ELASTIC MODULI AND SEISMIC WAVE ATTENUATION IN DRY AND SATURATED ROCK: MODULUS DISPERSION AND ATTENUATION

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Four different laboratory techniques were used to determine Young's modulus and extensional wave attenuation as a function of frequency for the same rock specimen while minimizing variations in other important parameters. The data were then pieced together and compared over a net frequency bandwidth ranging from $10^{-2}$ to $10^6$ Hz. Specimens of Sierra White granite and Berea sandstone were analyzed in dry and water saturated states.

Dispersion in moduli and attenuation in the dry samples are negligible. Frequency dependent attenuation and modulus dispersion are clearly evident in the saturated samples and attributed to flow of pore fluid, unconfined sample boundaries and sample dimensions.

A cyclic loading technique was used to examine the effects of strain amplitude and water saturation at constant frequency. The modulus and attenuation determined from hysteresis loops are strain amplitude dependent in both dry and saturated states. As strain amplitudes are increased above $5 \times 10^{-6}$, the modulus continually drops while attenuation increases. A significant reduction in modulus is observed in the saturated samples that is attributed to a reduction in surface energy at the fluid/rock interface and an unconfined sample radial boundary.
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MODULUS DISPERSION AND ATTENUATION

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Large changes in elastic moduli and seismic wave attenuation in the source region resulting from a nuclear explosion influence seismic pulse characteristics observed in the far field. It is concluded from a number of laboratory studies that dispersions in moduli and attenuation occur over several orders of magnitude in frequency. However, these data have been collected using several methods where frequency bandwidth, strain amplitude, sample dimension, boundary conditions and fluid type vary considerably between data sets.

In this study, four different laboratory techniques were used to determine Young's modulus and extensional wave attenuation as a function of frequency for the same rock specimen while minimizing variations in other important parameters. The data were then pieced together and compared over a net frequency bandwidth ranging from 10^-2 to 10^6 Hz. Specimens of Sierra White granite and Berea sandstone were analyzed in dry and water saturated states. Dispersions in moduli and attenuation in the dry samples are negligible. Frequency dependent attenuation and modulus dispersion are clearly evident in the saturated samples and attributed to flow of pore fluid, unconfined sample boundaries and sample dimensions.

A cyclic loading technique was used to examine the effects of strain amplitude and water saturation at constant frequency. The modulus and attenuation determined from hysteresis loops are strain amplitude dependent in both dry and saturated states. As strain amplitudes are increased above 5 x 10^-6, the modulus continually drops while attenuation increases. A significant reduction in modulus is observed in the saturated samples that is attributed to a reduction in surface energy at the fluid/rock interface and an unconfined sample radial boundary.
INTRODUCTION

Mechanical rock properties in the vicinity of a high strain seismic source significantly influence the pulse characteristics observed in the far field. Within the source region, strain amplitudes vary several orders in magnitude while the host rock exhibits components of inelastic, anelastic, and elastic behavior. As a result, dominant changes in the seismic pulse occur within a few wavelengths of the source itself. Dispersions in seismic wave velocities, elastic moduli, and seismic wave attenuations are exhibited that depend on the loading conditions and the response of the host rock to loading. These observations can be attributed to the time and strain amplitude dependencies in the anelastic response of rock to varying loading conditions.

Anelastic mechanisms in rock produce a phase lag of strain behind stress and affect velocity, moduli and attenuation. The modulus, $M$, is represented by a complex quantity (Toksoz and Johnson, 1981),

$$M = M_R + iM_I$$

where $M_R$ and $M_I$ are the real and imaginary parts of the modulus, respectively. The phase lag, $\phi$, is a direct measure of the intrinsic loss or attenuation, $Q^{-1}$. Attenuation can be determined from the energy input to the system relative to the energy lost to the system (Toksoz and Johnson, 1981),

$$Q^{-1} = M_I / M_R = \Delta W / 2\pi W = \tan \phi$$

where $W$ is the elastic energy stored at the maximum stress and strain and $\Delta W$ is the loss per cycle.

Laboratory measurements offer a convenient means to determine these properties with imposed conditions similar to those in the source region. It is important to define the correlations between properties determined in the field at seismic frequencies and those determined in the laboratory or overlapping seismic or at different frequencies. Significant differences in elastic moduli and wave attenuation have been observed for a given rock
measured with different laboratory and field techniques. A number of investigators have reported frequency, strain amplitude, and fluid saturation effects on $M$ and $Q^{-1}$ in crustal rocks (Gordon and Davis, 1968; Tittmann et al., 1977; Toksoz et al., 1979; Winkler et al., 1979; Spencer, 1981; Dunn, 1986; Paffenholz and Burkhardt, 1989; Lucet et al., 1991).

Spencer (1981) used a cyclic loading technique to examine the effects of frequency and fluid saturation in samples of sandstone, limestone, and granite for frequencies ranging from 4 to 400 Hz. He found that Young's modulus and extensional wave attenuation remained relatively constant over frequency in samples in a vacuum dry state. However, in saturated samples, the modulus and attenuation exhibited dependencies on fluid type and frequency. For all the fluids, similar trends in modulus dispersion and attenuation were observed. Reductions in moduli and increases in attenuation occurred due to the presence of pore fluid. The modulus increased with increasing frequency, eventually converging to the modulus of the dry condition. Attenuation increased with frequency more than an order of magnitude at frequencies above 100 Hz. Moreover, the degree of modulus reduction and attenuation increase showed a pronounced dependence on fluid type. The fluids had similar densities and viscosities but had very different chemical, electrochemical, and dielectric properties. He concluded that electrochemical interactions between the pore fluid and mineral surface were responsible for a reduction in surface energy at the fluid/pore wall interface. In turn, a frequency-dependent softening of the rock was generated. Spencer was able to predict the correlations between modulus dispersion and attenuation in fluid saturated and dry rocks using a Cole-Cole distribution (Cole and Cole, 1941) of stress relaxation times.

Tittmann et al. (1977), Tittmann et al. (1980), Bullau et al. (1984), and Lucet et al. (1991) used a resonant bar technique to measure modulus and $Q^{-1}$ as a function of water saturation at frequencies between $5 \times 10^3$ and $2 \times 10^4$ Hz for a wide variety of igneous and sedimentary rocks. In general, attenuation rose while moduli dropped as the partial pressure of water increased. For example, as the vapor pressure in a specimen of Berea sandstone was increased from vacuum to 35% relative humidity, $Q^{-1}$ rose by a factor of four while Young's modulus dropped 16% in magnitude.

Attenuation and elastic moduli determined at ultrasonic frequencies have been reported by several investigators (e.g. Gist, 1992; Toksoz et al., 1979; Coyner, 1984). At frequencies above $0.5 \times 10^6$ Hz, the Young's modulus in water saturated rocks is typically greater than that of dry rock. Extensional and compressional wave attenuations are often larger in dry rock than those of saturated. These observations are the opposite of those observed at frequencies less than $0.5 \times 10^6$ Hz. At ultrasonic frequencies, fluid content increases shear wave attenuation while
lowering the shear modulus. Toksoz et al., 1979 also noted that compressional, extensional, and shear wave attenuations decreased with an increase in confining pressure.

Recently, Tang (1991), developed a waveform inversion technique to measure moduli and attenuation in rock samples overlapping resonant bar and ultrasonic frequencies. He reported that attenuation began to increase above $5 \times 10^4$ Hz in dry samples of Sierra White granite. This increase was attributed to the decreasing wavelength and scattering interactions with sample grains, heterogeneities, and flaws.

Moduli and attenuation have been shown to change significantly with increasing strain amplitude in a variety of rock types and saturation conditions in experiments where frequency was held constant (Winkler et al., 1979; Johnson and Toksoz, 1980; Bulau et al., 1984; Martin et al., 1990). Winkler et al. (1979) used the cyclic loading technique at frequencies less than 1 Hz to examine the effects of strain amplitude and water saturation on stress-strain hysteresis loops. He observed that large changes in the modulus and attenuation occur in dry rock at strain amplitudes above $10^{-6}$; the modulus decreased while attenuation increased. Below this amplitude, the modulus and attenuation typically varied by a small percentage. For saturated samples, the same modulus and attenuation trends were observed. However, the onset of dominant changes in these properties occurred at a slightly lower strain amplitude than $10^{-6}$.

**Focus of problem**

These experimental data clearly demonstrate that elastic moduli and seismic wave attenuation depend on frequency, strain amplitude, fluid saturation, and confining pressure. However, these different laboratory methods are limited to specific frequency bandwidths with little to no overlap of each other. Using a single technique, only portions of certain important events could be measured for a given rock sample. Modulus dispersion and changes in attenuation occur over several orders of magnitude in frequency. For example, Spencer, 1981 was able to measure part of the low frequency end to 400 Hz of predicted attenuation peaks and modulus ramps that are defined from 1 to $10^5$ Hz. Other techniques have been used to measure these properties at higher frequencies but strain amplitudes, sample geometries and boundary conditions were considerably different than those of Spencer's experiments. Characterizing the complete frequency dependence for a specific rock by piecing existing data together becomes difficult, since these data come from such diverse sources.

To overcome this problem, it is necessary to perform a series of experiments that quantify rock properties over many orders of magnitude in frequency for a single rock specimen while
holding other conditions constant. The approach here was to use multiple techniques on the same rock core to determine Young's modulus and $Q_E^{-1}$ as a function of frequency, water saturation and strain amplitude.
EXPERIMENTAL PROCEDURE

Two basic studies were performed to examine the influence of frequency, strain amplitude, and water saturation on moduli and attenuation measured in laboratory experiments. The first study investigated the effects of frequency and water saturation over a large frequency band ranging from $10^{-2}$ to $10^{6}$ Hz. The cyclic loading, resonant bar, wavefrom inversion and ultrasonic velocity methods were employed to determine Young's modulus and $Q_E$ as a function of frequency within their respective bandwidths. In the second study, the cyclic loading technique was used to investigate the strain amplitude and stress dependence of moduli and attenuation in dry and water saturated rocks. Stress-strain hysteresis loops with peak strain amplitudes ranging from $10^{-7}$ to $10^{-4}$ were generated while holding loading frequency constant at $10^{-1}$ Hz.

In order to compare the data obtained with each of the methods over the complete bandwidth, it is necessary to account for differences in strain amplitude, sample geometry and sample size required for each technique. Variations in strain amplitude, saturation, preload, and ambient conditions were minimized as much as possible between the different methods. The ultrasonic velocity, resonant bar, and waveform inversion measurements utilize pulse propagation techniques with peak strain amplitudes on the order of $10^{-8}$ to $10^{-6}$. These data were compared with cyclic loading tests with strain amplitudes less than $5 \times 10^{-6}$. All measurements were carried out under room dry and fully water saturated states. The radial boundary of the samples was unconfined and at ambient room temperature, pressure and humidity. Table 1 lists the experimental techniques with corresponding sample dimensions and frequency bandwidths.

Measurements were made on representative specimens of two different and commonly found rock types: granites and sandstones. Each rock type has considerable differences in porosity, pore geometry, and other rock properties. Sierra White granite and Berea sandstone specimens from Raymond, CA and Berea, OH, were selected since they have been widely studied in many applications. The pore structure of Sierra White granite is characterized by populations of penny shaped microcracks (Martin et al., 1990). Berea sandstone contains spherical pores situated between cemented quartz grains with a percentage of clay between 10-20%. The porosities of the granite and sandstone samples were 0.9% and 17.8%, respectively.
<table>
<thead>
<tr>
<th>Technique</th>
<th>Frequency</th>
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<td>305</td>
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<td></td>
<td></td>
<td>Berea</td>
<td>305</td>
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<tr>
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<td></td>
<td>Berea</td>
<td>180</td>
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<td>Berea</td>
<td>165</td>
</tr>
<tr>
<td>Waveform inversion</td>
<td>1 - 200 kHz</td>
<td>Sierra White</td>
<td>150</td>
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<td></td>
<td>Berea</td>
<td>150</td>
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<td></td>
<td></td>
<td>Berea</td>
<td>40</td>
</tr>
</tbody>
</table>

**Sample preparation**

All samples were cored from a single block in the same orientation; the sample axis normal to bedding. Separate samples were dimensioned for each measurement technique. Test specimens were ground right circular cylinders. Some samples used in the resonant bar apparatus were square rods and are specified in Appendix C. All samples excluding those for the cyclic loading tests were initially vacuum dried for a period of 24 hours, then allowed to equilibrate to room humidity. The specimens were tested in the room dry state first, then they were saturated with water, then jacketed to minimize evaporation losses, and the measurement sequence was repeated. The large samples used in cyclic loading and those in ultrasonic tests were jacketed with cellophane. Resonant bar and waveform inversion samples were wrapped in Teflon tape.
Cyclic Loading (Hysteresis loop)

Cyclic loading tests were conducted on cylindrical samples housed in a servo-controlled loading frame. A continuous sinusoidal axial force was applied to one sample end with an axial ram operating in force-feedback mode. The force was measured with a load cell in series with the sample in the loading column to determine stress. Three axial LVDT displacement transducers were mounted to the sample with circumferential rings and used to determine strain. An axial preload of 1 MPa was applied to minimize mechanical losses across interfaces in the loading column.

Cyclic loads were prescribed from continuous sine functions at single frequencies ranging from $10^{-2}$ to $10^2$ Hz. For frequencies below 5 Hz, the modulus and $Q$ were determined from single loops, with 1400 points of data per cycle. The modulus was determined from the slope of the hysteresis loop (stress over strain) while attenuation was computed from the fractional energy loss per cycle (area inside the loop divided by the maximum stress). For frequencies at and above 5 Hz, up to 200 cycles were collected. Young's modulus and $Q_E^{-1}$ were determined from the Fast Fourier Transforms of the multiple cycle stress and strain sinusoids. The complex modulus was calculated from the ratio of the peak stress and strain spectra. $Q_E^{-1}$ was then computed from the ratio of the real and imaginary parts of the complex Young's modulus.

The experimental procedure was then duplicated for a low loss standard aluminum core with the same dimensions to calibrate the hysteresis of the experimental apparatus. The experimental apparatus, details of the experimental data acquisition, processing, and computation schemes are described further in Appendix A.

Resonant Bar

Benchtop resonant bar experiments were conducted on rock specimens suspended by a fine thread supported by a fixture in a modified desiccator bell-jar. Samples were thin cylindrical or square rods with a length to diameter ratio greater than 10. A standing extensional wave was generated from piezoelectric crystals mounted on the sample ends. A lock-in amplifier was then used to resonant the rod at fundamental resonances and at higher harmonics, up to the 18th harmonic in dry samples. As the mode or harmonic increased, the center frequency also increased. Young's modulus and $Q_E^{-1}$ were determined as a function of frequency from the resonance peaks. Extensional velocity was first computed from the center frequency of the resonant peak and the sample length. Young's modulus was then computed from the velocity and sample density. $Q_E^{-1}$ was computed from the center frequency and half-power bandwidth.
of the resonance peak. The mass of the piezoelectric crystals and jackets used for the saturated samples were accounted for in a correction factor described by Lucet et al., (1991). Further details on the resonant bar measurements, data processing, and calculations are presented in Appendix B.

**Waveform Inversion**

Elastic wave velocities and attenuation were determined jointly from frequencies ranging from $10^3$ to $2 \times 10^5$ Hz. The technique involves measuring waveforms in two cylindrical bars of the same material, but of different length. First, roughly estimated bar velocity and Poisson's ratio were used to propagate the waveform observed in the shorter bar to the length of the longer bar. Guided by the velocity model, the velocity dispersion curve was then computed from the model, the measured phase difference between the propagated waveform and the measured waveform within the longer bar. Using Rayleigh's solution for the low frequency phase velocities, the low frequency portion of the dispersion curve was inverted to obtain an accurate estimate of the bar velocity and Young's modulus.

$Q_E^{-1}$ was computed in the following steps. The waveform of the shorter bar was theoretically continued to the length of the longer bar to match the measured waveform. The waveform inversion then was performed to minimize the amplitude difference between the two waveforms, from which $Q_E^{-1}$ was computed. For further detail on the waveform inversion technique, see Tang (1992). Further details on the waveform inversion measurements are discussed in Appendix C.

**Ultrasonic Measurements**

A benchtop ultrasonic system was used to record compressional and shear wave travel times and waveforms propagating through the axis of rock cores. The sample core is placed in series between source and receiving piezoelectric ultrasonic transducers housed in the benchtop system. The source transducer was driven with a short electrical pulse with a center frequency of $10^4$ Hz. A pneumatic actuator controls a small ram which exerts an axial preload to the sample on the order of 1 MPa. This preload is comparable to that of the cyclic loading experiments. Compressional (P) and shear (S) wave velocities for the rock were obtained by measuring the one-way travel time of a P or S wave through the core and dividing by its sample length. Dynamic Young's modulus was then computed from the compressional and shear wave velocities and sample density.
$Q_{E}^{-1}$ was computed from values of $Q_{p}$ and $Q_{S}$ (P and S wave attenuation) determined with the spectral ratio technique using a low loss standard aluminum core for comparison with rock (Toksoz et al., 1979). P and S waves are recorded for an aluminum core with the same dimensions as the rock specimen. The Fast Fourier Transforms are computed from the P and S wave time series. The $Q$ of the rock is computed from the slope of ratio of the aluminum and rock spectra versus frequency. Further details on the ultrasonic measurement and data reduction are presented in Appendix D.
EXPERIMENTAL RESULTS

The effects of frequency, strain amplitude, and water saturation on Young's modulus and extensional wave attenuation have been examined in specimens of Sierra White granite and Berea sandstone. Four different laboratory techniques were used to acquire data over a large net frequency band ranging from $10^{-2}$ to $10^6$ Hz. Dry and water saturated states were then compared as a function of frequency. The cyclic loading technique was employed to determine rock property at strain amplitudes ranging from $10^{-7}$ to $10^{-4}$ at constant frequency. Dry and water saturated states were compared as a function of strain amplitude.

Frequency Tests

*Sierra White granite*

Young's modulus and $Q_E^{-1}$ are compared for room dry and water saturated rock as a function of frequency in Table 2 and shown in Figures 1 and 2. Moduli obtained from cyclic loading, resonant bar (first mode), and ultrasonic velocity measurements are shown in Figure 1. In the dry state, the modulus remains relatively constant with increasing frequency. The modulus determined from the waveform inversion technique at $25 \times 10^3$ Hz is more than 15% lower than those at other frequencies. A reduction in modulus is exhibited in saturated samples at frequencies less than 100 Hz. At $10^{-2}$ Hz, the modulus reduction from dry to saturated states is more than 20%.

As frequency is increased, the modulus of the saturated sample increases from 30.3 GPa at $10^{-2}$ Hz to 37.3 GPa at $10^2$ Hz, converging in magnitude to that of the dry modulus. The resonant bar dry and saturated values differ by less than 1%. At ultrasonic frequencies, the modulus in the saturated sample is considerably larger than that of dry rock modulus by approximately 25%. In addition, the modulus determined in the saturated rock at ultrasonic frequencies is almost double that of the cyclic load value measured at $10^{-2}$ Hz.

Extensional wave attenuation in Sierra White granite is presented as a function of frequency in Figure 2 for cyclic loading, all modes of the resonant bar, waveform inversion, and ultrasonic measurements. $Q_E^{-1}$ was only determined with the waveform inversion technique in the room dry condition. For dry granite, $Q_E^{-1}$ is consistently less than 0.01 over frequency. A slight increase is observed especially at ultrasonic frequencies. $Q_E^{-1}$ for the saturated condition exhibits a pronounced frequency dependence and is significantly greater than $Q_E^{-1}$ for the dry
<table>
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<th>Table 2  Frequency Dependence: Sierra White Granite</th>
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</table>


Sierra White granite

(o) dry
(●) saturated

Youngs modulus (GPa)

Frequency (Hz)

Figure 1
Sierra White granite

- (o) dry
- (●) saturated

**Figure 2**

Log-log plot showing attenuation as a function of frequency for Sierra White granite.

- Axis labels:
  - Y-axis: Attenuation (1000/Qe)
  - X-axis: Frequency (Hz)

- Key points:
  - Hysteresis loop
  - Resonance bar
  - Wavenumber inversion
  - Ultra
state at all frequencies less than $10^5$ Hz. Attenuation measured in the large sample with the cyclic loading technique continually increases between $10^{-2}$ and 50 Hz, as much as a factor of six. In contrast, the resonant bar data show a large decrease from 0.045 to 0.025 as frequency increases from $10^3$ to $10^5$ Hz. Attenuation in the saturated sample drops even further at ultrasonic frequencies, just slightly less than that of the dry rock.

*Berea Sandstone*

The effects of frequency on Young's modulus and $Q_E^{-1}$ for room dry and water saturated sandstone are presented in Table 3 and shown in Figures 3 and 4. A large data set was compiled for the dry state including cyclic loading, resonant bar, and ultrasonic measurements. However, cyclic loading data are only reported for the saturated sample. The modulus in the dry state remains relatively constant over frequency. A slight second order increase in modulus is observed, particularly at ultrasonic frequencies. A significant reduction in Young's modulus in saturated sandstone is also documented that is lower than the dry values by approximately 40% at all frequencies less than $10^2$ Hz. The modulus of the saturated sample increases less than 6% over frequencies to $10^2$ Hz.

Extensional wave attenuation in Berea sandstone is shown as a function of frequency in Figure 4. Values of attenuation in dry specimens were determined using all four techniques and compared. For a given technique, changes in $Q_E^{-1}$ are relatively independent of frequency. Cyclic loading and waveform inversion attenuation data are approximately in the same range, averaging 0.03. However, resonant bar attenuations are typically three times less in magnitude than cyclic loading and waveform inversion data. Attenuation determined with the ultrasonic technique is almost double than those of cyclic loads and waveform inversion data.

Attenuation measurements in saturated sandstone are reported only for the cyclic loading technique. In these data, $Q_E^{-1}$ is less than that of the dry state. For the large sandstone sample changes $Q_E^{-1}$ appear to be independent of frequency using this technique.

**Strain Amplitude Tests**

*Sierra White granite*

The effects of strain amplitude and water saturation are examined using the cyclic loading technique. Stress-strain hysteresis loops were generated at a frequency of $10^{-1}$ Hz and analyzed. Young's modulus and $Q_E^{-1}$ for both the room dry and saturated conditions are presented as a
<table>
<thead>
<tr>
<th>Technique</th>
<th>Frequency (Hz)</th>
<th>Attenuation (1000/Qf&lt;sup&gt;-1&lt;/sup&gt;)</th>
<th>Young's mod (GPa)</th>
<th>Frequency (Hz)</th>
<th>Attenuation (1000/Qf&lt;sup&gt;-1&lt;/sup&gt;)</th>
<th>Young's mod (GPa)</th>
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**Table 3: Frequency Dependence - Berea Sandstone**

**Dry State**

**Saturated State**
Berea sandstone

(o) dry

(●) saturated

Young's modulus (GPa)

Frequency (Hz)

Figure 2
function of strain amplitude in Table 4 and shown in Figures 5 and 6. In Figure 5, Young's modulus increases with strain amplitude in the room dry and saturated samples. The changes were most pronounced for strains above $5 \times 10^{-6}$. The modulus trends of the dry and saturated states are very similar. However, these trends are offset by a reduction in modulus from dry to saturated states.

An increase in strain amplitude also produces an increase in $Q_E^{-1}$ that has similar trends in both the dry and saturated specimens as depicted in Figure 6. $Q_E^{-1}$ remains constant below strain amplitudes of $5 \times 10^{-6}$ in dry granite, but shows a sharp increase above this amplitude. For the saturated condition, the sharp increase in $Q_E^{-1}$ is shifted to strain amplitudes as low as $10^{-7}$ to $10^{-6}$. It appears that the rate of increase at high strain amplitudes is approximately the same for dry and saturated states.

**Berea Sandstone**

Strain amplitude affects Young's modulus and $Q_E^{-1}$ in both the room dry and saturated sandstone sample. Results are presented in Table 5 and are shown in Figures 7 and 8. In Figure 7, the room dry and saturated samples exhibit a decrease in modulus with increasing strain amplitude, particularly above $5 \times 10^{-6}$. Data are also presented for 10 Hz hysteresis loops for comparison at a higher frequency in the saturated sandstone. The differences in moduli computed at $10^{-1}$ and 10 Hz are very small at a given strain amplitude. These data are consistently lower by as much as 40% at strain amplitudes of $10^{-6}$ compared to those of the dry state. This difference is less than 20% at larger strain amplitudes.

In dry and saturated sandstone, an increase in strain amplitude produces an increase in $Q_E^{-1}$, as depicted in Figure 8. For dry and saturated states, attenuation increases significantly at strain amplitudes above $5 \times 10^{-6}$. A small spread in the data is exhibited between the dry and saturated states. In these data, the attenuation in saturated sandstone is slightly less than that of dry rock above a strain of $5 \times 10^{-6}$. 
### Table 4: Strain Amplitude Dependence Sierra White Granite
(axial preload 1 MPa)

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### Table 5: Strain Amplitude Dependence Berea Sandstone
(axial preload 1 MPa)

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Sierra White granite

(o) dry
(c) saturated

0.1 Hz hysteresis loop

Youngs modulus (GPa)

Strain Amplitude

FIGURE 5
Berea sandstone: Effect of strain amplitude on modulus

Hysteresis loops
(x) dry sample (0.1 Hz)
(o) saturated sample (0.1 Hz)
(+) saturated sample (10 Hz)

FIGURE 7
Berea sandstone: Effect of strain amplitude on attenuation

Hysteresis loops

- (x) dry sample (0.1 Hz)
- (o) saturated sample (0.1 Hz)
- (+) saturated sample (10 Hz)

Figure 8
DISCUSSION

Changes in moduli and attenuation in granite and sandstone samples are observed as functions of frequency and strain amplitude. The magnitude and frequency dependence of these rock properties are strongly affected by the presence of pore fluid. Moreover, there are considerable differences exhibited between rock types. As the axial load is increased, producing strain amplitudes above $5 \times 10^{-6}$, a predominant strain amplitude dependence is observed in both rock types in dry and saturated states. These dependences are controlled by the pore structure, sample dimension, sample geometry, and loading conditions.

The effects of water saturation and frequency

Dry state

Modulus dispersion and changes in attenuation are independent of frequency to the first order in dry granite and sandstone. However, frequency dependences are exhibited above $10^5$ Hz. Trends in resonant bar, waveform inversion and ultrasonic data show an increase in attenuation. This increase can be attributed to scattering effects where the wavelength begins to approach the scale of sample grain size, pore size, and heterogeneities. This increase is more pronounced in the sandstone than the granite. The pores and sample inconsistencies in the sandstone are characterized to be larger than those in granite.

A large variation in attenuation is evident in Berea sandstone between the waveform inversion and resonant bar data at overlapping frequencies. Resonant bar attenuation measurements are particularly sensitive to changes in humidity (Tittmann, et al., 1977 and 1980). Humidity effects are also documented in Appendix C describing resonant bar experiments in greater detail. When vacuum dry samples are exposed to atmosphere conditions, attenuation increases by several orders of magnitude as the sample absorbs humid air over time. The associated modulus is also observed to decrease between 10 - 20% over the time span. Resonant bar and waveform inversion measurements were obtained in two different building locations and times of year. The resonant bar measurements for the sandstone were obtained in several winter months where humidity was low. The lower attenuations are then attributed to humidity and the ability of the sample to absorb moisture. Little to no differences in attenuation are observed between the waveform inversion and resonant bar data in dry granite. The granite samples have a much lower permeability than sandstone, in turn humidity infiltration into
granite may not play as much as a factor. The humidity dependence in waveform inversion measurements has not been investigated.

**Saturated state**

The presence of pore fluid, change in frequency and sample dimensions have significant and inter-related effects on the modulus and attenuation. A large reduction in Young's modulus is documented in the samples using the cyclic loading tests, especially at very low frequencies. The modulus reduction can be attributed to a decrease in surface energy along the grain boundaries, microcracks, and pores within the rock caused by the pore fluid (Murphy, 1982; Bulau et al., 1984; Murphy et al., 1984). Strain in rock causes solid bonds to break in the vicinity of crack tips and junctions between grain boundaries increasing the net surface area. Anions in pore fluid are adsorbed onto the newly exposed surfaces to complete the broken solid bonds. Cations are then hydrated to the highly viscous anion layer forming an electric double layer at the mineral/fluid interface. The bonds in the electric double layer are considerably weaker than solid bonds, in turn the surface energy has been reduced. Spencer, 1981 and Pride and Morgan, 1991 concluded that the surface energy of the double layer depends pore fluid chemistry, the mineral surface electrical charge, and electrochemical interactions occurring between the mineral and fluid. As a result, reductions in the modulus depend on the electrolyte species and concentrations contained in the pore fluid.

Tap water was used to saturate each specimen for all experiments. It is assumed that differences in electrochemistry for the experiments were negligible. Results show that the modulus reduction is more pronounced in sandstone than in granite. This is due, at least in part, to the pore geometry and solid-solid bond strength between grain boundaries. Cementation in sandstone is weaker than the quartz grains and strains more easily to stress. The addition of pore fluid softens the rock even further, especially compared to a crystalline rock.

A number of mechanisms are primarily responsible for an increase in wave attenuation in saturated rocks. Energy is lost to move pore fluid. Biot, 1956; Walsh, 1969; O'Connell and Budiansky, 1977; and Mavko and Nur, 1979, have developed fluid-mechanical models to describe viscous and inertial effects of pore fluid as dominant mechanisms of attenuation. The elastic deformation resulting from a stress produces pressure gradients in pore fluid and generates fluid flow. The viscous drag of the pore fluid also depends on the bond strength of the electric double layer that is controlled by fluid electrochemistry (Spencer, 1981, and Pride and Morgan, 1991). The stronger the bond strength between the fluid and mineral, the stronger
is the coupling between the fluid and mineral surface. Spencer demonstrated that attenuation varied considerably at a given frequency in the same rock specimen using fluids with different electrical properties and electrolytes but with similar densities and viscosities.

**Frequency dependence in saturated rock**

Both the modulus and attenuation exhibit a frequency dependence, especially in Sierra White granite. Young's modulus almost doubles over the entire frequency range in the granite specimens. A dramatic increase in the modulus is observed with increasing frequency in the large sample used in the cyclic loading tests. The frequency dependence of the modulus can be attributed to the apparent viscosity and fluid flow within the sample. When the cyclic loading rate exceeds the time constant for fluid diffusion within the pore space, the specimen begins to stiffen due to the compression of the pore water. In turn, the effective modulus increases. The point at which the relative stiffening occurs is directly related to the sample diameter, pore geometry, and the boundary conditions for fluid flow within the specimen (Spencer, 1981; Dunn, 1987).

At frequencies between $10^{-2}$ and $10^2$ Hz, attenuation increases from 0.012 to 0.060 in granite. This increase can be attributed to a number of mechanisms. Fluid-mechanical models also describe changes in attenuation with loading frequency. The viscous drag of the pore fluid is strongly dependent on frequency. As the loading frequency increases, the rate of fluid flow increases and viscous losses control the attenuation. Once the deformation rate is significantly greater than the pore fluid response, stiffening occurs and attenuation decreases. This decrease is clearly demonstrated in the resonant bar and ultrasonic data.

Fluid flow is also governed by the sample pore geometry, sample dimensions and boundary conditions imposed by the experimental set up. Relatively small increases in the modulus and attenuation in Berea sandstone are observed over frequency (only cyclic loads were examined in this study). Differences in modulus dispersion and attenuation in the granite and sandstone samples are primarily attributed to pore geometry. The permeability of the granite specimen is considerably smaller than that of sandstone. As a result, pressure gradients that drive the fluid flow build up at a faster rate in the granite with frequency. In turn, the rock stiffens at a faster rate and attenuation becomes significantly greater with frequency. Another critical factor is that the specimens were tested with a free radial boundary under drained conditions. Flow in and out of the sample is considerably larger for the sandstone sample than granite. The magnitude of the pressure gradients in sandstone are in turn smaller and stiffening of the fluid was
minimal. The rate of stiffening is also controlled by sample radius. Dunn, 1986, showed that stiffening occurred at lower frequencies with larger diameter samples.

**The effect of strain amplitude**

Young's modulus and seismic wave attenuation clearly exhibit a strain amplitude dependence in dry and saturated Sierra White granite and Berea sandstone. For rocks with low aspect ratio cracks (granite) or elliptical pores (sandstone), the modulus has been shown to decrease while attenuation increases with increasing strain amplitude (Winkler et al., 1979). Walsh, (1969), Mavko (1979), and Winkler et al., (1979) have proposed frictional mechanisms in which sliding on microcracks dissipates energy, observed as attenuation. Alternative mechanisms based on changes in surface energy and movement of fluid films in partially saturated rocks have been proposed by Spencer, (1981), Mavko and Nur, (1979), Bulau et al., (1984) and Murphy et al. (1984).

The dry and water saturated specimens showed similar trends in modulus and attenuation changes with strain amplitude. The modulus trends in dry and saturated granite mimic each other closely. The significant differences in these trends are in their magnitudes where the modulus in the saturated state is consistently lower than dry state by 25%. The attenuation is larger at a given strain amplitude in saturated granite. However, there are subtle but important differences that occur at different strain amplitudes.

Below 5 x 10^{-6} strain, the modulus and attenuation are relatively independent of strain amplitude in dry granite. Above this threshold, the modulus decreases while attenuation rises with strain amplitude. The deformation and strain due to loading is able to overcome frictional forces and initiate sliding of surfaces within microcracks. In turn, the strain becomes more nonlinear with increasing stress and effects the modulus and attenuation. These nonlinear effects are discussed in further detail by Winkler et al., 1979 and Martin et al., 1990. In saturated rock, attenuation begins to rise at a lower strain amplitude threshold. The hydrated layers on mineral surfaces reduce surface energy and the forces resisting microcrack sliding. This threshold occurs between 10^{-7} to 10^{-6} strain in the granite.

The strain amplitude dependence of the modulus and attenuation in Berea sandstone is similar to that of the granite. The modulus trends in the dry and saturated sandstone are different. The modulus in the dry rock drops at a faster rate than that in saturated rock with increasing strain amplitude. Attenuation in dry and saturated sandstone are approximately the
same value at a given strain amplitude. It is difficult to explain these results. One possible explanation is that the clay content in Berea sandstone may have an effect on the modulus and attenuation. Clay surfaces have a significant effect on the adsorption and desorption processes involving pore fluid. In turn, clays can swell when in contact with water and perhaps lower attenuation. As a result, the net effect of clay and fluid flow on attenuation may be equivalent to the dry state in clay rich sandstone.
CONCLUSIONS

Mechanical rock properties such as elastic moduli and seismic wave attenuation were determined experimentally in the laboratory. Four different techniques were used to compare moduli and attenuation over many orders of frequency. Data were examined at frequencies and strain amplitudes that are observed in the explosive source region. These data can then be incorporated into existing models to compute far field seismic pulse characteristics. By characterizing rock properties at a test site, it may be possible to discriminate the yields of nuclear explosions from natural teleseismic events using far field signals.

The moduli and attenuations determined from the experiments have been compared as a function of frequency, strain amplitude, and saturation. It is concluded from the data that:

(1) The changes in modulus and attenuation are small in the dry Sierra White granite and Berea sandstone. Modulus dispersion and frequency dependent attenuation are exhibited in the saturated samples. The frequency dependence is attributed to the flow of pore fluid caused by the loading stress and sample geometry. Increased stiffening occurs with frequency which causes an increase in modulus. At higher frequencies, the fluid motion cannot keep up with the rate of deformation. As a result, attenuation decreases and approaches that of dry rock.

(2) The effect of saturation in the unconfined samples produces a significant reduction in Young's modulus and increase in attenuation at low frequencies. A reduction in surface energy caused by the presence of pore fluid softens the rock lowering the modulus. Fluid flow and viscous drag increase attenuation.

(3) Both the modulus and attenuation are strain amplitude dependent in dry and saturated samples. The modulus decreases while attenuation decreases significantly with increasing strain amplitude, especially above $5 \times 10^{-6}$ strain in dry rock and is attributed to nonlinear strain in response to linear stress. This threshold shifts to $10^{-7}$ to $10^{-6}$ in saturated granite. The presence of clay may raise these thresholds in saturated sandstone.
REFERENCES


Coyner, K.B., 1984, Effects of stress, pore pressure, and pore fluids on bulk strain, velocity, and permeability in rocks: Ph.D. thesis, Massachusetts Institute of Technology


APPENDIX A

CYCLIC LOADING
CYCLIC LOADING (Hysteresis Loop)

Young's modulus and $Q_E$ were determined from steady-state stress and strain data measured during unconfined cyclic loading tests. A continuous sinusoidal axial load was applied to cylindrical rock samples. The cylindrical sample dimensions were 6 inches in diameter by 12 inches in length. Measurements were performed on room dry and saturated samples. The sample was mounted in a servocontrolled loading frame. Axial stress was imposed by a piston in the loading column and controlled by a Hewlett Packard 203A function generator. Cyclic loads were prescribed from continuous sine functions at single frequencies ranging from 0.01 Hz to 100 Hz. Three axial LVDTs are used to measure strain. The LVDTs are set 120 degrees apart from another and held to the sample with fixed rings. The experimental and sample configurations are shown in Figure 7. The three LVDT strain measurements are collected on separate channels, then averaged. If the strains vary outside 10% from another, the LVDTs are repositioned and the experiment is repeated (assuming the sample is isotropic).

The sample was initially loaded to 10 bars, then cycled approximately 0.5 bar around the 10 bar preload. This load was chosen to minimize mechanical losses across interfaces in the loading column. The experimental procedure is duplicated for an aluminum core with the same dimensions to calibrate the hysteresis of the experimental system.

Data Acquisition and Processing

Data were collected digitally at a sample resolution up to 50 microseconds. In particular, this minimum time step is necessary to resolve small phase shifts on the order of 5 milliradians and to adequately sample the 100 Hz cycles. The voltages from the load cell and three LVDTs are collected via an A to D card set in a PC. The modulus and $Q$ were determined from different techniques depending on the frequency of the sinusoidal load. For frequencies below 5 Hz, the modulus and $Q$ were determined from single loops, with 1400 points of data per cycle. For frequencies at and above 5 Hz, many cycles (up to 200) were collected and analyzed to compute the modulus and $Q$. The number of points per cycle decreased with increasing frequency. The signal noise level is handled by applying a Butterworth filter to the raw data, then compute stress and strain values. The filtered data is represented in Figure 8.

System Calibration

The energy losses of the system are critical in determining the modulus and $Q$ of the sample of interest. The system includes the interfaces in the loading column, the mechanical features and responses in the loading frame and sample column, electronic filters which remove signal noise, and signal amplifiers. The system loss (phase angle) was determined by performing a duplicate experiment at each frequency and preload using a low loss aluminum reference sample. The system phase angle was then subtracted from the net phase angle measured for the sample of interest. The stress-strain hysteresis loop for aluminum measured at 0.5 Hz is shown in Figure 9. The data has
Figure 7. Schematic diagram of the experimental apparatus for low-frequency hysteresis loop modulus and attenuation measurements. The sample is mounted in a servo-controlled loading frame.
Raw Data: 5 Hz Hysteresis Loop: Aluminum/calibration

Filtered Data: 5 Hz Hysteresis Loop: Aluminum/calibration

FIGURE 8
been filtered to remove signal noise. The calibration phase angle is then subtracted from the phase angle of the rock phase angle as depicted in the bottom panel in Figure 10. Once the phase angle is computed, the Q value is determined from the tangent to that corrected angle.

Determining Young's Modulus

Below 5 Hz:

Young's modulus was determined from the slope of the line which bisects the averaged stress-strain loop at a single frequency. The 5 Hz hysteresis loop for the aluminum calibration is shown in Figure 9. The dotted line is a plot of the measured data (including system hysteresis).

At and Above 5 Hz:

Complex Young's modulus was determined by computing the Fast Fourier Transforms of the stress and strain sinusoids. The amplitude spectra for stress and strain are shown in Figure 10. The complex modulus was next computed from the ratio of the peak stress and strain spectra at the input frequency. The modulus amplitude was then calculated from the complex modulus and compared with moduli at lower frequencies. As the frequency was increased above 10 Hz, an apparent increase in Young's modulus was observed. Since the modulus of aluminum is independent of frequency below 100 Hz (Spencer, 1981), this increase was attributed to system losses. It was assumed that the strain amplitude was under estimated above 10 Hz. A correction factor was required to process the data to obtain Young's modulus. To account for the amplitude loss, a scaling factor was computed and applied to the apparent modulus. First, the apparent modulus of aluminum was fitted with a polynomial as a function of frequency from 0.01 to 100 Hz. The modulus fit for aluminum is constant at frequencies below 10 Hz. The scaling factor was then computed from the polynomial by reducing the aluminum signal loss to the constant value below 10 Hz. The scaling factor was then applied to the rock data.

Determining Attenuation

Attenuation was determined from the phase angle measured between the stress and strain sinusoids.

Below 5 Hz

Two sinusoidal signals of the same frequency can be compared to determine the phase shift in time. For example, there are 1400 points per cycle or period, T, in a single loop.
5 Hz Hysteresis Loop: Aluminum/calibration

- 50 bar preload point
- Measured loop with system lag
- Loop - system lag

**Figure 9**
Amplitude Spectra - Sierra White granite - 30 Hz cyclic load

- stress
- strain

FIGURE 9
Figure 10: Berea sandstone (saturated) : Hysteresis Loop

0.1 Hz cyclic load

Axial stress (bars)

Axial strain (microstrain)
The phase shift is then the difference in the number of points, $\Delta n$ for the maximum stress and strain. The phase in radians is:

$$\phi = 2\pi \Delta n / T$$

The phase angle and $Q_E^{-1}$ were also computed from the area of the hysteresis loop for comparison purposes. The error was within 7%.

At or above 5 Hz

$Q_E^{-1}$ was determined directly from the complex Young's modulus by computing the ratio of the imaginary and real modulus components. The system attenuation increased with frequency above 10 Hz. The calibration phase angle for the system was determined at each frequency point and subtracted from the measured sample phase angle (sample + system).
APPENDIX B

RESONANT BAR EXPERIMENTS
RESONANT BAR EXPERIMENTS

Benchtop resonant bar experiments were conducted on cylindrical or square rods of rock specimen. The samples were cored from a 12" cubic block and ground to this shape with the ends flat and parallel to within 0.001 inches. Two identical, 0.25 inch diameter, 1 MHz, compressional, PZT, piezoelectric crystals were epoxied to the sample using Epotek 301 epoxy, with one crystal arbitrarily designated as the excitation source and the second as the sensing receiver. During the epoxy cure schedule a small load was applied to the crystals with a small screw-driven loading frame and a stiff spring. After curing, number 38 gauge wires were tightly coiled and soldered to the piezoelectric crystal chrome-gold electrodes.

Crystals epoxied to the sample ends were used to drive the sample in extensional (also referred to as longitudinal) mode. The crystals epoxied to the sample ends were also used to measure a mixed compressional wave/bar velocity in an ultrasonic pulse experiment.

Resonant bar experiments were conducted in a modified dessicator bell-jar that isolated the sample in a controlled atmosphere. The sample was supported at its midpoint with a thin, cotton string suspended from a fixture. The coiled leads from the piezoelectric crystals were clasped with spring-loaded test hooks. The leads from these test hooks were brought out of the chamber through a sealed fitting and into coaxial connectors. A lock-in amplifier was used to resonate the rod at the fundamental resonances and at higher harmonics, up to the 18th harmonic for the extensional resonance of the dry sample. The corrected fundamental extensional resonance was approximately 6.0 kHz. Both source and sensing piezoelectric crystals were wired to the amplifier, one to the reference sine wave output and the second to the differential amplifier input. The differential amplifier input from the sensing crystal was able to "lock-in" to the phase shifted sine wave frequency being used to resonate the rod. A PC-controlled D/A converter was used to externally control the lock-in amplifier reference sine wave output to sweep through the entire frequency band of interest, and an A/D input was used to measure the amplified signal from the sensing crystal.

Computation of Velocity, Modulus, and Attenuation

Extensional velocity, Young's modulus, and extensional wave attenuation are determined from the half-power width frequency band, where \( F \) is the resonant or center frequency. A correction is applied to the measurement accounting for the effects of the crystal mass and sample jacketing. \( F_{corr} \) is the value of the resonant frequency after it is corrected for the effect of end loading by the piezoelectric crystals. The correction used is from Lucet and others (1991):

43
Berea sandstone: Resonant Bar

Amplitude vs. Frequency (kHz)

Figure 11
where $F_{\text{meas}}$ is the measured frequency and $M_{\text{crystals}}$ is the mass of the piezoelectric crystals and $M_{\text{bar}}$ is the mass of the bar without the piezoelectric crystals.

$Q$ is the quality factor, $\Delta F$ is the frequency half power band width:

$$Q = \frac{F_{\text{corr}}}{\Delta F}$$

$V_e$ is the extensional velocity in m/s.

$$V_e = F_{\text{corr}} \cdot \frac{2}{\text{peak}}$$

where $l$ is the length of the bar in meters and $2/\text{peak}$ corrects for the wavelength of the resonant frequency. For example, the fundamental is one-half wavelength, so the length of the bar needs to be doubled to get the correct velocity.

$E$ is the Young's Modulus in GPa.

$$E = V_e^2 \rho$$

where $\rho$ is the density in kg/m$^3$. For the flexural measurements,

$$V_s = F_{\text{corr}} \cdot \frac{2}{\text{peak}}$$

$$G = V_s^2 \rho$$

For the measurements made on samples wrapped in teflon tape, the data was corrected using the technique of Lucet and others (1991). First, $F_{\text{corr}}$ was calculated and $V_e$ was determined from it. This corrects for the end-loading of the sample. Next the $V_e$ was corrected for the effect of the jacketing using the equation:
\[ V_{e, \text{corr}} = V_e \sqrt{\frac{M_{\text{wrapped sample}}}{M_{\text{unwrapped sample}}}} \]

where \( V_e \) is the calculated \( V_e \) from the \( F_{\text{corr}} \). Then, a new center frequency was calculated using the relation

\[ F_{\text{corr, new}} = \frac{V_{e, \text{corr}}}{1 - \alpha} \]

An error of less than 5% can be associated with the attenuations and moduli from the resonant bar experiment.

Table 1 contains results of extensional resonant bar measurements on the room dry specimen of Berea sandstone sample. The measured center frequencies have been corrected using the procedure of Lucet and others (1991). The measured frequency was corrected by dividing by \( 1 - \alpha \), where \( \alpha \) is the ratio of the mass of the crystals epoxied to the sample to the mass of the sample. For a crystal mass of 1.139 g and a bar mass of 34.045 g, \( \alpha \) is 0.0335. In the table are columns of corrected center resonant frequency, the harmonic of the resonance, the velocity in km/s, Young's modulus in GPa, and extensional quality factor \( Q_e \).
TABLE 1

Attenuation Measurements in Berea sandstone Bar Utilizing the Resonant Bar Technique.

<table>
<thead>
<tr>
<th>CORRECTED FREQUENCY (Hz)</th>
<th>HARMONIC</th>
<th>VELOCITY (km/s)</th>
<th>YOUNG'S MODULUS (GPa)</th>
<th>QUALITY FACTOR (Qe)</th>
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</thead>
<tbody>
<tr>
<td>6064</td>
<td>0</td>
<td>2.176</td>
<td>10.41</td>
<td>160</td>
</tr>
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<td>12220</td>
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<td>10.56</td>
<td>153</td>
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<tr>
<td>17895</td>
<td>2</td>
<td>2.141</td>
<td>10.07</td>
<td>157</td>
</tr>
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<td>23955</td>
<td>3</td>
<td>2.149</td>
<td>10.15</td>
<td>150</td>
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<td>143</td>
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<td>53883</td>
<td>8</td>
<td>2.149</td>
<td>10.15</td>
<td>139</td>
</tr>
<tr>
<td>60287</td>
<td>9</td>
<td>2.164</td>
<td>10.29</td>
<td>143</td>
</tr>
<tr>
<td>65500</td>
<td>10</td>
<td>2.137</td>
<td>10.03</td>
<td>136</td>
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<tr>
<td>72240</td>
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<td>2.161</td>
<td>10.26</td>
<td>136</td>
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<td>12</td>
<td>2.171</td>
<td>10.35</td>
<td>146</td>
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<td>84174</td>
<td>13</td>
<td>2.158</td>
<td>10.23</td>
<td>114</td>
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<td>89637</td>
<td>14</td>
<td>2.145</td>
<td>10.11</td>
<td>102</td>
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<td>95447</td>
<td>15</td>
<td>2.141</td>
<td>10.07</td>
<td>132</td>
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<td>101330</td>
<td>16</td>
<td>2.139</td>
<td>10.05</td>
<td>90</td>
</tr>
<tr>
<td>107545</td>
<td>17</td>
<td>2.144</td>
<td>10.10</td>
<td>117</td>
</tr>
<tr>
<td>111449</td>
<td>18</td>
<td>2.105</td>
<td>9.74</td>
<td>109</td>
</tr>
</tbody>
</table>
Attenuation Measurements without Confining Pressure

The first experiments of Tittmann and others (1972) were performed using an extensional-mode excitation at room-temperature and pressure. The lunar samples were cut into long thin bars. Then a small tab of iron was glued onto each end of the bar. One end was driven electromagnetically and the electromagnetic response of the system was measured at the other end. Tittmann and others (1972) found that a room-dry lunar basalt (Sample 14310) yielded a \( Q_E \) of approximately 90. The same sample, when exposed to hot water vapor for thirty seconds, yielded a \( Q_E \) of approximately 10. Longer exposure to hot water vapor produced a \( Q \) so low that it could not be measured. When a vacuum of \( 5 \times 10^{-8} \) Torr was applied, \( Q_E \) was found to increase to between 130 and 150. A \( Q_E \) of approximately 800 was obtained under simultaneous exposure to a vacuum of \( 5 \times 10^{-8} \) Torr and temperature of \(-180^\circ \)C.

These experiments were followed by further extensional-mode experiments (Tittmann et al, 1974) that documented higher Q values with more extensive outgassing procedures. A lunar basalt (Sample W-8) was outgassed to \( 2 \times 10^{-8} \) Torr at room temperature and a \( Q_E \) of 688 was obtained at a resonant frequency of 15,531 Hz. The sample was then heated to 200°C at \( 10^{-8} \) Torr to further outgas it. It was cooled and then exposed to N2-enriched air for about two hours. This produced a \( Q_E \) of 1478 at a resonant frequency of 15,335 Hz. The sample was again heated to 200°C at \( 10^{-8} \) Torr and allowed to cool under a \( 10^{-8} \) Torr vacuum to room temperature. Under a \( 10^{-8} \) Torr vacuum, a \( Q_E \) of 1890 was measured two hours after the heat treatment; a \( Q_E \) of 2010 was measured five and one-half hours after the heat treatment. Both measurements had a resonant frequency of 15,294 Hz. The changes in \( Q_E \) were found to be reversible: reexposing the sample to laboratory air lowered the \( Q_E \) to approximately 100. The changes in \( Q_E \) were attributed to the presence of H2O in the sample.

Another set of extensional measurements (Tittmann et al, 1975) was carried out in a new vacuum chamber. The new chamber was constructed from stainless steel and was designed to allow a better vacuum to be attained in the system. Experiments on a lunar basalt (Sample 70215.85) in the new chamber yielded the following results. A \( Q_E \) of 60 at a resonant frequency of 19,969 Hz was obtained at laboratory temperature and pressure. The sample was then placed in the vacuum chamber and a pressure of \( 10^{-3} \) Torr was established. A \( Q_E \) of 340 at 20,335 Hz was obtained. The first of three heating cycles to 300°C at \( 10^{-3} \) Torr produced a \( Q_E \) of 400 at 20.353 Hz. The new two cycles produced a \( Q_E \) of 800 at 20.253 Hz. A fourth heating cycle to 200°C was followed by lowering the pressure to \( 10^{-6} \) Torr. This produced a \( Q_E \) of 2420 at 20.138 Hz. Continued pumping, for one week, further reduced the pressure to \( 10^{-7} \) Torr and yielded a \( Q_E \) of 3130. A \( Q_E \) of 142 at 20.154 Hz was obtained after long term reexposure to laboratory air.
Measurements of Attenuation under Confining Pressure

Tittmann and others (1976, 1977) began a program to measure attenuation in outgassed samples as a function of confining pressure. In order to make these measurements, a system utilizing torsional resonance was developed, in order to prevent attenuation due to coupling between the sample and the confining pressure medium. They also developed a system that allowed rock samples to be cased in thin-walled copper tubing under a vacuum of 1x10⁻⁵ Torr. This allowed measurements to be obtained on outgassed samples. These experiments on terrestrial basalts (Tittmann et al., 1977) documented large differences in QT between vacuum-outgassed samples and samples with 0.04 and 2.0% H₂O. As before, the outgassed samples had a Q that was substantially higher than the wet samples. The variation of QT with pressure seems to depend on the porosity of the sample. A terrestrial basalt with 10% porosity had a QT that increased as a function of pressure from 0 to 1 kbar. Two samples of terrestrial basalt with a porosity of 1% had QT's that increased up to about 50-100 bars; then QT remained constant or decreased as the pressure was increased.

Tittmann and others (1978) developed a system for measuring QE under confining pressure. They rounded the end-plugs used to jacket the samples in an effort to produce an aerodynamic shape that would interact as little as possible with the confining pressure fluid. Using this technique, they measured QE under confining pressure for an outgassed terrestrial basalt with 1% porosity. They found that QE behaved similarly to QT for the low porosity basalt (Tittmann et al., 1977). QE was high and decreased slightly as confining pressure was increased.

Tittmann and others (1979) measured QE as a function of temperature and pressure on a terrestrial basalt that had been outgassed and encapsulated at 1x10⁻⁶ Torr. The measurement was designed to simulate conditions along the lunar selenotherm. As the confining pressure was increased from 0.25 kbar to 1.25 kbar and the temperature was simultaneously increased from 20°C to 120 °C, QE was found to increase from approximately 800 to near 1400. They concluded that these values were consistent with the high Q obtained for the lunar crust using seismic observations and that the high lunar Q was due to the extremely dry condition of the lunar crust.

Temperature Dependence of Attenuation

Tittmann and others (1976) conducted a series of experiments on the temperature dependence of the Q of terrestrial basalts to determine if Q was changed to a higher value by welding of the grains at high temperature. A torsional mode experiment was conducted near 55 Hz using temperatures from 25°C to 600°C under a 10⁻⁷ Torr vacuum. The heating/cooling rate was near 60°C/hour. The QT increased from 500 to 920 at room temperature after the heating cycle, but the resonant frequency returned to nearly its original value. The small hysteresis in
Qr was thought to be due to the relatively fast heating/cooling rate. Two extensional mode experiments were conducted under a vacuum at frequencies between 15 kHz and 17 kHz and temperatures between -40°C and 170°C. The heating/cooling rates were 30°C/hour and 1°C/hour. For the 30°C heating/cooling rate, the resonant frequency increased from about 15.4 kHz at 25°C to about 16.3 kHz at 150°C. For the 1°C heating/cooling rate, the resonant frequency increased from about 15.6 kHz at -40°C to about 17.6 kHz at 170°C. In both cases the resonant frequency returned to its original value on cooling. The lack of hysteresis in the resonant frequency for all three experiments indicates that there was no permanent change in the elastic moduli. Therefore, Tittmann and others concluded that no permanent welding of the grains had occurred. They suggested that the increase in the elastic moduli may be due to a thermally-induced partial closing of microcracks in the rock.

Tittmann and others (1978) measured the temperature dependence of Q using the torsional method. Their results are given below. For an outgassed terrestrial basalt, Qr decreased as temperature increased from -100°C to 100°C. Then Qr began to increase from 150°C to 450°C.

<table>
<thead>
<tr>
<th>Temperature</th>
<th>Qr</th>
</tr>
</thead>
<tbody>
<tr>
<td>-100</td>
<td>1670</td>
</tr>
<tr>
<td>-50</td>
<td>1430</td>
</tr>
<tr>
<td>0</td>
<td>1330</td>
</tr>
<tr>
<td>50</td>
<td>800</td>
</tr>
<tr>
<td>100</td>
<td>620</td>
</tr>
<tr>
<td>150</td>
<td>700</td>
</tr>
<tr>
<td>250</td>
<td>900</td>
</tr>
<tr>
<td>350</td>
<td>950</td>
</tr>
<tr>
<td>450</td>
<td>900</td>
</tr>
</tbody>
</table>

Tittmann and others (1979) conducted a series of thermal cracking experiments on Westerly Granite in order to determine the relative importance of thermal cracking versus adsorbed water content. In this set of experiments, the sample was heated to the required temperature at less than 2°C/minute under vacuum. It was then cooled to room temperature, backfilled with dry nitrogen gas and iron driving tabs were attached. The sample was placed into the torsional apparatus and a vacuum was reestablished. The resonant frequency was found to decrease smoothly from about 14 kHz at 0°C to 8 kHz at 480°C; then a large downward shift occurred between 500°C and 600°C and the resonant frequency again decreased smoothly from 4.2 kHz at 600°C to 3.8 kHz at 750°C. Qr was found to constantly increase from about 400 to 1400 over the same range of temperature. Tittmann and others (1979) attributed the drop in resonant frequency between 500°C and 600°C to the α–β phase transition in quartz. They concluded that high degrees of thermal cracking are not inconsistent with high Q when the sample is dry.
Effect of Volatiles at Room Pressure

Tittmann and others (1975) also conducted a detailed set of experiments to determine what effect different volatiles would have on the Q. The experiment consisted of placing a terrestrial basalt sample into the chamber and outgassing it until the Q was approximately 500. Then the sample was exposed to 1 atm of the volatile of interest and the Q was measured. They reported the results of these experiments as percent decrease from the original Q. I have recalculated their figures to a Q value assuming the original outgassed Q is always 500.

<table>
<thead>
<tr>
<th>Gas</th>
<th>Q</th>
</tr>
</thead>
<tbody>
<tr>
<td>He</td>
<td>485</td>
</tr>
<tr>
<td>CO₂</td>
<td>460</td>
</tr>
<tr>
<td>O₂</td>
<td>425</td>
</tr>
<tr>
<td>H₂</td>
<td>375</td>
</tr>
<tr>
<td>H</td>
<td>300</td>
</tr>
<tr>
<td>CO</td>
<td>275</td>
</tr>
</tbody>
</table>

With the exception of helium, all the gases caused a reduction in Q. Tittmann and others (1975) suggest that the adsorption of gas is responsible for the decrease in Q.

The test chamber was attached to a gas-source mass spectrometer to analyze the residual gas that remained after outgassing the sample. The dominant gases that remained in the chamber were H₂O and either CO or N₂. Tittmann and others (1975) suggested that it was primarily the presence of H₂O that prevented the achievement of higher Q values in the outgassed terrestrial basalt.

Tittmann and others (1976) also continued their systematic study of the effect of different volatiles on the Q of a terrestrial basalt. The procedure involved outgassing the sample at 400°C and 10⁻⁸ Torr and then allowing 1 atm of the gas to equilibrate with the sample. The results were again reported as a percent reduction in Q. I have recalculated their figures to a Q value assuming the original outgassed Q is always 735. These values are reported below.

<table>
<thead>
<tr>
<th>Gas</th>
<th>Q</th>
<th>Viscosity</th>
<th>Partial Pressure</th>
<th>Boiling Point</th>
<th>Dipole Moment</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>(cP)</td>
<td>(mm Hg)</td>
<td>(°C)</td>
<td>(Debye)</td>
<td></td>
</tr>
<tr>
<td>N₂</td>
<td>691</td>
<td>760</td>
<td>-198.5</td>
<td>0</td>
<td></td>
</tr>
<tr>
<td>Trichloroethylene</td>
<td>441</td>
<td>68</td>
<td>86.7</td>
<td>0.9</td>
<td></td>
</tr>
<tr>
<td>Benzene</td>
<td>419</td>
<td>90.8</td>
<td>80.1</td>
<td>0</td>
<td></td>
</tr>
<tr>
<td>Dichloromethane</td>
<td>390</td>
<td>382</td>
<td>40.7</td>
<td>1.60</td>
<td></td>
</tr>
<tr>
<td>Freon 14</td>
<td>672</td>
<td>760</td>
<td>-130</td>
<td>0</td>
<td></td>
</tr>
<tr>
<td>H₂O</td>
<td>294</td>
<td>21.1</td>
<td>100.0</td>
<td>1.82</td>
<td></td>
</tr>
</tbody>
</table>
Any reduction of Q due to the viscosity of the condensed liquid is masked by other effects, since liquid trichloroethylene has a viscosity larger than liquid dichloromethane; yet the reduction in Q due to trichloroethylene is less than dichloromethane. Dichloromethane has a vapor pressure that is larger than trichloroethylene. Both dichloromethane and trichloroethylene, like water, have a non-zero dipole moment. The dipole moment of dichloroethylene is larger than trichloroethylene. Tittmann and others (1976) report that the changes in Q are completely reversible; the original high Q can be restored by a bake-out under vacuum. They concluded that the reduction in Q is due to an absorption mechanism.

Tittmann and others (1979) measured the dependence of Q on the partial pressure of water in a terrestrial basalt. They constructed a chamber in which the partial pressure of water could be controlled. The sample was placed in the chamber and Q was measured for a range of partial pressures from 0 to 0.8. The weight of the sample was continually monitored. Even at a partial pressure of 0.72, the sample did not absorb more than 0.2 mg of H_2O per gram of rock. The Q decreased from 1450 to 950 over this range. A similar effect was observed for methanol over a range of partial pressures from 0 to 0.9. The Q did not decrease monotonically over this range. Tittmann and others (1979) noted two regimes in which the behavior of Q was different. They attributed the change in behavior of Q as reflecting a region, at low partial pressures, where water or methanol was adsorbed onto grain surfaces and a region, at high partial pressures, where water or methanol condensed in capillary space between grains.

**Attenuation in crystalline rocks**

Gordon and Davis (1968) measured the attenuation of a variety of crystalline rocks using the resonant bar technique. They used a composite piezoelectric resonator comprised of a rock sample glued to a quartz crystal to measure the extensional attenuation. They measured the attenuation at 90 kHz at a variety of strain amplitudes. Below, I have compiled their data on dry samples at a strain amplitudes between 2.3 x 10^{-7} and 3.0 x 10^{-7}. These samples were cleaned using pentane and alcohol and then placed under vacuum prior to being measured. The relatively high Q_E measured is consistent with Tittmann and others (1972, 1974, 1975, 1976, 1977, 1978, 1979) work on the attenuation of dry crystalline rock samples at similar strain amplitudes (10^{-8} range) (Tittmann, 1977).

<table>
<thead>
<tr>
<th>Sample</th>
<th>Attenuation</th>
<th>Q_E</th>
</tr>
</thead>
<tbody>
<tr>
<td>Single Quartz Crystal</td>
<td>5.37 x 10^{-5}</td>
<td>18622</td>
</tr>
<tr>
<td>Quartzite (15)</td>
<td>1.00 x 10^{-3}</td>
<td>1000</td>
</tr>
<tr>
<td>Dunite (28)</td>
<td>1.45 x 10^{-3}</td>
<td>690</td>
</tr>
<tr>
<td>Basalt (29)</td>
<td>3.31 x 10^{-3}</td>
<td>302</td>
</tr>
<tr>
<td>Pyroxenite (8)</td>
<td>3.80 x 10^{-3}</td>
<td>263</td>
</tr>
<tr>
<td>Rock Type</td>
<td>Q \times 10^{-3}</td>
<td>\text{QE}</td>
</tr>
<tr>
<td>-------------</td>
<td>------------------</td>
<td>-----------</td>
</tr>
<tr>
<td>Basalt (25)</td>
<td>4.57</td>
<td>219</td>
</tr>
<tr>
<td>Anorthosite (11)</td>
<td>5.89</td>
<td>170</td>
</tr>
<tr>
<td>Granite (3)</td>
<td>1.38 x 10^{-3}</td>
<td>72</td>
</tr>
<tr>
<td>Diopsidite (33)</td>
<td>1.74 x 10^{-2}</td>
<td>57</td>
</tr>
<tr>
<td>Amphibolite (5)</td>
<td>2.04 x 10^{-2}</td>
<td>49</td>
</tr>
<tr>
<td>Quartzite (Q-2)</td>
<td>2.95 x 10^{-2}</td>
<td>34</td>
</tr>
</tbody>
</table>

It is not clear whether the large range in Q is due to intrinsic differences between the samples or it is the result of different, uncontrolled moisture content. The fact that samples of quartzite had both the highest and lowest Q's further complicates the analysis.

Gordon and Davis (1968) also measured QE in quartzite (sample 15) as a function of temperature.

<table>
<thead>
<tr>
<th>Temperature (°C)</th>
<th>Attenuation</th>
<th>QE</th>
</tr>
</thead>
<tbody>
<tr>
<td>30°</td>
<td>1.03 x 10^{-3}</td>
<td>970</td>
</tr>
<tr>
<td>41°</td>
<td>1.07 x 10^{-3}</td>
<td>935</td>
</tr>
<tr>
<td>60°</td>
<td>1.15 x 10^{-3}</td>
<td>870</td>
</tr>
<tr>
<td>80°</td>
<td>1.10 x 10^{-3}</td>
<td>910</td>
</tr>
<tr>
<td>100°</td>
<td>1.37 x 10^{-3}</td>
<td>730</td>
</tr>
<tr>
<td>151°</td>
<td>1.57 x 10^{-3}</td>
<td>637</td>
</tr>
<tr>
<td>199°</td>
<td>1.90 x 10^{-3}</td>
<td>526</td>
</tr>
</tbody>
</table>

These results are similar to those found by Tittmann and others (1978) using torsional resonance on an outgassed terrestrial basalt. In both experiments, Q decreased as a function of temperature up to 200°C. However, Gordon and Davis did not perform their experiment under vacuum like Tittmann and others. Note that some researchers (Toksoz and Johnston Q-book, Johnston and others, 1979) cite this experiment as demonstrating that Q is relatively constant as a function of temperature up to 150°C. It is important to note the relatively high values of Q found by Gordon and Davis (1968) at elevated temperatures, especially in light of the frequently held assertion that this experiment shows a decrease in Q due to thermal cracking of the sample at temperatures above 150°C. Any decrease in Q due to thermal cracking is evidently a small effect compared to the change to extremely low Q due to water-saturation.

Gordon and Davis (1968) also conducted an experiment comparing the effect of water-saturation on granite (sample 3). They measured Q before and after repeated cycles to high strain amplitude on both wet and dry samples.

53
<table>
<thead>
<tr>
<th>Sample</th>
<th>Attenuation</th>
<th>$Q_E$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Dry Before</td>
<td>$6.86 \times 10^3$</td>
<td>146</td>
</tr>
<tr>
<td>Dry After</td>
<td>$8.11 \times 10^3$</td>
<td>123</td>
</tr>
<tr>
<td>Wet Before</td>
<td>$5.87 \times 10^2$</td>
<td>17</td>
</tr>
<tr>
<td>Wet After</td>
<td>$7.11 \times 10^2$</td>
<td>14</td>
</tr>
</tbody>
</table>

Gordon and Davis (1968) attributed the difference in the $Q$ before and after the repeated cycles to high strain amplitude to sample fatigue. The large differences in $Q$ between the wet and dry samples are again consistent with the work of Tittmann and others (1972, 1974, 1975, 1976, 1977, 1978, 1979).

**Summary**

These experiments on basalt and other crystalline rock indicate that the $Q$ of a dry sample is highly dependent on the nature and amount of volatiles contained in the sample. The presence of water, even in minute amounts, is especially effective at reducing the $Q$. $Q$ increases dramatically, in both lunar and terrestrial rocks, as the sample is outgassed under vacuum. Heating the sample under vacuum increases the amount of outgassing and causes further increases in $Q$. Backfilling the outgassed sample with anhydrous gases at atmospheric pressure can also reduce the $Q$, although not all are as effective as water vapor. The reduction in $Q$ appears to be due to the adsorption of vapor on the grains within the sample. Even at high partial pressures of $H_2O$, the sample does not gain enough weight to be more than a few percent saturated. However, the identification of two regions of behavior of $Q$ as a function of the partial pressure of $H_2O$ suggests that the two attenuation mechanisms may exist.

The increase in $Q$ of dry samples as temperature increases seems primarily to reflect the effect of outgassing the sample. Therefore, high degrees of thermal cracking are not inconsistent with high $Q$ when the sample is dry. The effect of temperature on a fully-saturated sample have not been reported.

The effect of confining pressure on a dry sample varies, apparently depending on the porosity. However, the $Q$ is very high for all reported measurements on dry samples under confining pressure. Here again, the $Q$ appears to depend more strongly on the amount and type of vapor present than on the applied confining pressure.
APPENDIX C

WAVEFORM INVERSION TECHNIQUE
WAVEFORM INVERSION TECHNIQUE

The main purpose for measuring seismic wave attenuation in the laboratory is to infer the rock properties at in-situ scales and seismic frequencies. The pulse transmission technique (Toksoz et al., 1979) operates in the megahertz frequency range, while the resonant bar technique operates at sonic frequencies of a few kilohertz. It is desirable to measure attenuation in a range between sonic and ultrasonic frequencies. For this purpose, New England Research has designed a new technique that operates between 10 kHz and 150 kHz, depending on the sample dimensions.

The Technique:
The technique consists of measuring low-frequency waveforms using cylindrical bars of the same material but of different lengths. The bar is essentially a waveguide which allows the low-frequency fundamental mode to be propagated without having to use rock blocks of large sizes. The procedure of measuring attenuation consists of two major steps. In the first, the waveform received with the short bar is theoretically propagated to the distance of the long bar, in which the dispersion effect on the waveform due to the waveguide is removed. The second step is the inversion of attenuation \((1/Q_E)\) or \(Q_E\) of the bar material through minimizing the difference between the propagated waveform and the actually measured waveform with the long bar. Since the waveform inversion is performed in the time domain, the waveforms can be appropriately truncated to avoid multiple reflections due to the finite size of the (short) sample, allowing attenuation to be measured at long wavelengths or low frequencies. For a detailed description of the technique and inversion formulation the reader is referred to the paper by Tang (1991).

Experimental Procedure

We used the waveform inversion technique to measure extensional \(Q_E\) for Berea Sandstone samples of different dimensions. The first set B-90-C, had diameters of 0.524 inches and lengths of 1.5 inches and 6.83 inches. The second group of samples are labeled as 7 inch and 16 inch; the diameters are 1.25 inches and the lengths are 7.177 inches and 15.70 inches. All the samples are room dry. The relative humidity in lab was ~ 50%. During the measurement, the source and receiver transducers are mounted at the two ends of the bar sample and a signal generator is applied to drive the transducer at various frequencies. The waveforms are recorded with a digital oscilloscope.
Results

To illustrate the application of the technique, we show the comparison of inverted waveform from the short sample with the waveform measured from the large sample. Figure 1 shows the result at 10 kHz for the 1.25 inch diameter sample and Figure 2 shows the result at 80 kHz for the 0.524 inch diameter sample. The matches are very good. Furthermore, the $Q_E$ values obtained from the two measurements of very different frequencies and sample sizes are fairly consistent (38.9 vs. 36.1), showing the constant $Q_E$ behavior of the dry sample. Table 1 shows the inverted $Q_E$ values for the two pairs of samples for the measured frequencies.

Table 1. $Q_E$ Values From Waveform Inversion

<table>
<thead>
<tr>
<th>Sample Set</th>
<th>Frequency (kHz)</th>
<th>Extensional $Q_E$</th>
</tr>
</thead>
<tbody>
<tr>
<td>7&quot; &amp; 16&quot; (1.25&quot;)</td>
<td>10</td>
<td>38.9</td>
</tr>
<tr>
<td>7&quot; &amp; 16&quot; (1.25&quot;)</td>
<td>20</td>
<td>33.8</td>
</tr>
<tr>
<td>7&quot; &amp; 16&quot; (1.25&quot;)</td>
<td>30</td>
<td>33.4</td>
</tr>
<tr>
<td>B-90-C (0.524&quot;)</td>
<td>60</td>
<td>34.6</td>
</tr>
<tr>
<td>B-90-C (0.524&quot;)</td>
<td>80</td>
<td>36.1</td>
</tr>
</tbody>
</table>

Future Directions

The waveform inversion technique can be easily adapted to measure attenuation under confining pressure appropriate to the in-situ condition, if the length of the pressure vessel is modified to accommodate the bar geometry and the low-frequency transducers are designed to operate under pressure. Temperature apparatus may also be added as necessary. Since the frequency of the waveforms is adjustable within a certain range, this technique can be used to measure $Q_E$ as a function of frequency within this range. Furthermore, the waveforms of flexural and tortional modes in the bar can also be inverted to find the shear attenuation of the bar. This would be the direction of future research.
a. Waveforms Measured Within Berea Sandstone Bars

b. Inverted $Q_e = 32$

FIGURE 12
a. 

![Graph showing phase velocity vs. frequency with initial model, measured data, and best fit to data. Bar velocity is 3536 m/s and Poisson's ratio is 0.09.]

b. 

![Graph showing c(w) vs. [w/2c(w)]^2.]

FIGURE 13
Waveform Inversion Method for Measuring Attenuation Using Cylindrical Bars

Sample: Berea Sandstone
Diameter: 1.25 inches
Length: 7.177", 15.700"
Frequency: 10 KHz.
Inverted Q: 38.9

Inverted Waveform of 7.177" matched with 15.7"

Waveform Measured at 15.7 inches

Truncated

FIGURE 14
Waveform Inversion Method for Measuring Attenuation Using Cylindrical Bars

Sample: Berea Sandstone
Diameter: 0.524 inches
Length: 1.5", 6.83"
Frequency: 80 KHz.
Inverted Q: 36.1

Inverted Waveform of 1.5" matched with 6.83"

Waveform Measured at 6.83 inches
Truncated

FIGURE 15
APPENDIX D

ULTRASONIC VELOCITY MEASUREMENT
ULTRASONIC VELOCITY MEASUREMENT

The compressional (P) and shear (S) wave velocities for the rock, at any stress condition, were obtained by measuring the one way travel time of a P or S wave through the core and dividing by the sample length. The wave propagation direction was parallel to the core axis. The ultrasonic transducer consisted of a piezoelectric source and receiver pair of like crystals (P or S) positioned at opposing ends of the cylindrical core. Only one crystal pair could be activated at a time. The source crystal was driven with a short electrical pulse (center frequency of 1 MHz) generated with a Panametrics 5055PR pulser-receiver.

The signal from the receiver crystal was amplified, high-pass filtered at 0.3 MHz, and displayed on a LeCroy 9400 digital oscilloscope. The travel time was measured on the oscilloscope screen with a cursor control marking the first break of the signal. The break was defined by a threshold voltage that was 1.25% of the peak-to-peak amplitude of the first three half-cycles of the signal. The arrival time was picked at the point when the threshold voltage was exceeded relative to the baseline voltage and had a resolution of ± 0.01 microsecond. The travel time of the wave through the end pieces is subtracted from the total time measured on the oscilloscope. The travel times of P and S wave propagation through the titanium end pieces or zero times were measured as a function of confining pressure by placing the end pieces head to head.
Well: BENCH TOP
Sample: CALIBRATION  Length: 0.000 inch  Fluid: DRY
Conf Pr: 79.00 psi  Pore Pr: 0.00 psi  Temp: 21.00 C
File: BTHTH.1  Pick: Manual  12-4-91 14:24:50

**P Wave**
- Velocity(ft/s): 0.00
- ArrTm(microSec): 9.76
- Ampltd (mVolt): 19.51
- PkPk (mVolt): 1569.03
- HtoH (microSec): 0.00

**S1 Wave**
- Velocity(ft/s): 0.00
- ArrTm(microSec): 17.27
- Ampltd (mVolt): 20.66
- PkPk (mVolt): 1582.42
- HtoH (microSec): 0.00

**S2 Wave**
- Velocity(ft/s): 0.00
- ArrTm(microSec): 17.27
- Ampltd (mVolt): -2.44
- PkPk (mVolt): 162.21
- HtoH (microSec): 0.00

**FIGURE 16**
Well:
Sample: ALUMINUM  Length:1.019 inch  Fluid: DRY
Conf Pr:79.00 psi  Pore Pr :0.00 psi  Temp :21.00 C
File: ALUMINUM1.1  Pick: Manual  12-4-91  15:27:26
OPERATOR= TIMOTHY HILL

FIGURE 17
Well: SW-15A-90
Sample: GRANITE
Length: 1.754 inch
Fluid: DRY
Conf Pr: 80.00 psi
Pore Pr: 0.00 psi
Temp: 21.00°C
File: GRANITE.1D
Pick: Manual
12-16-91 12:18:16

**FIGURE 18**

P Wave
- Velocity (ft/s): 13546.49
- ArrTm (microSec): 20.55
- Amplt (mVolt): 5.72
- PkPk (mVolt): 299.47
- HtoH (microSec): 9.76

S1 Wave
- Velocity (ft/s): 8633.59
- ArrTm (microSec): 34.20
- Amplt (mVolt): 3.63
- PkPk (mVolt): 293.02
- HtoH (microSec): 17.27

S2 Wave
- Velocity (ft/s): 8747.26
- ArrTm (microSec): 33.98
- Amplt (mVolt): -13.89
- PkPk (mVolt): 944.20
- HtoH (microSec): 17.27
Well:
Sample: SIERRA WHITE Length: 1.013 inch Fluid: DISTILLED WATER
Conf Pr: 79.00 psi Pore Pr: 0.00 psi Temp: 21.00°C
File: SIERRA.1 Pick: Manual 12-6-91 17:15:12
SATURATED IN DISTILLED WATER UNDER A VACCUM

FIGURE 19
Well: BEREA SANDSTONE
Sample: BS/91/1-6/1 HP Length: 4.379 cm Fluid: DRY
Conf Pr: 20.00 Pore Pr: 0.00 Temp: 22.00 C
File: BS161HPD.4 Pick: Manual 7-31-91 7:49:00
ISR/ASTM rocks at Qh pressure with small diameter subsamples; Dry.
Orientation plane is parallel to polarization plane of S2.

7-31-91 7:49:00

FIGURE 20

68
Well: BEREA SANDSTONE
Sample: BS/91/1-6/1 HP Length: 4.379 cm Fluid: WATER
Conf Pr: 1.00 Pore Pr: 0.00 Temp: 22.00 C
File: BS161HPS.1 Pick: Manual 8-1-91 9:30:16
ISR/ASTM rocks at high pressure w/ small diam. subsamples; Saturated.
Orientation plane is parallel to the polarization plane of S2.
8-1-91 9:30:16

P Wave
Velocity(m/s): 3009.62
ArrTm(microSec): 27.27
Ampltd (mVolt): 0.00
PkPk (mVolt): 1.55
HtoH (microSec): 12.72

S1 Wave
Velocity(m/s): 1785.16
ArrTm(microSec): 47.51
Ampltd (mVolt): 0.19
PkPk (mVolt): 5.86
HtoH (microSec): 22.98

S2 Wave
Velocity(m/s): 1788.08
ArrTm(microSec): 47.44
Ampltd (mVolt): -0.26
PkPk (mVolt): 4.69
HtoH (microSec): 22.95

FIGURE 21

69
ULTRASONIC Q AND PROCEDURE

Values of ultrasonic $Q_F$ were determined for Sierra White granite and Berea sandstone in dry and saturated conditions during hydrostatic loading tests. Compressional and shear wave time series are collected for specific pressure increments for the rock and duplicated for a reference sample with a very high $Q$, such as aluminum. The $Q$ value for aluminum is approximately 150,000 (Toksoz et al., 1979). The value of $Q$ at a given pressure can then be computed from the spectral ratio of the rock and aluminum. Moreover, in this technique, the extrinsic loss mechanisms of the transducer apparatus can be isolated and accounted for.

The procedure is as follows:

1) Time Series
The received time series through the rock and aluminum are edited for Fourier analysis in the following order:
   1. DC value is first removed
   2. Taper the signal end with a $\cos^2$ function to zero
   3. Pad signal on front and end with zeroes (4096 total pts) for FFT

Unedited time series for Berea (dry) and aluminum are shown in Figure 1.

2) FFT
The Cooley-Tukey FFT is used to compute the spectrum of each time series and is the IEEE approved algorithm (Brenner, 1967).

3) Derivation and computation of $Q$
$Q$ is derived from the spectra of the rock and aluminum signals in the following manner. The amplitude spectrum of the rock, $R(f)$ and aluminum, $A(f)$ are shown below:

$$A(f) = G_A(x)e^{-\alpha_A(f)x_k}e^{i(2\pi ft - k x)}$$

$$R(f) = G_R(x)e^{-\alpha_R(f)x_k}e^{i(2\pi ft - k x)}$$

where $f$ is frequency, $x$ is the sample length, $k$ is the wave number, $v$ is velocity, and $\alpha$ is the attenuation coefficient. $G(x)$ is the geometrical factor which depicts the extrinsic loss mechanisms of the transducer and wave spreading. The subscript $A$ and $R$ refer to aluminum and rock, respectively. The wave number can be described in terms of frequency and velocity as:
Ultrasonic P-Wave time series for Berea (dry) & Alum: CP = 1088 psi

Ultrasonic S-Wave time series for Berea (dry) & Alum: CP = 1088 psi

Figure 1
\[ k = \frac{2 \pi f}{v} \]

In the frequency range from 0.1 to 1.0 MHz, \( \alpha \) is assumed to be linear (Toksoz et al., 1979) thus,

\[ \alpha(f) = \gamma f \]

where \( \gamma \) is a constant related to the quality factor of attenuation.

Taking the spectral ratio reduces to

\[ \frac{A(f)}{R(f)} = \frac{G_A}{G_R} e^{-(\gamma_A x_A \cdot \gamma_R x_R) f} \]  

(3)

Taking the natural log (3) becomes

\[ \ln \left( \frac{A(f)}{R(f)} \right) = (\gamma_R x_R - \gamma_A x_A) f \ln \left( \frac{G_A}{G_R} \right) \]  

(4)

and

\[ \gamma = \frac{\pi}{Q v} \]

where \( Q \) is the quality factor. Substitute \( Q \) into equation (4) yields

\[ \ln \left( \frac{A(f)}{R(f)} \right) = \left( \frac{x_R}{Q_R v_R} - \frac{x_A}{Q_A v_A} \right) \pi f + \ln \left( \frac{G_A}{G_R} \right) \]  

(5)

For the aluminum reference, \( Q_A v_A \) is very large (\( Q_A = 150,000 \)), thus \( \gamma_A \) is approximately zero. Equation (5) further reduces to

\[ \ln \left( \frac{A(f)}{R(f)} \right) = \frac{x_R \pi f}{Q_R v_R} + \ln \left( \frac{G_A}{G_R} \right) \]  

(6)

The ratio of \( G_A \) to \( G_R \) is approximately constant over frequency, thus taking the derivative with respect to frequency will drop that term, accounting for the extrinsic losses. The value
of $Q$ can then be determined from equation (6) where

$$\frac{d}{df} \left[ \ln \left( \frac{A(f)}{R(f)} \right) \right] = \frac{x_R \pi}{Q_R \nu_R} \tag{7}$$

$Q_R$ is best approximated by fitting a straight line to the left hand side of (7) defining the slope of the ratio with respect to frequency. Thus

$$Q_R = \frac{\pi x_R}{(\nu_R) \text{ slope}} \tag{8}$$

This method is further exemplified in Figure 2. The spectra of recorded time series for dry Berea and aluminum are displayed in the top panel of Figure 2. The data were acquired during hydrostatic compression tests. The natural log of the spectral ratio as depicted in equation (6) is represented by the star points in the bottom panel of Figure 2. A least squares fit was applied to the ratio data and is represented by the solid line.

Only a portion of the spectral ratio is fit. The spectral window length is determined by the following criteria:

1. The front end of the window is determined by the aluminum spectrum. This point is chosen 3, 10, or 20 db down from the maximum aluminum amplitude.
2. The back end of the window is determined by the rock spectrum. This point is chosen 3, 10, or 20 db down from the maximum rock amplitude.

In the example in Figure 2, the end points are determined 3 db from the maximums. To note, 3 db down is the half power point. The ultrasonic pulse and spectral ratio technique to determine $Q$ is further explained in detail by Toksoz et al. 1979 and Spencer, 1979.

ULTRASONIC $Q$ RESULTS

Ultrasonic velocities and time series were measured during hydrostatic compression tests up to 13,750 psi. Results are shown for dry and saturated Berea sandstone samples. Dry compressional and shear wave velocities, $V_p$ and $V_s$ are shown in Figure 3. The signal quality was good to excellent and there is a high degree of confidence in the first arrival picks. The signal times series are also of good quality allowing for reliable spectral data. $Q_p$ and $Q_s$ are plotted in Figure 4 as a function of pressure. The actual calculations of $Q$ from the measured data are represented by the point markers. A polynomial curve fit is applied to the data and also shown. $V_p$ and $V_s$ are shown in the top panel of Figure 5 for saturated Berea. The pore pressure was held constant at 145 psi. $Q_p$ and $Q_s$ are shown
Spectra of Berea Sandstone (dry) and Aluminium: CP = 1088 psi

Figure 2
Ultrasonic P and S Wave Velocities: Berea Sandstone (dry)

Figure 3
Ultrasonic P-Wave Attenuation: Berea sandstone (dry)

Ultrasonic S-Wave Attenuation: Berea sandstone (dry)

Figure 4
Ultrasonic P and S-Wave Velocities: Berea (saturated)

Attenuation of Ultrasonic P and S-Waves: Berea (saturated)

Figure 5
in the lower panel.

$Q_E$, (from anelastic Young's Modulus) is calculated from $Q_s$ and $Q_p$ in order to compare to the Hysteresis Loop and Resonant Bar Experimental values of $Q_E$. This ultrasonic parameter is as follows:

\[
\frac{1 + v}{Q_E} = \frac{(1 - v)(1 - 2v)}{Q_p} + \frac{2v(2 - v)}{Q_S}
\]

(9)

where $v$ is Poisson's Ratio and is determined from the ultrasonic velocities, $V_p$ and $V_s$ below:

\[
v = \frac{V_p^2 - 2V_s^2}{2(V_p^2 - V_s^2)}
\]

Poisson's Ratio is shown for the dry and saturated cases in the top panel of Figure 6. $Q_E$ is shown for the dry and saturated cases in the bottom panel.

The ultrasonic Q is strongly dependent upon pressure. In the dry condition, $Q_p$, $Q_s$, and $v$ are very comparable i.e. value at each pressure increment. The rock strain is on the order of millistrain to 10's of millistrain due to the hydrostatic load (Martin et al., 1990, Haupt and Martin, 1991). As pressure increases, the porosity decreases with pore collapse and crack closure. Note that the rock strain due to the ultrasonic pulse is less than 1 microstrain. Thus, the rock is changing considerably with pressure and in turn strongly affects the attenuation and velocity. For the saturated condition the presence of water and pressure, both greatly affect $Q_E$. The effect of saturation suppresses Q with pressure as compared to the dry condition, especially for $Q_s$ and in turn $Q_E$. 

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Poisson's Ratio - from Ultrasonic Velocity: Berea Sandstone

- saturated
- dry

Ultrasonic QE: Berea Sandstone

- dry
- saturated

Figure 6
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