Variations in the Seismic Quality Factor Q for the Mississippi Embayment Sediments Utilizing Spectral Analysis of Reflection Data

Michael Wilson Towle

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To the University Council:

The Thesis Committee for Michael Wilson Towle certifies that this is the final approved version of the following electronic thesis: “Variations in the Seismic Quality Factor Q for the Mississippi Embayment Sediments Utilizing Spectral Analysis of Reflection Data.”

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VARIATIONS IN THE SEISMIC QUALITY FACTOR Q FOR THE MISSISSIPPI EMBAYMENT SEDIMENTS UTILIZING SPECTRAL ANALYSIS OF REFLECTION DATA

by

Michael Wilson Towle

A Thesis
Submitted in Partial Fulfillment of the Requirements for the Degree of Master of Science

Major: Earth Sciences

The University of Memphis

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Endless love and appreciation go to my family who has always supported my pursuits. I would also like to thank Jennifer for her love, patience, intelligence, and companionship.
ABSTRACT


A natural log spectral ratio technique is used to calculate Qp for the post-Paleozoic, unconsolidated sediments of the northern Mississippi Embayment. The method is applied to high-resolution marine seismic reflection data acquired along the Mississippi River. The method assumes a frequency independent Q over a certain frequency band. The frequency bands inspected range from 10-150 Hz and 10-200 Hz based on observed linear trends in the spectral ratio data. The large volume of data available allows for increased accuracy in Q estimates. Values of Qs are then approximated using an empirical Qp/Qs relationship for the embayment. Results show that average Qp=100-160 and average Qs=50-80. These values are nearly twice those calculated in previous studies, therefore suggesting that attenuation within the unconsolidated sediment column may not cancel out site amplification effects. These effects have important implications for seismic hazards and ground displacement predictions in the Central U.S. region.
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CHAPTER 1
INTRODUCTION

The seismic quality factor $Q$ is a physical property that represents energy lost per unit cycle for a wave propagating through a medium. It is inversely related to attenuation; therefore, if the attenuation of a medium is small, the $Q$ factor will be large. Knowing the seismic quality factor is essential for the study of large ground motions and earthquake hazards modeling because it provides an indication of how well or how poorly an area will transmit earthquake energy. Distance attenuation relationships and site amplification effects both consider the seismic quality factor as a parameter in their calculations. This has meaningful implications for damage predictions, engineering concerns, and ultimately reducing loss due to large magnitude earthquakes.

The purpose of this study is to calculate the lateral variation of the seismic quality factor $Q$ for $P$ waves in the unconsolidated sediments of the Mississippi Embayment, in the Central U.S. This is done by utilizing the spectral ratio technique (Chapman, 2008) on a large dataset of high-quality, seismic marine reflection data. $Q_p$ is then converted to $Q_s$ by means of an empirical relationship determined from the Embayment Seismic Excitation Experiment (ESEE) (Langston et al., 2005). It is the hypothesis of this thesis that $Q$ values for the Mississippi Embayment unconsolidated sediments are larger than previous studies suggest (Liu et al., 1994; Wang et al., 1994; Chen et al., 1994; Pujol et al., 2002). This has important ramifications for hazards modeling.

Many studies have measured attenuation, and therefore $Q$, for the Mississippi Embayment. Al-Shukri and Mitchell (1990) used a spectral decay technique $t^*$ on high-
frequency earthquake P-wave data and calculated that \( 63 < Q_p < 250 \) for the upper 5 km of crust beneath the New Madrid Seismic Zone (NMSZ). Liu et al. (1994) found that \( Q_p \sim 59 \) and \( Q_s \sim 36 \) by using a spectral ratio technique on converted Sp/S phases of earthquake body waves. In another P and S-wave study, Kang and McMechan (1994) separated intrinsic \( Q \) from scattering \( Q \) using wide-aperture reflection data and obtained values of \( Q_p=220 \) and \( Q_s=68 \). However, the ray paths in their study sampled not only the unconsolidated sediments, but also the basement rocks in the embayment. Wang et al. (1994) incorporated a pulse width technique on S wave refraction data and discovered that \( Q_s=10-60 \). Chen et al. (1994) used a converted Sp/S spectral ratio technique on microearthquake data and found \( Q_p=25-60 \) and \( Q_s=25-30 \). Pujol et al. (2002) utilized spectral analysis of vertical seismic profiling data in order to find \( Q_s=22-34 \) in the near surface sediments less than 60 m in depth. Langston et al. (2005) obtained \( Q_p=200 \) and \( Q_s=100 \) from using the spectral distance decay from explosion source P and Rayleigh wave data.

It is apparent from the previous studies that there is ambiguity in determining attenuation for the Mississippi Embayment unconsolidated sediments. In an Sp/S body wave attenuation study, Langston (2003b) describes potential biases in earthquake body wave spectral analyses. He finds that near-surface resonance phenomena due to shallow low-velocity zones and wave interference effects due to free surface reflections create spectral peaks and troughs in the spectral ratio that mimic attenuation. This is evidenced in his synthetic data analysis, in which \( Q_p \) and \( Q_s \) are infinite, and the near-surface resonance effects create a slope in the spectral ratio that corresponds to \( Q_s=42 \) (assuming infinite \( Q_p \)). Langston (2003b) also claims that there is no evidence that the
unconsolidated sediments are highly attenuating. Furthermore he states that no previous study has calculated a Q value for the entire sedimentary column that is independent of these earthquake wave propagation biasing effects. If this is the case, then previous studies have overestimated anelastic attenuation, and therefore Q values should actually be higher for the unconsolidated sediments.

Through the use of layer-over-half-space modeling, Langston (2003b) also shows that scattering effects due to heterogeneous velocity structure can cause spectral falloff which mimics attenuation; however, despite the spectral falloff, the scattering effects can still cause large site amplification. He claims that given the scenario of high attenuation, as previous studies claim, the near surface resonance effects would possibly overcome any attenuation that may occur, resulting in large site amplifications. In an explosion-source seismic study of the distance attenuation of surface waves and trapped sediment P waves in the Mississippi embayment, Langston et al. (2005) claims that the low Qs values of prior studies do not allow for surface waves to travel the distance that they are actually detected. This would suggest a higher value for Q in the embayment. These studies highlight the necessity of finding more methods to constrain Q in the Mississippi embayment.

While many of the previous studies use stationary receivers to interpret Q over different ray paths, this study is unique in the sense that sources and receiver locations were distributed over and sampled a geographically broad area. The large area covered and the data redundancy allow us to estimate lateral variation of Q within the embayment sediments. The spectral analysis technique used in this study follows a method that Chapman (2008) used on seismic reflection data in the Virginia Coastal Plain.
Additionally, this technique addresses many of the biasing concerns addressed by Langston (2003b). This study finds an average $Q_p=100-160$ and an estimated average $Q_s=50-80$ for the 800 m thick column of unconsolidated sediments in the Northern Mississippi Embayment.
CHAPTER 2
GEOLOGY

A prominent regional feature of the Central United States is the Mississippi Embayment (Fig. 1). The embayment is a regional, synclinal trough that is geographically bounded by the Ozark uplift to the west, the Nashville dome to the east, and tapers up to the north in southern Illinois. The northern basin has a basement of Paleozoic sedimentary rocks that is overlain by a thick layer of Cretaceous and Tertiary unconsolidated sediments. The sediments increase in thickness to the southwest, reaching a depth of about 1 km near Memphis, Tennessee (Stearns, 1957; Dart, 1992).

Geophysical studies show that the embayment is home to multiple gravity and magnetic potential anomalies (Hildenbrand et al., 1977; Kane et al., 1981; Hildenbrand, 1985). The Reelfoot Rift is one of the most prominent features that appear in the geopotential data and is characterized as having regional magnetic lows (Hildenbrand et al., 1977; Kane et al., 1981; Hildenbrand, 1985). The rift trends northeast-southwest (Figs. 1 and 2) and is approximately 300 km long and 70 km wide (Kane et al., 1981). This feature is interpreted as a failed Paleozoic rift, or aulacogen, that formed during the opening of the Iapetus Ocean (Ervin and McGinnis, 1975). Many positive gravity and magnetic anomalies flank the edges of the Reelfoot Rift (Hildenbrand et al., 1977). These anomalies are interpreted as Cretaceous mafic intrusions associated with the reactivation of the rift and stretching during the breakup of Pangaea. Wide angle refraction studies (Ginzburg et al., 1983; Mooney et al., 1983) show that the average crustal thickness under the embayment is 41 km. These studies, along with gravity potential studies, also
find a large (~13 km thick), high-density (3.0 g/cm³), high-velocity (Vp=7.3 km/s) layer about 28 km deep, located directly underneath the New Madrid Seismic Zone (NMSZ). This layer is interpreted by Ginzburg et al. (1983) to be lower-crustal material which was altered by an intrusion of material from the upper mantle. This upper mantle material was most likely emplaced during an initial period of Late-Precambrian rifting, thus providing further evidence for the rifting that occurred in the region. This thick layer of altered lower-crustal material is therefore also known as the rift pillow.
FIG. 2. Topography map of the Northern Mississippi Embayment in the Central US showing the 300 km long marine seismic reflection survey acquired in 2008 along the Mississippi River (thick black and red lines). Red line highlights section of seismic marine reflection survey analyzed in this study. New Madrid Seismic Zone (NMSZ) seismicity from 2000-2008 with M>1.5 is shown as white circles. Yellow lines represent main suspected and known faults. ERRM: East Reelfoot Rift Margin; WRRM: West Reelfoot Rift Margin; FZ: Fault Zone (Magnani, 2008). Red stars indicate cities.
The Mississippi Embayment hosts one of the most active seismic zones in the Central and Eastern United States (Johnston, 1996a), the New Madrid Seismic Zone. The NMSZ is situated in the northern embayment and is comprised of multiple arms of seismicity (Figs. 1 and 2). A series of three large magnitude earthquakes occurred in the NMSZ during the winter of 1811-1812 (Nuttli, 1973; Johnston and Schweig, 1996). The magnitudes of these earthquakes have been debated but are generally accepted to range from M7.0-M8.1 (Hough et al., 2000; Hough and Martin, 2002; Bakun and Hopper, 2004; Johnston, 1996b). Paleoseismic evidence shows that these large magnitude events have a recurrence interval of approximately 500 years (Tuttle et al., 2002). The large historic events, along with recent seismic activity, are the main reasons why the NMSZ and the surrounding region have the highest hazard in the Central United States.
CHAPTER 3

DATA

In the summer of 2008, a joint experiment was conducted in collaboration between the Center for Earthquake Research and Information, the United States Army Corps of Engineers, the University of Texas Institute for Geophysics, and the Ground Water Institute at the University of Memphis. The experiment, entitled the Mississippi River Project, was a seismic marine reflection study conducted along nearly 300 km (200 miles) of the Mississippi River from Caruthersville, Missouri to Helena, Arkansas (Fig. 2). The scope of the experiment was to image the subsurface features of the Mississippi Embayment and identify Quaternary deformation and faulting (Fig. 11).

The survey was conducted by piloting a barge downstream with a seismic energy source and a linear array of hydrophones towed from the stern. The seismic source was a 245/245 cm$^3$ (15/15 in$^3$) mini GI airgun fired at 13.790 MPa (2000 psi), and recorded by a 24-channel, 75 m-long active streamer with 3.125 m group spacing and a 12 m nominal shot interval (Magnani et al., 2008). The sampling rate of the data was 0.5 ms resulting in a Nyquist frequency of 1000 Hz. The signal to noise ratio of the data so acquired is exceptionally high in part because of the unique method of acquisition: the barge was piloted backward through the river with the streamer and source leading the way. The vessel faced upstream, drifting downstream at a speed less than that of the river, so to reduce the speed through the water of the hydrophone array, while maintaining a straight geometry. This method significantly reduced the noise caused by the recording array traveling at high speed through the water.
Over the course of the three-week-long survey, over 40,000 shots were fired, resulting in over 960,000 traces of raw data (Magnani et al., 2008). However, for the purpose of this study, a subset of data was used, located in the northern section of the Mississippi Embayment south of the NMSZ, between Caruthersville, MO and Wilson, AR (Fig. 2). This subset (Table 1) includes six days of acquisition, consisting of nine seismic lines and 16,545 shots, resulting in a total of 397,080 traces of raw data. The first profile (Line 131) crosses the southern arm of the NMSZ seismicity, and the last profile (Line 139) approaches the Eastern Margin of the Reelfoot Rift (EMRR) (Fig. 2).

### TABLE 1
**SUBSET OF MISSISSIPPI RIVER PROJECT DATA**

<table>
<thead>
<tr>
<th>Julian Day</th>
<th>Line Number</th>
<th>Number of Shots</th>
<th>Number of Traces</th>
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<tr>
<td>167</td>
<td>131</td>
<td>2841</td>
<td>68184</td>
</tr>
<tr>
<td>168</td>
<td>132</td>
<td>623</td>
<td>14952</td>
</tr>
<tr>
<td>170</td>
<td>133-135</td>
<td>1205</td>
<td>28920</td>
</tr>
<tr>
<td>171</td>
<td>136-137</td>
<td>3946</td>
<td>94704</td>
</tr>
<tr>
<td>172</td>
<td>138</td>
<td>4018</td>
<td>96432</td>
</tr>
<tr>
<td>173</td>
<td>139</td>
<td>3912</td>
<td>93888</td>
</tr>
<tr>
<td><strong>Totals</strong></td>
<td><strong>9</strong></td>
<td><strong>16545</strong></td>
<td><strong>397080</strong></td>
</tr>
</tbody>
</table>
The Mississippi River Project was highly successful in imaging the top of the Paleozoic, Cretaceous, and Cenozoic sedimentary deposits of the Mississippi Embayment (Fig. 11) over the entire profile. This allows for the determination of the thickness of the sedimentary column, which is essential information for the attenuation calculations. The reflection profiles confirm both the sediment thicknesses as mapped by Dart (1992) and their increasing thickness toward the south. Assuming an interval Vp= 2 km/s for the unconsolidated sediments from the base of the alluvium to the top of the Cretaceous, as provided by well log data from Wilson well 2-14 used by Langston (2003a), we calculated a thickness of the sedimentary cover ranging between 750 m- 900 m from north to south (Fig. 11).
CHAPTER 4

METHOD

Basic filter theory dictates that the amplitude spectrum of a wave is simply the product of its frequency domain operators. These operators can typically be classified into three general groups: source properties, path properties, and instrument properties (Båth, 1974). A wave that is detected at a receiver is shaped by all three of these factors. The reflection data collected in the Mississippi River Project is assumed to have amplitude spectra that are affected by a source operator, receiver operator, geometrical spreading term, ground effects from the sedimentary layers, and attenuation. The calculation of the amplitude spectrum of a trace of reflection data is

\[ |Y(\omega)| = |S(\omega)|I(\omega)|G(r)|E(\omega, r)|A(\omega, r), \]  

where \( S(\omega) \) is the source operator, \( I(\omega) \) is the receiver operator, \( G(r) \) is the geometrical spreading operator, \( E(\omega, r) \) is the effect due to the earth sediment layers, and \( A(\omega, r) \) is the attenuation operator.

Chapman (2008) utilized a natural log spectral ratio technique in order to determine Q values for the Virginia Coastal plain. Chapman (2008) outlined the method as follows: the Q factor can be reasonably approximated by a horizontally layered velocity structure with frequency-independent Q. We can divide the seismogram into two time windows of equal length with the first starting at \( \tau_p \) and the second starting at \( \tau_q \) (Fig. 3). Each time segment has \( N \) sample points with sample interval \( \Delta \tau \). For \( N \Delta \tau \)
sufficiently small, the Fourier amplitude spectra of each time window \(|Y_p(\omega)|\) and \(|Y_q(\omega)|\) can then be estimated by assuming average values for the two-way travel time for each segment. The ratio of the two spectral amplitudes can then be represented by

\[
\frac{|Y_p(\omega)|}{|Y_q(\omega)|} = \left[ \frac{\sum_{j=p}^{p+N} \alpha_j \exp(i\omega\tau_j)}{\sum_{j=q}^{q+N} \alpha_j \exp(i\omega\tau_j)} \right] \exp\left[ -\omega(\tau_p + N\Delta\tau / 2) / 2Q \right].
\]

(2)

The effects of the sediment column on the spectrum are accounted for by the \(\alpha_j\) terms, which contain a product of transmission and reflection coefficients. The \(1/\nu_j\tau_j\) terms are representative of the geometrical spreading with \(\nu_j\) as the rms velocity for the jth reflection and \(\tau_j\) is the two-way travel time. The source and receiver effects are the same for both amplitude spectra and therefore factor out in the ratio.

Taking the natural log of Equation 2 yields

\[
\ln \left( \frac{|Y_p(\omega)|}{|Y_q(\omega)|} \right) = C + (\tau_q - \tau_p)\omega / 2Q.
\]

(3)

where \(C\) is a constant that contains the geometrical spreading and earth structure effects, \(Q\) is the seismic quality factor, the term on the left-hand side is the natural log of the ratio of spectral amplitudes for time windows \(\tau_p\) and \(\tau_q\), and \(\tau_p - \tau_q\) is the time window length.

As a first order approximation, \(Q\) is assumed to be constant over every layer. This
assumption allows Equation 3 to represent the equation of a line. Therefore, if the slope of the best fit line to the natural log spectral ratio can be found, then $Q$ can be extracted. After a simple substitution of $\omega$ from radians to frequency in Hz, $Q$ can be calculated by Equation 4

\[
Q = \frac{\tau_q - \tau_p}{\mu} \pi,
\]

(4)

where $\mu$ is the slope of the best fit line to the natural log spectral ratio over the frequency band of interest. The frequency band for the linear fit can be determined by inspection of the natural log spectral ratio. In order for this method to work, a frequency independent $Q$ must be assumed over this frequency band. The nature of the dependence of $Q$ on frequency is an important, outstanding issue but is beyond the scope of this study.

Simply stated, this method works because the frequency content of an unattenuated source signal is compared to the frequency content of signal that has been attenuated by the sedimentary column. The first time window used for analysis is dominated by an unattenuated direct wave, which contains the source spectrum. The water allows the direct wave to pass relatively unattenuated to the receiver. The second time window contains the spectral content of the attenuated wave content. In essence, the method compares the spectral amplitude of one snapshot in time to the spectral amplitude of a later snapshot in time. By comparing the frequency content and amplitude of the two time windows, the attenuation of the wave due to the sediment column can be inspected. All effects due to the source and receivers cancel out in the ratio; effects to geometrical spreading and scattering effects due to heterogeneous structure are taken into account via
the intercept of the spectral ratio; and the effect due to the attenuation is what remains and is represented by the slope of the spectral ratio over the frequency band of interest.

**FIG. 3.** Example of typical marine reflection data collected from the Mississippi River Project. Represents data collected from one trace. \( \tau_p \) represents the first time window. \( \tau_q \) represents the second time window. The windows are of equal length 0.4 s.

This method addresses some of the biases outlined by Langston (2003b). Additionally, Morozov (2008; 2009) suggests separating the geometrical spreading in order to gain an unbiased value of attenuation. Any geometrical spreading or earth structure effects are taken into consideration by the intercept \( C \) in Equation 3. This addresses the issue of near surface resonance effects that may have skewed prior earthquake wave spectral studies. Additionally, since the source and receiver effects for
each measurement are the same for each data trace, this eliminates contamination of frequency content introduced by the seismic source or by the individual hydrophones. The most common noise element that is introduced to the spectrum is caused by water bottom multiples, which are easily recognized in the data due to their arrival time and frequency content.
CHAPTER 5
SYNTHETICS

In order to verify that the spectral ratio method provides accurate Q values, the technique was tested on synthetic seismograms. The synthetic seismograms were created using a velocity structure determined from the Wilson 2-14 acoustic well log data (Fig. 4) used by Langston (2003a). The well location is near Keiser, AR, which is in close geographical proximity to the end of Line 139 of data and is therefore an appropriate velocity model with which to create synthetic reflection data.
FIG. 4. An example of application of the Chapman et al. (2008) spectral technique to synthetic data. The data are shown in the lower left hand corner with the chosen time windows. The natural log of the spectral ratio of the amplitude spectra of these time windows is shown in the upper left. The linear regression from 10 to 150 Hz is shown. The model comes from an acoustic well log for the Wilson 2-14 well near Keiser, AR. The model shown in the upper right has 1002 thin layers including a water layer. The spectral falloff beyond 150 Hz in the natural log spectral ratio shown above is due to numeric noise introduced by the time impulse function assumed in the production of the synthetic data.

A variety of synthetics were created with assumed constant values of $Q_p$ for the sediment column. The different input values of $Q_p$ were 50, 100, and 200. The synthetics were created by using a Thompson-Haskell propagator matrix method used by Langston (2004). The method assumes that an impulsive, plane pressure wave strikes a plane-layered, elastic halfspace with the known velocity structure applied. The impulse function used is a Gaussian impulse function with a 0.1 s time width. The models created have 1002 thin layers with a water layer on top and a half space below. The plane pressure
wave incident on the layered surface closely resembles the experimental setup in the sense that near vertical offset can be assumed when comparing the 75-meter-long streamer to the nearly 1 km of depth analyzed from the marine seismic reflection data.

The spectral analysis method was then applied to the resulting synthetic reflection data. Two time windows of 0.5 s were selected beginning at 0 s. This time window length was chosen because it is comparable to sampling a sediment depth of ~1 km. The linear regression for the natural log spectral ratio was conducted over the frequency band 10-150 Hz. For $Q_p=50$, the spectral analysis technique recovered $Q_p$ to within 6% of the assumed value (Fig. 4). Table 2 summarizes the results of the synthetic tests. Interestingly, for $Q=200$, the percent difference was noticeably higher than the percent differences found for the $Q=50$ and $Q=100$ scenarios. These findings indicate that the method can determine reasonable values of $Q$.

<table>
<thead>
<tr>
<th>Input</th>
<th>Output</th>
<th>Percent Difference</th>
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<tbody>
<tr>
<td>$Q_p$</td>
<td>$Q_p$</td>
<td></td>
</tr>
<tr>
<td>50</td>
<td>47.43</td>
<td>5.28</td>
</tr>
<tr>
<td>100</td>
<td>97.48</td>
<td>2.55</td>
</tr>
<tr>
<td>200</td>
<td>243.66</td>
<td>19.68</td>
</tr>
</tbody>
</table>

**Table 2**

RESULTS FOR SYNTHETIC SEISMOGRAM TRIALS

$T_p=0$ s, $T_q=0.5$ s, WINDOW LENGTH=$0.5$ s, LINEAR REGRESSION FROM 10-150 Hz
CHAPTER 6

RESULTS

The spectral ratio technique was applied to all nine lines of data in a variety of ways. In general, the data were processed as follows: the original, raw reflection shot gathers (which are composed of 24 hydrophone channels) were converted from SEGY format to SAC format for ease of data processing. Each trace was normalized by its maximum energy, and then divided into two adjacent time windows of equal length. These windows are labeled \( \tau_p \) and \( \tau_q \) with the former signifying the start of the first time window and the latter signifying the start of the second window. Time windows were chosen so as to contain energy from the sediment column alone and not the underlying Paleozoic basement. For every data rendering, \( \tau_p \) always started at 0 s. A Fast Fourier Transform (FFT) was taken on each time window, after a Hanning taper was applied in order to reduce spectral lobes, and each amplitude spectra was saved. The natural log of the ratio of the two spectral amplitudes was calculated, resulting in 24 individual natural log spectral ratio files for each shot gather. All 24 natural log spectral ratio files were then averaged together to give a single, averaged natural log spectral ratio for the single shot gather.

The resulting averaged natural log spectral ratio was then fitted with a standard linear regression over a frequency bandwidth. The frequencies analyzed were limited only by the Nyquist frequency; however, the bandwidth chosen to perform the linear regression was selected based on the interval of frequencies over which the slope of the natural log spectral ratio is steepest. Many natural log spectral ratios were visually
inspected to determine the general trend of steep slope. This range was typically 10-150 Hz and 10-200 Hz. Both ranges were explored, as will be discussed later. Because of the large amount and consistency of data, processing was simplified by applying a single frequency bandwidth to all of the single shot spectral ratio files. The bad regression picks could later be eliminated by finding correlation coefficients that were less than a chosen tolerance level.

Once the bandwidth was selected, the linear regression was applied to the averaged single shot spectral ratio file over the chosen bandwidth. The slopes of the best fit lines (µ), as well as their associated errors, and correlation coefficients of the fits were recorded for subsequent analysis. These processing steps were performed for every shot gather (the average of 24 corresponding trace spectral ratios) over the entire data set (9 lines). Different averaging intervals were also tested. Once µ was calculated for a single shot, the natural log spectral ratio files for 10 shot gathers (240 trace spectral ratios) and 100 shot gathers (2400 trace spectral ratios) were also averaged. A linear regression was then applied to the resulting averaged spectral ratio files for these different intervals, and µ values with their associated statistics were also recorded. This was done to bypass any small-scale changes and search for large-scale trends, as each shot gather corresponds to a distance of about 12 meters along the river. The variance of the natural log spectral ratio files is greatly reduced when the natural log spectral ratios from multiple traces are averaged together (Fig. 5). This reduction of variance provides a better linear fit and therefore a more representative Q value.
FIG. 5. Averaging multiple natural log spectral ratios reduces the variance and provides a more accurate Q estimate. The natural log spectral ratios are shown for a single trace, a single shot gather (24 traces), the average of 10 shot gathers (240 traces), and the average of 100 shot gathers (2400 traces). Respective slopes and standard deviations are also shown.

The values of Q resolved from the data are largely based on the time window length and frequency bandwidth over which the linear regression is calculated. Therefore, the data were processed using different, appropriate time window lengths and frequency bandwidth values, resulting in a total of five different processing parameter sets. Each processing set uses all of the data listed in Table 1 (Lines 131-139). The names and parameters for each data processing set are provided in Table 3.
The primary difference between the D and E datasets is the time window length used for processing. The D sets use a 0.5 s time window, and the E sets use a 0.4 s time window. The difference between the time windows corresponds to different depths in the sediment column, with the 0.5 s time windows sampling the first kilometer or so of sediments and the 0.4 s time windows sampling the first 800 m of sediments.

It is also important to note that even though D2 and D3 use the same processing parameters, they were rendered with different trace averaging methods. The single shot gathers in D2 were processed as listed above, by averaging the 24 natural log spectral ratio files computed from the 24 trace files—the Log Spectral Ratio averaging scheme (LSR in Table 3). In order to explore the robustness of different averaging techniques, D3 was processed in a different manner—the Numerator-Denominator averaging scheme (ND in Table 3), given its title because it separately averages the numerator and denominator terms in the left-hand side of Equation 3 before taking the ratio and natural log. For D3, for each shot gather, the 24 amplitude spectra from the first time windows were averaged together, and the 24 amplitude spectra from the second time windows were averaged together.

### Table 3
PARAMETERS FOR DATA PROCESSING

<table>
<thead>
<tr>
<th>Dataset Name</th>
<th>Win. Length(s)</th>
<th>Linear Fit (Hz)</th>
<th>Averaging Scheme</th>
</tr>
</thead>
<tbody>
<tr>
<td>D1</td>
<td>0.5</td>
<td>10-150</td>
<td>LSR</td>
</tr>
<tr>
<td>D2</td>
<td>0.5</td>
<td>10-200</td>
<td>LSR</td>
</tr>
<tr>
<td>D3</td>
<td>0.5</td>
<td>10-200</td>
<td>ND</td>
</tr>
<tr>
<td>E1</td>
<td>0.4</td>
<td>10-200</td>
<td>LSR</td>
</tr>
<tr>
<td>E2</td>
<td>0.4</td>
<td>10-150</td>
<td>LSR</td>
</tr>
</tbody>
</table>
were averaged together. The natural log spectral ratio was then calculated from the two resulting files. Likewise, the natural log spectral ratio for 10 shot gathers was calculated by averaging 240 amplitude spectra from the first time window and 240 amplitude spectra from the second time window. The processing beyond this point was the same for D3 as for the other processing sets.

As previously mentioned, since a single frequency bandwidth was chosen for every linear regression, the imposed frequency range did not necessarily result in the best fit for some of the data. The selected frequency range may overshoot or undershoot the appropriate range for which the slope is constant in the natural log spectral ratio. However, these “bad picks” can easily be discerned because they are characterized by a large standard deviation and thus a low correlation coefficient. For the purposes of this study, any regression with a correlation coefficient less than 0.5 was discarded. From inspection of the data, a tolerance level of 0.5 rids the dataset of the extreme outliers while leaving about 95% of the data points for every parameter set, greatly reducing the scatter of values for μ.

The applied method has good resolution in the sense that it easily discriminates between different spectral ratio slopes, which in turn relate to different Q values. For example, the slopes associated with Q=50 and Q=100 are distinctly separate and can each be distinguished from the other. Figure 6 shows a stack of spectral ratio files that have been averaged over ten shot gather intervals. The slopes associated with different Q values (solid, colored lines) are also shown, demonstrating the good resolution of the method.
FIG. 6. Stack of 62 averaged natural log spectral ratio files from Line 132 and processed with parameter set E2. Each average spectral ratio file is comprised of the average of ten shot gather spectral ratios. The slopes associated with Qp values of 10, 50, 100, 200, and 1000 are shown for the linear regression in the frequency band 10-150 Hz. The intercept used for all slopes is 2.5.
Since the employed method directly calculates the slope (μ) of the natural log spectral ratio, the μ values determined by analyzing single shot gathers are presented first. The μ values having correlation coefficient tolerance levels above 0.5 have been plotted in histogram form (Fig. 7 a-e). The histograms are shown in the same order as the processing sets in Table 3. Table 4 shows the lognormally determined means (discussed below), one standard deviation from the mean μ, average Q values for the associated average μ, and minimum and maximum Q based off of one standard deviation from the mean. The average Q (Q_{avg}) values shown in Figure 7 a-e were calculated by substituting the average μ (μ_{avg}) for μ in Equation 3. The average values of Qp found for single shot gathers ranged from Qp=100-144.

<table>
<thead>
<tr>
<th>Data Set</th>
<th>μ_{avg}</th>
<th>Std</th>
<th>Q_{avg}</th>
<th>Qmin</th>
<th>Qmax</th>
</tr>
</thead>
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<tr>
<td>D1</td>
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<td>.0041</td>
<td>106</td>
<td>83</td>
<td>147</td>
</tr>
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<td>E2</td>
<td>.0125</td>
<td>.0037</td>
<td>100</td>
<td>78</td>
<td>143</td>
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</table>
FIG. 7. Histograms of $\mu$ values for single shot gathers for parameter sets (a) D1, (b) D2, (c) D3, (d) E1, (e) E2. Histograms have 50 bins each. $\mu_{\text{avg}}$, standard deviation of the slope, and $Q_{\text{avg}}$ is calculated for each parameter set. Values shown are for correlation coefficient tolerance levels above 0.5.
Visual inspection of the histograms in Figure 7 suggests a lognormal distribution of \( \mu \) values. Further analysis also indicates that the slope distributions exhibit lognormal behavior. A probability plot is a graphical representation of a given distribution (Ang and Tang, 2007). The cumulative frequencies, or probability, of the data are plotted versus the data. Different scales representing varied distributions are prepared depending on the type of distribution to be tested. The linearity of the data trend, when observed on the distribution-specific scale, indicates whether or not the data follows that particular distribution. If the trend is linear, the data matches the distribution; a lack of linearity suggests a different distribution. Figures 8a and 8b show normal and lognormal probability plots respectively of parameter set D1 for single shot gather intervals. The linearity observed on the lognormal probability plot further suggests that the distribution is lognormal. Figure 8c shows a histogram of D1 for single shot gathers with normal and lognormal distribution curves fit to the data. It is also apparent from this that a lognormal distribution is likely for the data. Lognormal statistics were therefore incorporated for every data set.

As spectral ratios from greater numbers of shot gathers were averaged (Figs. 9 and 10), the distributions became increasingly difficult to tell between normal and lognormal. However, since the largest amount of data came from the single shot gather spectral ratios and were lognormally distributed; lognormal statistics were maintained throughout all data sets for consistency of analysis. Regardless, interpreting the data with normal statistics makes an insignificant change of about 0.1Q when compared to the lognormal analysis (Fig. 8c).
FIG. 8. Comparison of normal versus lognormal statistical analysis. (a) Normal probability plot for single shot gathers of D1. (b) Lognormal probability plot for single shot gathers of D1. (c) Histogram of single shot gathers of D1. Best fits of normal and lognormal distributions are shown.
The slope values $\mu$ determined by averaging the spectral ratios of 10 shot gathers are presented next (Fig. 9 a-e). Again, only the good picks are shown and are in 25 bins for the histograms. Table 5 shows the respective statistics. Values of Qp for averages of ten shot gathers ranged from $Qp=108-159$. The same is done for the values determined by averaging the spectral ratios of 100 shot gathers (Fig. 10 a-e). The histograms for 100 averaged shot gathers are grouped in 20 bins. Table 6 shows the statistics for these Qp values, which range from 110-160.
### Table 5

$Q_{avg}$ VALUES AVERAGED BY 10 SHOT GATHERS

<table>
<thead>
<tr>
<th>Data Set</th>
<th>$\mu_{avg}$</th>
<th>Std</th>
<th>$Q_{avg}$</th>
<th>$Q_{min}$</th>
<th>$Q_{max}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>D1</td>
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<td>87</td>
<td>144</td>
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<tr>
<td>D2</td>
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</tr>
<tr>
<td>D3</td>
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<td>.0025</td>
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<td>120</td>
<td>195</td>
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<tr>
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<td>E2</td>
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<td>.0032</td>
<td>133</td>
<td>105</td>
<td>182</td>
</tr>
</tbody>
</table>

### Table 6

$Q_{avg}$ VALUES AVERAGED BY 100 SHOT GATHERS

<table>
<thead>
<tr>
<th>Data Set</th>
<th>$\mu_{avg}$</th>
<th>Std</th>
<th>$Q_{avg}$</th>
<th>$Q_{min}$</th>
<th>$Q_{max}$</th>
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<td>D1</td>
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<tr>
<td>D2</td>
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<td>139</td>
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</tr>
<tr>
<td>D3</td>
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<td>.0020</td>
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<td>.0116</td>
<td>.0022</td>
<td>136</td>
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<td>169</td>
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</table>
FIG. 9. Histograms of $\mu$ values for shot gathers averaged every 10 for parameter sets (a) D1, (b) D2, (c) D3, (d) E1, (e) E2. Histograms have 25 bins each. $\mu_{avg}$, standard deviation of the slope, and $Q_{avg}$ is calculated for each parameter set. Values shown are for correlation coefficient tolerance levels above 0.5.
FIG. 10. Histograms of $\mu$ values for shot gathers averaged every 100 for parameter sets (a) D1, (b) D2, (c) D3, (d) E1, (e) E2. Histograms have 20 bins each. $\mu_{\text{avg}}$, standard deviation of the slope, and $Q_{\text{avg}}$ is calculated for each parameter set. Values shown are for correlation coefficient tolerance levels above 0.5.
Figure 11 shows the lateral variation of Qp with river distance, and the statistics for the individual parameter sets are listed in Table 6. Each Qp value shown is derived from the averaging of the 100 shot gather natural log spectral ratios. All five parameter sets are shown. For brevity and visual clarity, the Q profiles determined from the single shot gather averages and the 10 shot gather averages are not shown. However, they follow a similar trend as the data shown. No data were eliminated in this set because every correlation coefficient from the linear regression analysis was above 0.5. In fact, when averaging the natural log spectral ratios from 100 shot gathers (2400 traces), 87% of the data had correlation coefficients that were greater than 0.8. This is in relation to 32% of single shot gather averaged data and 61% of 10 shot gather averaged data had correlation coefficients above 0.8 respectively. This indicates that averaging by greater quantities of shot gathers reduces the variance in the data and provides a more representative value of Q over a given distance (approximately 1.2 km river distance). The Q profile is spatially correlated with the subsurface features shown below. The record section shown is the interpreted result from the Mississippi River Project (Magnani, 2008).

Based upon interpretation of all three groupings of data shown from the histograms (single shot gathers, 10 shot gather averages, 100 shot gather averages), average values range from Qp=100-160. Langston et al. (2005) determined Qp=200 and Qs=100 from sediment trapped P- and Rayleigh wave data in the Mississippi Embayment. This corresponds to an empirically determined ratio of Qp/Qs=2. This relationship was chosen to apply to this thesis because of the claim in the study that both Qp and Qs were independently calculated for the entire sediment column. Using this
relationship as a first order approximation to determine Qs, this thesis determines a range of Qs=50-80. It must be acknowledged that other Qp/Qs relationships for the embayment exist; however, it is not the intent of this study to directly calculate Qs.

In Figure 11, the maximum value of Qp for all the parameter sets is 297 and the minimum value is 66. The mean value of Q=132 for all of the data averaged by 100 shot gathers shown. Many relative spikes in Qp can also be seen from the profile, particularly those located to the left of Line 132 and the region of relative peaks located to the right of Line 138.
FIG. 11. Lateral variation of Qp compared with interpreted subsurface features. Individual Qp values are derived from the natural log spectral ratio of the average of 100 shot gathers. Record section interpreted by Magnani (2008). Paleozoic basement (purple), Cretaceous sediments (green), Mesozoic sediments (orange), and Quaternary alluvium (yellow), and interpreted faults (vertical, dark red lines) are shown. Two way travel times of 0.8 s and 1.0 seconds correspond to approximately 800 meters and 1 km respectively. These times are associated with the top of the Paleozoic layer.
CHAPTER 7
DISCUSSION

It is obvious from Figure 11 that there are similarities in the Q estimate trends. This indicates that regardless of the parameters used in the processing, the relative changes in Q along the profile are similar. The primary difference between the parameter sets is the value of Q itself. This highlights the subjectivity and importance of choosing the data processing parameters. The data shown in terms of the averaged single shot gathers and the averaged 10 shot gathers show similar trends with minute, small-scale variations in Q and do not contribute to the overarching trend. Therefore, in order to clearly interpret the overall trend, data from the 100 averaged shot gathers shown in Figure 11 are analyzed in the following paragraphs.

For the D parameter sets that use a time window of 0.5 s, the Q for D1 is generally lower than for D2 and D3, while the latter two sets are very similar in amplitude. D1 uses a frequency bandwidth of 10-150 Hz while both D2 and D3 use 10-200 Hz. Typically, the natural log spectral ratio increases with constant slope up to a certain frequency at which point it flattens out, thus creating a ‘knee’ (Figs. 4, 5, and 6). It is ideal to take the linear regression in the portion of constant slope and ignore the region after the spectral knee. If the regression is taken past the spectral knee, the calculated slope is therefore underestimated (Q overestimated). Since for D2 and D3 the linear regression is calculated over a wider frequency range (10-200 Hz), it is likely that the slope is being underestimated, resulting in higher Q values than those shown for D1, which uses the 10-150 Hz bandwidth for regression. This may suggest that the \( \mu \) values
calculated for D1 are more accurate because they do not underestimate the slope; however, inspection of the correlation coefficients for the D sets shows very similar values that are predominantly above 0.8, indicating that there is not a set that fit the spectral ratios better than the others. It is also evident that a relative spectral dip around 150 Hz in the natural log spectral ratio skews the slope lower (Fig. 6). Even though the trend is apparent from 10-200 Hz, this spectral dip misrepresents the slope over the frequency band of interest. Therefore, the linear regression analyzed from 10-150 is more representative of the trend as it does not contain the spectral dip.

It is interesting to note that Chapman (2008) found a similar spectral falloff after 150 Hz. He speculated that single scattering is the primary mechanism below approximately 150 Hz, and multiple scattering takes precedence at higher frequencies. The spectral falloff in the current results is not interpreted to be a function of poor signal-to-noise ratio beyond the spectral knee. Figure 12 shows the signal to noise ratio typical in the data. It is evident that the signal does not dip into noise until about 800 Hz, well beyond the region of the spectral knee (150-200 Hz).
FIG. 12. Signal to noise ratio for the average of the amplitude spectra from four traces. Spectral content from the pre-signal noise (0-0.04 s), the first first time window $T_p$ (0.04-0.44 s), and the second time window $T_q$ (0.44-0.88 s) are shown as thick blue, red, and green lines respectively.

However, there is another source evident in the data. High frequency signal-generated noise was generated by river revetment reflections during data acquisition and are probably responsible for the spectral knee as demonstrated below.
Let

\[
\hat{Y}_p(\omega) = \hat{S}(\omega) \left[ \hat{R}_p(\omega) + \hat{\eta}(\omega) \right] \\
\hat{Y}_q(\omega) = \hat{S}(\omega) \left[ \hat{R}_q(\omega) + \hat{\eta}(\omega) \right]
\]

(5)

where \(\hat{S}(\omega)\) is the source function, \(\hat{R}_p(\omega)\) is the reflection response of the earth in the first time window, \(\hat{R}_q(\omega)\) is the reflection response of the earth in the second time window, and \(\hat{\eta}(\omega)\) is the common high frequency signal generated noise in both data time windows. Taking the natural log of the amplitude spectral ratio gives

\[
\ln \left| \frac{\hat{Y}_p(\omega)}{\hat{Y}_q(\omega)} \right| = \ln \left[ \frac{\hat{R}_p(\omega)}{\hat{R}_q(\omega) + \hat{\eta}(\omega)} + \frac{\hat{\eta}(\omega)}{\hat{R}_q(\omega) + \hat{\eta}(\omega)} \right].
\]

(6)

At high frequency \(\hat{R}_q(\omega)\) goes to zero faster than \(\hat{R}_p(\omega)\) because of attenuation. This gives

\[
\ln \left| \frac{\hat{Y}_p(\omega)}{\hat{Y}_q(\omega)} \right| = \ln \left[ \frac{\hat{R}_p(\omega)}{\hat{\eta}(\omega)} + 1 \right].
\]

(7)

If the signal-generated noise, \(\hat{\eta}(\omega)\), is a flat spectrum, then the natural log spectral ratio should decrease as frequency increases because \(\hat{R}_p(\omega)\) also decreases through anelastic attenuation.
High frequency signal-generated noise is easily seen in the seismic reflection data at all time intervals, appearing as “chevron”-shaped events that span the reflection image of Figure 11. This signal is much larger than the ambient noise level. The stacked spectral ratio data of Figure 6 display a roughly linear amplitude decrease with frequency after 150 Hz, suggesting that this noise model explains the general character of the data.

The similarity between the D2 and D3 parameter sets (Fig. 11) can be explained by their similarity in processing. As previously mentioned, the only difference between the two sets is that D2 and D3 used the LSR and ND averaging schemes respectively (Table 3). Additionally, D2 and D3 lie within the error bounds of each other, suggesting that the different averaging schemes do not significantly change calculated Q values.

For the E parameter sets that use a time window of 0.4 s, the Q for E1 is generally higher than that of E2. Again, this can be explained by the frequency ranges used for linear regression. E1 uses a 10-200 Hz bandwidth, and E2 uses a 10-150 Hz bandwidth; therefore it is assumed that E1 underestimates μ (overestimates Q) because the frequency range continues past the spectral knee or is lowered by a spectral dip misrepresentative of the trend. Inspection of correlation coefficients again show that there is no preferable choice over which set has a better fit to the spectral ratios. Both sets have correlation coefficients predominantly above 0.8. These two sets also lie within one standard deviation of each other and are interpreted as not being significantly different from each other.

The trend of Q over the whole profile for all sets lie within a relatively cohesive band between about Qp=100-200. Areas of relative highs, to the left of Line 132 for
instance, are characterized by larger variance in their calculations. Furthermore, these areas do not seem to correspond to any defining subsurface features.

It is known that the presence of interstitial fluids greatly reduce Q values in rock samples (Gardner et al., 1964; Toksöz et al., 1979; Winkler and Nur, 1979; Johnston, 1981; Tittmann et al., 1981; Winkler and Nur, 1982). It has also been hypothesized that regions of fluid filled cracks and faults, when saturated, show signs of lowered Q values because of this phenomenon (Al-Shukri and Mitchell, 1990). If this is the case, it may be conceivable to identify regions of faulting by interpreting localized low Q structure for an area. Figure 11 shows shallow deformation and interpreted recent deformation along the Mississippi River. The Q profile above the record section is spatially correlated to the structures below. Careful inspection of the two profiles in tandem do not show strong evidence of Q related with subsurface sedimentary structure. Where there are interpreted faults, there is a general lack of relative low Q. Perhaps the whole study area is uniformly saturated, thereby reducing Q. This would imply that the Q values calculated with this study are a lower bound, as Al-Shukri and Mitchell (1990) conclude that the saturated, localized, and low-Q areas are surrounded by higher Q regions. Additionally, since Al-Shukri and Mitchell (1990) were interpreting the upper 5-km of crust for the New Madrid seismic zone, it is more likely that Q in the current study is not a good indicator of subsurface deformation in the sediments.

However, it is fascinating to note that previous studies analyzing earthquake body wave data use frequencies typically below 10 Hz. This study uses higher frequency content and calculates values of Qp that are similar to previously determined values
(Langston et al., 2005). This may have implications for the nature of the frequency
dependence or independence of attenuation, an outstanding scientific issue which is not
fully understood.
CHAPTER 8
CONCLUSIONS

In order to calculate Qp for the shallow, post-Paleozoic sediments of the Northern Mississippi Embayment, a natural log spectral ratio technique has been applied to marine seismic reflection data collected along the Mississippi River. The spectral ratio technique, following Chapman (2008), addresses concerns discussed by Langston (2003b), including apparent attenuation effects caused by near surface resonance effects in earthquake body wave data. Q is assumed to be constant over the sediment column and independent of frequency over a given bandwidth; this study indicates a constant slope for frequencies ranging predominantly from 10-150 Hz. Average values of Qp are determined for the profile ranging from Qp=100-160. Using an empirical relationship taken from Langston (2005), average values of Qs are approximated to range from Qs=50-80. These values are higher than previous earthquake body wave studies have found (Wang et al., 1994; Chen et al., 1994; Pujol et al., 2002; Liu et al., 1994), yet are lower or within a similar range of what other comparable studies have calculated (Langston et al., 2005; Al-Shukri and Mitchell, 1990). Interpretation of the laterally varying profile of Qp along the river indicates that there is not a significant correlation between Qp and subsurface deformation and faulting. Key processing parameters, such as time window length and frequency bandwidth for linear regression, are found to preferentially affect the value of Q but not the overall trend of variation along the profile. These findings contribute to the body of knowledge on attenuation in the seismically
active Northern Mississippi Embayment and can ultimately be applied to hazards modeling and ground shaking scenarios in the future.
REFERENCES


