber-optic cables are delicate and break easily. We begin repairing using the fiber-optic fusion splicer provided by Dr. Kowalski (Department of Physics, Colorado School of Mines). However, controlling the fiber length after repair is hard, and therefore the arms of the interferometer are not the same length. When the interferometer is undisturbed (unstressed) the light intensity could be at its minimum (Figure 5.3) if



Figure 5.8. Third setup for the FOS to estimate the maximum light intensity by stretching the fiber cable with the sample to the same (interference) length as the cable with the aluminum.

the length of the fiber is so that the merging light from both arms interfere destructively. Before any stress-strain measurements we first would like to increase the light intensity of the undisturbed fiber. We modify the interferometer as shown in Figure 5.8 to try to estimate the maximum light intensity. Here we use the piezo not as a compensator but as a fiber *stretcher*. By applying a voltage to the piezo, we expand its diameter so that both arms interfere to a maximum intensity. When we expand the fiber by more than one wavelength we jump from one fringe to the next. Therefore to find the light intensity maximum for a single fringe we have to expand the fiber by a maximum of 633 η m (one wavelength). For a cylindrical piezo crystal the relation between the change in the diameter (δd) and the applied voltage (V) is:

$$\delta d = D_{31} \frac{d}{T} V \tag{5.1}$$

where d is the piezo diameter, D_{31} is the piezo coefficient and T is the piezo thickness. The fiber is wrapped N times around the piezo crystal so we convert δd into δl (fiber length change) by: $\delta l = \pi \delta dN$ - assuming that the fiber expands exactly as the diameter of the piezo. The factors for this particular piezo crystal and fiber are: $D_{31} = -26410^{-12}m/V$, d = 0.0192m, T = 0.0037m, N = 17. To stay within one fringe, the maximum expansion for δl should be less than 633 η m, which translated into a maximum applied voltage of 8.65 V.

To estimate the maximum fringe intensity and for the setup as shown in Figure 5.8, a sinusoidal voltage from 1 V to 17 V (peak-to-peak voltage, the amplitude of the sine function is half this voltage) is applied to the piezo crystal alone while maintaining the sample and aluminum undisturbed. The recorded maximum fringe amplitude as a function of voltage applied to the piezo crystal is plotted in Figure 5.9. Two measurements are taken at each input voltage with a time difference of about 3 minutes. First note that there is no systematic increase in the amplitude of the fringe as we sweep through piezo voltage. Second,



Figure 5.9. Fringe amplitude as a function of the voltage applied to the piezo. The experiment consists of estimating the maximum fringe amplitude by expanding the piezo no more than λ .

for the same piezo voltage but different recording times, the fringe amplitude is largely variable. This probably results again from the instability of the amplitude of fringes over several cycles. To try to reduce the fringe amplitude variations the entire interferometer is setup inside a large aluminum chamber with a lid. The experiment to obtain the maximum fringe amplitude was repeated without significant improvement from the observations in Figure 5.9.

Thermocouples are used to measure temperature over time inside the aluminum chamber with the laser on and off, but no temperatures fluctuations were found that could account for the fringe instability.

5.4 The road ahead

Though out the process of building and testing the fiber-optic strainmeter we were successful in observing fringes at the same frequency as the input signal (stress). Velocity and attenuation was overall in agreement with data from strain gages from a preliminary test on a Plexiglas sample. We focused our efforts to compute attenuation rather than moduli because phase angle estimation is not dependent on the amount of fiber in contact with the fiber. Fringe instability is probably not resulting from air movement or ground vibrations, but probably due to temperature gradients and variable light polarization. Fusing fiberoptic cable is necessary not only for fixing broken fiber, but also in the future to implement the system in a pressure vessel. Having a compensator (feed-back loop) to record the strain data has its advantages (directly record deformation) and complications (does not lock if fringes are unstable).

The biggest challenge is to stabilize the fringes or to find a correction. Thermally isolating the system and using polarized fiber can be tested. When measuring strain on the aluminum and sample we have three fiber cables. If three systems of laser/detector combos are available it is possible that the ambient effects will be reduced because the measurements are performed simultaneously. The fist test on the Plexiglas sample was performed on the large shaker used to measure strain gages, later we switched to a smaller shaker. Evaluating the accuracy and efficiency of the small shaker is important to address.

Young's attenuation could be measured directly with a setup as in Figure 5.8 but without the piezo. The two arms of the fiber are glued vertically on the aluminum and sample. From the interference fringes we would like to directly estimate the phase shift (ϕ) between the strain on the aluminum and the strain on the Plexiglas sample $(Q_E^{-1} \approx \phi)$. The combined light intensity from the two interferometer arms is: $I = E_1^2 + E_2^2 + 2E_1E_2\cos\alpha$, where E_1 and E_2 are the field of the electromagnetic waves from each interferometer arm. For the fiber-optic interferometer the phase α is the combination of different phase shifts: $\alpha = \Delta \theta + \delta + \phi$. If the light coming out of each arm are not identically polarized we have the contribution of $\Delta \theta$: the angle between the light polarization of each arm. The single mode fiber used does not guarantee that the light from each arm has the same polarization. δ is the shift due to the electronic components in the setup and finally ϕ is the phase shift due to intrinsic attenuation. Although some preliminary tests were performed, additional equipment is required to estimate attenuation with this method.

CHAPTER 6

TIME-LAPSE Q

6.1 Introduction

Production and enhanced oil recovery (EOR) over time create changes in the reservoir, which are monitored mostly by studying amplitude (reflection coefficient, AVO), velocity or travel time differences. However, changes in seismic wave attenuation can be significantly greater than for these rock properties as a result of fluid substitution in the reservoir. To monitor variations in reservoir rock properties due to fluid production or injection, seismic surveys are acquired separately over time. These data sets are called time-lapse or 4D seismic surveys. In this chapter we study changes in seismic wave attenuation for a reservoir undergoing EOR.

Intrinsic attenuation is the loss of amplitude due to elastic energy conversion into heat as a wave passes through a medium. Fluid in, saturation and permeability of a rock is related to intrinsic attenuation. Scattering attenuation is the loss of seismic energy when waves deflect out of the direction of propagation. In this chapter we create synthetic seismograms to study contributions of scattering and intrinsic attenuation for time-lapse attenuation studies. We base our modeling of synthetic seismograms on measured laboratory data on carbonate rocks (chapter 3). Our experimentally measured data show that intrinsic attenuation can increase on average by 275% when brine replaces a light hydrocarbon in these rocks. By comparison, the average compressional-wave velocity changes by 7% for measured carbonate samples. At seismic frequencies, in the laboratory we measure only the intrinsic attenuation, but for field measurements the amplitude of a propagating seismic wave is attenuated by both intrinsic and scattering mechanisms.

Estimating attenuation from seismic data, whether it is from a vertical seismic profile (VSP) or surface seismic survey, gives an apparent attenuation. This type of attenuation is the combination of scattering and intrinsic losses. In this chapter attenuation is defined as the inverse of the quality factor Q, and we use the subscripts s for scattering, i for intrinsic and a for apparent. Estimating Q from seismic data has gained popularity in recent years. Dasgupta and Clark (1998) and Clark et al. (2001) present an innovative approach to estimating Q_a at zero-offset from surface seismic data by analyzing Q_a versus offsets. They find qualitative correlations between gas coming out of solution and Q_a changes for a North Sea reservoir. Zhang and Ulrych (2002) describe a methodology to estimate Q_a from CMP gathers and frequency shift. Azimuthally anisotropic Q_a from CMP gathers is studied by Vasconcelos and Jenner (2005). Changes in apparent attenuation can also

be estimated from a zero-offset VSP geometry. For this acquisition geometry, Michaud (2001) correlates changes in P- and two S-wave apparent attenuations to fluid mobility and closure of cracks. Several studies describe ways to separate intrinsic from scattering attenuation (Parra et al., 2007; Mateeva, 2003; van Wijk, 2003; van Wijk et al., 2004; Fogg, 1995). However, intrinsic attenuation is estimated by modeling the scattering loss and subtracting this loss from the apparent attenuation. In other cases, specific data is acquired to separate both attenuations (e.g., full waveform sonic log or seismograms including the coda). Because attenuation can be described as a preferential loss of frequencies with distance for a propagating wave, other techniques (e.g. wavelet transforms) are available to extract frequency information and relate the frequency losses to attenuation. However, these methods often rely on dictionaries of wavelet types, so quantitative Q extraction is not straight-forward. And finally, full-waveform procedures also exist to estimate Q_a (Hicks and Pratt, 2001). As we can see, few studies have focused on estimating intrinsic attenuation from seismic data. The goal of this chapter is to study the feasibility of extracting intrinsic attenuation information from synthetic seismic data.

Seismic wave attenuation is commonly estimated from the amplitude spectrum of recorded waves. Intrinsic attenuation can be described by an exponential decay, where the exponent can be written in terms of Q_i or an absorption mean free path ℓ_i (van Wijk et al., 2004). Scattering attenuation can also be described by an exponential decay, but the makeup of the exponent is more complicated: O'Doherty and Anstey (1971) show that the exponent is made of the average power spectrum of a reflectivity series, while for ensembles of realizations (again, some averaging procedure!), the decay of the coherent intensity is described by the scattering mean free path in the exponent. In either case, averaging is important, and it is not obvious that for a single realization of the subsurface there is an equivalent exponential decay described by scattering Q: Q_s . It does seem intuitive that scattering and intrinsic attenuation are additive processes, together producing apparent attenuation:

$$\frac{1}{Q_a} = f\left(\frac{1}{Q_s}\right) + \frac{1}{Q_i}.$$
(6.1)

Schoenberger and Levin (1974); Spencer et al. (1982) and Richards and Menke (1983) confirm this additive relationship of scattering and intrinsic attenuation to apparent attenuation, in a qualitative manner, and observe for their synthetic seismograms that

$$f\left(\frac{1}{Q_s}\right) = 1/Q_s$$

The focus here is to analyze time-lapse changes in attenuation instead of absolute attenuation. We define a time-lapse change in attenuation between two acquisition times as $\Delta \frac{1}{Q} = \frac{1}{Q(t_2)} - \frac{1}{Q(t_1)}$. We propose the idea that if we have two 4D seismic surveys taken at two different times, between which fluid substitution occurs in the pore space, taking

the difference in apparent attenuation results in the difference of intrinsic attenuation. This idea works if the change in scattering attenuation at two acquisition times is negligible compared to changes in intrinsic attenuation.

We create synthetic traces in VSP and surface seismic (shot gather) geometries for an earth model with intrinsic attenuation. In this chapter we discuss the VSP synthetic seismograms and leave the analysis and discussion of shot gathers for appendix D. We estimate interval apparent attenuation from direct arrivals (in a VSP) or reflectors of interest in synthetic seismograms. To model changes in intrinsic attenuation we assume an EOR process, where brine substitutes for a light oil in a carbonate reservoir and we use our laboratory measured Q_i in carbonate samples as the Q parameter guideline when modeling seismograms. To estimate interval Q_a we apply the most common Q estimation method; namely, analyzing spectral ratios (Hauge, 1981; Toksoz et al., 1979). Another popular Q estimation technique involves estimating frequency shifts (Quan and Harris, 1997). In Appendix C we compare the spectral ratio and frequency shift methods and give a summary on their derivation and estimation approach. Both methods estimate Q from the primary (e.g direct arrival, reflection). Although their basic assumptions are the same, the computation of Q is different.

In the following sections we will model synthetic VSP seismograms and estimate attenuation under two circumstances: 1) that attenuation is only due to scattering (elastic modeling), and 2) that attenuation results from scattering and intrinsic attenuation (viscoelastic modeling). For the latter scenario we will test receiver separation, noise level in the traces, and how changes in velocity, density and Q influence $\Delta 1/Q$ estimates. At the end of the chapter we will give some insight on the challenges in estimating and interpreting apparent time-lapse attenuation.

6.2 Velocity, density and Q synthetic model

We began by modeling a zero-offset VSP, because the direct arrival in the downhole receivers usually has a good signal-to-noise ratio. Also, Mateeva (2003) showed that Q at the near surface principally affects late arrival times in a VSP. However, for reflection seismology, near surface Q could have a large contribution on all arrivals in the seismogram. In a timelapse monitoring reservoir program, the near-surface rock properties can change (e.g., wet vs. dry near surface). Therefore, analyzing time-lapse $1/Q_a$ changes due to variations in intrinsic Q in the reservoir layer from direct arrivals in a VSP has an advantage over the same estimates from surface seismic data.

To model scattering losses, the earth model should have contrasting elastic rock properties. A blue reflection coefficient series (RCS) is one in which reflection coefficients have alternating sign, typically due to cyclic changes in lithology. This type of RCS model is observed in well logs. In addition, alternating of absolute reflection coefficients between large and small values, creates scattering attenuation (O'Doherty and Anstey, 1971). The FFT



Figure 6.1. P-wave and S-wave velocity, and density synthetic logs. The section between 500 and 1000 m is generated from pseudo-random numbers to create a blue reflection coefficient series. The circles are the locations of the two VSP receivers. The model has 106 layers.

of a blue RCS shows that amplitude increases with frequency (Figure 6.2). For a transmission experiment and a blue RCS, scattering Q_s is positive, so that $1/Q_a > 1/Q_i$, because high frequencies are redirected in space. However, Mateeva (2003) shows that for reflection seismic data from a blue RCS, scattering attenuation could either contribute to the loss of high frequencies $(1/Q_a > 1/Q_i)$ or actually boost them $(1/Q_a < 1/Q_i)$. Because we will model a VSP geometry we will be dealing with an over-estimation of intrinsic attenuation by measuring apparent attenuation.

We created two velocity-density-Q models (logs): 1) with thick layers (greater than λ , the dominant wavelength), and 2) the same model, with thin layers between 500 and 1000 m with realistically high contrasting reflection coefficients (Figure 6.1). Velocity and density profiles were created by using average values for sedimentary water saturated rocks as compiled by Mavko et al. (1998). We generated a pseudo-random log at a 5 m interval from uniform distributions of P- and S-wave velocity as well as bulk density (ρ). Between 500 m and 1000 m, the average P- and S-wave velocity and density are 4500 m/s, 2400 m/s and 2.4 g/cm³, respectively. The model has 106 layers in total. P-wave intrinsic Q is included for the interval between 500 and 1000 m. The P-wave Q for the rest of the section and S-wave attenuation for all depths is high (Q=10⁴). Therefore, in this analysis all references to Q are for P-wave losses. The zero-offset VSP was modeled with a free surface. We used a compressional point source at 25 m from the well head with a dominant frequency of 60 Hz.



Figure 6.2. Spectra of the reflection coefficient series from the velocity and density logs in Figure 6.1.

Frequencies were modeled from 0 to 150 Hz, and the time sampling interval is 2 ms.

We model synthetic VSP traces by using the OASES code (Schmidt and Tango, 1986) developed for a horizontally stratified Earth. The code models seismograms for elastic and visco-elastic media. The output seismograms are modeled vertical particle velocities (in m/s), for simplicity referred to here as amplitude.

6.3 Elastic modeling

Based on the log with random RCS (Figure 6.1), we modeled synthetic seismograms for a zero-offset VSP with two receiver depths, R1=500 m and R2=1000 m. To compare the effect of layering, we also compute VSP seismograms for a six-layer model. This model is the same as in Figure 6.1 but with a homogeneous layer between 500 and 1000 m with the following constant velocities and density: $V_P=4500 \text{ m/s}$, $V_S=2400 \text{ m/s}$, $\rho=2.4 \text{ g/cm}^3$.

We plot the seismic signature at one receiver depth to compare the seismic trace with and without layering. Figure 6.3 shows the seismogram for the deeper receiver. Observe how the number of arrivals increases after the direct arrival when we include a RCS with high contrast. Most of these arrivals are waves that have bounced multiple times between the layers; i.e., the coda. The direct arrival for this receiver is attenuated by scattering. These observations are at the moment qualitative, so let's compute the interval apparent attenuation between the receivers.

From the recorded direct arrivals in the VSP receivers we estimated interval apparent attenuation. We applied a 100 ms Hanning-type window centered at the direct arrival for the top and bottom receivers (Figure 6.4) and computed a discrete Fourier transform. Figure 6.5a is the amplitude spectrum of the windowed direct arrival modeled with a random



Figure 6.3. Synthetic VSP traces for the deep receiver based on the elastic modeling in Figure 6.1. The zero-offset VSP synthetic traces are plotted for: a) homogeneous log (6 layers) and b) a blue RCS log (106 layers).



Figure 6.4. Complete VSP traces (black solid lines) and windowed traces (dashed red lines) for R1 and R2 for the elastic model in Figure 6.1.



Figure 6.5. a) Amplitude spectra of the windowed traces in Figure 6.4, and b) amplitude ratio and linear fit. Scattering attenuation: $1/Q_s=0.0153\pm0.0006$ is estimated from the slope of the line.

RCS. The base 10 logarithm of the ratio of the two spectra as a function of frequency is shown in Figure 6.5b, and we apply spectral ratios to estimate Q_a (Appendix C). We use Equation C.1 to relate the line slope to Q. From the linear fit we also estimate the variance in the random error for Q and 1/Q as described in Appendix A. In this chapter we will plot and refer uncertainty as one standard deviation from the mean. We estimate $Q_a=65.5\pm2.5$ ($1/Q_a=0.0153\pm0.0006$) for this elastic model and receiver positions. Because we are modeling elastic seismograms, $Q_i \gg Q_s$, the measured Q_a results only from scattering attenuation. Hereafter, we will limit estimates of 1/Q and the associated standard deviation to three decimal places. Thus, if there is no reference to the uncertainty in 1/Q, it is because one standard deviation is less than 0.0005.

The frequency range used to estimate Q is important to understand. Consistently using the same frequency interval for computing Q is critical, because estimates of scattering attenuation depend on the frequency window (Mateeva, 2003). For our synthetic data, the amplitude ratios for frequencies greater than 150 Hz are noisy. We choose an optimal S/N level from 0 to - 15 dB, corresponding to a frequency range between 12 and 134 Hz. We consistently used this frequency interval to compute the spectral ratios in this chapter.

6.4 Visco-elastic modeling

In this section we will first test quantitatively whether Equation 6.1 is valid for synthetic VSP traces. Next, we will model synthetic seismograms at two acquisition times, simulating an EOR process, for which intrinsic attenuation has changed. If Equation 6.1 holds true, and scattering attenuation does not change because the elastic rock properties remain constant at the two times, taking the difference of apparent attenuation should yield the change in intrinsic attenuation. And finally, we will study the influence of receiver separation and random noise on time-lapse 1/Q estimates. For better understanding of the nomenclature and attenuation values, we summarize the equations involved.

If apparent attenuation is the sum of intrinsic and scattering attenuation, and Q_s is constant over two acquisition time intervals:

$$\Delta \frac{1}{Q_a} = \frac{1}{Q_{a,2}} - \frac{1}{Q_{a,1}} = \left(\frac{1}{Q_{s,2}} + \frac{1}{Q_{i,2}}\right) - \left(\frac{1}{Q_{s,1}} + \frac{1}{Q_{i,1}}\right) = \frac{1}{Q_{i,2}} - \frac{1}{Q_{i,1}} = \Delta \frac{1}{Q_i} \quad (6.2)$$

where 1 and 2 refer to the data acquisition at two time intervals.

Intrinsic attenuation changes used to model synthetic seismograms are obtained from laboratory data in carbonate rocks. At seismic frequencies and reservoir conditions we measured intrinsic attenuation in three carbonate samples (chapter 3). These data shows that for our carbonate samples compressional-wave intrinsic attenuation is greater than shearwave attenuation. For one of the samples Q_i for P-waves decreases from 100 (hydrocarbon saturated carbonate) to 50 (brine saturated carbonate). These intrinsic P-wave Q values

$1/Q_{a,1}$	$1/Q_{a,2}$	$\Delta 1/Q_a$	$\Delta 1/Q_{i,true}$
$0.024 {\pm} 0.001$	$0.034{\pm}0.001$	$0.010 {\pm} 0.001$	0.010

Table 6.1. Attenuation estimates for receivers at 500 and 1000 m.

are used to modeled synthetic seismograms for time-1 ($Q_i=100$) and time-2 ($Q_i=50$). We will refer to 1/Q for most of the results and discussion, because it is 1/ Q_i and 1/ Q_s that are additive. If 1/ Q_s cancels in a time-lapse experiment, the true (predicted) change in apparent attenuation is:

$$\Delta \frac{1}{Q_{a,true}} = \Delta \frac{1}{Q_{i,true}} = \frac{1}{Q_{i,2,true}} - \frac{1}{Q_{i,1,true}} = 1/50 - 1/100 = 0.010.$$

The predicted apparent attenuation based on the estimated Q_s (section 6.3) and modeled Q_i under the assumption that $f\left(\frac{1}{Q_s}\right) = 1/Q_s$:

$$1/Q_{a,1,true} = 1/Q_{s,est} + 1/Q_{i,1,true} = 1/65.5 + 1/100 = 0.025$$
$$1/Q_{a,2,true} = 1/Q_{s,est} + 1/Q_{i,2,true} = 1/65.5 + 1/50 = 0.035.$$

The estimated $1/Q_a$ values from direct arrivals between the two receivers are shown in Table 6.1. Within one standard deviation, the predicted and estimated apparent attenuation at time-1 and time-2 match with the predicted $1/Q_a$. This quantitative result agrees with our hypothesis that apparent attenuation equals the sum of scattering and intrinsic $1/Q_a$ and that $f\left(\frac{1}{Q_s}\right) = 1/Q_s$ describes the scattering loss contribution to the overall $1/Q_a$. Moreover, the difference in apparent attenuation between two surveys acquired at different times is the time-lapse change of the intrinsic attenuation: 0.010. This result confirms that, although $1/Q_{a,1}$ and $1/Q_{a,2}$ cannot independently give quantitative information about the intrinsic attenuation, their difference does describe the visco-elasticity created by fluid exchange.

Fluid substitution in a reservoir not only changes the visco-elastic properties of the reservoir, but the elastic rock properties as well. However, our laboratory measurements from chapter 3 indicate that for our carbonates fluid substitution has a much larger effect on intrinsic attenuation than on density and velocity. These parameters change on average 7%, while attenuation changes 275%. Thus, the difference is at least an order of magnitude. It is velocity and density that define the acoustic impedance, and therefore scattering strength. Because changes in scattering strength are small compared to those in attenuation, we can safely assume that our reservoir changes in apparent attenuation are dominated by changes in intrinsic Q.



Figure 6.6. Apparent attenuation and time-lapse apparent attenuation sensitivity to receiver separation. Receiver depth in meter: 500-1000 (C1), 600-900 (C2), 700-800 (C3) and 730-760 (C4). In Case 1 $(1/Q_{a,1})$ Q is 100; in Case 2 $(1/Q_{a,2})$ Q is 50.

6.4.1 Receiver separation

Scattering attenuation will also depend on the location of the receivers, because waves interfere differently as a function of receiver depth, and in this section we will test 4D apparent attenuation sensitivity to receiver separation. Synthetic seismograms are modeled by varying the receiver separation in meters as: 500-1000, 600-900, 700-800 and 730-760. The remaining modeling parameters were the same as for the previous section. Figure 6.6 shows the estimates for apparent attenuation for time-1 and time-2, and their difference. Note that although the apparent attenuation at time-1 and time-2 varies with receiver separation, the 4D apparent attenuation is constant (~0.01). The variation in the estimate of $1/Q_{a,1}$ and $1/Q_{a,2}$ probably results from $1/Q_s$: at different receiver depths the layers contribute to scattering loss differently.

As the receiver separation decreases, the apparent attenuation increases linearly. This observation agrees with a synthetic VSP modeled by Mateeva (2003) in the case when a strong RCS overlays a weak RCS between receivers. This RCS geometry is similar to our earth model where the strong RCS is above the deep receiver. As the receiver separation decreases, the error increases, because the direct wave amplitude interferes with reflections from the local stratigraphy and from the stratigraphy between receivers (Mateeva, 2003; Spencer et al., 1982; Shapiro and Zien, 1993).

6.4.2 Random noise

Another parameter that can affect the estimate of apparent attenuation is noise. Next, we will study the effect of random noise on the estimate of time-lapse apparent attenuation.



Figure 6.7. Original trace at R1 and time-1 and with added noise at three levels.

Based on work from Soroka et al. (2007) on a time-lapse VSP with fixed geophones, the trace repeatability (background noise) is 1% of the original trace amplitudes. This VSP is from the same field from where samples 100, 200 and 300 were measured, and as mentioned before, the attenuation change from light hydrocarbon to brine is estimated from laboratory measurements on these samples (Chapter 3). Landrø (1999) shows that for a 3D VSP, the repeatability error (root-mean-square error) in a North Sea VSP is 8% of the original trace amplitude.

To reproduce noise in the field, we added pseudo-random Gaussian noise to both receivers and times, and model the visco-elastic parameters from section 6.4. The applied noise was from a pseudo-random Gaussian distribution with zero mean and for which one standard deviation is 1, 3, 5, 10 and 15% of the maximum amplitude of the direct arrival. Figure 6.7 shows the VSP trace for R1 with different levels of added noise. Figure 6.8 is the estimated apparent attenuation for the time-lapse experiment and noisy traces. For a low noise level (n<5%), random noise does not significantly affect the estimate of $\Delta 1/Q_i$. On average, $\Delta 1/Q_a=0.009$, and the true value is 0.010. However, larger levels of noise (n \geq 5%) affect the shape of the direct arrival (Figure 6.7), and in turn the shape of the amplitude spectra, biasing $\Delta 1/Q_a$ estimates.

Coherent noise interfering with the direct arrival, such as tube waves in a VSP, can also bias apparent attenuation. Coherent noise can change the shape of the amplitude spectra of the events of interests, changing amplitude spectra ratios. Therefore, contributions of



Figure 6.8. Estimated apparent attenuation as a function of random noise added to the traces. At low noise levels (n<5%) the estimate of $\Delta 1/Q_a$ agrees with the expected $\Delta 1/Q_a$.

coherent noise on VSP direct arrivals can modify the slope of the logarithm of the amplitude spectra ratio as a function of frequency, affecting the apparent attenuation estimates.

We have performed a similar analysis on time-lapse attenuation for a surface seismic geometry described in Appendix D. For a simple forward model of a shot gather seismogram, the window size applied to the event of interest significantly controls the apparent attenuation estimate. Nonetheless, with a window size that cleanly isolated the event of interest, we are able to recover the reservoir time-lapse intrinsic attenuation from the timelapse apparent attenuation estimates.

6.5 Interpreting time-lapse apparent attenuation

In this section we will explore possible pitfalls when estimating time-lapse intrinsic attenuation. We will first describe the non-uniqueness of the $\Delta 1/Q_i$ estimate. Second, if Q_i varies significantly with depth between the VSP receivers, then the estimate of $\Delta 1/Q_a$ will not be a good estimate of $\Delta 1/Q_i$. By a significant Q_i variation with depth, we mean that between the VSP receivers there are several layers with distinctive Q_i , but only one of them experiences fluid substitution and thus a change in Q_i , while the others remain the same. Finally, we will describe the difference in wavelet spectra that might help us decide whether to estimate Q from spectral ratios or frequency shift.

Figure 6.9 shows all possible combinations for time-lapse $1/Q_i$ based on intrinsic Q varying from 5 to 200. Observe that there will be several combinations of $1/Q_{i,1}$ and $1/Q_{i,2}$ that produce, for example, $\Delta 1/Q_i=0.010$. However, from laboratory data the range of possible $1/Q_i$ can be narrowed. Also, when a fluid is substituted for, replaces another fluid, $1/Q_{i,2}$ may be either smaller or larger than $1/Q_{i,1}$; yielding a negative or positive $\Delta 1/Q_a$ respectively, based on our definition of $\Delta 1/Q_a$ (Equation 6.2). And as we discussed in



Figure 6.9. Estimated time-lapse changes in $1/Q_i$ for all possible combinations of intrinsic attenuation between 0.005 and 0.2.

previous sections, $\Delta 1/Q_a$ has only contributions from intrinsic rock losses, thus, the rock fluid dynamics.

To analyze the contributions of layers with different Q_i , we positioned the receivers at 300 and 1100 m for a VSP geometry. We modeled synthetic traces with $Q_i=100$ and 50 only for the layers in the random sequence (500-1000 m), while the rest of the depth section has a $Q_i=10^4$. Thus, the recorded wave in the deep receiver propagates through a low (0-500 m and 1000-1100 m) and high (500-1000 m) attenuating section. Table 6.2 shows the estimated apparent 1/Q between 300 and 1100 m. The $\Delta 1/Q_a$ does not agree with the true $\Delta 1/Q_i$, because we have contributions of low and high attenuating layers. Therefore, $\Delta 1/Q_i$ cannot be correctly estimated from $\Delta 1/Q_a$ if the layers between the receivers have significantly different $1/Q_i$ and only a sublayer experiences changes in Q_i over time. In this example, the ratio of low attenuation to high attenuation rocks, 3:5, is significant.

Interestingly, the estimate of interval apparent attenuation at a particular time follows the rule of a weighted average of intrinsic attenuation. The weights (w) are governed by the layer thickness or travel times in the contrasting Q_i regions between the receivers. For n parallel layers with different Q_i we propose that interval $1/Q_a$ can be estimated from:

$$\frac{1}{Q_a} = \frac{1}{Q_s} + \sum_{j=1}^n \frac{1}{Q_{i,j}} w_j \tag{6.3}$$

with j representing each of the layers with different Q_i between the VSP receivers.

The previous relation was also found valid for a six-layer model (Figure 6.1 with no

$1/Q_{a,1}$	$1/Q_{a,2}$	$\Delta 1/Q_a$	$\Delta 1/Q_{i,true}$
0.015	0.021	$0.005 {\pm} 0.001$	0.010

Table 6.2. Estimates of $1/Q_a$ for receivers at 300 and 1100m. The Q_i between the receivers varies with depth from 10^4 to either 100 or 50.



Figure 6.10. Amplitude spectra from the direct arrival of VSP receivers. The blue curves represent a Gaussian function fit to the spectra.

random layers) with three different Q: 120, 50 and 100 in layers 2, 3 and 4. Applying Equation 6.3 gives $1/Q_{a,true}=0.016$, while the estimated $1/Q_a$ from spectral ratios is 0.014. A similar approach to estimate effective 1/Q is was proposed by Spencer et al. (1982).

Finally, we show why we choose spectral ratios (SR) over frequency shift (FS) to estimate Q_a in this chapter. Both methods have the same assumptions, but the methodology for estimating Q is different (Appendix C). Figure 6.10 shows the amplitude spectra of the traces modeled with a log and a $Q_i=50$. Interference from near stratigraphy affects the direct arrival. Therefore, even a small window (80 ms) has interfering events, creating amplitude spectrum asymmetry (Figure 6.10). When we apply the FS method, we assume that the spectrum can be modelled by a Gaussian function, but Figure 6.10 clearly shows that the spectrum is not Gaussian-like. The Gaussian fit to the spectra of the top and bottom receivers is shown in blue. Q_i from FS is estimated at 74.9±8.3; that from SR is 54.2±3.1 ($Q_{i,true}=50$). In Appendix C the FS approach showed to be more stable under noise when the spectra are Gaussian shaped, that is not the case for an asymmetric spectra as in Figure 6.10. Deciding which methodology to follow could be assessed by studying the shape of the amplitude spectrum of the trace. Nonetheless, if the spectrum is Gaussian shaped, SR would probably be stable as a Q estimator method, because the amplitude ratios with frequency will behave linearly.

6.6 Conclusions

Although estimating attenuation from seismic data is still a challenging procedure, with increasing seismic data quality, estimation of attenuation could be common practice in the near future. From this study we conclude that time-lapse apparent attenuation has the potential to be used as an attribute in reservoir fluid monitoring. We have quantitatively shown that apparent attenuation is the exact sum of scattering and intrinsic attenuation (in 1/Q terms). Therefore, scattering attenuation cancels when taking the difference in apparent attenuation between two acquisition times, so that we are left with changes in intrinsic attenuation alone, which are directly related to the type of fluid in the rock. Even though individually estimated $1/Q_a$ for each acquisition time may vary with receiver separation or random noise (< 5% for VSP), this variability cancels or becomes less significant when we take the difference in apparent attenuation, and yields the predicted 4D difference in intrinsic attenuation.

CHAPTER 7

CONCLUSIONS

The elastic properties of the measured carbonate samples in the laboratory under varying fluids and pressure conditions show that:

- The carbonate rock frame weakens when the pore space is saturated with brine a polar fluid - compared to the same samples dry (humidified) and fully-saturated with liquid-butane (light hydrocarbon). This frame weakening is observed by the decrease in the rock shear modulus at seismic frequencies due to fluids. At ultrasonic frequencies, rock-fluid mechanisms at these high frequencies actually increase the shear moduli of brine saturated carbonates compared to dry. For our measured samples this frequency dispersion dominates over the weakening of the rock frame.
- Shear modulus weakening is greater at low differential pressures when compliant pores are open and steadily decreases as we reach the reservoir differential pressure. This pressure dependence of the frame weakening implies that matrix softening is related to the amount of open soft pores.
- The rock shear modulus weakening is reversible in our controlled experimental environment. Therefore possible mechanisms that explain the weakening due to a polar fluid are the bound breakage of grain or cement contact areas or sub-critical crack growth. Mineral dissolution due to brine is not a reversible process.
- Gassmann's theory is used to model the saturated bulk modulus of rocks and is derived for low frequencies (zero frequency). At seismic frequencies the butane-saturated rock bulk modulus of all our measured samples is predicted by Gassmann's. However, for a lower compressibility fluid such as brine the correlation between measured brine-saturated carbonates and Gassmann's prediction is not straightforward. For our sample set we find a match between the measured and Gassmann's predicted brinesaturated rock bulk modulus for samples with high aspect ratio pores: round or more sandstone-like grains.
- At ultrasonic frequencies the saturated bulk modulus of some carbonates is predicted by Gassmann's relation. This is an apparent correlation because the bulk and shear modulus are dispersive with frequency from a few Hz to MHz. And while the same sample at low frequency is not predicted by Gassmann, the use of this zero-frequency theory at high frequencies can be misleading.

- The modulus and velocity frequency dispersion from seismically to ultrasonically measured samples can be reduced if the ultrasonic V_P/V_S ratio or time-lapse absolute reflection coefficients are analyzed for seismic reservoir characterization.
- Sample heterogeneity can affect modulus and velocity estimates especially at ultrasonic frequencies. Heterogeneity can result from rock frame variability (e.g. mineralization, texture or porosity contrast) or patchy saturated regions in the core sample.
- The low frequency measured rock properties could also be sensitive to rock heterogeneity. Therefore we develop and test a fiber-optic strainmeter. Instead of the commonly used strain gages to measure deformations we used fiber-optic cable with an interferometry setup to estimate visco-elastic properties of samples. With limited success we measured a Plexiglas sample at room conditions. However, the system's strength is also it weakness: large sensitivity. Measurements on a rock sample have not been accomplished because of the instability of the interference fringes from which the visco-elasticity of the material is estimated. However we document of trials and steps to improve the system.

By studying the visco-elasticity from the laboratory samples we conclude that:

- Contrary to popular belief, for fully- and partially saturated carbonate samples with either a light hydrocarbon or brine show that the bulk modulus intrinsic attenuation (1/Q) dominates over the shear wave attenuation. This observation holds for seismic and ultrasonic frequencies for our carbonate sample set.
- A change from brine to liquid-butane in the pore space increases the bulk attenuation by a factor of 1.5 to 4, compared to an increase of P-wave velocity in average by a factor of 1.07 for these samples.
- The bulk attenuation is more sensitive to fluids than the other modes of attenuation. This information can help on the decision on which seismic data volume (P-wave or S-wave) is more sensitive to extracting fluid information from intrinsic attenuation.
- Modeled bulk and shear attenuation from the laboratory measured elastic moduli by using the Cole-Cole relation verifies the above three conclusions: bulk losses are greater than shear losses, bulk attenuation is more sensitive to brine than shear attenuation, and the change in bulk attenuation due to fluid substitution in at least a factor of 2.
- Within the exploration seismic frequency band, the measured samples show that attenuation can be assumed frequency independent. However, over the whole range of measurement 3-1000 Hz and the ultrasonic data, 1/Q is frequency dependent for some of the measured samples.

We model synthetic seismic seismograms to study the feasibility of estimating timelapse intrinsic attenuation and conclude that:

- The estimated apparent attenuation from synthetic seismograms is quantitatively the addition of intrinsic and scattering attenuation.
- For our model and assumptions, because the contribution of scattering attenuation over-time is repeatable, the time-lapse difference in apparent attenuation results only from changes in the intrinsic attenuation.
- For a VSP geometry, apparent attenuation is sensitive to receiver separation, velocity and density changes or random noise; at a particular acquisition time. However, the 4D apparent attenuation is equal to the time-lapse intrinsic attenuation because scattering attenuation is repeatable and cancels over time.
- For a non-stacked zero-offset synthetic surface seismic trace, the time-lapse apparent attenuation is also equal to the 4D intrinsic attenuation. However, the window size applied to isolated the reflection of interest influences the estimate of apparent attenuation, especially small windows. Random noise significantly affects the estimate of apparent attenuation from a zero-offset trace for noise level greater than 10%. But when correcting by NMO and stacking the near offsets traces, the random noise cancels and the estimate of apparent attenuation is accurate.
- The synthetic seismogram modeling and analysis concludes that time-lapse apparent attenuation has the potentially to be a powerful attribute to analyze the fluid changes in dynamic reservoirs.

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APPENDIX A

DATA ESTIMATION AND UNCERTAINTY ANALYSIS

At seismic frequency moduli and velocities are measured with the strain-stress method. Vertical strain is defined as: $\epsilon = \Delta L/L$, where L is the sample length and ΔL is the deformation. We measure strain in the vertical (*vert*) and horizontal (*hrz*) directions on our samples. The samples are cylindrical and we mount them in series with cylindrical aluminum standards, which in turn house ultrasonic transducers. Aluminum serves as our calibrating material with vertical strain gages, while our rock has vertical and horizontal mounted gages (Figure A.1). We apply a sinusoidal vertical stress (σ) for a range of frequencies (3-3000). The vertical stress on the aluminum is the same as on the rock under the assumption of linear elasticity. $\sigma_{vert} = E\epsilon_{vert}$ and E is the Young's modulus. By equating the vertical stress on the rock (no subscript) and the aluminum (*al*) we obtain the Young's modulus of the rock:

$$E = E_{al} \frac{\epsilon_{vert,al}}{\epsilon_{vert}}; \tag{A.1}$$

 E_{al} is 70 GPa and $\epsilon_{vert,al}$, ϵ_{vert} are measured strains. The rock Young's modulus is our first data. Our second data set is Poisson's ratio relating the horizontal and vertical strains on the rock:

$$\nu = -\frac{\epsilon_{hrz}}{\epsilon_{vert}}.$$
(A.2)

Both the rock Young's modulus and Poisson's ratio are thus measured as a function of frequency. Assuming isotropy, the bulk (K) and shear (μ) moduli of the rock are:

$$K = \frac{E}{3(1-2\nu)}; \quad \mu = \frac{E}{2(1+\nu)}.$$
 (A.3)

From these moduli and the rock density (ρ) velocity follows:

$$V_P = \sqrt{\frac{K + \frac{4}{3}\mu}{\rho}}; \quad V_S = \sqrt{\frac{\mu}{\rho}}.$$
 (A.4)

In the stres-strain experiment we first estimate the bulk and shear moduli and later include the rock density to estimate velocity. Thus, our moduli estimates are independent of the rock density. As we will see next, for ultrasonic data the rock density *is* needed to



Figure A.1. Sample setup sketching the location of the strain gages and ultrasonic transducers. The low frequency waveforms are also sketched for the three different strain gages.

estimate the bulk and shear moduli.

For ultrasonic data we measure the time a wave takes to propagate from the top of the sample to the bottom. The velocity, either P- or S-wave, is estimated by:

$$V = \frac{L - \delta L}{T_m - T_0},\tag{A.5}$$

where L is the sample length measured at atmospheric pressure, δL is the change in sample length due to pressurization, T_m is the measured travel time, and T_0 is a time correction. δL is ignored because the change in length, which we can estimate from the low frequency experiment, is very small. T_0 , the travel time through the aluminum material between the ultrasonic transducer and the sample, is known and constant for all measured samples. Therefore we can rewrite Equation A.5 as simply:

$$V = \frac{L}{T},\tag{A.6}$$

where T is the corrected travel time. Using the S-wave velocity and rock density, we calculate the ultrasonic shear modulus by Equation A.4, and we compute the bulk modulus from the P-wave velocity, estimated shear modulus and density.

A.1 Data uncertainty

To estimate the uncertainty in our data we will assume that the *true* model for the low frequency data is a line and a 2^{nd} order polynomial for the ultrasonic data. Both data errors are assumed random, Gaussianly distributed and with zero mean.

The equation of a line with random noise ϵ is:

$$y_i = \beta_0 + \beta_1 x_i + \epsilon_i, \tag{A.7}$$

where y_i is the *i*-th measured data (e.g., Young's modulus) and x_i is *i*-th the predictor (e.g., frequency). β_0 and β_1 are the model parameters (Draper and Smith, 1981). When fitting the data by least-squares for example, estimators (labeled with $\hat{}$) of the data and model parameters are:

$$\hat{y}_i = \hat{\beta}_0 + \hat{\beta}_1 x_i, \tag{A.8}$$

An estimator of the variance of the random error is:

$$\hat{\sigma_{\epsilon}}^{2} = \sum_{i=1}^{n} \frac{(y_{i} - \hat{y}_{i})^{2}}{n - p},$$
(A.9)

where n is the number of data points and p is the number of model parameters we are estimating in the regression, which for a straight line is equal to 2.

We are interested in obtaining an estimator of error of the fitted values (\hat{y}_i) . We can rewrite Equation A.7 in matrix from, where \mathbf{y} , β and ϵ are vectors and A is a matrix (operator):

$$\mathbf{y} = A\beta + \epsilon; \quad A = \begin{pmatrix} 1 & x_1 \\ 1 & x_2 \\ \vdots & \vdots \\ 1 & x_n \end{pmatrix} \quad \beta = \begin{pmatrix} \beta_0 \\ \beta_1 \end{pmatrix}. \tag{A.10}$$

The variance of the fitted values $(\hat{\mathbf{y}})$ depends on both the variance of the random error of the data and on the assumed model (structure of A). We obtain the variance estimate for each fitted value as Draper and Smith (1981):

$$\hat{\sigma}_{\hat{y}_i}^2 = \hat{\sigma}_{\epsilon}^2 [A(A^T A)^{-1} A^T]_{ii}.$$
(A.11)

Applying a least squares methodology means minimizing the residual between data and fitted values about the data mean. The vector $[A(A^TA)^{-1}A^T]_{ii}$ weights the errors for the fitted values so that the $\sigma_{\hat{y}_i}^2$ is larger as we get away from the predictors associated with the data mean. Errors can also be systematic (bias) instead of random, but at this point we assume we have no bias in the data or have corrected for it (e.g., our corrections for Poisson's ratio). We propagate the variance of the fitted values into other variables (e.g., bulk and shear moduli). For a general variable U = f(X, Y, ...Z), where X, Y, ...Z are independent random variables with small variances, an estimate of the variance of U is approximated by:

$$\hat{\sigma}_U^2 \approx \left(\frac{\delta f}{\delta X}\right)_{\hat{\mu}_X}^2 \hat{\sigma}_X^2 + \left(\frac{\delta f}{\delta Y}\right)_{\hat{\mu}_Y}^2 \hat{\sigma}_Y^2 + \dots + \left(\frac{\delta f}{\delta Z}\right)_{\hat{\mu}_Z}^2 \hat{\sigma}_Z^2, \tag{A.12}$$

where $\hat{\mu}_{(X,Y,Z)}$ is an estimate of the mean of the random variables.

This relation is applied to each experimental method to obtain errors on the estimated values: modulus and velocity.

A.1.1 Uncertainty analysis on the modulus and velocity

Our data sets consist of Poisson's ratio and Young's modulus as a function of frequency and differential pressure (seismic frequency), and travel time as a function of differential pressure (ultrasonic frequency). We assume that the Poisson's ratio and Young's modulus relation to the logarithm base 10 of frequency is linear, while the travel time with differential pressure follows a second order polynomial. These are our *true* models. We also assume that the error on our data is random, Gaussianly distributed and with zero mean. In Chapter 2 our core analysis is performed under the assumption that all requirements for Gassmann's theory applicability are satisfied. If our samples and experimental setup violate one (or more) of the assumptions of Gassmann's theory we introduce a bias (systematic error) in our estimates, and we give an interpretation to why some results on the samples do not obey Gassmann's assumptions.

A.1.2 Stress-strain methodology

In Figure A.2 we plot data for the stress-strain experiment (E and ν) showing a linear trend with \log_{10} of frequency. We fit a straight line to our data and estimate the variance of our random error. We use the variance of the random error to compute the error of estimates of Young's modulus and Poisson's ratio (Equation A.11), and later propagate this error into the estimates of bulk and shear modulus (Equation A.12). The Young's modulus of aluminum equals 70 GPa, and we assume this value is error free for the uncertainty analysis. On average, our estimates of the standard deviation of the estimated bulk modulus is 1.2 GPa, and that of the shear modulus is 0.3 GPa for seismic frequencies.

We acquire data for 100 frequency points, but sometimes we limit our study to 10, 100, 1000 Hz which are representative seismic frequency points. These data are obtained from the fitted \hat{y}_i and are used to simplify the analysis between ultrasonic and low frequency data.



Figure A.2. Error analysis on sample C at a differential pressure of 17.5 MPa and butane saturation. From the least-squares fit we estimate the variance in our estimate of E and ν .

A.1.3 Ultrasonic pulse propagation

In addition to low frequency measurements, we measure wave propagation travel times at 0.8 MHz as a function of differential pressure. Travel time decreases with increasing differential pressure (higher velocity). Figure A.3 shows this dependence, resulting from open cracks and compliant pores at low differential pressures. A second order polynomial is fit to the ultrasonic travel time data as a function of pressure (dashed and solid lines in Figure A.3), from this data estimate we obtain the variance of the random error. We then compute the error of our estimated travel times.

The ends of our samples are machine flattened and when the length is measured repeatedly no significant variability is observed. Also the change inlength due to differential pressure is minimal. Therefore, we consider the sample length error free. We also assume there is no error in the differential pressure measurements (no errors in x_i). The estimated error in P- and S-wave velocities is obtained by propagating the estimate of the travel time error. To estimate the bulk and shear modulus we need the rock density which depends on porosity, grain density, and fluid density. We will assume that the variance of the rock density is 0.5% (which is in the lower end of errors reported in core measurements). The next steps are to propagate the error in S-wave velocity and rock density into the shear modulus; and then propagate the P-wave velocity, rock density and shear modulus variance estimators into the bulk modulus. On average, one standard deviation of the estimated Pand S-wave travel times is small ($\hat{\sigma}_t = 0.06 \eta_s$). Still, a small error in the rock density (0.5%) significantly affects the error of the bulk and shear modulus estimates ($\hat{\sigma}_K = 2.4$ GPa and $\hat{\sigma}_{\mu} = 0.8$ GPa) compared to the errors for data from the stress-strain experiment.



Figure A.3. Second order polynomial fit to ultrasonic travel times as a function of differential pressure for sample D under butane saturation.

A.2 Attenuation estimates

For an ideal elastic material, the stress and strain are related to each other by the modulus of elasticity, which is a real quantity. For a visco-elastic material, the strain response to an applied stress is delayed in time compared to the stress, which makes the modulus complex. Because the modulus is complex, the resulting strain signature will be phase shifted when compared to the applied sinusoidal stress. O'Connell and Budiansky (1977) defined the quality factor as:

$$\frac{1}{Q} = \frac{M_I}{M_R} = tan\phi \approx \phi, \tag{A.13}$$

here M_I and M_R are the imaginary and real parts of the complex modulus, and ϕ is the phase angle shift between stress and strain (usually very small).

Aluminum has a high Q (10⁵), thus we can assume that the phase of the strain represents the phase of the stress. We therefore estimate Young's modulus attenuation $(1/Q_E)$ from the phase shift between the vertical gages on the aluminum and on the rock.

To obtain the shear-wave attenuation we use the following relation (White, 1965):

$$\frac{1}{Q_S} \simeq \frac{(1+\nu)\frac{1}{Q_E} - \nu tan\theta_{\nu}}{1+\nu}$$
(A.14)

where ν is Poisson's ration and θ_{ν} is the phase lag between the vertical and horizontal strain on the rock sample.

The bulk and P-wave attenuation are estimated from (Winkler and Nur, 1979; Winkler, 1979):

$$\frac{(1-2\nu)}{Q_K} = \frac{3}{Q_E} - \frac{2(1+\nu)}{Q_S}$$
(A.15)

$$\frac{(1-\nu)(1-2\nu)}{Q_P} = \frac{1+\nu}{Q_E} - \frac{2\nu(2-\nu)}{Q_S}$$
(A.16)

Gautam (2003) obtained the standard deviation for phase measurements with our lock-in amplifier (repeatability of the phase), which we then propagate to estimate the errors in Q^{-1} at seismic frequencies. The estimated Poisson's ratio standard deviation also contributes to the errors in Q_S^{-1} , Q_K^{-1} and Q_P^{-1} .

A.2.1 Uncertainty on the line slope

In Chapter 6 we estimate attenuation from the logarithm of the amplitude ratio of two events vs. frequency. By fitting a line we estimate the slope and attenuation (1/Q) is proportional to the slope. The error bars in the plots in Chapter 6 are the estimate of the uncertainty in the model parameter (slope). In this analysis the data (y_i) is $log_{10} \frac{A_{top}}{A_{bottom}}(f)$ and with an associate random error. From Equation A.9 we estimate $\hat{\sigma}_{\epsilon}$ and the estimate of the variance of the random error in the slope of a line is obtained from Draper and Smith (1981):

$$\hat{\sigma}_{slope}^2 = \frac{\hat{\sigma_e}^2}{\sum_{i=1}^n (x_i - \hat{\mu_x})} \tag{A.17}$$

x is frequency for this case and this estimate of error in the slope is propagated into 1/Q by applying Equation A.12.

APPENDIX B

FIBER-OPTIC COMPENSATOR

The compensator is a Proportional-Integral-Derivative (PID) controller, which is a system with a feedback loop that keeps a process running around a set-point (e.g. amplitude of a fringe) by taking a corrective action whenever there is a deviation from the desired set-point. The PID receives a signal from sensors (light detector) and corrects by: error computation (proportional), sum of previous errors (integral) and rate of error change (derivative).

The system detects the light from the combined two fiber arms of the interferometer. When the sample is strained, the detected light represents the interference fringes. The system is set to compensate by applying a sinusoidal voltage to a piezo crystal so that instead of recording fringes we detect a constant light intensity. The piezo voltage is proportional to the sample deformation - similar to a film negative. We do not need to convert the piezo voltage into deformation if we take the voltage ratio: vertical on sample and calibrating aluminum and, vertical and horizontal on the sample. However there will always be some uncompensated signal which is not possible to get lower than 40 mV. The compensator has a low-pass filter with a cut-off frequency of 200 Hz Mason (2005).

APPENDIX C

Q ESTIMATION METHODS

In this appendix two methodologies are compared to estimate Q: spectral ratios (SR) and frequency shift (FS). The amplitude of a plane wave can be written as: $A(f) = A_o G(x)e^{-\alpha(f)x}e^{i(2\pi ft-kx)}$. A_o is the initial amplitude and G(x) is the geometric factor that includes amplitude changes due to wavefront spreading, transmission/reflection and source and receiver signature and is assumed frequency independent (Toksoz et al., 1979). x, t and f are distance, time and frequency, respectively. α is the attenuation coefficient, which can be assumed linearly dependent on frequency Toksoz et al. (1979), and we can write $\alpha(f) = \gamma f$, where γ is the attenuation constant. The quality factor is related to γ as: $Q = \frac{\pi}{\gamma v}$, and v is velocity. Q is assumed frequency independent which based on a narrow seismic band of the laboratory data is a reasonable statement. The attenuation for an interval from spectral ratios is estimated from the following equation (Hauge, 1981; Toksoz et al., 1979):

$$ln\left[\frac{A_t(f)}{A_b(f)}\right] = (\gamma_b - \gamma_t)xf + ln\left[\frac{G_t(x)}{G_b(x)}\right]$$
(C.1)

where A(f) is the amplitude spectrum for the event: top (t) or bottom (b) of a layer or group of layers. Equation C.1 is the equation of a line for data in the form of the logarithm of amplitude ratio vs. frequency, and the line slope is $(\gamma_b - \gamma_t)x$. Interval Q (Q_{int}) is proportional to the line slope: $Q_{int} = \pi(time_b - time_t)/slope$.

The frequency shift method assumes the amplitude spectrum can be described by a Gaussian function with central frequency f_c and variance σ^2 . Q and G(x) are assumed frequency independent as for SR. Interval Q is estimated by Quan and Harris (1997):

$$\frac{1}{Q_{int}} = \frac{v_{int}}{\pi} \frac{1}{\sigma_t^2} \frac{f_{c,t} - f_{c,b}}{x_b - x_t}$$
(C.2)

In this section we study three aspects of estimating intrinsic Q: 1) analyze the signalto-noise level of the amplitude spectrum. 2) compare SR to FS and, 3) study the resolution in estimating Q for a wedge model. In these three subsections, scattering Q does not contribute to the estimated Q because the modeled layers are thick with respect to the wavelength ($\lambda \approx 50$ m). Therefore when estimating Q we estimate intrinsic Q rather than apparent Q. Scattering attenuation is included in section Section X.



Figure C.1. Zero offset surface seismic synthetic seismograms for a 4-layered model with a high Q (a) and Q=50 between R2 and R3 (b). c) are the windowed events from the trace in plot b).

C.1 Estimating signal-to-noise level

We first create a synthetic zero offset surface seismic trace for a homogeneously horizontal 4-layered model. Figure C.1a shows the traces with a high Q (10^9) and with an intrinsic Q=50 between reflectors R2 and R3 (b). Figure C.1c are the traces after applying a Hanning-type window of 100 ms . Figure C.2 a shows the amplitude spectrum of the two events. However, even for synthetic and clean traces, especially when using SR we have to decide on a signal-to-noise level (S/N). For that purpose, we convert the amplitude spectrum into decibels where 0 means the best S/N, and away from zero (more negative) the S/N decreases (Figure C.2 b). For the modeled traces, the noise probably results from small numerical artifacts and ratios of small numbers. Figure C.2 c shows the optimum dB level applied on the amplitude ratios as a function of frequency (circles). The red line is the linear fit to the amplitude ratios as described by Equation C.1, and from the slope we obtain Q. When selecting a larger dB level (larger S/N) the estimate Q is 69, 90 and 117 for a -10, -20 and -30 dB level respectively.



Figure C.2. a) Amplitude spectrum of the events windowed in Figure C.1c. b) Amplitude in dB: each box represents the accepted noise level (amplitude data used to estimate intrinsic Q) from the amplitude ratio between top and bottom events. c) Estimate of Q for a -5 dB noise level.

Q_{true}	Q_{dB} $_{-5}$	$Q_{dB - 10}$	Q_{all}
Spectral ratio			
50	53	55	193
100	107	106	153
ΔQ	54	51	-40
Frequency shift			
50	53	54	62
100	107	107	110
ΔQ	54	53	48

Table C.1. Interval Q estimates for layer 5 in Figure C.3 (between A and B events). We compare SR and FS methodologies and different level of noise suppression. A low negative dB means a high signal to noise ratio. Note how SR are significantly affected by the (numerical) noise in the data when the whole data (all) is analyzed compared to FS. The true ΔQ is 50.

C.2 Estimating intrinsic Q

To compare the spectral ratio and frequency shift methods we model a 10 layer reservoir and create a synthetic common mid-point (CMP) seismogram. We generate two seismograms by changing the P-wave Q-log in which the quality factor for one layer changes from Q=100 (rock fully-saturated with a light hydrocarbon) to Q=50 (rock fully-saturated with brine). These traces resemble a time-lapse surface seismic acquisition at time 1 and time 2. The S-wave quality factor is set high (10^6) to only model P-wave attenuation. Figure C.3 shows the velocity, density and Q model and the generated synthetic seismogram with a 60 Hz Ricker wavelet. The synthetic seismogram includes: spherical divergence, transmission losses, higher order multiples and conversions. We model a CMP gather with a maximum offset of 200 m at 10 m receiver separation. The gathers are corrected by NMO (no stretching) with the true velocities. The stacked trace is analyzed for offsets less than 100 m. After selecting two events (top and bottom of the reservoir) and applying a 100 ms Hanning-type window, the discrete fast Fourier transform (DFFT) is computed to obtain the frequency spectrum. At the end both methodologies SR and FS are used to estimate Q. We use Hampson-Russell (Geoview version 4.3 CE7/R4) to model CMP anelastic gathers.

We generate two stacked traces for the two Q values in layer 5 (between A and B events Figure C.3), and obtain the time-lapse Q (Δ Q) by subtracting the two estimates for Q. The computed values for a Q of 100 and 50 are shown in Table C.1. Observe how the estimates even for this synthetic data are within 10% of the true value for high S/N. Our goal is to estimate time-lapse changes in Q rather than absolute Q. The true Δ Q is 50 and the data up to -5 dB estimates a Δ Q=54 for both methods. In this first test we can estimate the correct time-lapse Q within 10%.



Figure C.3. P-wave, density and Q_p logs for a homogeneous 10 layer model (left). Synthetic CMP gather corrected for NMO (right). Traces with offsets up to 100 m are stacked and this trace is used to estimate Q between the A and B events.

$\Delta x (m)$	$Q_{SR,-5dB}$	$Q_{SR,all}$	$Q_{FS,-5dB}$	$Q_{FS,all}$
646	65	300	61	71
574	50	229	50	62
502	55	2692	55	67
431	54	138	55	65
359	57	-106	58	63
287	54	82	54	58
215	53	48	54	56
144	50	73	$\overline{52}$	54
72	30	-207	$\overline{39}$	55

Table C.2. Q estimates for a wedge model by using two methodologies and different levels of noise suppression. The window is 100 ms. The true Q is 50.

Also note that when using all the data from the amplitude spectrum the spectral ratio estimate is largely affected, while for the frequency shift the estimate is more robust. The increase instability from SR comes from the nature of the methodology. SR is a point to point method: the amplitude ratio of two events are estimated for each frequency; and the data at each frequency contributes to the minimization of the difference between data and fit (least-squares fit to Equation C.1). By comparison, the FS methodology *fits* a Gaussian function to the amplitude spectrum of each event, estimate the central frequency and variance, and then take the ratio of these quantities to obtain Q. Actually, the FS methodology computes the central frequency and variance by integrating the amplitude spectrum rather than by least-square fitting a Gaussian function to the amplitude versus frequency data.

C.3 Wedge model

We generate a zero-offset synthetic seismogram of a wedge model to study the sensitivity of estimating Q to the layer thickness. Layer 5 in Figure C.3 decreases in thickness from 646 m to 72 m. λ is 50 m, thus the smallest modeled thickness is above tuning, but still has the interference between top and bottom reflectors. Modeling parameters and window is the same as in Section XX.

Table C.2 shows the estimates of Q with thickness for $Q_{true}=50$ (Q=100 shows similar results). We compare the frequency shift and the spectral ratio for a noise level up to -5 dB, and when including the whole data (*all*).

For all thicknesses, the estimate of Q is mostly within 10%. As before, when we do not limit the noise level in our data, the SR Q estimate is significantly distorted. However, for all thicknesses the FS estimate using the whole data is within 25% of the true value. The differences at 646 m and 72 m are mainly due to the choice of window. For the 646 m case, the estimates are affected because the window includes a strong multiple or conversion. When choosing a smaller window (40 ms), the Q estimate is 54 for both methods. For a thickness of 72 m, the top and bottom of the layer are resolved but both contained in the 100 ms window. Again, limiting the window (20 ms) gives a Q=58 for SR at -5 dB S/N.

APPENDIX D

TIME-LAPSE APPARENT ATTENUATION FROM SURFACE SEISMIC

In this section our surface seismic analysis focuses on three aspects: 1) window size effect on estimated time-lapse apparent attenuation, 2) $\Delta 1/Q_a$ estimates from zero-offset vs. stacked traces, and 3) random noise influence on attenuation estimates. We start by briefly describing the modeling approach. A synthetic shot gather is modeled with the same elastic and visco-elastic code as for the VSP from the previous section. The velocity, density and Q log model are shown in Figure D.1. The seismogram was modeled with frequencies between 0 and 150 Hz, with a dominant frequency of 40 Hz. The time sampling interval is 2 ms. The source and receivers are at the free surface. The minimum and maximum offsets are 25 m and 520 m respectively, at a 5 m interval. Between 500 and 1000 m, depth the layers have a strong contrast in velocity and density. It is in this depth interval that we varied the P-wave intrinsic Q from 100 to 50 as in the VSP study.

Figure D.2 is the surface seismic elastic response (shot gather) for the model in Figure D.1. The interval apparent attenuation is estimated between the events at 0.33 and 0.54 s, reflections corresponding to geological interfaces at 500 and 1000 m. Let's first analyze the zero-offset trace. Figure D.3 plots the elastic zero-offset trace with windowed top (a) and bottom (b) events for a 64 ms window. The computed spectra of the events is shown in Figure D.4. The estimated $1/Q_a$ (i.e. $1/Q_s$) by spectral ratios is 0.014. However, for surface seismic data, the window size can significantly affect the estimate of apparent attenuation. The contribution of P-to-S wave conversions and multiples can be part of the windowed trace when targeting a main reflection event.

Figure D.5 is the estimated apparent attenuation as a function of window size from the elastic zero-offset trace (triangles), and when including visco-elasticity with $Q_i=100$ (time-1, squares) and $Q_i=50$ (time-2, circles). The error bars are one standard deviation of the estimate of the random error in 1/Q (see Appendix A). The smallest window is 28 ms long which corresponds to about one dominant wave period (25 ms) and the largest window is 180 ms. Observe in Figure D.5 that for a window from 28 to 64 ms the estimate of apparent attenuation is sensitive to the window size. This sensitivity occurs because the event of interest is not recovered completely, as a result of applying a short Hanningtype window. However, our main interest is in the apparent time-lapse attenuation that, under the assumption that scattering attenuation is repeatable and cancels, is equal to the intrinsic $\Delta 1/Q$.

Figure D.6 shows the estimate of $\Delta 1/Q_a$ between the events at 0.33 and 0.54 ms. The true time-lapse intrinsic attenuation is plotted as a dashed line at 0.01. The $\Delta 1/Q_a$



Figure D.1. Velocity and density model used for generating synthetic surface seismic. Q_i is added between 500 and 1000 m ($Q_i=100$, time-1 changes to $Q_i=50$, time-2).

estimate significantly increases with window size, and gets more stable after 0.64 ms. For small window sizes (< 64 ms), the main reflection event windowed with a Hanning-type function has an incomplete spectrum producing low values of attenuation. Nonetheless we are still able to estimate the time-lapse intrinsic attenuation from 4D apparent attenuation if the selected window is appropriate.

Next we will compare the zero-offset and stacked traces. The elastic gather in Figure D.2 and the gathers with intrinsic Q were corrected for normal move-out (NMO) with the known (modeled) V_{rms} (Figure D.7) and stacked up to an offset of 225 m. The zerooffset and stacked traces are compared in Figure D.8. Because the modeled synthetic traces have no noise, and the offsets are short enough to not include amplitude vs. offset (AVO) effects, there is no difference between zero-offset and stacked traces at the 0.33 and 0.54 ms events. Therefore, the estimate of $\Delta 1/Q_a$ from these noise-free zero-offset or stacked traces are the same. If traces are noisy, stacking may improve estimates of time-lapse apparent attenuation.

As a last test, we add pseudo-random noise to the traces in the elastic and visco-elastic seismograms as described in the VSP noise section. The noise level was estimated with respect to the shallow reflector (at 0.33 ms). Figure D.9 shows part of the original elastic zero-offset trace along with traces obtained when three levels of noise are added. The estimated time-lapse apparent attenuation for the zero-offset trace is plotted in Figure D.10a. The estimate of apparent attenuation is consistent up to 5% noise, but having 10% noise on the trace can significantly affect the estimated attenuation. If we now stack the noisy



Figure D.2. Elastic response for the model in Figure D.1 for a surface seismic geometry (shot gather). The first offset is 25 m from the source and the offset interval is 5 m (maximum offset is 520 m). Interval $1/Q_a$ is estimated between reflections shown by the arrows (at 0.33 and 0.54 s).



Figure D.3. Elastic zero-offset trace (solid black line) with two windows of 64 ms each (red dashed lines) to isolate the top (a) and bottom (b) reflections.



Figure D.4. Amplitude spectra of the windowed events in Figure D.3.



Figure D.5. Interval apparent attenuation estimate for the elastic seismogram (triangles) and for visco-elasticity (Q=100, squares or Q=50, circles) between 0.33 and 0.54 s.



Figure D.6. Time-lapse apparent attenuation. The true $\Delta 1/Q_a = \Delta 1/Q_i$ is represented by the dashed line.



Figure D.7. Elastic shot gather with offsets limited to 225 m. The gather is NMO corrected with V_{rms} velocities estimated from the V_P log in Figure D.1.



Figure D.8. Zero-offset (solid black) and stacked (dashed pink) traces. The stacked traces result from stacking the CMP gathers up to an offset of 225 m. Each plot is for different Q_i models.



Figure D.9. Zero-offset noise-free trace (original) and with three levels of noise added.

traces up to 225 m offset, the time-lapse apparent attenuation is stable for all noise levels (Figure D.10b). This stability is no surprise, as random noise is expected to cancel when traces are stacked. As long as the AVO influence is limited (short offsets), and noise can be assumed random, working with stacked traces over zero-offset traces from randomly noisy gathers is advantageous.

Although there is a long list of possible studies and tests to be done on surface seismic data for time-lapse attenuation, here we only wanted to answer the two main questions posed at the beginning of this chapter: 1) Is $1/Q_a=1/Q_s+1/Q_i$? and 2) if $1/Q_s$ is repeatable, is $\Delta 1/Q_a=\Delta 1/Q_i$? For modeled zero-offset and surface seismic traces, these two questions are answered in the affirmative, and the next steps in future research would be to perform similar analysis on real seismic data. However, there are challenges in time-lapse intrinsic attenuation studies which we will examine next.



Figure D.10. Apparent estimates of attenuation for individual times and time-lapse difference as a function of random noise from: a) zero-offset trace and b) NMO corrected and stacked trace up to 225 m offset.