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**Low Strain Shear Wave Damping (Q_s) Measurements in Champlain Sea Deposits
Using Downhole Geophysical and Lab Techniques**

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ABSTRACT

In Ottawa, Ontario, unusually high amplification ratios have been measured in clayey silts (known as 'Leda Clays') at low levels of earthquake-induced ground shaking. Therefore, low-strain shear wave damping and velocity properties of Leda Clays are tested in field and lab (resonant and non-resonant column) analyses using new and traditional approaches. Downhole testing in two deep Ottawa-area boreholes (96m, 54m) was carried out using a monofrequency seismic source over a frequency band of 10-100Hz, and two identical downhole triaxial geophones. Monofrequency signals also allowed for the measurement of dispersion (variation of velocity with frequency). Negligible dispersion was observed over the 10-100Hz frequency band and elevated interval Q's ranged from 100-200 (material damping of 0.2% to 0.4%). This indicates that in relatively homogenous Champlain Sea deposits, very little elastic energy is lost at low strains. This contributes significantly to the observed elevated amplifications experienced during small strain earthquakes in the Ottawa area.

TABLE OF CONTENTS

Abstract.....	i
Table of Contents	ii
List of Tables.....	v
List of Figures.....	vi
Acknowledgements.....	viii
 1.0 Introduction.....	 1
 2.0 The Problem of Soft Soil Amplification.....	 7
2.1 Seismic Site Classification	12
2.2 Shear Wave Database for Site Classification in Ottawa, ON	13
2.3 V_{s30} Map for Ottawa, ON	16
2.4 Fundamental Site Period Map for Ottawa, ON	21
 3.0 Attenuation in Soils	 24
 4.0 Geophysical/Geotechnical Properties of Champlain Sea Deposits Near Ottawa	 32
4.1 John Shaw Rd Borehole: Kinburn, ON (BH-GSC-JSR-01)	34
4.2 Anderson rd. Borehole: Ottawa, ON (BH-AR93-1)	41
 5.0 Geotechnical Testing of Dynamic Properties	 43
5.1 Stress and Strain.....	45
5.2 Inherent Soil Properties.....	48
5.3 Resonant Column (RC) Testing.....	50
Review of previous work.....	50
Test Procedures for this study	53
Results - RC.....	60

5.4 Non-Resonant (NR) Column Testing.....	63
Review of Previous work.....	64
Test Procedures for this study	65
Results - NR	67
6.0 In situ Measurements of Q – Processing Approaches.....	71
6.1 Pulse Broadening.....	74
6.2 Spectral Ratio Method.....	77
6.3 Relationship Between Attenuation, Velocity Dispersion, and Q.....	81
7.0 Attenuation Measurements – Field Work.....	86
7.1 Calibration of Downhole Tools	88
7.2 Experimental Setup in the Field	93
7.2.1 Dispersion Experiments	95
7.2.2 Amplitude Experiments.....	98
8.0 Data Analysis and Results.....	100
8.1 Pulse Broadening	100
8.2 Spectral Ratio - Analysis.....	105
8.3 Spectral Ratio - Results.....	120
8.4 Dispersion.....	135
9.0 Discussion.....	143
9.1 <i>In situ</i> Processing Approaches for Very Small Strain	145
9.1.1 Testing Considerations - Monofrequency Approach	148
9.2 Lab vs. <i>In Situ</i> Data	151
9.3 Intrinsic vs. Effective Q.....	152
10.0 Conclusions	154

11.0 Future Research.....	161
Bibliography.....	162
Appendix A - Spatial Correlation of Vsav Depth Functions	168
Appendix B - Geotechnical RC Lab Data.....	197
Appendix C - Formulae for Non-resonant Column Testing.....	205
Appendix D – Additional Downhole Shear Wave Testing.....	210

LIST OF TABLES

Table 2.1 – 2005 NBCC Seismic Site Classification Scheme.....	13
Table 5.1 – Soil Behaviour at Different Strain Levels.....	45
Table 5.2 – RC Test Results from Previous Studies.....	52
Table 5.3 – Shear Strength and Soil Moisture Content Tests From This Study	55
Table 5.4 - Calculated Confining Pressures for the RC/NR Testing.....	57
Table 5.5 – Summarized RC Test Results From This Study.....	60
Table 5.6 – Summarized NR Test Results From This Study	68
Table 6.1 – Q_s Measurements From Previous Studies	73
Table 7.1a-c – Results of Geophone Calibration Tests.....	91
Table 8.1 –Travel Time Error Analysis Between Two Downhole Tools.....	118
Table 8.2 – Slope Error Analysis of Spectral Ratio Method.....	119
Table 8.3 - Parameters of the Monofrequency Amplitude Data Sets	121
Table 8.4 - Parameters of the Monofrequency Dispersion Data Sets.....	136
Table 8.5 - Sample Dispersion Table From the JSR Borehole.....	138
Table 8.6 – Linear Equations of V_s vs. Frequency	141
Table 8.7 – Calculated Variation in V_s vs. Frequency	141
Table 8.8 – Calculated Q_s and Damping Using Dispersion Approach.....	142
Table 9.1 – Summarized Q and Damping Results from This Study.....	144

LIST OF FIGURES

Figure 2.1 – Amplification Effects.....	10
Figure 2.2 – Ottawa-area Resonance Amplification	11
Figure 2.3 – Seismic Test Site Locations	15
Figure 2.4 – Borehole Locations.....	16
Figure 2.5 – 2005 NBCC Seismic Site Class Map of Ottawa.....	18
Figure 2.6 – Ottawa Spectral Accelerations Pre- and Post- V_{s30} Scaling Factors	20
Figure 2.7 – Fundamental Site Period Map of Ottawa	22
Figure 2.8 – Influence of Soil Damping on Resonance Amplification.....	23
Figure 3.1 – Simplified Surficial Geology of Ottawa.....	30
Figure 4.1 – Downhole Geophysical/Geotechnical Logs from BH-GSC-JSR-01.....	36
Figure 4.2 – Q Test Intervals in BH-GSC-JSR-01	39
Figure 4.3 - Downhole Geophysical/Geotechnical Logs from BH-AR93-1	42
Figure 5.1 – Hysteretic Stress-Strain Behaviour in Soils	46
Figure 5.2 – Sample Modulus Reduction/Damping Curves.....	47
Figure 5.3 – Resonant Column Schematic and Data Sample	54
Figure 5.4 – Resonant Column System Components.....	56
Figure 5.5 – Resonant Column Drive Plate Assembly	59
Figure 5.6 – Schematic of Frequency Effects on Damping and Velocity	65
Figure 5.7 – Idealized Transfer Function for NR test.....	67
Figure 5.8 – NR Results from 10m Sample	69
Figure 5.9 – NR Results from 30m Sample	70
Figure 5.10 – NR Results from 70m Sample.....	70
Figure 6.1 – Definition of Pulse Width.....	75
Figure 6.2 – Definition of Debye Peak.....	84
Figure 7.1 – Monofrequency Downhole Testing Configuration.....	88
Figure 7.2 – Geophone Calibration with 100Hz Source	89

Figure 7.3 – Geophone Calibration with Impulsive Source	92
Figure 7.4 – Field Experimental Setup.....	94
Figure 7.5 – Signal Taper Design.....	97
Figure 8.1 – Filtering of Tube Waves for Pulse Broadening Dataset.....	102
Figure 8.2 – Results of Pulse Broadening - BH-GSC-JSR-01.....	103
Figure 8.3 - Results of Pulse Broadening - BH-AR93-1	105
Figure 8.4 – Effect of Window Length on Spectral Ratio Approach.....	106
Figure 8.5 – Reflections from Bedrock and Glacial Horizons	109
Figure 8.6 –Time Traces in BH-JSR Showing Bedrock/Glacial Reflections	111
Figure 8.7 – Impact of Time Window Length on Frequency Spectrum.....	112
Figure 8.8 – Impact of Time Window Length on Spectral Ratio Slopes	114
Figure 8.9 – Q and Damping Calculations for JSR Interval 2 (30m-50m).....	122
Figure 8.10 – Spectral Ratios $\pm 2\text{Hz}$ on Either Side of Main Frequency Peak.....	124
Figure 8.11 – Q and Damping Calculations for JSR Interval 3 (30m-70m)	126
Figure 8.12 – Q and Damping Calculations for JSR Interval 4 (30m-90m)	127
Figure 8.13 – Q and Damping Calculations for JSR Interval 1 (20m-30m)	129
Figure 8.14 – Q and Damping Calculations for AND Interval 1 (10m-30m)	133
Figure 8.15 – Q and Damping Calculations for AND Interval 2 (10m-45m)	134
Figure 8.16 – Q and Damping Calculations for AND 30-45m Interval	135
Figure 8.17 – Sample Waveforms from Dispersion Experiments.....	137
Figure 8.18 – Dispersion Results - BH-GSC-JSR-01	139
Figure 8.19 – Dispersion Results - BH-AR93-1	140
Figure 9.1 – Idriss’ 1990 Damping Curve for Clay	147
Figure 10.1 – Improved Monofrequency Test Results vs. Broadband.....	157

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1.0 INTRODUCTION

The damage caused in urbanized areas by earthquakes when seismic waves reach the surface is due to several parameters: source characteristics (magnitude, mechanism, depth), travel path and distance to the site, and rock types along the travel path. However, the ground motions at surface are very strongly influenced by the materials encountered during the last few hundred meters of the travel path. In studies of damage resulting from numerous worldwide earthquakes (Niigata, Japan, ($M_w 7.5$) 1964; Anchorage, Alaska, ($M_w 9.2$) 1964; Mexico City, Mexico, ($M_w 8.0$) 1985; Loma Prieta, California, ($M_w 6.9$) 1989; Northridge, California, ($M_w 6.7$), 1994), it was observed that the greatest destruction tended to be seen in areas where soft soils were present in the near surface, causing a ground motion amplification effect (Anderson 1986; Martin 1994; Anderson et al. 1996).

Amplification of seismic waves due to soils near surface has been recognized since the 1970's, and early soil factors which accounted for site effects were introduced into building codes of the day. Over the last few decades, the link between soft soil amplification and shear wave velocity of the subsurface has been established (Kramer 1996). In a medium of low velocity (low density) materials overlying higher velocity (higher density) materials, an upward travelling wave will undergo a shortening of shear wavelength and an increase in shear amplitude across a wide frequency band as it travels across the boundary between the two materials (Shearer and Orcutt 1987). The ratio of the acoustic impedance contrast (the product of density and velocity) between the two layers has a strong effect on

the resulting amplitude of the ground motions, and a greater contrast will result in an increased amplification effect. Other factors also contribute to the amplification effects, including resonance (trapped wave phenomenon), focusing/defocusing effects, and the generation of surface waves in buried valleys.

Acting to reduce the level of ground motion are the natural damping characteristics of the materials. A certain amount of the energy contained by a seismic wave will be lost due to anelastic attenuation (energy converted to heat through grain friction) as it travels from the source. A medium with elevated damping characteristics will act to reduce the overall effect of amplification, while soils exhibiting low damping characteristics will effect little change on amplification at a given site.

The Ottawa-Gatineau area is ranked third in Canada (behind Vancouver and Montreal) for large Canadian cities at seismic risk, based on earthquake hazard coupled with population size (Adams et al. 2002). Ottawa is located within the Western-Quebec seismic zone (WQSZ) and local seismometers record, on average, 1-2 low magnitude ($<M3.0$) earthquakes per week, making it an excellent location to study the elevated amplification effects from frequent, local, small-strain earthquakes.

A frequency spectrum is a representation of a seismic signal in the frequency domain, where both the amplitude and/or the phase of the signal can be plotted against frequency. A ratio of the spectral amplitudes at two different sites can be taken to look at the relative increase in amplitude versus frequency from one site to

another. In Ottawa, the spectral ratio of soil-to-rock (SSR) between the hard Paleozoic/Precambrian bedrock and the soft Champlain Sea deposits can yield spectral amplifications of 80+ times during weak motion events (Motazedian and Hunter 2008). Although damping is occurring, the level at which it attenuates soil motion is currently not well understood.

Since 2005, Geological Survey of Canada (GSC) in collaboration with Carleton University has been working within the Eastern Canada Geohazards Assessment Project (a sub-activity of the Reducing Risk from Natural Hazards Program), studying seismic amplification in the Ottawa area, where soft, silty soils are known to overlie high-velocity bedrock. To quantify the relative levels of predicted ground motion amplification based on shear wave velocity, the project participants have produced two seismic microzonation maps of the Ottawa area. The first depicts the 2005 National Building Code of Canada (2005 NBCC) seismic site classes, and the second depicts fundamental site period. In addition to producing the map outputs, the program's aim was to demonstrate the usefulness of geophysical techniques for the estimation of ground response due to near surface conditions.

Contribution to Microzonation Project for the City of Ottawa

As both a member of the GSC and an M.Sc. student at Carleton University, I have contributed to the earthquake hazard program through research focused in two areas:

- i) Seismic site category (V_{s30}) and fundamental site period mapping, and
- ii) damping in Champlain Sea muds ("Leda Clays").

Vs30 maps: I contributed to the development of the seismic site category map by participating in the collection and analysis of shear wave velocity data across the city. With these data (over 700 sites in total) and stratigraphic information from a pre-existing database of 20,000 boreholes, I applied geostatistical methods to investigate the spatial variation of soil thickness (and thus shear wave velocity) versus depth within the Ottawa area. This analysis resulted in a more accurate depiction of the estimated shear wave velocity profile down to 30m depth (V_{s30} , as required by the 2005 NBCC) in areas where surface velocity measurements did not exist. These results are discussed in Chapter 2, and the spatial analysis methods are presented in Appendix A.

Fundamental Site Period Map: Recently, it has been recognized that V_{s30} may not represent the entire seismic soil amplification phenomenon, and there is a trend towards inclusion of fundamental period (T_0) or fundamental frequency ($f_0 = 1/T_0$) of a site in the calculation of seismic soil amplification factors. T_0 is simply the two-way travel time down to the major acoustic impedance boundary (bedrock or till in Ottawa) and can easily be calculated from our seismic site data.

The high acoustic impedance constant in the Ottawa area can increase the possibility of significant amplification at a site's natural period due to the matching of the fundamental frequency of the soil with the frequency content of the earthquake ground motion. Thus, investigation of fundamental period is particularly important for Ottawa and areas with similar geological settings in Eastern Canada. In addition, damping plays a role in the level of amplification, as

soils with low levels of damping at low strains will not be able to significantly reduce the level of amplification at the fundamental site period.

Damping in Champlain Sea Muds (“Leda Clays”): As the *in situ* damping response, or attenuation, of soft soils at low-strains in the Ottawa area is not well understood, I undertook a study of dynamic properties (damping and shear wave velocity) of local soft soils in two Ottawa-area deep boreholes using a new approach: a monofrequency shear wave source and two downhole triaxial geophones. Lab testing was also carried out on three soil samples collected at one of the boreholes using a resonant column (RC) apparatus, and the results are compared with those collected in the field.

In Chapter 3, I present a literature review of attenuation theories and discuss in more depth the objectives of my research with regard to attenuation. I follow with a summary of the properties of the Champlain Sea muds in the Ottawa area in Chapter 4. The lab methods used to analyze the soil samples and the resulting damping measurements are presented in Chapter 5.

Chapter 6 presents the theory behind the methods I chose to process the *in situ* attenuation data. The field work is described in Chapter 7, and the analyses and results are presented in Chapter 8. I compare the results of all the testing methods – lab and field – in Chapter 9, and discuss the relative importance of the soil parameters, as observed, on attenuation.

This work contributes new understanding to why we have observed such elevated seismic amplifications in Ottawa-area soft soils during frequent low-strain

earthquakes, and additionally, how damping levels vary throughout the soil column within different frequency bands.

2.0 THE PROBLEM OF SOFT SOIL AMPLIFICATION

Seismic soil amplification has been described in literature for well over a century. MacMurdo (1824) noted from reports written during the Cutch, India earthquake of 1819 that structures founded in rock did not suffer as much damage as structures built on soil. The installation of strong motion seismograph networks over the last several decades has made it possible to make quantitative measurements of local site effects due to varying geologic conditions. This first occurred in 1989 in North America when the $M_w 6.9$ Loma Prieta earthquake struck the San Francisco Bay area, providing unprecedented strong motion data on soft soil sites, as well as on rock sites. The Loma Prieta data were used to calibrate earlier empirical derivations for ground motion amplification on soft soils (Rinne 1995). These observations of soil-to-rock amplification led to the development of the amplification factors which have been adopted in current North American building codes.

The phenomenon of amplification is strongly affected by the change in acoustic impedance, Z , as a wavefront travels upward from a material of higher density and shear wave velocity material into a material of lower density and shear wave velocity (Figure 2.1a). In absence of damping, the amplification (A) across the band of earthquake frequencies can be approximated by

$$A \sim \sqrt{\frac{\rho_{rock} V_{s_{rock}}}{\rho_{soil} V_{s_{soil}}}} \quad (2.1)$$

where ρ_{rock} is the average density of the rock, V_{Srock} is the shear wave velocity of the rock at the soil-rock interface, ρ_{soil} is the average density of the soil, and V_{Ssoil} is shear wave velocity of the soil at the ground surface (NRC 2005b). As this acoustic impedance ratio becomes larger, the amplification effect increases. This effect is approximate, however, as it neglects damping, heterogeneity in the overburden, velocity reversals or significant gradients.

Contributing to the broadband amplification effect is a resonance amplification, which occurs at the resonant frequency of the soil deposit (Figure 2.1b). A resonance amplification is set up when a wave which has travelled up from the source reaches the free surface and is reflected back toward the impedance boundary. As the reflection coefficient for Ottawa-area type soils predicts that most of the energy will be reflected back upward from the impedance boundary, the shear wave is effectively trapped in the lower velocity material at the fundamental frequency of the deposit, until the waves are eventually attenuated by geometric spreading and anelastic attenuation. The lowest frequency at which this resonance will occur is called the fundamental frequency, f_0 , and is given as

$$f_0 = \frac{V_{\text{Sav}}}{4H} \quad (2.2)$$

where V_{Sav} is the average shear wave velocity down to the impedance boundary, and H is the thickness of the soil column above the impedance boundary (Kramer 1996). The harmonics of fundamental frequency ($n > 0$), are given by

$$f_n = \frac{V_{\text{Sav}}}{4H} * (1 + 2n), n=1,2,3,4... \quad (2.3)$$

The amplitude of the resonance peak for the fundamental frequency, f_0 , is estimated by

$$A_{res} \sim \frac{\rho_{rock} V_{s_{rock}}}{\rho_{soil} V_{s_{av_{soil}}}} \quad (2.4)$$

where ρ_{soil} is the average density of the soil, and $V_{s_{av_{soil}}}$ is the travel-time weighted average shear wave velocity of the soil column. The resonance amplification predicted by 2.4 is larger than the broadband amplification predicted by 2.1, but both contribute to spectral amplification. The greatest resonance amplification generally occurs at the fundamental frequency. Earthquakes recorded in Ottawa that were rich in higher frequencies have produced harmonics which have higher amplitudes (sometimes many times higher) than the fundamental frequency.

Other site conditions complicate the prediction of ground motions, due largely to the shape of local bedrock topography. Soil amplification can also occur due to the shape of the bedrock topography (Figure 2.1c). Soils deposited in narrow basins can cause focusing and defocusing of seismic waves, resulting in increased (or decreased) ground motion (Bard and Bouchon 1985). Finally, surface waves produced as a result of seismic waves reflecting from bedrock on the edges of buried valleys (Figure 2.1d) can increase the level of ground shaking experienced at a site (Lomnitz et al. 1999).

Figure 2.1. Illustrations of the major amplification effects associated with soft soil overlying firm ground or bedrock. In the Ottawa area, such effects occur for large seismic acoustic impedance contrasts between rock and soils (Figure after Hunter et al. 2010).

Figure 2.2 compares the horizontal motions of rock versus soil from the same earthquake, and shows the effects of broadband and resonance amplification at the soil site in the frequency domain. The event was recorded in the Ottawa suburb of Orléans by the ORHO and ORIO stations of the Polaris seismograph network. These stations are located 1.6 km apart and are equipped with identical broadband seismometers. Station ORIO is installed on bedrock and station ORHO is installed

over 81m of soft soil. They both recorded the M3.0 Val-des-Bois earthquake on December 31, 2008, which had an epicentral distance of 77km from Ottawa. The station on soil shows that at frequencies from 0.5Hz to 20Hz, the motions experience significant spectral broadband amplification, which is even higher at the resonant frequency and harmonics of the soil. Due to the proximity of the earthquake, it was still rich in higher frequencies which can be seen in harmonics F_2 and F_3 which are slightly higher than the fundamental frequency.

Figure 2.2 A comparison of spectral components showing both broadband and resonance amplification in soil.

Although 2.3 predicts harmonics at equal intervals of period or frequency, in practice, the periods of the harmonics may be slightly different from one another. This effect could be influenced by local geologic conditions (e.g. presence of basin structures), heterogeneities of the soil column, or seismic signal interference.

2.1 SEISMIC SITE CLASSIFICATION

The common element to predicting these effects is knowledge of the shear wave velocities throughout the soil/rock column. This has been recognized by the 2005 National Building Code of Canada (NRC 2005) which has adopted (and slightly modified) the 1994 National Earthquake Hazard Reduction Program (NEHRP) guidelines from the United States for seismic amplification factors.

The NEHRP site classes are divided into six categories, A through F. Site classification is based on the time-weighted average shear wave velocity through the upper 30m (V_{S30}) of soil for classes A – E, with site F requiring site-specific geotechnical testing (Table 2.1). The provisions were later modified to also allow site classification based on the number of normalized blow counts, N_{60} , or undrained shear strength measurements, S_u , at a site to be used as an acceptable replacement for V_{S30} (Finn and Wightman 2003). The 2005 NBCC provides look up tables which contain amplification factors, F_a and F_v , based on the site category and on the spectral acceleration of the site at periods of 0.2 (short period) and 1.0 seconds (long period).

Site Class	Site class name and generic description	Geotechnical Properties of the Upper 30m		
		V_{s30}	N_{60}	S_u
A	Hard rock	$V_{s30} > 1500$ m/s		
B	Rock	$760 < V_{s30} \leq 1500$ m/s		
C	Very dense soil and soft rock	$360 < V_{s30} \leq 760$ m/s, or	$N_{60} > 50$, or	$S_u > 100$ kPa
D	Stiff soil	$180 < V_{s30} \leq 360$ m/s, or	$15 \leq N_{60} \leq 50$, or	$50 \leq S_u \leq 100$ kPa
E	Soil profile with soft clay	$V_{s30} < 180$ m/s	plasticity index $PI > 20$, water content $w > 40\%$, and	$S_u < 25$ kPa
F	Site-specific geotechnical investigations and dynamic site response analyses required: (i) soils vulnerable to potential failure or collapse under seismic loading (liquefiable soils, quick and highly sensitive clays, collapsible weakly cemented soils, etc.); (ii) peats and (or) highly organic clays ($H > 3$ m of peat and (or) highly organic clay, where H is thickness of soil); (iii) very high plasticity clays ($H > 8$ m with $PI > 75$); (iv) very thick "soft – medium-stiff clays" ($H > 36$ m)			

Table 2.1 – Site classification scheme to define site response categories (after NRC 2005). Site categories A-E can be defined based solely on average shear wave velocity measurements to 30m depth.

2.2 SHEAR WAVE DATABASE FOR SITE CLASSIFICATION IN OTTAWA, ON

To map the variation in potential ground motion amplification, the GSC and Carleton undertook the mapping of V_{s30} site categories (A through E) across the City of Ottawa using surface and borehole geophysical shear wave surveys. It was anticipated that the map would be useful in two ways: as a planning tool for city officials on the placement of critical infrastructure and emergency preparedness, and as an input layer for risk-based software analyses such as HAZUS-MH, a FEMA software product which combines geologic and microzonation layers with GIS information to predict hazard-related damage prior to an earthquake event. Risk assessments using seismic site class maps have been conducted in Canada in

Victoria, Vancouver and New Westminster, BC, for the insurance industry (Finn et al. 2004) and as part of pilot study in downtown Ottawa (Ploeger et al. 2008).

The V_{s30} map was created with several sources of data. The first source came from shear wave velocity measurements made at hundreds of sites across the city using reflection/refraction techniques (686 sites), borehole shear wave velocity measurements (16 boreholes), and 25km of landstreamer reflection profiling, all collected during the 2005-2008 field seasons by the GSC and Carleton field crews. Figure 2.3 shows the locations of these measurements. The shear wave data from these sites allowed us to develop average velocity-depth (V_{sav} -depth) functions in the post-glacial soils using the reflection measurements, and estimate interval velocities in the glacial and bedrock materials using the refraction data.

Figure 2.3 Location of seismic test sites within the city of Ottawa, collected between 2005 and 2008.

The second data source was a borehole database containing 20,000 water wells and engineering borings within the City of Ottawa which is part of the 'Urban Geology of the National Capital Region' Database (Bélanger 2008). The descriptions of the soil and bedrock units were broken down into three simplified geotechnical units based on their shear wave velocity and density characteristics: post-glacial materials (soft, marine clays deposited by the Champlain Sea after the most recent glaciation ~11,000 years BP), glacial materials (till and diamicton deposited during glaciation), and bedrock. Each of these units was assigned shear wave velocity characteristics based on interpretations of the shear wave data collected. Waterwell and engineering borehole locations are shown in Figure 2.4

Figure 2.4. Location of 20,000 water well and engineering boreholes used in the creation of the Ottawa seismic site category map.

2.3 V_{S30} MAP FOR OTTAWA, ON

My role in this study included analyzing approximately 200 of the early seismic sites collected by the GSC, training/supervising students in subsequent data analyses, and compiling all the shear wave reflection velocities in the post-glacial material to develop a city-wide V_{sav} -depth equation. This was determined to be

$$V_{sav} = 123.86 + 0.88z \pm 20.30 \text{ m/s} \quad (2.1)$$

where V_{sav} is equal to the average shear wave velocity at depth z with an error of one standard deviation (1σ). This equation results from a dataset of over 9000

reflection measurements from the surface and downhole seismic measurements within the post-glacial layer. In addition to the city-wide $V_{s_{av}}$ -depth function, I reanalyzed 260 of the geophysical sites where reflection data were interpreted to develop site-specific $V_{s_{av}}$ -depth functions. By establishing acceptance criteria for the quality of reflection data at a site, I found that sufficient data existed to create high-reliability $V_{s_{av}}$ -depth functions at 160 sites. I then applied geostatistical approaches to the spatial analysis of lateral shear wave velocity variation to predict how $V_{s_{av}}$ -depth equations would vary across the city in areas of thick post-glacial soils. These $V_{s_{av}}$ -depth equations, coupled with knowledge of glacial and bedrock velocities and thicknesses allowed us to compute a $V_{s_{30}}$ value at the 20,000 locations where borehole stratigraphy was available from the borehole database. These $V_{s_{30}}$ measurements were then binned into their respective site categories (A through E) and gridded using a natural neighbour interpolation technique to produce the seismic site category map. Details of the spatial analysis of shear wave velocities in the post-glacial soils can be found in Appendix A. The seismic site category map was released as Open File 6191 by the Geological Survey of Canada (Hunter et al. 2009). A more detailed report of the entire work has been released as Open File 6237 which contains the fundamental site period map and additional detail on the production of the maps, including data collection, analysis, and interpretation methodologies (Hunter et al. 2010).

The $V_{s_{30}}$ map is shown in Figure 2.5. All five of the NBCC site classes are present, indicating that in the event of an earthquake, the level of shaking across the city is expected to vary. The locations of thick soft soil (>20m) can be seen in areas

of red and orange contouring (Zones E and D, respectively) which can be found underneath urbanized areas of the city, notably Orleans. Areas of green (Zone A) or blue (Zone B) are those with bedrock near or at surface with elevated V_{S30} values greater than 760m/s. Due to the presence of buried steep-sloped valleys filled with soft glaciomarine soils, the transition from Zone A to E can happen over the span of only a few hundred meters.

Figure 2.5. 2005 NBCC seismic site class map of Ottawa based on shear wave velocity ranges of upper 30m (V_{S30}) of ground surface (Hunter et al. 2009).

The 2005 NBCC provides tables of scaling (amplification) factors for the spectral acceleration expected at a site based on seismic site class. Spectral accelerations are probability based, and are used in the 2005 NBCC to estimate what horizontal shearing a structure will experience (as a percent of gravity (%g)) during

the design earthquake with a return period of 2475 years. Spectral accelerations can be determined using the Natural Resources Canada hazard calculator (NRCan 2009) and are always given for site class C (“firm ground”) at any particular location. By applying the scaling factor for a site (generally speaking: amplification for sites D and E, and deamplification for sites A and B) engineers are able to assess the correct level of reinforcement required for a structure to withstand the design earthquake. Figure 2.6a shows the spectral acceleration (0.2 second period) with a recurrence time of 2475 years for firm ground condition (Class C) for the city of Ottawa. Figure 2.6b shows the same map when amplification factors are factored in based on the seismic site class (V_{s30}) map. The difference in spectral accelerations between Figures 2.6a and 2.6b demonstrates the influence of site classification and soil amplification factors on the seismic hazard map for city of Ottawa.

Spectral Accelerations(Sa) for 0.2sec period (2% in 50 yr probability)
Zone C - "Firm Ground"

Spectral Accelerations(Sa) for 0.2sec period (2% in 50 yr probability)
Zones A - E

Figure 2.6. Spectral accelerations pre and post scaling factors from 2005 NBCC, based on V_{s30} site categories.

The amplification factors may not take into account the complexity of local site effects in areas of complex bedrock topography, which may act to increase or decrease the level of shaking experienced at a specific location. The pattern of shaking during a strong motion earthquake undoubtedly will be more variable than is implied by the distribution of NBCC seismic site classes and the respective spectral accelerations modified for amplification. Nevertheless, the basic trend is that the areas of soft soils (D/E) will experience greater shaking than the stiffer soils or bedrock.

2.4 FUNDAMENTAL SITE PERIOD MAP FOR OTTAWA, ON

The second map type produced was the fundamental site period (FSP) map, presented in Figure 2.7. As given in the 2005 NBCC, the fundamental site period can be computed using

$$T_o = \frac{4H}{V_{sav}} \quad (2.5)$$

where H is the depth to the major impedance boundary, and V_{sav} is the average shear wave velocity down to this impedance boundary. The fundamental period predicts the period (or frequency, $f_0 = 1/T_o$) at which resonance amplification will occur.

Figure 2.7 Fundamental site period map of Ottawa based on thickness of soft soil and average shear wave velocity down to the base primary impedance layer, as recommended by the 2005 NBCC (Hunter et al. 2009).

Structures founded on soils with periods close to that of the soil may undergo more intense shaking due to a quasi-resonance set up between structure and soil. This building-soil interaction is important because the 2005 NBCC amplification factors may not be sufficiently high at the resonance period of a site (Canadian Geotechnical Society 2006). This makes attenuation in soils a particularly important property to understand, as the resonance effect in soft soils will be more long lasting if soil dissipation of trapped waves is low, as demonstrated in Figure 2.8. The following chapters will discuss the field and lab studies carried out to measure small strain damping (an important dynamic property) in the Ottawa area soft soils.

Figure 2.8. Simple schematic illustrating influence of damping ratio (ξ) on soft soil response at a site's fundamental frequency (here, 0.5Hz) and subsequent harmonics of 1Hz. (after Kramer 1996).

3.0 ATTENUATION IN SOILS

The study of dynamic properties is concerned with the behavior of soils when subjected to cyclic loading (i.e. earthquake-induced motions). Geotechnical engineers are primarily interested in the shear modulus and damping ratio of a soil to perform response analyses to cyclic loading and predict ground motions at surface. In addition to the traditional lab approaches used to determine these parameters, shear wave velocity and attenuation are properties which can also be measured *in situ* using geophysical techniques. In this study, new and traditional geophysical field and lab techniques are applied to investigate the dynamic properties of the Champlain Sea muds at varying depths and low strains (10^{-6} – $10^{-5}\%$).

A seismic shear wave will lose energy in various ways as it propagates through the earth, resulting in changes to the shape of the waveform over time. Amplitude loss in a signal is estimated by

$$A = \frac{A_0}{R} e^{-\alpha R} \quad (3.1)$$

where A is the signal amplitude at a distance, R , from the source, A_0 is the amplitude at the source, and α is the attenuation coefficient. The two basic forms of energy loss which contribute to 3.1 are:

- A geometric spreading term, $1/R$, and

- an anelastic (intrinsic) energy loss term ($e^{-\alpha R}$) attributed to conversion to heat through particle friction. (This term has also been called attenuation, absorption (Telford et al. 1991), dissipation (Spencer et al. 1982), and damping).

As the wave radiates outward from the source, geometric spreading accounts for the decrease in energy flux in the earth due to the enlargement of the wave front (Telford et al. 1991). As the waves move away from the source, body wave amplitude drops off at $1/R$, and energy content decreases at $1/R^2$.

Damping, or attenuation, describes the complex process of anelastic energy loss of a wave as it travels from a source. The attenuation factor is represented by

$$\alpha = \frac{\pi f}{Qv} \quad (3.2)$$

where f is the frequency and v is the velocity of the wave, and Q is the seismic quality factor. (Another term used is the 'seismic attenuation factor', represented by $1/Q$). The type of Q measured should be further specified by the type of wave; compressional waves (p-waves) provide a Q_p and shear waves (s-waves) provide a Q_s . In seismology, Q_p and Q_s are of interest, whereas in geotechnical engineering, Q_s is the quantity of primary interest because most damage which occurs during earthquakes is due to shearing forces. In this study, Q_s is the term referred to, and will be referred to simply as Q unless otherwise specified.

In a stress-strain model, Q is most often defined as the maximum energy (E) stored during a cycle divided by the energy lost during the cycle (ΔE) (Kramer 1996), defined as

$$Q = \frac{2\pi E}{\Delta E} \quad (3.3)$$

Processes such as phase distortion (or velocity dispersion) and scattering also affect anelastic energy loss (attenuation) as the higher frequency content of a wave is dissipated as it moves away from the source. Scattering decreases the amplitude of seismic phase by shifting the energy in the direct arrival into later scattered arrivals caused by the wave front's interaction with small scale heterogeneities (Lay and Wallace 1995). These processes may provide some level of information on the Q value(s) in a medium. Wuenshel (1965) wrote that the attenuation coefficient of rock had been more intensively studied than dispersion. In recent decades, studies have indicated that dispersion can be important in attenuation studies (Sun et al. 2009), but as it is difficult to measure in the seismic frequency band of 5-200Hz, it has often been neglected in Q estimation. Measurements of both phase distortion and amplitude attenuation of a seismic wave provide complementary information about rock properties (Quan 1997). Further discussion on the relationship between dispersion, attenuation, and Q can be found in Chapter 6, Section 3.

Given the complex effect of soil structure on wave attenuation, Q is usually separated into two components. A frequency-independent component, called Q_{in} , or intrinsic, accounts for the friction between grain boundaries, and a frequency-dependent term, Q_{sc} , takes scattering into account (Campbell 2009). The effective, or total Q , is termed Q_{ef} where

$$Q_{ef}^{-1} = Q_{in}^{-1} + Q_{sc}^{-1} \quad (3.4)$$

Debate continues on whether Q is frequency-dependant or independent within certain frequency ranges and materials, but there seems to be general consensus that Q in fine grained soils (clays/silts) is predominantly independent of frequency in some bands (0.1-10Hz) (Shibuya et al. 1995), and more strongly dependant across others (10-100Hz) (Rix and Meng 2005; Shibuya et al. 1995). This does suggest an overall decrease in Q_{ef} (or increase in $1/Q_{ef}$) due to the effect of dispersion and scattering at increasingly higher frequencies. Q_{sc} may be low in low frequencies ranges where wave lengths are too long to be attenuated by small scale structure, but as wavelengths decrease, Q_{sc} becomes more of a factor. This degree of frequency dependence from the Q_{sc} term will ultimately be controlled by the unique structural characteristics of the soil type (Campbell 2009; Dainty 1981).

The quality factor is of interest to researchers for many reasons and has been studied in depth since the early 1960's. Reverse Q filters are of use in the study of earthquake source characteristics recorded by seismograph stations at various

distances from the epicentre. In the near surface, Q estimates can be used to understand amplitude attenuation in shallow seismic records, where preferential attenuation of high frequencies lengthens the dominant signal wavelength (and period) and causes a reduction in resolution (Dasgupta and Clark 1998). Estimation of near-surface Q has also been a factor in hydrocarbon or groundwater exploration as the S -wave Q to P -wave Q ratio, Q_S/Q_P , has been found to be a better indication of fluid saturation than the velocity ratio V_P/V_S (Jeng et al. 1999).

However, the effect of Q_S in soft soils ($V_S < 200$ m/s) near surface has received less attention, and has important implications for the field of seismic hazard studies. Since most economic structures are built on unconsolidated sediments, the determination of Q is an important physical property in the prevention of potential site hazards from earthquake seismic loading. Anderson et al. (1996) suggested that the varying influence of Q within the near surface (i.e. < 100 m) may have a greater impact on ground motion prediction than the current building code's reliance on 30m depth (V_{S30}).

Numerous methods have been developed to estimate seismic Q , many of which have been customized for in situ testing using downhole vertical seismic profiling (VSP) techniques in near surface soils and rock. In the time domain, pulse broadening, rise time, and amplitude decay techniques (Tonn 1991) have been developed using the shape of the first arrival pulse at different depths to estimate the attenuation of the signal. The more broadly accepted approaches use frequency domain analysis techniques, including spectral ratios developed by Redpath (1982)

and numerous variations thereof (Stewart 1992), frequency downshift (Quan and Harris 1997), and others. Seismic data sets used in spectral ratio analyses have included surface seismic reflections, downhole vertical seismic profiles (VSP), seismic cone penetrometers (SCPT), and spectral analysis of surface waves (SASW). Earthquake data sets have also been used to compare the spectral content on soil to that of nearby rock stations.

To measure Q for this study, a variation of the spectral ratio technique is proposed with a monofrequency source, which proved to be advantageous in high- Q environments. For comparison, two datasets are also analyzed using the time domain pulse broadening approach. Amplitude and dispersion experiments were carried out in two deep Geological Survey of Canada (GSC) boreholes within the city of Ottawa: the Anderson Rd borehole (BH-AR93-1; 54 meters of clayey silt), and the John Shaw Rd borehole (BH-GSC-JSR-01; 96+ meters of Leda Clay). Figure 3.1 shows the location of these two boreholes over a simplified surficial deposit map of the city of Ottawa. Both sites are located in deep pockets of Champlain Sea muds; BH-AR93-1 in the Mer Bleue region, and BH-GSC-JSR-01 in a deep basin along the Carp Fault.

Figure 3.1. Simplified surficial geology map showing locations of deep boreholes. The bold outline shows the extent of the City of Ottawa. Post-glacial materials are primarily the soft Champlain Sea muds.

Geotechnical testing was also performed on three preserved samples of clayey silt taken from the JSR borehole at the time of drilling in 2008. Resonant and non-resonant column tests were performed at the University of Waterloo Civil Engineering lab over two visits in November 2009 and January 2010. The resonant approach measures the shear velocity and damping ratio of the specimen at the sample's resonant frequency. The non-resonant approach uses the same resonant column apparatus, but evaluates the frequency-dependence of damping and shear wave velocity. The results of these tests are compared with the results of the downhole shear wave experiments.

The goals of my study are to answer the following questions with regard to small strain soft soil response in the Champlain Sea deposits:

1. whether *in situ* Q is constant or varies with frequency and soil properties,
2. whether a monofrequency *in situ* approach reduces uncertainty in Q estimation versus traditional broadband techniques,
3. whether velocity dispersion is occurring *in situ* within the seismic exploration band (10-100Hz), and
4. how field measurements of Q relate to lab measurements of damping ratio.

It is intended that the results of the study will enhance our current understanding of dynamic soil properties and ground motion in the regionally extensive post-glacial soils, and contribute to future studies in prevention of seismic hazards from larger strain earthquake loading.

4.0 GEOPHYSICAL/GEOTECHNICAL PROPERTIES OF CHAMPLAIN SEA DEPOSITS NEAR OTTAWA

Muds deposited by the Champlain Sea, sometimes called 'Leda Clays', have long been recognized as materials which present challenges for engineering projects. The term 'clay' is a misnomer: the soil is predominantly composed of silt particles and can be highly sensitive under certain conditions. The Leda Clay is unique in Canada, but has been compared geotechnically to the Norwegian marine clays which exhibit many of the same sensitivity characteristics. The behavior of the muds is particularly important to earthworks projects as disturbance or loading of the soil can potentially cause collapse, leading to concerns for the stability of embankments and slopes. In an undisturbed state, the soil can withstand substantial loads, but when leached of its saline porewater which provides inter-particle strength, the remoulded soil can flow like a liquid (Law and Bozozuk 1988). For earthquake geotechnical engineers, this causes concern for the stability of the soil under seismic cyclic loading conditions. In this chapter, general geotechnical and geophysical properties of the Champlain Sea deposits are summarized, and the results of sample tests carried out by the GSC in 2008 at BH-GSC-JSR-01 are presented.

Champlain Sea deposits are regionally extensive, covering areas of the St. Lawrence Lowlands between Pembroke, ON and Quebec City, QC. The muds (a geologic term indicating the mixing of silt- and clay-sized particles) were deposited in the Champlain Sea, a large inland waterbody formed after the last glacial retreat approximately 11,500 to 10,000 – 14C years before present (Anderson et al. 1985).

An invasion of the Atlantic Ocean, followed by isostatic rebound, trapped the waters causing deposition to occur in a saline environment. Mixing of the sea water with fresh-water glacial inflow caused varying levels of salinity during deposition. The saline porewater provides chemical bonding strength to the soil, which is composed of silt and clay sized particles ('rock flour') eroded during the advance of the glaciers over the Canadian Shield. Gadd (1968, 1986) initially identified three broad units in the Champlain Sea deposits - basal rhythmites or varves, a thick interval of massive muds (the 'Leda Clays'), and red and gray rhythmically laminated muds at the top of the sequence. Fluvial outwash was later deposited in certain areas of the St. Lawrence Lowlands, composed of larger silt, sand, and gravel particles. When the muds are capped by coarser fluvial materials, surface water can be trapped above them causing 'leaching' or replacement of the saline pore waters with fresher water, leading to potential instability (Hunter, pers comm. 2009).

The mineralogy of the soil is complex and varies from region to region. Rock-forming minerals such as feldspars, quartz and amphibole constitute a large portion of the clay and silt size particles. Others report illite and chlorite minerals as common, with moderate amounts of vermiculite and smectite (Law 1988; Berry and Torrence 1998). The elevated water content (and porosity) plays a large role in the behavior of the soil, both in its short term response to remolding, and its longer term response to settlement. The liquidity index (LI) of the clay is greater than 1.0, which causes the 'quick' (sensitivity ≥ 100) condition of the clay when remolded. The plasticity index (PI) ranges between 15 to 40% from location to location (Law and Bozozuk 1988).

4.1 JOHN SHAW RD BOREHOLE: KINBURN, ON (BH-GSC-JSR-01)

In 2008, the GSC drilled a deep 96m borehole (BH-GSC-JSR-01) near Kinburn, Ontario to study the depositional history of the Champlain Sea in the region, and the dynamic and static properties of the soil. A 4.5km-long high-resolution seismic profile collected by the GSC was key in selecting the location of the boring. The profile, which crossed the Carp fault, identified a thick package of Champlain Sea deposits (up to 100+m) lying above a sand and gravel deposit of glacio-fluvial origin within the fault basin. The borehole stopped short of intersecting the gravels, which were known locally to have extremely high artesian pressures.

Soil moisture content and fall cone tests were carried out by the GSC on 64 samples within 48 hours from the time of collection at the drill rig. Grain size analyses were later completed on dried samples following the moisture content testing. Water content tests showed that the range varied from 40-85%. Drop cone tests were carried out in the GSC soil lab using freshly cut surfaces from the samples. The method follows the procedures developed for undrained cohesive soils by the Swedish Geotechnical Institute, and uses the penetration distance (in mm) of various weights into an undisturbed, and then remolded sample to assign shear strength values (C_u , C_r) to the soil. The ratio of C_u to C_r provides a sensitivity value for the soil.

These results, alongside the downhole geophysical logs and the core log, are shown on Figure 4.1. Natural gamma, apparent conductivity and magnetic susceptibility downhole tools were run to detect stratigraphic changes in the soils

and variation in bulk properties. Downhole seismic shear-wave surveys provided accurate travel time and shear wave velocities at 0.5m intervals downhole. Shear velocity is also a method of obtaining the small strain shear modulus of the soil, a measure of the soil's maximum stiffness. Combining this suite of geotechnical and geophysical data for analysis of the geotechnical properties of the Champlain muds is critical. The geophysical logs identify locations where chemical changes in the soil may not be evident to the eye during core logging, but which may affect the strength of the material, or indicate a change in the depositional environment/provenance the soil.

Figure 4.1. Geophysical, geotechnical, and geological logs from BH-GSC-JSR-01, shown next to monofrequency test intervals. ¹ Undrained shear strength, ² Remoulded shear strength (both derived from Swedish Drop Cone Tests), ³ Sensitivity ratio of Cu:Cr, ⁴ Moisture content collected by the GSC in 2008, ⁵ Geological log completed by the GSC, 2010 (Medioli et al. 2010).

The natural gamma log, in absence of elevated levels of naturally occurring potassium, uranium, and thorium, generally follows a trend of increasing counts in finer soils and decreasing counts in coarser soils. In this borehole, an upward decreasing count per second (cps) level indicates an upward coarsening sequence (i.e. clay to silt or sand sized particles), and an upward increase in counts indicates a fining upward sequence (i.e. silt to clay-sized particles). These trends are confirmed by the grain size results. The magnetic susceptibility log can also be used as a relative grain size predictor. Sands and coarser grained materials are more likely to contain magnetic materials, whereas soft soils (silts and clays) exhibit very low susceptibility readings as the heavy magnetic minerals have already dropped out of suspension prior to deposition. The increased susceptibility readings in this borehole appear in areas where sand content is relatively elevated. The apparent conductivity log measures the bulk conductivity of the soil. In the presence of saline porewater, the conductivity of the fluid overwhelms the inherent conductivity of the soil matrix, and becomes a relative indicator of increasing or decreasing salinity – an important connection to the mud's sensitivity.

At the base of the borehole from 96m-90.6m, a raised sand content level indicated by the grain size analysis is reflected in the elevated magnetic susceptibility log. A drop in natural gamma counts also reflects the presence of coarser materials in the lower two units. A sensitivity value over 100 indicates the soil at this depth is considered 'quick' and may flow if disturbed. Between 86.6 – 90.8m, sensitivity levels are greater than 20, indicating the soil in this interval is 'sensitive'. Within this sensitive to quick zone, core recovery was limited. Note that

up to approximately 62m, the sensitivity level hovers around the 'sensitive' level (20), which corresponds with the gradual decreasing apparent conductivity levels. Whether the saline porewater has been 'leached' from the soil through a process of molecular diffusion or was deposited in an increasingly saline environment is unknown at this time.

Relatively low magnetic susceptibility levels, coupled with elevated gamma counts and conductivity levels suggest a rather massive unit of clayey silt above the basal sandier zone from approximately 80m up to 20m depth: the deep water 'Leda Clays'. Grainsize results indicate gently decreasing then increasing levels of clay within this package. Velocity decreases and porosity increases uniformly upward with decreasing overburden pressures. This zone of relatively massive soils presented an excellent location to carry out attenuation measurements, as data could be recorded with minimized frequency-dependant energy losses in the 10-100Hz frequency band due to few textural changes in the soil column. In fact, the greatest visible grainsize variation seen in the soil column was in the base of the borehole (>90m) where interbedded sand-silt layers thicker than 10cm were not observed (Medioli, B. pers. comm.).

Four intervals were tested in this borehole: 20 to 30m, 30m to 50m, 30m to 70m, and 30m to 90m were tested with the downhole shear wave tools and the monofrequency source, as shown in Figures 4.1 and 4.2. In these intervals, the effects of porosity, grain size, and *in situ* shear strength on anelastic attenuation

could be evaluated with relatively few other varying parameters (i.e. significant changes in lithology types or stratigraphic layers).

As shown on the seismic profile in Figure 4.2, the soil above 30m forms a distinct reflective boundary which can be followed along the entire 4.5km seismic profile collected on John Shaw Rd. The soil fluctuates with rising and falling levels of silt and clay, but contains <1% sand between 5-30m depth. A series of attenuation measurements were preformed between 20m and 30m to investigate any frequency dependence of attenuation in a region of varying soil properties, as identified by the seismic section.

Surface Seismic Reflection Profile over Borehole Location

Figure 4.2. 400m segment of the high-resolution seismic 'landstreamer' profile collected adjacent to BH-GSC-JSR-01. Selected test intervals for the attenuation measurements are shown. Test interval 1 was chosen to bracket a zone where downhole logs and sample tests indicate numerous variations in soil properties. Test intervals 2, 3 and 4 were chosen to bracket zones of relatively massive Champlain Sea muds.

Within the 20m-30m test interval (JSR-1), moisture content increases upward from 45% to 60%, corresponding to a porosity of 1.2 – 1.6. Shear strength decreases upward evenly from 60 to 30kPa, and grain size remains stable (~70% silt, 30% clay). Within the 30m - 50m test interval (JSR-2), silt content dominates (62-70%), with clay particles (28-36%), and sand (1-2%) making up the rest. Porosities were stable, ranging from only 0.55 to 0.58, and shear strengths and shear wave velocities decrease gradually upward with increased vertical overburden stresses. This range represents a texturally massive interval. Interval JSR-3 (30m – 70m) encompasses varying levels of silt and clay; while silt dominates, the clay content rises upward to 40%. Porosity increases slightly in the clay zones (58-62m) and shear wave velocity and strength undergo a slight decrease/increase cycle. Variations in the magnetic susceptibility readings indicate an increase in heavier minerals and very thin layers of sand may be present. Within Interval JSR 4 (30m – 90m), the soil characteristics change enough below 70m to introduce structural effects visible to the eye. Sand layers (mm-to-cm scale) are visible below 85m, and porosities drop, indicating a change in material depositional environment.

The choice of test intervals based on the material properties from the downhole logging and geotechnical testing permitted an evaluation of damping versus certain key criteria (but not all) which have been determined to affect damping in the lab (Hardin and Black 1968): soil type and grain size, confining pressures (effective vertical stress, σ'_v), porosity (or void ratio), saturation, and strain amplitude (which will be $\sim 10^{-6}$ % for shear wave testing). The use of wide

testing intervals was intended to investigate bulk damping properties of the soil mass.

4.2 ANDERSON RD. BOREHOLE: OTTAWA, ON (BH-AR93-1)

Attenuation measurements were also carried out at a second deep borehole with 53m of Champlain Sea muds and 5m of till overlying bedrock. This site on Anderson Rd. in Ottawa serves as a calibration hole for the GSC and has been logged with the GSC's natural gamma, conductivity, and magnetic susceptibility probes. Geotechnical and geological data are derived from Woodley (1996). The results are presented on Figure 4.3, along with the geological log and selected downhole test intervals.

The geophysical/geotechnical log responses at Anderson Rd are, in general, similar to those at John Shaw Rd. Elevated moisture content (>90%) in the top 8m drops to an average 60% in soil below. Undisturbed shear strength ranges between 30-80kPa decreasing upward with decreasing overburden stresses. Similarly, shear wave velocity decreases smoothly upward. Sensitivities increase upward, correlating well with decreasing conductivity values. Very little variation in the natural gamma and magnetic susceptibility logs between 10m and 45m indicate soils are fairly massive in texture. Clay content varies between 20-40%.

The two intervals chosen for downhole testing were between 10m and 30m (AND-1) and 10m and 45m (AND-2), due to their uniform soil properties.

Figure 4.3. Geophysical, geotechnical, and geological logs from BH-AR93-1, shown next to test intervals for monofrequency test intervals. ¹ Undrained shear strength, ² Remoulded shear strength, ³ Sensitivity, and ⁴ Moisture content, ⁵ Clay content (Woodley, 1996) ⁶ Geological log (Woodley, 1996, after Gadd, 1986)

5.0 GEOTECHNICAL TESTING OF DYNAMIC PROPERTIES

Characterizing dynamic properties of soft soils is an important aspect of site response assessment during earthquake loading. As such, the field of dynamic soil testing using geotechnical lab techniques to evaluate shear modulus (G) and material damping (ξ) is vast, and a comprehensive treatment of these topics is not intended to be covered in this thesis. However, it is of interest to generally review the methods used to gather G and ξ in the lab and how they relate to the measurement of Q . As part of this study, I took three soil samples from BH-GSC-JSR-01 to the University of Waterloo Civil Engineering lab to conduct resonant column (RC) and the newly proposed non-resonant column (NR) tests on the samples for comparison with the in situ attenuation results. These methods are discussed in this chapter.

Geotechnical engineers use the term material damping ratio (ξ) to describe a soil's ability to dissipate energy, instead of the seismic quality factor, Q . Material damping is generally thought to be attributable to energy loss due to hysteresis, caused by particles sliding past one another (friction). In addition to the damping ratio, variations of the term are used to define this form of intrinsic attenuation: the loss factor (η), and the logarithmic decrement (δ) which provides a measure of decay over successive cycles for a medium undergoing vibration. These terms, and Q , can all be related to one another by Equation 5.1.

$$\frac{\delta}{\pi} = \frac{1}{Q} = 2\xi = \eta \quad (5.1)$$

In geotechnical terms, the shear modulus (G) is often the parameter of interest, as it represents the ratio of shear stress to shear strain. When field testing is carried out with a seismic source, the soil strains induced are very small (i.e. $<10^{-6}\%$) and the G measured is considered to be at its stiffest ($G_{\max}$). Shear modulus, G , and shear velocity, V_s , are related to one another through Equation 5.2

$$V_s = \sqrt{\frac{G_{\max}}{\rho}} \quad (5.2)$$

where ρ [kg/m^3] is the density of the specimen.

Numerous factors have been identified which influence the testing of dynamic properties in soils. Among them, inherent properties, such as grain size, mineralogy, void ratio, water content, age (and thus, stress history), and saturation are all important factors. Knowledge of these properties, as discussed in Chapter 4, plays an important role in the analyses of field and lab studies carried out in this research. External forces controlling confining pressure and strain amplitude are also extremely important and generally have the greatest degree of control over the response of a soil to loading (Hardin and Black 1968; Hardin and Drnevich 1970; Hamilton 1972).

The frequency-dependence of damping is generally not addressed in traditional lab testing methods (i.e. RC testing). However, the NR method proposes a method of obtaining frequency dependent damping (and shear wave velocity) characteristics of a soil sample. These techniques are discussed in sections 5.4 and Appendix C.

5.1 STRESS AND STRAIN

Of all the factors affecting dynamic properties, experimental research over the last few decades has shown that the most important external variable is the magnitude of applied stress, which acts as a measure of the level of shear strain induced in soil mass (Lai et al. 2001). Table 5.1 describes the expected soil behavior at different strain levels during cyclic loading (Khan 2007; Vucetic 1994)).

Level	Very Small	Small	Moderate	Large
Shear Strain, γ	$<10^{-7}$	$10^{-6} - 10^{-3}$	$10^{-3} - 1$	>1
Soil Behavior	Linear Elastic	Non-linear Viscoelastic	Non-linear Viscoplastic	Failure

Table 5.1 – Generalized soil behaviour in various strain ranges based on energy loss mechanism (after Vucetic, 1994, and Khan, 2007).

Figure 5.1a represents the hysteretic stress-strain behavior of a soil in the very small and low-end small strain ranges subjected to one cycle. The response is linear, and the soil experiences no permanent deformation. When a soil is subjected to a very small strain ($\gamma \leq 10^{-7}$), the stiffness is constant (i.e. $G=G_{\max}$) and the response of soils is linear, irrespective of the number of cycles. Even within this range, very small amounts of energy are lost during the time lag between cyclic strain and stress, which is more characteristic of viscoelastic behavior. Within the small strain range ($10^{-6} < \gamma < 10^{-3}$), pore water pressures can develop in undrained conditions, but there is very little change in soil properties. Although this range is termed non-linear, there is very little degradation of the soil with number of cycles. The non-linearity of the small strain range can be important when comparing

damping values. Campbell (2009) pointed out that damping measured at strains of 10^{-6} and 10^{-4} in the same material can produce significant differences.

As strains increase, G (and thus stiffness) rapidly decreases, leading ultimately to viscoplastic behavior in the soil which results in deformation. In cases of moderate to large cyclic stresses, such as large earthquake loading, the soil will follow the backbone curve and become weaker with successive cycles (Figure 5.1b).

Figure 5.1 – a) Loop showing the stress-strain behavior of a soil subjected to one cycle of a very small strain. The soil remains in the linear range, meaning that the response is elastic and permanent deformation does not occur. b) At large strains, the soil responds non-linearly to the strain, resulting in deformation. During cyclic loading, the shear modulus may be reduced further with each cycle. ΔE represents the area within the entire hysteresis loop (dissipated energy), and the triangular area, E , represents the peak strain energy of the cycle.

The moderate zone ($10^{-3} < \gamma < 10^{-0}$) characterizes the soil as it reaches a pre-failure state. Large deformations can occur and the energy losses are termed plastic (Vucetic 1994).

Material damping can be determined directly from a hysteresis loop as the ratio of the energy dissipated through one cycle of loading (ΔE) to the peak strain energy stored during the cycle (E) times a factor of 4π . Under moderate to relatively high strains, triaxial compression tests, simple shear tests, or torsional shear tests can be used. As the stiffness of a soil decreases, its ability to attenuate seismic waves increases. Therefore, damping and modulus reduction curves are often used together and are collectively called linear equivalent models (Figure 5.2). Hysteretic lab tests have been carried out to develop modulus reduction curves for various material types to predict the decrease of G with increasing strain (Seed and Idriss 1970; I. M. Idriss 1990).

Figure 5.2 – The relationship between modulus reduction (G/G_{max}) and damping versus strain are shown in this simplified schematic. Boundaries between strain ranges and linear/non-linear behavior are approximate, and different for all soil types. As the stiffness of a soil decreases with increasing strain, the damping in the material increases. All work in this study is carried out in the very small strain viscoelastic range.

Damping is commonly distinguished to be either hysteretic or viscous. When damping is determined from stress-strain curves, it is said to be hysteretic, where the energy lost in each cycle is independent of the loading frequency. Viscous damping is said to occur when the energy lost in each successive cycle increases proportionally with frequency. This is the assumption made when the logarithmic decrement method is used in the resonant column device (Zavoral 1990).

5.2 INHERENT SOIL PROPERTIES

Soil properties will also affect response to dynamic excitation. Studies carried out by Hamilton (1972) in deep sea marine clays off the coast of San Francisco led him to draw numerous conclusions about the influence of soil properties on attenuation in saturated fine grained soils. Although he looked at compressional (P) waves in kilohertz ranges, he noted the following relationships in viscoelastic mediums:

1. Lack of significant porewater movement in a soil (impermeable fine-grained silt and clays) relative to the solids leads to no or negligible dispersion, meaning that attenuation is not dependent on porewater properties in viscoelastic media.
2. Saturated, flocculated clay sediments (marine clays) respond to shear wave energy as linearly viscoelastic media, as given by the equation $\alpha = k f^n$ where n is an exponent of frequency, $n \approx 1$, and k is a constant. (Note: k would be equivalent to π/Q_v as defined in Eq. 3.2.)

3. Approximate frequency-independence of Q in rock can be extrapolated back to natural water-saturated viscoelastic sediments (i.e. $k=\pi/Qv=\text{constant}$).
4. Attenuation increases with decreasing grain size up to the fine sand/coarse silt grainsize, where interparticle friction is maximized due to greatest surface area for grains.
5. Beyond a grain size of 0.044mm (coarse silt to clay) relative movement between grains is controlled by complex interparticle forces (cohesion). Saturated clay and fine silt particles may not be in contact, but surrounded by an adsorbed layer of water. This lack of interparticle friction significantly reduces the attenuation of fine silts and clays. Attenuation drops off exponentially for soils with a grain size smaller than 0.044mm. The smallest k value in marine sediments occurs in fine silts and clays ($>0.002\text{mm}$). McCann and McCann (1969) found that 0.044mm is where interparticle forces become dominant.
6. Increased porosity has the effect of reducing the number of interparticle contacts. In fine grained soils, the effect of increasing porosity further reduces the attenuation.
7. Mineralogy (and thus interparticle forces) may affect attenuation in soils – calcareous oozes may attenuate P-waves more rapidly than clay minerals of the same porosity and grain size.

Although defined for P-waves, these observations show that the soil properties which are common to the Champlain Sea muds (silt and clay-sized grains, high porosity, and complete saturation) can exhibit low levels of attenuation based on

grain size, and influenced by porosity. As was seen in Chapter 4, the core sampled in BH-GSC-JSR-01 is composed approximately ~99% of grains smaller than 0.044mm with porosities almost always greater than 50%. This elevated level of porosity may act to decrease interparticle friction, thereby contributing to lowered damping levels.

5.3 RESONANT COLUMN (RC) TESTING

The basic principle of the RC device is to vibrate a cylindrical sample in first mode torsional resonance and compute the shear wave velocity and damping ratio separately by solving the equation of motion for a column mass system (ASTM D4015-92 2000). Once the resonant frequency is determined, the sample dimensions and the conditions of end restraint provide the necessary boundary information to solve the wave equation and compute the velocity. Dynamic soil properties are therefore evaluated at the resonant frequency of the sample. The solution for the RC test is calculated using a visco-elastic Kelvin-Voigt model for a material with weak energy dissipation (low damping), which is why low strains are used for RC testing. Measurements of material damping and shear modulus with the RC apparatus are generally limited to strains of less than $10^{-3}\%$, at testing frequencies larger than 30Hz. Detailed derivation of the analytical solution for the RC method can be found in numerous texts, including the ASTM standard.

REVIEW OF PREVIOUS WORK

Summarized results from previous RC tests in soft fine grained soils can be found in Table 5.2. Of particular interest were dynamic soil tests performed on Champlain Sea mud samples during a National Research Council study in 1984-5 (Law et al. 1985). Resonant column and triaxial shear testing were carried out on 12 cores sampled with an Osterburg piston sampler within 13m of ground surface. The authors concluded that the samples were undisturbed based on good agreement between shear wave velocities measured in the lab and shear wave velocities measured during crosshole testing in situ. Samples were tested at strains of 6×10^{-3} % to 2×10^{-1} % confined at *in situ* pressures, and damping values range from 1.75% at the lower strains to 5.25% at the higher strains. Shear moduli ranged from 32.9MPa to 44.6MPa (approximately 140 – 160 m/s). Water contents and shear strengths reported are similar to the results obtained in the JSR borehole in the top 13m, including the high water contents, low shear wave velocities, and elevated porosities near surface.

RC testing has also been carried out in Holocene silts of similar composition to the Champlain Sea deposits in the Fraser River Delta near Richmond, BC. Results show very low levels of damping were measured at strains ranging from 1×10^{-4} – 10^{-3} %. A best fit line through 3 RC tests showed that damping ranged from 0.75 – 1.0% (Stewart 1992).

Location	Soil Type	Testing Method	Strain Level (%)	Sample Confinement	Velocity (m/s)	Shear Modulus (Mpa)	Damping Ratio (%)	Source
Ottawa, ON	Leda Clay	RC	6×10^{-3} - 2×10^{-1}	150 KPa	140 - 160	33-44	1.75 - 5.25	Law et al. 1985
Richmond, BC	Silt, Holocene	RC	10^{-4} - 10^{-3}		179-202		0.75 - 1	Stewart, 1992
Richmond, BC	Clay, Holocene	RC	10^{-3}				0.9-2.4	Zavoral, 1990
Wallaceburg, ON	Wallaceburg Clay, Soft grey silty clay	RC	2×10^{-3} - 3×10^{-1}	404 KPa			1.2 - ~17	Kim and Novak, 1981

Table 5.2. Test results from RC tests in Leda Clay or materials of similar composition at strains ranging between 10^{-4} to $10^{-1}\%$. Of note is the low damping measured at strains = 1×10^{-3} ($<1\%$).

TEST PROCEDURES FOR THIS STUDY

The RC equipment I used at the University of Waterloo Geomaterials testing lab was the fixed-free type, with the bottom of the specimen fixed and torsional excitation applied to the top of the sample (Figure 5.3a). Damping is computed either by the half power method ($D=(f_2-f_1)/2f_r$) surrounding the resonant frequency curve (shown in figure 5.3b), or the logarithmic decrement method (δ) (Stokoe, et al. 1999). The shear wave velocity is computed as shown in Equation 5.3 (Cascante et al. 2003):

$$V_s = \sqrt{\frac{\left[\left(\frac{I}{3} + I_o\right)H\right]}{J_p \rho}} \omega_o^2 \quad (5.3)$$

where ω_o is the resonant frequency of the sample, J_p is the area polar moment of inertia and I is the mass polar moment of inertia of the sample, I_o is the mass polar moment of inertia of the driving system, H is the height and ρ is the density of the sample. The shear modulus is then derived from the V_s using equation 5.2.

Figure 5.3. a) Schematic of a typical fixed-free RC testing apparatus. b) Result of an RC test from the 70m specimen. The resonant frequency of the sample is 63.88 Hz. Analysis of the shape of the curve around the resonant frequency using the half power method will produce a damping and velocity value for the test.

The three samples I brought to Waterloo were collected from depths of 10m, 30m, and 70m in the BH-GSC-JSR-01 borehole. The samples had been sealed, but were stored at ambient temperature for a year prior to testing. Before the RC testing, moisture content of these samples was redetermined and compared to test results from 2008 to assure they had not been affected by loss of water content. The samples were also retested for undisturbed and remoulded shear strength. Results of the retests are shown in table 5.3.

	Depth (m)	Undisturbed Shear Strength (kPa)	Remoulded Shear Strength (kPa)	Sensitivity Cu/Cr	Moisture Content (w)
Original, Above	8.84	4.41	0.87	5.1	65.4%
Retest	9.14	23.54	1.57	15.0	81.0%
Original, Below	9.98	23.54	0.83	28.2	82.2%
Original, Above	28.37	57.88	6.38	9.1	45.6%
Retest	29.41	37.28	3.43	10.9	47.0%
Original, Below	30.05	58.86	5.89	10.0	46.3%
Original, Above	69.80	320.30	15.77	20.3	45.7%
Retest	70.1	86.33	9.02	9.6	41.0%
Original, Below	71.14	435.56	21.19	20.6	40.4%

Table 5.3 – Results of the original (grey) and retested (black) shear strength and moisture content values in the specimens brought to the Waterloo testing lab. Specimens could not be collected for the RC/NR tests from the exact depths of the original samples, so the original test results above and below the specimen depths are shown. Although depressurization has decreased the shear strength, this can be compensated for by confinement in the RC device. Moisture contents appeared nearly unchanged.

Shear strengths in the 30m and 70m samples decreased significantly from the original testing, which was completed within 48 hours of the core recovery. Both samples had experienced significant depressurization during the year since collection, which predictably affected the shear strength. This could, in part, be corrected in the lab by repressurizing the samples up to in situ vertical confinements. Moisture content was unaffected in all samples, which was important for the lab testing, as moisture conditions could not be recreated in the lab.

Figure 5.4 shows the primary components of the RC/NR system, recently developed by Giovanni Cascante and Zahid Khan at the University of Waterloo. This system is unique, and electronically enhanced from commercial RC systems to accommodate the non-resonant (NR) testing which will be described in the

following section. Prior to preparing the sample, an LVDT (linear varying displacement transducer) was calibrated, which monitored the vertical consolidation of the sample under confinement. The RC system was also calibrated with aluminum rods to measure the mass polar moment of the driving system, I_o .

Figure 5.4. Several components of the RC/NR testing system at the University of Waterloo. Pictured: MP (mounting plate), SR (support ring for drive plate), BP (base plate), VC (voltage converter), PS (DC voltage power supply for LVDT), OS (oscilloscope), SA (signal analyzer), DM (digital multimeter).

A confining pressure was calculated for each sample to recreate the mean normal effective overburden pressures (σ'_v) in situ. Table 5.4 shows the water contents ($w\%$) and void ratios (e) calculated for each specimen during the sample retest. With these values, bulk (γ_B) and dry (γ_D) densities were computed for the samples. Known parameters include: water table $<10\text{m}$ ($\therefore$ saturation (S) = 100%), density of water ($\gamma_w = 9.81\text{kN/m}^3$), and specific gravity ($G_s = 2.65$). Downhole shear wave measurements are shown in column V_s^* .

	Depth (z)	w	e	γ_B	γ_D	σ'_v	Vs*
	m	%		kN/m ³	kN/m ³	KPa	m/s
Sample 1	9.14	0.81	2.15	14.95	5.14	47.02	90.9
Sample 2	29.41	0.47	1.25	17.02	7.21	193.13	204.1
Sample 3	70.10	0.41	1.09	17.57	7.76	508.80	285.7

Table 5.4. Calculated confining pressures for the RC/NR testing used to recreate the effective normal overburden pressures for each sample. $\gamma_B = [((1+w)/(1+e))\gamma_w G_s]$; $\gamma_D = \gamma_B - \gamma_w$; $\sigma'_v = Z\gamma_D$. Downhole shear wave measurements collected at the sample depth are shown in column Vs* for comparison with the calculated Vs during the testing.

Each test required a 70mm diameter by 140mm long cylindrical sample. The diameter of the core was approximately 73mm, so the samples were prepared by gently removing the outer 2-3mm of soil. Five to six filter paper strips were used around the long axis of the sample to facilitate wicking of water from the sample and to decrease consolidation time. Cuttings of the sample were collected, weighed, and baked to calculate a pre-test water content. Pre and post-test weights of the sample were also made to assess changes in the sample. Each sample was prepared only minutes before loading into the column and beginning the consolidation.

After preparation, a geomembrane was placed over the sample, which was then affixed on the base pedestal. The top cap was attached and suction was immediately applied to keep the sample vertical. Porous stones were placed between the sample and the top cap and bottom pedestal. In theory, the longitudinal axis of the sample must be kept perpendicular to the base plate, and the top platen must be perfectly horizontal when the driving plate is attached. However, the softness of the clay made it challenging to keep the sample completely vertical. In practice, sample deviations from vertical can take place, which excites

unwanted flexural modes. This can be assessed during the testing phase after the consolidation has taken place to ensure the sample has remained vertical.

After the driving plate was installed and the confining chamber locked in place, consolidation was initiated by opening the drainage valves and gently increasing the pressure on the pressure control panel, and turning off the suction. During the consolidation phase which should last at least until the sample consolidation stabilizes (or 1000minutes), the LVDT was carefully monitored. It was expected that the sample consolidation would be fairly significant (especially for the 30 and 70m samples) and possibly affect the vertical position of the sample, so flexural and torsional RC tests were occasionally run during the consolidation phase to assess the shape of the resonant peaks. The burette levels on the pressure control panel were also recorded to note any fluid variation in the top and bottom fluid lines connected to the top cap and bottom pedestal of the samples; this information could be used to monitor any changes in the volume of the sample.

When confinement was reached, low-strain torsional shear was applied to the sample. A top-down view of the drive plate sitting above the support ring can be seen in Figure 5.5. A Hewlett Packard HP35670A signal generator was used to input a swept-frequency signal to the drive coils which caused the magnets (which rest inside the coils but are attached to the drive plate) to move the drive plate. The strain applied to the sample can be controlled by the current supplied to the coils, here $\leq 10^{-5}\%$. The movement of the driving plate attached to the sample creates a torsional shearing motion in the sample. The response of the sample to the motion

is recorded by the two accelerometers (PCB 353B65, PCB Piezotronics Inc., NY) fixed on either side of the drive plate.

Figure 5.5 – A top-down view of the drive plate prepared for sample testing. The support ring (SR) rests on the mounting plate, and the driving plate (DP) is connected to the sample below. A user selected current is run through the drive coils (DC) causing movement in the magnets which are connected to the drive plate (DP). This causes torsional rotation in the sample. The sample response to the motion is recorded by the two accelerometers (AM) attached to the drive plate. The LVDT is used to record the consolidation of the sample from the beginning of consolidation to the end of the testing.

The material damping ratio can then be computed by curve fitting the output signal from the accelerometers in the frequency domain with either the corresponding theoretical equations (Cascante et al. 2003), or the half power bandwidth method (see figure 5.3b). The shear wave velocity of the sample can then be computed using Equation 5.3 by knowing the resonant frequency. In the lab, the damping ratio and the shear wave velocity were computed by the signal

analyzer using the transfer function in the frequency domain between the input and the output signals.

RESULTS - RC

Results of the three tests carried out can be found in Appendix B in graphical form, but a summary of the damping ratios and shear wave velocity measurements at maximum consolidation are shown in Table 5.5.

Sample	Actual Confinement	Strain, %	Resonant Frequency	Damping	Velocity
10m	60 KPa	3.11E-05	25.0 Hz	1.36%	47.5 m/s
30m	236 KPa	4.85E-06	39.9 Hz	1.84%	135.2 m/s
70m	413 KPa	9.37E-06	63.88 Hz	1.23%	200.6 m/s

Table 5.5. Summary of the damping ratios and velocity measurements from the RC tests.

Strains for all the tests remained in the small strain range. Consolidation was not carried out for the entire 1000 minutes if the LVDT readings indicated consolidation was leveling off, and particularly that the resonant frequencies and damping ratios were not varying from reading to reading. Confinement did not reach the desired 500KPa for the 70m sample. Pressures were leveled off at 413KPa to take a series of resonance readings, and when increased confinement was applied, the sample began to lean, introducing flexural modes. Consequently, manual adjustment of the sample would require depressurizing the chamber which would disturb the sample. It was decided to continue the test at the 413KPa level and not compromise the sample further. Therefore, the damping would likely have

continued to decrease and the velocity continued to increase at the true confining pressure.

The shear wave velocities calculated from the RC tests indicate unexpectedly low values. The velocity discrepancy is likely explained by the state of the samples. The drilling program had not originally been designed for undisturbed soil sampling, and the method used to collect the soil was with a wireline core barrel and not a thin walled Shelby tube. In addition, the core had been stored in ambient temperatures for a year, during which time it had depressurized. Oxydization of the outer centimeter of the samples could be observed by the extent of the hydrotroilite (dark iron sulphide often found in deep marine clays) disappearance on a fresh face of the core.

However, examination of the total consolidation of the samples during the testing using the LVDT data did not suggest the samples were highly disturbed – all samples consolidated less than 0.8% over the testing time. (The sample at 70m consolidated 1.07mm over 1485min, @ 30m, 0.26mm over 1149 minutes, and @ 10m, 1.17mm over 744 min). However, disturbance or stress in a soil sample mimics the effect of elevated strains during testing, which would result in increased damping and reduced shear wave velocity. In addition, only mean normal stresses were considered in the calculation of confinement levels for the tests, but in situ stresses would include a component of horizontal stress. Under-confinement of the samples in the lab versus their true state in the field would further reduce the velocity measured in the lab. Kim and Novak (1981) conclude in their report on RC

testing of seven clays around southern Ontario that during low strain testing performed at constant confining pressures, shear velocity (and thus modulus) will continue to increase with time beyond 1000 minutes, and that response versus time depends on the material. Therefore, it is likely that if the samples were allowed to consolidate over a longer period of time, the velocity would have approached that of the in situ measurements, and the damping would have decreased.

The results indicate that damping in these soils is relatively low, and if sample disturbance has played a role, in situ damping is quite likely even lower than measured. As shown in Table 5.2, RC tests at low strains in the Fraser Delta silts have yielded very low damping ratios ($\sim 0.75\%$), and RC tests in Champlain Sea deposits at higher strains (10^{-3}) have shown values of $\sim 1.75\%$. It is therefore reasonable to expect that damping is likely $\leq 1\%$ based on these values and the damping curves for clay given by Idriss (1990) at low strains, but this will need to be confirmed with more sampling at each depth.

Too few samples were tested to draw strong correlations between grainsize, confining pressure, porosity, and damping ratio. However, in general, we would expect to see a decrease in damping with an increase in confining pressure (i.e. with depth), and an increase in damping with an increase in porosity. Save for the measurement at 30m, we do see a general decrease in damping between 10m and 70m, and 30m and 70m. The increased damping value at 30m suggests more sample disturbance, and greater reliability in the 10m and 70m samples.

5.4 NON-RESONANT (NR) COLUMN TESTING

The dependence of damping (or Q) on frequency in low frequency ranges (0.1-20Hz) is of interest in earthquake engineering for site assessment and prediction of earthquake-derived ground motions. However, frequency dependant properties of soils are difficult to measure with the RC device. Although a multi-frequency sweep is used in the RC test, the excitation voltage is constant, which results in varying shear strain levels imposed on the sample for the different frequency components of the signal.

Non-resonance (NR) methods have been recently developed to evaluate frequency-dependant soil properties in the resonant column device at different strain levels (Lai, et al. 2001, Khan, et al. 2008). The method is based on the solution to the equation of motion for the forced vibration of a cylindrical soil specimen which is assumed to be homogenous, continuous, and viscoelastic. An important aspect of the method is the use of a viscoelastic medium. In order to fulfill the principle of causality, a wave travelling in a viscoelastic medium at low strains must have a phase velocity and attenuation which satisfy the Kramers-Krönig equation (further discussed in Chapter 6). This states that the real and imaginary parts of a body wave's complex wave number are Hilbert transforms of one another, and are therefore not independent quantities (Lai et al. 2001). This principle forms the basis of the non-resonant method, which allows for the calculation of the shear wave velocity and the damping ratio simultaneously, as a function of frequency.

Equations used to solve for velocity and damping in the NR method can be found in Appendix C.

REVIEW OF PREVIOUS WORK

Non-resonant testing was carried out by Rix and Meng (2005) on an undisturbed sample of natural sandy silty clay from Oplika, Alabama, USA. Prior to testing, the mass density (1.72 Mg/m^3), the void ratio (1.34), the moisture content (49%) and the degree of saturation (99%) were all measured. Unlike the Champlain Sea muds, with a plasticity index of 15-40% (Law and Bozozuk 1988), the plasticity index of this sample was 4%. The sample was confined to a vertical effective stresses of 68KPa caused by in situ overburden pressures. Testing was carried out at low strains of 8×10^{-6} over a frequency range of 0.01 to 30Hz. Results for this soil indicated that there was a rise in the shear modulus from 45 to 51 MPa (approximately 162-172 m/s) over the 0.01Hz to 30Hz frequency band. The damping had a broad 'U'-shaped curve: the damping ratio dropped from 2% to 1% between 0.01 and 0.1Hz, remained relatively stable between 0.1 and 1Hz, then increased from 1% to ~3% between 1Hz to 30Hz. In this sandy silty clay soil, the results indicate a slight dependence of damping on frequency in the earthquake band, (0.01 to 10Hz), and a greater dependence in the seismic exploration band (10 to ~200Hz).

Figure 5.6. Conceptual diagram of frequency effects on damping and shear wave velocity on a sandy silty clay (after Rix and Meng 2005).

In a finer grained material, Khan (2007) reports a negligible variation of damping with frequency in a clay soil (location not mentioned) between 10 and 50Hz. The soil was confined at pressures of 33KPa, and strains of 10^{-6} % were applied. Damping remained relatively constant at $\sim 3.3\%$, decreasing slightly to $\sim 3.1\%$ at 50Hz. In the range between 1Hz and 5Hz, the damping actually decreases from 3.9% to 3.3%.

These results indicate that at the same strain levels and low consolidations, material properties (grain size, in particular, although moisture content (void ratio), and saturation are also important) appear to play a large role in the frequency dependence of damping ratios and, of course, the range of damping. More data is required to investigate this observation.

TEST PROCEDURES FOR THIS STUDY

The same experimental set up as the RC column was used, and NR data was collected immediately after the low-strain RC testing was completed. During the testing, the complex shear modulus (G^*) is measured, which is a measure of the

amount by which stress lags behind (a harmonic) strain. G^* can be determined once the amplitude of the applied torque and the geometry of the specimen and resonant column are known.

RC testing must be first completed to determine the resonant frequency of the sample. Then, during the NR testing, monochromatic sine waves are applied to the sample via the signal generator. The voltage applied to the coils was pre-calculated so each input sine wave would achieve a common level of strain for the test – a problem with the RC method which is rectified in the NR approach. Frequencies were applied to the sample at 10Hz intervals (from 10 to 130Hz), except at 1Hz intervals for a 10Hz band across the resonant frequency, where greater sample density was desired. The peak of the input sine wave ('v1' – signal generator) and the peak of the output sine wave ('v2' – accelerometer response) were recorded for each frequency. The transfer function, TR, or its inverse, $H(w)$, was calculated as the ratio of v2 to v1, and an idealized plot of this ratio is shown in Figure 5.7. The angle at which v1 and v2 were out of phase with one another was calculated by the signal analyzer. These values, along with the frequency and torque, were used to compute the variation of damping and velocity with frequency using MatLAB software code developed by Lai and adapted by Khan to MathCAD (Cascante 2009, pers. comm.). Input tables for the MatLAB code for each of the tests can be found in Appendix C.

RESULTS - NR

The results of the NR testing likely reflect the level of disturbance in the samples. The NR method is very sensitive to heterogeneity in the sample, or uneven coupling of the sample with the top and bottom plates (Khan 2010, pers. comm.). In an undisturbed sample, one should be able to reconstruct a transfer function curve, as shown below.

Figure 5.7. Idealized transfer function (TR) (Khan 2010, pers. comm.).

However, of the samples tested, only the 10m sample produces a transfer function of this shape, and only in the range of 55-68Hz do the velocities seem realistic when compared to the downhole testing (Figure 5.8). The velocity measured in situ at this depth is 90.9m/s. The sample velocity in the 55-68Hz frequency band ranges from 80 m/s – 103m/s, and the damping decreases from 1.2% - 0.08% - encompassing the range of damping values measured in the field. According to damping curves published by Idriss (1990), we would not expect a damping value of greater than 0.3% for clay soils at strains of less than $10^{-4}\%$.

However, the variation of damping and velocity with frequency is much greater than that measured in the field, and could be a result of sample disturbance, poor coupling of the sample in the RC chamber, or potentially, insufficient consolidation time. Table 5.6 summarizes the results of the NR testing. These results indicate the importance of using undisturbed samples (through such means as block sampling) to reduce uncertainty in the relationship between damping and frequency.

Sample	Frequency Range	Damping	Shear Velocity	Downhole Vs
1 - 10m	55 - 68 Hz	0.5 - 0.1 %	80.2 - 103.4 m/s	90.9 m/s
2 - 30m	24 - 36 Hz	0.3 - 3.0 %	85.1 - 146.4 m/s	204.1 m/s
3 - 70m	26 - 44 Hz	1.0 - 3.0 %	99.8 - 218.9 m/s	285.7 m/s

Table 5.6 Results of NR testing. Downhole Vs is shown for comparison with lab tests. Sample 1 shows best relationship with downhole Vs and damping measurements.

Figure 5.8. NR results from the 10m sample. Left hand figures show the damping and Vs results across the entire 10-100Hz frequency test band, while the figures on the right hand side present the results in a band of 55-68Hz. The results do not represent good low strain results, and the velocities and damping vary considerably with frequency even within the narrow range presented. The velocities measured in the lab are also 50m/s to 100m/s slower than those measured in the field, potentially indicating disturbance or underconsolidation.

Figure 5.9. NR results from 30m sample.

Figure 5.10. NR results from 70m sample.

6.0 IN SITU MEASUREMENTS OF Q – PROCESSING APPROACHES

In this chapter, the approaches chosen for calculating Q from *in situ* (field) data are discussed in sections 6.1 and 6.2. The relationship between Q, velocity dispersion and attenuation in field measurements is also reviewed in section 6.3.

As with lab techniques for the measurement of material damping, the number of publications reporting on Q estimation from seismic data is quite extensive. As early as the 1950's, it was recognized that attenuation could be a valuable property in the interpretation of seismic reflection data for gas reservoir exploration, and research interest grew. Much of what is known of Q today is based on estimates from rock, but literature diverges on the appropriate processing techniques. In crustal rock, Q is generally quite high (i.e. >100) and an assumption of constant-Q conditions is often made between 10^{-2} to 10^7 Hz (Johnston et al. 1979) based on lab tests. It should be mentioned, however, that from a regional earthquake perspective, Q is a frequency dependent parameter. This is based on large-scale seismology studies, and is mainly due to the fact that when a seismic wave travels long distances from the source of an earthquake to a remote site, it goes through different types of rock materials with different damping characteristics, in addition to the frequency dependency of seismic wave scattering, which is not a concern of this thesis.

In near surface soils, however, there is no consensus in literature on how best to measure Q, and the reported ranges of Q vary quite widely, even within the

same material types. Table 6.1 presents a summary of *in situ* Q_s values measured in soft soils found within seismically active regions of North America.

As we shall see, the effect of strain on the measurement of Q in the field is a significant factor when compared to lab or earthquake data. The heterogeneities of soil, the range of soil properties, and the complexities of deposition history are also complicating factors in an *in situ* assessment of Q .

Location	Soil Type	Depth		Q_s	Freq Range	Sediment V_s Range	Method	Author
		from	to					
Mississippi Embayment, AK	sand, silt, some gravel	10 m	50 m	34 - 44	10 - 50 Hz	250 - 300 m/s	VSP, Spec Ratio	Pujol et al. 2002
Mississippi Embayment, AK	sand silt, some gravel			100	0.4 - 4 Hz	250 - 300 m/s	Surface Wave Atten.	Langston et al 2005
Gilroy, CA	sands, silts, gravels	10 m	115 m	10 $\pm$ 3	0 - 80 Hz	200 - 700 m/s	VSP, Spec Ratio	Gibbs et al 1994
San Francisco, CA	silty clays, sands	6.1 m	24.4 m	13 $\pm$ 5	20 - 80 Hz	265 - 315 m/s	VSP, Spec Ratio	URS 1986
San Francisco Bay, CA, Treasure Isl.	silty sands with clay zones	~1.5 m	10.5 m	25 - 208 (layer Q_s 's)		145 - 160 m/s	Surface Wave Atten.	Rix et al. 2000
Fraser Delta, BC	Holocene-aged Silts	21.9 m	34.9 m	125 - 157	50 - 70 Hz	179 - 202 m/s	SCPT, Spec Ratio	Stewart 1992
Fraser Delta, BC	Holocene-aged Clays	4.7 m	12.7 m	63 - 71	50 - 70 Hz	101 - 119 m/s	SCPT, Spec Ratio	Stewart 1992

Table 6.1 Q_s calculations made from *in situ* V_s measurements in soil deposits from other seismically active regions in North America. These results demonstrate the range of Q values interpreted from soil materials. VSP = vertical seismic profile, SCPT = seismic cone penetrometer.

Many researchers have extracted Q estimates from downhole seismic data in the time or frequency domains and numerous processing approaches have been developed. Tonn (1991) reported on ten techniques of Q estimation in rock and compared them using synthetic seismograms of varying noise levels. In the time domain, he reviewed amplitude decay, wavelet modeling, phase modeling, frequency modeling, rise time, and pulse amplitude and broadening methods. Today, these approaches have largely fallen out of favor versus frequency domain techniques because time domain methods require true amplitude recordings. In the frequency domain, Tonn describes the matching technique, spectral modeling, and the spectral ratio technique. He concludes that although the spectral ratio method and the analytical signal (under true amplitude conditions) prove the most reliable, noise levels and recording conditions dictate which method is the most appropriate and no clear winner in all cases was found. To this day, the spectral ratio method, and variations thereon, remain among the most popular techniques for Q estimation.

In the following sections, I describe the techniques I chose for Q estimation: the spectral ratio method in the frequency domain, and for comparison, the pulse broadening time domain approach.

6.1 PULSE BROADENING

The time-domain pulse broadening approach is based on changes in the shape of the first arriving pulse at increasing distances from a source. The rate at which a signal reaches its peak is based on frequency content, and changes in signal

shape can therefore be associated with the amount of loss in the signal. The shape of the first arrival pulse is characterized by its rise time (Gladwin and Stacey 1974) and pulse width (Ricker 1953). Rise time, τ , is the maximum amplitude of the pulse divided by its maximum slope, and pulse width, w , is defined by the time interval between the intercept of the maximum slope with the time axis and the first zero crossing after the peak (Figure 6.1). Both τ and w generally lengthen with distance from the source as the higher frequency components in the signal are attenuated. The pulse broadening approach estimates Q by quantifying an increase in pulse width, w , with increasing distance from a source. This technique has been used with both earthquake records (Wu and Lees 1996; Giampiccolo et al. 2003) and downhole seismic data (Badri and Mooney 1987; Liu et al. 1994).

Figure 6.1. Definition of the pulse width (w) of a seismic wavelet (Ricker 1953).

The method is based on the assumption that the broadening effect is due to anelastic attenuation, and that Q is approximately frequency-independent. As

proposed by Gladwin and Stacy (1974) and Kjartansson (1979), the broadening is proportional to the travel time, t , and can be used to calculate Q by

$$w = \left(\frac{C}{Q}\right) t + w_o \quad (6.1)$$

where w is the pulse width at an arbitrary depth and w_o is the width at the source. Equation 6.1 is simply a linear relationship between travel time and pulse width, with the C/Q term as its slope. The C term is a constant which has been reported as having various values between 0.1 and 0.5, depending on the source. Badri and Mooney (1987) and Wu and Lees (1996) use 0.5, as recommended by Gladwin and Stacey (1974), and this will also be used here. However, I am less interested in obtaining an absolute value for Q as I am in observing the trend of the plot of pulse width versus travel time. Based on equation 6.1, the slope of this plot will be equivalent to C/Q . If the slope is very small, there is little pulse broadening implying very large Q 's. If the slope is significant, the Q value will be lower.

Numerous methods have been developed to measure a pulse's width. The first is to measure the first quarter cycle, which is the time lapse between the first break and the first peak amplitude (Badri and Mooney 1987). The second is defined by the time interval between the intercept of the maximum slope with the time axis and the first zero crossing after the peak (Ricker 1953). A third approach, as suggested by Wu and Lees (1996), is a slight variation of the previous approach. Here, pulse width is found by measuring the time between a linear fit of the rising slope at the half-peak which is extrapolated back to the zero crossing, and the first

zero crossing after the peak. For consistency, I chose this approach for the pulse width measurements.

The advantage of the broadening approach is the short length of first arriving signal required, which is relatively free of other interfering waveforms. As will be seen with frequency domain approaches, the numerous complications of windowing in the time domain and the effects of amplitudes and spacing in the frequency domain are not of concern with the pulse broadening approach. However, the criticisms of this approach are that true amplitude comparisons must be made, which requires geophones to be identical, and oriented in the same direction. In addition, rotation of the waveforms as energy travels through the earth, a process which has been recently identified at the GSC and studied in Champlain Sea muds (Pugin et al. 2009), will affect the rise time and pulse width. For this reason, I have analyzed the pulse broadening of both horizontal components of the seismic pulses, and plot them both against travel time.

6.2 SPECTRAL RATIO METHOD

The spectral ratio method, developed by Redpath (1982) and furthered by others (e.g. Badri and Mooney 1987; Tonn 1991; Stewart and Campanella 1993; Gibbs et al. 1994), calculates a constant-Q value based on the observed amplitude attenuation in the frequency domain between two geophones placed a distance apart. Numerous derivations exist for the approach, but in general terms, the spectral amplitude of a wave, A , at a distance R from the source at frequency, f , can be thought of as the product of

$$A(R, f) = S(R, f)C(R, f)W(R, f)\frac{1}{G(R)}e^{-\alpha R} \quad (6.2)$$

where S is a source term, C describes the coupling of the geophone to the casing, and the casing to the surrounding formation, and W accounts for distortions due to wave propagation (scattering, etc). The G factor accounts for geometric spreading (i.e. $1/R$) and changes in amplitude due to variations in acoustic impedance (layering) along the ray path (Gibbs et al. 1994). This expression forms the Fourier domain equivalent of the equation for anelastic energy loss

$$A = \frac{A_0}{R}e^{-\alpha R} \quad (6.3)$$

For all practical purposes, C and W are assumed to be equal to 1 in the spectral ratio approach. Realistically, there is probably a frequency dependence for both of these terms, but this can be minimized during careful experimentation by verifying the casing coupling and, in the case of downhole experiments, bracketing intervals where stratigraphic variation (layering) is minimal (Spencer et al. 1982; Gibbs et al. 1994).

Taking the ratio of the spectral amplitudes (A_2/A_1) of two records collected a distance R from the source ($R_2 > R_1$), equation (6.2) becomes:

$$\frac{A_2(R_2, f)}{A_1(R_1, f)} = \frac{\frac{1}{G(R_2)}e^{-\alpha R_2}}{\frac{1}{G(R_1)}e^{-\alpha R_1}} \quad (6.4)$$

Taking the natural logarithms of both sides we obtain:

$$\ln \frac{A_2}{A_1} = \alpha(R_2 - R_1) + \ln \frac{G_1}{G_2} \quad (6.5)$$

Substituting equation (3.2) for α , we obtain:

$$\ln \frac{A_2}{A_1} = \frac{\pi \Delta R}{Q_s V_s} f + \ln \frac{G_1}{G_2} \quad (6.6)$$

where (6.6) is an equation of linear form ($y=mx+b$), which allows us to calculate Q from the slope of the natural logarithm of the ratio of the spectral amplitudes, where

$$m = \frac{\pi \Delta R}{Q_s V_s} \text{ or, } Q_s = \frac{\pi \Delta R}{m V_s}, \text{ or } Q_s = \frac{\pi \Delta t}{m} \quad (6.7)$$

where Δt is the travel time between the two receivers. The effect of the geometric spreading term is conveniently accounted for as the intercept of the trend at $f \approx 0$ Hz. A linear plot of $\ln(A_2/A_1)$ vs. f indicates that Q is constant, while a non-linear plot shows that the assumption of constant Q is incorrect. It has been observed in this study that the inherent point scatter in the $\ln(A_2/A_1)$ vs. f plot makes the interpretation of a linear or non-linear plot subjective in traditional broadband spectral ratio approaches.

Criticisms of this approach include the difficulties in choosing an appropriate window length in the time domain to obtain optimal frequency content for ratio calculation. With impulsive source data, the passing wave is very short (often <100 ms) and if combined with noise, may become unreliable or unusable. If the first pulse is isolated, the range of useable frequencies for the ratio may be quite limited. As with the pulse broadening method, if reflections and scattering from a layered media are incorporated into the signal, measurement errors may be included.

Despite these arguments, the spectral ratio approach is still commonly used today, both for seismic surface and downhole testing and earthquake spectral analysis (e.g. Lermo and Chavez-Garcia 1993; Borchardt 1994). A significant advantage of this method is that spectral amplitudes of the two horizontal geophone components (H1 & H2) in each tool can be combined using a root mean square (RMS) calculation. This eliminates the concern of comparing single unequal components in the upper and lower instruments (i.e. $H_1(R_1)$ with $H_1(R_2)$, or $H_2(R_1)$ with $H_2(R_2)$) caused by tool compasses not orienting identically, or any naturally occurring rotation of the waves between the two geophones.

In a study of Fraser Delta Holocene-aged silts and clay soils, a spectral ratio approach was used with seismic cone penetrometer (SCPT) data to calculate damping ratio, ξ (where $\xi = 1/2Q$). The results (i.e. very low damping ratios $\sim 0.75\%$) were found to agree with RC data collected at the same sites, and also with previously published results (Stewart 1992, Stewart and Campanella 1993).

The approach used in this study does not modify the basic spectral ratio calculations, but uses monochromatic signals, rather than a broadband source, to better observe amplitude attenuation in narrow frequency bands. Advantages of this approach include lengthening of the pulse time to improve the spacing density in the frequency domain, which will significantly reduce the scatter in the amplitude ratio plots, thereby improving the estimate of Q . A new dataset was collected specifically for these spectral ratio analyses and is described in Chapter 7.

A second advantage of the monofrequency data is the ability to look at whether the velocity of the signals change with frequency (velocity dispersion). The spectral ratio and pulse broadening approaches both look at the amplitude decay between two signals to estimate Q . Section 6.3 looks at the theory behind Q estimation using dispersion data, and discusses whether this approach is feasible in these soils.

6.3 RELATIONSHIP BETWEEN ATTENUATION, VELOCITY DISPERSION, AND Q

As discussed in Chapter 3, no perfectly elastic media exists in the near surface. Seismic waves are attenuated, in part, by anelastic energy loss due to friction which can be observed as phase velocity dispersion, in addition to amplitude loss. As energy travels through the earth, different frequencies travel at different velocities (phase velocities) and are attenuated at different rates. Higher frequencies are attenuated more quickly as the increased particle motion causes faster dissipation through friction. In theory, the entire wave, travelling at a group velocity composed of many frequencies, experiences a frequency downshift with time and distance from the source. In the last five decades, numerous theorems have been developed to understand the relationship between attenuation and dispersion, and the feasibility of estimating Q from dispersion observations.

As early as the 1950s, scientists were experimenting with attenuation in solids and liquids using seismograms (Gutenberg 1958). In 1958, Knopoff and MacDonald presented conclusions from the study of several different attenuation

models. They saw that in a linear medium (i.e. where stress is linearly related to strain), wave velocity *and* Q are frequency dependant. Since most strain induced by passing seismic waves is small, most earth materials can be reasonably assumed to be linear.

In the 1960's, Futterman published his influential paper on the relationship between attenuation and velocity dispersion, basing his theorem on the Kramers-Krönig relation. In physics, Kramers-Krönig relationships can be used to describe how a physical system responds to a small applied force; in the case of earthquake energy, a stress wave (real component) will travel through the earth with a strain wave (imaginary component) lagging behind by a certain phase. The imaginary part of the system would show how the medium attenuates energy, based on the degree that it is out of phase with the energy source, but the system must obey the laws of causality in order for this to be correct. Futterman (1962) reasoned that without a dissipative medium, the requirement of causality would not be fulfilled, since pulses travelling at different frequencies in a medium without attenuation could arrive at the receiver before the source is released. His theorem produced a method of calculating Q from velocity dispersion observations by using the Kramers-Krönig pair of real M_R and imaginary M_I parts of the elastic modulus (M), which is the ratio of stress to strain.

$$Q(w) = \frac{M_R(w)}{M_I(w)} \quad (6.8)$$

This approach was later found to be difficult to implement due to the very small levels of dispersion observed in seismic data. During the last decade, this approach has been revisited through the development of the non-resonant (NR) lab testing approach (Lai et al. 2001), which is applied to three samples from the John Shaw Rd borehole.

In contradiction to previous observations, Knopoff later observed that Q could be constant over a broadband of frequencies (Knopoff 1964). Liu (1976) published an explanation of this phenomenon, maintaining that if dispersion exists, Q could be constant within the medium at some frequency range. He reasoned that by superposing a series of Debye peaks unique to the material, a nearly constant Q -model within a certain frequency band could exist (Figure 6.2). The peak of attenuation (or damping peak) in a material is called a Debye peak and results from interactions at a grain boundary level for earth materials. At certain frequencies of motion between a material's relaxed and unrelaxed states, grain boundaries will move, causing maximum dissipation of energy. As earth materials are not uniform, different relaxation processes occur at different frequencies within the same material. When a series of Debye peaks are superimposed over one another, a constant Q model can be achieved in theory (Lay and Wallace 1995). These observations have large implications for studies of earth velocity structure due to the low frequency ranges under observation ($f_L=2.8 \times 10^{-4}$ to $f_H=1\text{Hz}$, or 1cycle/hr to 1cycle/sec) (Dahlen and Tromp 1998), but within the seismic exploration band, this 'constant Q band' only reaches the lowest end of the range, and its applicability in soil is unexplained.

Figure 6.2. (a) Debye peak where Q^{-1} is a function of frequency in a linear solid. (b) Superposition of Debye peaks form the basis for a constant Q model, where velocity (V) increases as a function of frequency across the constant Q band (after Lay and Wallace 1995; Liu et al. 1976).

Kjartansson reasoned in the late 70's that velocity increases with frequency in a constant-Q model and developed the following equation relating to Figure 6.2 (Kjartansson 1979):

$$\frac{V(f_2)}{V(f_1)} \approx 1 + \frac{1}{\pi Q} \ln \frac{f_2}{f_1} \quad (6.9)$$

where f_1 and f_2 are found within the f_L and f_H frequency band. Sun et al. (2009) found this equation to be valid in the exploration frequency band (5Hz-200Hz) when $Q > 10$ in dispersive media using shear wave velocities > 2500 m/s.

Equation 6.9 implies that negligible dispersion leads to elevated Q and thus very small damping, regardless of observed amplitude attenuation. Also, the linear-Q equation is extremely sensitive to very small changes in $V(f_2)$ and $V(f_1)$. The dispersion approach for Q estimation was developed for low loss materials of elevated velocity (rock) that fit a constant-Q model. However, literature provides measurements of damping in consolidated silts and clays which are also very low:

ranging from 0.4% in silts (Stewart, 1992 – seismic methods) to 1% for clays and silts (Campbell 2009; Zavoral 1990: lab methods). Given such elevated Q 's (50 - 125), would this hold in a low-loss low-velocity environment over the seismic exploration band?

These questions led to combined amplitude *and* dispersion attenuation experiments in the Champlain Sea muds. By using monochromatic pulses from 10Hz to 120Hz, both the relative amplitude decay and variation in velocity (dispersion) could be measured over an order of magnitude. Observing little to no increase in ratios of the signal amplitudes with frequency would support the idea of using the Kjartansson approach in low-loss soils within the exploration frequency band. In addition, if the spectral ratios at each frequency are varying non-linearly, valuable information will be revealed about the frequency-dependence of Q .

7.0 ATTENUATION MEASUREMENTS – FIELD WORK

Downhole seismic data were acquired in the Ottawa area to analyze the variation of pulse shape, spectral amplitude, and velocity dispersion for soft soil Q estimation, as discussed in Chapter 6. Although traditional techniques for these analyses have used broadband seismic data (e.g. 5-200Hz), the data collected for this study has primarily been monofrequency pulses ranging from 10Hz to 120Hz. In this chapter, the survey system components, calibration of survey tools, survey design, and collection of downhole shear waves are discussed.

The monofrequency approach has two advantages: to look at the variation of velocity with frequency (dispersion effects), and to look at the variation of signal amplitude with frequency (attenuation). This will allow us to approach the dispersion and attenuation aspects of Q and evaluate whether one reasonably predicts the other in our geologic conditions.

Field data were acquired over several visits to GSC-BH-JSR-01 and BH-AR93-1 during the summer and fall of 2009. The testing equipment consisted of three main components: downhole seismic receivers (geophones), a seismic source (Minivib), and a seismograph (Figure 7.1).

Downhole geophones: The downhole instruments chosen for this survey are manufactured by Geostuff Inc. of California. The BHG-3 model is a wall-locking downhole tool with three 15Hz geophone elements oriented vertically (V) and horizontally in the longitudinal (H1) and transverse (H2) orientations. The tool has a motor-driven bowspring arm which can be extended using a control console at

surface when the tool is lowered to a desired measurement depth. When the arm is extended, the tool is solidly clamped to the borehole wall, and the downhole cable can be slackened to prevent surface vibrations from being transmitted to the tool through a taut cable. The tool contains a fluxgate compass and a rotating motor to orient the geophones once the tool is clamped to the borehole wall. When oriented to magnetic north, component H1 is in-line with magnetic north, and H2 in-line with the E-W direction. The fluxgate is reportedly accurate within 5°. Two BHG-3 instruments were used for the survey, one clamped as a reference, and a second positioned at varying deeper positions in the borehole.

Minivib Seismic Source: The GSC's Minivib (pictured in Figure 7.3) is manufactured by Industrial Vehicles Inc. (IVI) of Houston, TX. The minibuggy Minivib provides a low-impact vibrating seismic source which can be operated in both P- (vertical) and SH- (horizontal shear) modes. Vibrating sweeps are programmable in time (seconds) and frequency (10-550Hz). For these Q surveys, the Minivib was operated in a monofrequency mode, which is not standard operating procedure for collection of seismic profile data.

Seismograph: A Geode seismograph mounted inside the Minivib buggy was used to record the 7 channels of data – three from each downhole tool, and the pilot trace input by the Minivib.

Figure 7.1. Schematic of the two-instrument test configuration. The upper instrument was clamped in place and provided a reference, while the lower instrument was moved in the hole to various positions, bracketing zones of varying geophysical and geotechnical properties (see Chapter 4). The control box and the seismograph were connected to a computer in the Minivib which controlled the input signal and recorded data.

7.1 CALIBRATION OF DOWNHOLE TOOLS

A critical component of amplitude measurements was that they record amplitude content identically. To accomplish this, six new 15Hz geophones were ordered prior to the surveys and tested in the lab to ensure they performed within specifications. The geophones selected were the GEO OMNI 15-2400 series made by GeoSpace Inc of Houston, TX. The OMNI models are operational at any degree from horizontal, so that a borehole slightly off vertical won't compromise the response of

the geophone. The tolerance for the amplitude of the geophones was $\pm 5\%$, meaning we should not expect the amplitudes to differ from one another by more than 10% when subjected to the same input.

The six geophones were tested in a variety of positions. Initially, they were firmly fixed upon a 15cm diameter vibrating 'shake surface' producing a 100Hz sine wave output. Figure 7.2 below shows the time and frequency domain response of the geophones. The amplitudes of the frequency peak at 100Hz were recorded and the difference between the two highest and lowest values was 10.9%.

Figure 7.2. Response of the six identical geophones affixed to a surface vibrating at 100Hz. The variation in the amplitude response was 10.9%. No phase lag was identified in the tests.

As this source was found to be slightly noisy and suspected of unevenly exciting the plate surface, a second test was carried out to determine the response of the geophones to an impact source generating lower frequencies. All six phones were affixed together and solidly glued to the concrete basement floor of the testing lab. Noise level tests indicated the surface was free of significant background noise.

A 16lb sledge hammer was stuck on a metal plate in vertical (V) and shear horizontal (SH) orientations at distances of approximately 10m-20m from the geophones. Tests were carried out in the three orientations the geophones would occupy in the geophone block of the downhole tool (V - upright, H1 – on side, in line with source, H2 – on side, orthogonal to source).

At various frequency peaks across the band from 10 to 100Hz, the results were analysed and the minimum and maximum peaks compared. Results indicated that all geophones were operating well within the specified range of $\pm 5\%$. Additionally, the highest and lowest values were not consistently seen on the same two geophones, indicating a noticeable bias did not exist among the phones. Tables 7.1a-c present the results of the amplitude tests, and Figures 7.3a-c show the time and frequency domain results in each orientation.

Vertical	Amplitude							Average Difference
Peak Freq	10Hz	22Hz	37Hz	52Hz	61Hz	76Hz	90Hz	
Geophone 1	285.673	5607.23	7353.08	1788.27	1385.11	1198.7	2528.29	
Geophone 2	285.839	5582.04	7551.93	1811.88	1403.14	1195.65	2574.3	
Geophone 3	285.175	5556.77	7128.44	1766.13	1364.32	1201.4	2472.38	
Geophone 4	274.2	5516.97	7659.34	1802.95	1396.73	1165.93	2581.28	
Geophone 5	275.791	5561.68	7070.05	1787.12	1382.53	1240.2	2480.42	
Geophone 6	289.482	5649.27	7655.91	1753.5	1356.77	1105.31	2515.02	
Greatest difference	5.3%	2.3%	7.7%	3.2%	3.3%	10.9%	4.0%	5.2%

Horizontal - 0°	Amplitude							Average Difference
Peak Freq	13Hz	23Hz	29Hz	39Hz	52Hz	73Hz	99Hz	
Geophone 1	211.434	1237.77	1463.33	640.133	218.042	185.903	258.137	
Geophone 2	203.603	1265.76	1462.12	649.322	224.708	189.519	260.383	
Geophone 3	219.973	1171.52	1409.27	625.65	202.983	184.318	258.305	
Geophone 4	205.82	1206.65	1443.08	639.972	215.061	188.408	262.695	
Geophone 5	202.137	1271.93	1449.1	664.035	226.886	197.549	267.085	
Geophone 6	218.35	1278.18	1453.68	650.334	216.995	188.564	260.638	
Greatest difference	8.1%	8.3%	3.7%	5.8%	10.5%	6.7%	3.3%	6.6%

Horizontal - 90°	Amplitude							Average Difference
Peak Freq	11Hz	20Hz	25Hz	30Hz	58Hz	70Hz*	95Hz	
Geophone 1	158.291	310.026	606.859	565.893	430.785	139.66	168.161	
Geophone 2	154.442	301.541	599.544	562.026	436.539	135.1	170.154	
Geophone 3	158.377	296.831	588.458	554.451	427.794	128.898	170.174	
Geophone 4	151.278	297.515	587.543	557.274	432.782	133.066	168.969	
Geophone 5	153.296	291.798	566.668	553.701	445.538	119.997	176.93	
Geophone 6	162.38	305.089	598.472	555.329	411.675	122.229	166.724	
Greatest difference	6.8%	5.9%	6.6%	2.2%	7.6%	14.1%	5.8%	7.0%

Tables 7.1a-c – Results of the amplitude tests in three orientations, using an impulsive source. The maximum (red) and minimum (green) amplitude values at each peak indicate that the geophones are working within the rating of $\pm 5\%$, as the greatest difference between the responses is $< 10\%$.

Vertical Orientation

Source Distance from Geophones 9.2 m

Horizontal 1 - 0°

Source distance from geophones 10.8m
Plate struck in SH mode

Horizontal 2 - 90°

Source distance from geophones 10.8m
Plate struck in SH mode

Figure 7.3a-c. Results of the impulsive source tests in the time and frequency domains.

Three new geophones were installed in each probe. An epoxy sealant was used to seal each geophone inside the geophone block which holds them in their V, H1, and H2 positions. Based on the results of the tests, it was decided that a reasonable and conservative error limit for the amplitude experiments was 10% peak to peak, or $\pm 5\%$.

7.2 EXPERIMENTAL SETUP IN THE FIELD

At each borehole, the Minivib plate was lowered 4m away from the borehole and the vibrating mass was oriented initially to be in line with magnetic north (Figure 7.4a, b). Thus, it was expected that H1 would receive the largest signal component and H2 would receive very little energy. At John Shaw Rd, this set up was used, and then modified so that the source was vibrated 45° from magnetic north allowing both horizontal components to receive similar amounts of energy. In fact, analysis would show that the recorded data did not show a clear effect of the source orientation.

The Minivib was configured to transmit horizontal shear motion (SH) with a ground force of 2000 lb/ft^2 ($9800 \text{ kg-force/m}^2$). In this environment, SH energy above 120Hz is attenuated (Pugin, pers. comm.), therefore, although data were collected up to 120Hz, the cut off for interpretable amplitude data was deemed to be 100Hz. In reality, the Q measured in these experiments is a Q_{SH} – a distinction which has not yet been made in literature. These tests did not incorporate the

contribution from a Q_{sv} component (Pugin, pers. comm.), which is a topic for future research.

The downhole geophones were connected through the Geostuff control box to a Geode seismograph mounted in the Minivib cab. The control box (Fig 7.4c) acts as a multiplexer for the six incoming channels from the downhole instruments. The two downhole tools were deployed, clamped, and geophones oriented, and the cables were then slackened prior to testing to reduce the transmission of surface vibrations to the geophones (Figure 7.4d).

Figure 7.4. The experimental set-up. The Minivib was driven up to the borehole (a) so the plate, when lowered (b), would sit 4m from the borehole. (c) The two BHG-3 downhole instruments with 6 identical 15Hz geophone elements. The seismic cables are connected to the yellow control box which lead to the seismograph in the

cab of the Minivib. (d) The deployed instruments; upper and lower cables are kept slack during the testing to reduce surface vibration transmission to the geophones.

The experimental frequency band ranged from 10Hz to 100Hz. While it is of interest to experiment at frequencies as low as 0.1Hz, the Minivib source is not capable of vibrating at frequencies below 10Hz. On the upper end, 100Hz is considered well beyond the uppermost limit of damaging earthquake energy, but was used to collect enough data to look at velocity and attenuation trends with increasing frequency over an order of magnitude. As mentioned, in SH mode, energy is significantly attenuated beyond $\sim 100\text{Hz}$. The monochromatic testing was run at 5Hz increments, and the Geode was configured to collect the signals with a sampling rate of 0.125ms, or 8000 samples/sec.

To select a depth for the upper phone, downhole vertical seismic profiles were consulted, which had been collected at the boreholes in 2008 using the Minivib or a hammer source. It was determined from observation of these logs and by experimentation at Anderson Rd that 15m was the shallowest depth the upper instrument could be placed to avoid surface wave and surface noise contamination. The lower depths were chosen based on variation of soil properties observed in the geophysical and geotechnical logs, as discussed in Chapter 4.

7.2.1 DISPERSION EXPERIMENTS

Both the amplitude and dispersion experiments acquired data with monochromatic input signals, however, each experiment required different input wave parameters so the experiments were run separately. In the upcoming

sections, I discuss the staking and taper parameters used for the dispersion experiments.

SIGNAL STACKING

During this stage of the data collection, one to three stacks were recorded at each frequency. Staking data adds traces together to improve the signal-to-noise ratio and increase the amplitude of reflected events. The high sensitivity of the new geophones was sufficient to produce a clear trace with one only stack, even at 90m, and generally this was adequate. If the initial stack was not clear, it was cleared and re-recorded.

SIGNAL TAPER

The Minivib is capable of generating swept- and mono-frequency input signals using a linear or non-linear sweep, and tapering the front and back end of the signal to gently or abruptly ramp up or down the input wave. For the dispersion experiments, a single frequency signal of 1 second length was used with a Hanning-type taper of 0.01 seconds. A short taper such as this would cause the shape of the transmitted wavelet to be sharp, making the interpretation of the arrival times clearer for the dispersion measurements. However, a sharp taper in the time domain would cause a broader peak in the frequency domain, and lead to a smaller amplitude peak. Figure 7.5 demonstrates this effect and shows 20 Hz waveforms collected at BH-GSC-JSR-01 with differing taper parameters. The sharp taper is preferable for dispersion observations as the onset of the wave is more clearly

identified, while the longer taper provides a sharper peak in the frequency domain which is better for the monochromatic amplitude measurements.

Design of Signal Taper

Comparison between 0.01 and 0.3 sec taper at 30m depth using a 20Hz signal

Figure 7.5. The experiments required different signal parameters. Dispersion experiments are improved by choosing a short signal taper to improve picking of the first arrival of the wave.

At the John Shaw Rd borehole, the reference tool was positioned at 30m, below the base of the low velocity zone. The lower tool was moved to positions at 50m, 70m, and 90m, as discussed in Chapter 4. To also collect data within the low velocity zones, the upper tool was placed at 20m, and the lower tool was placed at

30m. Using a taper of 0.01sec and an input signal length of 1 second, records were collected at 10Hz intervals from 10Hz to 120Hz.

At the Anderson Rd borehole, the depth intervals chosen were 10m & 30m, and 10m & 45m. Again, using a taper of 0.01sec and an input signal length of 1 second, records were collected at 10Hz intervals from 10Hz to 120Hz.

7.2.2 AMPLITUDE EXPERIMENTS

The monochromatic amplitude experiments were designed to focus energy in narrow ($\sim 4\text{Hz}$) frequency bands over an order of magnitude. Monochromatic records were collected at 5Hz intervals from 10Hz to 100Hz.

TAPER

As discussed in the previous section, signal taper settings were increased to 0.3s for the amplitude data collection, as a sharper peak in the frequency domain was desired around the center frequency.

SIGNAL STACKING

During this stage of the data collection, one to three stacks were recorded at each frequency. The high sensitivity of the new geophones was sufficient to produce a clear trace with one only stack, and this was generally adequate, although three stacks were collected at times at the 90m position to improve the signal-to-noise ratio. If the stacked signal was not clear, it was deleted and re-recorded. Once the

two downhole tools were clamped in place at the selected depths, the stack number was maintained across the whole frequency range.

SIGNAL LENGTH

A one second monofrequency pulse was used for the input signal. Although longer signals were tested, 1 second was an adequate length to define the amplitude spectrum in the frequency domain. The significance of the length of the signal window is discussed in Section 8.2.

8.0 DATA ANALYSIS AND RESULTS

This chapter contains a description of the processing steps used to analyze the downhole shear wave data sets in BH-GSC-JSR-01 and BH-AR93-1 using pulse broadening, spectral ratio, and dispersion approaches. For the pulse broadening, I used two pre-existing GSC broadband downhole shear wave datasets. For the spectral ratio method, I used the monofrequency amplitude data set, and for the dispersion experiment, I used the monofrequency dispersion data set, both collected for this study and described in Chapter 7.

8.1 PULSE BROADENING

BH-GSC-JSR-01: The dataset in BH-GSC-JSR-01 was collected in 2008 using the GSC's Minivib vibratory source. A broadband 20-300Hz input signal was swept over 7 seconds in shear horizontal (SH) mode and readings were collected at 0.5m increments up the borehole using a triaxial downhole tool (with 15Hz geophones). Signal processing was carried out using software developed by the Kansas Geological Survey (Pugin, pers. comm). The swept signal was correlated over the 20-300Hz band using a synthetic trace. This produces a minimum phase wavelet where the maximum peak is equivalent to the 'first break' or arrival of the shear wave with an impulsive source, such as a hammer or air gun. With correlated Minivib data, the width of the pulse is measured on either side of the maximum peak. As the Minivib source was vibrated 45° from magnetic north (allowing both horizontal components to receive similar amounts of energy), the width and travel times of H1 and H2 were averaged.

Prior to wavelet analysis (as shown in Figure 6.1), it was recognized that the downhole data contained a high frequency tube wave component which needed to be removed. Tube wave energy propagates directly along the borehole wall at the interface between the fluid and the PVC casing; consequently, it generally retains high amplitudes and interferes with the interpretation of the later arriving signal passing through the soil. Figure 8.1 shows the wavelets from the H1 and H2 components at 93 meters downhole, and the frequency spectrum of the wavelets. The high frequency component visible on the initial pulses distorts the pulse shape. I applied a bandpass filter to remove frequency components greater than 70Hz, which allowed me to more consistently measure the pulse broadening effect of the lower frequency shear waves (20-40Hz) from the top to the bottom of the borehole.

Figure 8.1 The H1 and H2 shear wave components at 93m downhole are shown before and after the filtering out of frequencies above 70Hz. The high frequency tube wave noise distorted the shape of the shear wave pulses and created the appearance of a narrower pulse. Once removed, the width of the minimum phase wavelet could be interpreted more reliably.

Figure 8.2. Variation of pulse width over the length of the JSR borehole. The trend indicates a gradual 0.5ms widening of the shear pulse between 10m and 95m, equivalent to a loss of approximately 2Hz over the entire borehole. These results indicate there is very little energy lost to anelastic attenuation in these soils.

Results from the JSR borehole in Figure 8.2 indicate that the pulse width trend increases over the borehole from approximately 11.3ms at 10m to approximately 11.8ms at 96m, following the equation $w = 11.18 + 0.0014t \pm 0.24\text{ms}$. This is equivalent to an overall loss of frequency content of the shear wave of approximately 2Hz (from 44Hz at the top of the borehole to 42Hz at the base).

Slight fluctuation of the trend between 25-35m likely indicates some influence from stratigraphy/soil conditions on the shape of the pulse. Assuming a C value of 0.5, the slope of the pulse broadening data provides a Q value of 357, which is unrealistic, but does indicate that the losses in the soil are so low that they are likely beyond the detection limit of the approach.

BH-AR93-1: At BH-AR93-1, the SH dataset was collected in 1996 using a 16lb sledge hammer. The signals were again filtered to remove high-frequency components (e.g. tube wave) above 70Hz.

The analysis results are shown in Figure 8.3. In this case, the majority of the seismic energy was in-line with the H1 component, and the low signal-to-noise ratio of the H2 component tended to cause fluctuations of several milliseconds when the two pulse widths were averaged together. Therefore, only the H1 component is shown. The change in pulse width over the soil column is so small that it is essentially interpreted to be below the detection level of the method with this data set, producing a very small negative slope. The results do indicate, however, that the method predicts very low levels of anelastic attenuation in these soils, likely greater than $Q \sim 150$.

Figure 8.3. Variation of pulse width over the length of the Anderson Rd borehole. The geological log shown on the graph indicates where changes in material type may have localized effects on pulse width.

8.2 SPECTRAL RATIO - ANALYSIS

In the following section, I describe the processing of the monofrequency data using the spectral ratio method, and present the results of the test intervals selected. A sample from each test interval is included, but additional figures can be found in Appendix D.

Window Length: One of the most important choices in the application of the spectral ratio method is the selection of window length. The effect of signal windowing is a significant one which has been discussed in Q literature for decades. Time domain windowing has consequences for the shape of the frequency spectrum, which is especially important for the spectral ratio method (Sams and Goldberg 1990). A longer window produces a smaller sample interval in the frequency domain, improving the definition of the frequency spectrum (Figure 8.4).

Figure 8.4. Sample data from the John Shaw Rd borehole showing a 95Hz signal in the time domain. A 0.1 second window results in a frequency domain sample spacing of 3.9Hz ($\Delta f=3.9\text{Hz}$), whereas a 0.2 second window results in $\Delta f=1.95\text{Hz}$. There is greater data density in the frequency domain with the longer window which creates a better defined peak around the signal's frequency. This improves the Q estimate when using a spectral ratio approach.

In the traditional spectral ratio technique, a broadband impulsive source is used to generate a wavefront. Before converting the time series into the frequency

domain, this waveform must be windowed to remove unwanted signal beyond the passing wave. This signal may be a fraction of a second long ($\sim 0.1\text{s}$), resulting in limited amplitude and frequency content across a broad spectrum which degrades Q estimates. In addition to the impact of the window length, when a boxcar window is applied in the time domain, a convolution with a sinc function results in the frequency domain. When the ratios are taken, the convolution effect does not cancel out, leading to further impacts on the Q estimate.

An article by Ingram et al. (1985) written after studying the errors introduced by windowing showed that errors are minimized when a longer window is used, and when the same portion of the two wavelets from the different tools are windowed (i.e. no misalignment). In the present study, I use the same window length for the upper and lower signals, but offset them by the travel time difference between them.

A monochromatic approach benefits from a longer time series, as the Minivib can be configured to set up a steady state signal for a desired length of time with tapers on either end. This allowed me to window several cycles of the signal, and to choose zero crossings as our start and end points. The proper balance between signal length in the time domain and signal amplitude in the frequency domain was found through experimentation. A one second monofrequency pulse was used for the input signal, although longer signals were tested.

Unwanted Events: Despite the advantages of the monofrequency approach, a long window could also capture interference from later arriving events (reflections from

deeper horizons, multiples, etc) affecting the amplitude and shape of the spectrum. A consideration with this approach, therefore, must be that a longer time signal could experience interference from later reflections from the bedrock (BR) and glacial (GL) layers. Spencer et al. (1982) discussed the interfering effects of stratigraphy on the spectral ratio technique. Figure 8.5 shows shear wave travel time sections from BH-GSC-JSR-01 (collected using a broadband Minivib signal and correlated over 10-300Hz) and BH-AR93-1 (collected using a hammer source). This allowed us to observe the travel of a single pulse and the times of the returning reflections from significant reflectors at depth. As the tools were lowered deeper into the borehole during the monofrequency experiments, the times of these reflections would arrive earlier in the record.

For the John Shaw Rd. borehole, according to the travel time plot at 90m, the arrival of the reflection from the glacial layer would be almost instantaneous after the passing of the wavelet, and 150ms later, the bedrock reflection would arrive. For the geophones placed at shallower depths, these reflections would arrive later in time at a reduced amplitude. For the Anderson Rd. borehole, at the 10m geophones, the reflection and multiple from the glacial layer arrive 530ms and 730ms, respectively, after the first arrival. At the 30m geophones, the reflection and multiple from the top of the glacial horizon arrive 200ms and 720ms, respectively, after the first arrival. At the 45m geophones, the glacial reflection arrives within 100ms of the direct wave, followed by a small reflection from the bedrock, and a multiple of the glacial reflection 720ms after the first arrival.

(a)

(b)

Figure 8.5. (a) Shear (SH) wave travel time section at the John Shaw Rd borehole using a broadband correlated Minivib signal, and (b) at the Anderson Rd borehole using a hammer source. These time sections provide insight into the arrival times of reflections resulting from a single pulse. Laid over the time section (yellow bars) is the window length of the steady state monochromatic pulses collected in the field. The depths represent the geophone positions for the test intervals chosen in Chapter 4 (Figures 4.2 and 4.3).

How and if the BR and GL reflections would impact the amplitude of a steady state signal (constructive or destructive interference) at both boreholes was an important consideration. Figure 8.6 shows the travel time traces from 30m, 50m, 70m, and 90m, pulled from the JSR profile in Figure 8.5(a) identifying the bedrock and till reflections. The amplitudes of the reflections in the 30m and 50m traces are negligible. As these events would not contribute significantly to the amplitude of a steady state signal of longer duration, the length of the window for the spectral ratio calculations was chosen to be 0.9s (as the 0.05s at the start and end of the 1s signal

were significantly suppressed by the taper). In the 70m trace, the till reflection is barely visible and does not appear to be significant relative to the amplitude of the passing signal. In this case, a 0.2s window (time elapsed between first arrival and till reflection) are compared to a longer 0.9 second window during the spectral ratio calculations. The effect at 90m from the glacial till reflection is difficult to assess as it becomes mixed with the main signal and some roll behind the first arrival related to the Minivib signal. An option would be to select a very small window (<0.1 m) of data from the onset of the wave, and compare it with increasingly larger windows. As a window of <0.1 s would produce very large Δf in the frequency domain, the ability to produce a stable spectral amplitude ratio curve is compromised.

Figure 8.6. Determining the presence of significant reflectors and deciding on window length is an important consideration with the monochromatic approach as large amplitude events from deeper reflections could increase or decrease the amplitude of a 1s monofrequency signal. Here, travel time traces from depths of 30m, 50m, 70m, and 90m in the JSR borehole from 10-300Hz signals show the arrivals of reflections from the till and bedrock layers. It was determined that the amplitudes of these GL and BR reflections are not significant relative to the amplitudes of the signals.

After reviewing the amplitudes of the GL and BR reflections for the Anderson Rd. borehole from Figure 8.5(b) in a similar fashion to those of the Anderson Rd

borehole, the reflections and multiples are very small at the 10m and 30m geophones (<10% of the amplitude of the primary wave), but at the 45m geophone, the primary and reflection are too intermixed to be separated for windowing of the primary wave. Therefore, a window of 900ms was used for the analysis of both data sets (10m-30m, 10m-45m), but the confidence in the results of the 10m-45m test is lower relative to the 10m-30m pair, as data point scatter is expected.

Figure 8.7 shows the resulting frequency spectrum from an increasingly larger window size. At the 0.1 sec window, the Δf is so large (5Hz) that the spectral peaks are no longer centered over the input frequency, resulting in a lower spectral ratio and a greater degree of data scatter in the resulting graphs.

Figure 8.7. Sample 30Hz signal recorded at 90m depth. Images show the impact of time windows in the frequency domain – although a 0.1 second window would be affected by fewer reflections from the bedrock and glacial layers, the resolution in the frequency domain around the central frequency is poor due to too few samples. A sharper definition around the central peak is desired for the ratio calculations.

As can be seen in Figure 8.8, an increase in data point scatter resulted from the reduction of window size. Although the drop from a 900ms to a 500ms window did not make a significant difference to the slope of the spectral ratios, the drop to

100ms did. In the case of the 30m-70m geophone pair, a drop to a window of 200ms to avoid possible interference from the glacial reflector impacted the slope value, lowering it from 0.004 to 0.002, and increasing the Q estimation at the site. The 100ms window also caused a reversal of the sign of the slope from negative to positive. It is hard to know whether this is due to the reduction of resolution in the frequency domain or an actual amplitude interference effect, but the amplitude of the reflection is so small compared to the amplitude of the passing wave that the reflector at this depth is not predicted to impact the amplitude value when using a steady state signal for 1 second.

Interestingly, a 100ms window would be a typical length for a passing wavelet from an impact source, showing the inherent problem in assigning a slope to the natural logarithm of the spectral ratios when the window is too short to permit good resolution in the frequency domain.

Figure 8.8. Graphs showing the increase in data point scatter of the spectral ratios when the time window is reduced; a result of the decreased resolution in the frequency domain. Another contributing factor is the increase in distance between the lower geophone and the source –a predictable effect as the signal-to-noise ratio decreases with increasing depth.

The distance between the source and receiver also had a significant impact on the spectral amplitudes. In the three panels of Figure 8.8, an increase in the data scatter occurs as the lower receiver moves further from the source. In the upper

panel, the 30m-50m ratios are quite linear, while the scatter increases in the 30m-70m pair, and more so in the 30m-90m pair. This is attributed to the overall decrease in signal-to-noise ratio as the lower geophones get further from the source. The scatter also tends to increase with frequency in all cases, likely due to effect of small scale stratigraphy scattering waves which travel at higher frequencies. This effect was studied by Spencer et al. (1984) in his research on the effect of stratigraphy on Q estimates. Although a large interval Q is of interest for earthquake hazard studies, a large separation (here, greater than 40m) increases the scatter and reduces the effectiveness of this method. In a material which is highly layered, this minimum source-geophone and geophone-geophone separation would need to decrease.

Signal Filtering: The monofrequency signals in this study are quite simple in form, and therefore, the windowed segments of the traces were truncated where amplitudes were equal to zero. Under these conditions, the discrete Fourier transform is equal to the continuous transform within a scale factor. This meant that side lobes derived from the transform of a rectangular window would be equal to zero, and therefore not a factor in the frequency domain (Pujol et al. 2002).

As the Minivib is not able to produce large amplitude signals at frequencies lower than 15Hz, the signal-to-noise ratio of the signals <15Hz was low. Therefore, the 10Hz and 15Hz signals were filtered using high-cut filters. The corner frequencies of the high cut filter were 12-17Hz for the 10Hz signal, and 17-22Hz for the 15Hz signal. Signals above 20 Hz did not require filtering.

Q Calculation: Once the time series were filtered (if necessary) and windowed using Signal Analyzer software (developed by Dew Research Inc.), signals were then converted to the frequency domain using a Fourier transform, and exported to Excel. Each exported data file contained five columns: a frequency column, and four spectral amplitude columns ($H1_u$, $H2_u$ (upper tool), and $H1_L$, $H2_L$ (lower tool)). The root mean square (RMS) value of the spectral amplitudes of the H1 and H2 components for the upper and lower tools were calculated and plotted on a composite plot at 5Hz intervals, as shown in Figure 8.9a. I chose to perform the RMS calculation to account for any misalignment of the compasses within the two downhole tools, and any rotation of the waveforms which occurred during transmission.

The spectral ratio of the lower to upper amplitude peak could then be calculated for each frequency (see Figure 8.9a). The natural logarithm of these ratios ($\ln(A2:A1)$) were computed and plotted against frequency (see Figure 8.9b). Using Equation 6.7, Q can then be calculated by knowing the difference in travel time between the downhole tools, and the slope of the best fit line through the $\ln(A2:A1)$ vs. f graph.

Q Error Analysis: According to equation 6.7, the quantities required for the calculation of Q_s are π (π), the travel time between the two receivers (Δt), and the slope of the natural logarithm of the spectral ratios, plotted against frequency (m). With the exception of π , each of these quantities has an error, ∂ , associated with it which has been quantified and carried through the Q calculations.

The error in delta time, $\partial\Delta t$, was assigned based on the standard deviation of the difference in travel times, Δt , measured between depth intervals using the waveforms from the monofrequency dispersion experiments. Errors in delta time ($\pm\Delta t$) are shown in Table 8.1. Sources of error I identified in the travel times could include potential depth positioning errors of the tools in the borehole, mispicking of the exact zero crossing at the onset of the shear wave, and any trigger delays related to the source. To minimize these sources of error, depth marks were carefully taped along the cable at 0.50m increments. As the tools were moved in the borehole, the tape marks were measured against a reference meter marker fixed to the borehole casing. Therefore, I felt that the error in depth positioning between the tools could have been up to 4cm, or 0.2% on the shortest interval of 20m. Errors in picking of zero crossing and maximum peaks were minimized by using automated picking software (Signal Analyzer) which snapped to a zero crossing or maxima of the wavelet. Triggering errors were not observed with the Minivib source, as the computer times the trigger. For the hammer source, when used, a contact closure trigger wire was employed to minimize any error associated with the triggering. Although the lower frequency hammer source created larger error in Δt (up to 5% using the $\frac{1}{4}$ cycle error of the broadband wavelet (Hunter 2010, pers. comm.)), the monofrequency Q and dispersion experiments were all run with the Minivib source.

Borehole	Depth Interval	Freq Range	Δt	$\pm \Delta t$	$\pm \Delta t$ (%)
JSR	20-30m	10-120Hz	0.0527	0.0008	1.5%
	30-50m	10-120Hz	0.0882	0.0006	0.7%
	30-70m	10-120Hz	0.1603	0.0005	0.3%
	30-90m	10-120Hz	0.2239	0.0009	0.4%
AND	10-30m	10-90Hz	0.1270	0.0007	0.6%
	10-45m	10-90Hz	0.2017	0.0010	0.5%

Table 8.1. Table of error (1σ) in delta time (Δt) between two downhole tools used in Q calculations.

The error in slope, ∂m , was assigned based on a straight line fit through the points using a linear regression calculation. This approach estimates the potential variation of the slope given the degree of scatter in the experimental dataset. The LINEST routine in Excel calculates a linear regression through the data points (giving slope, m , and intercept, b) and also the slope and intercept errors, ∂m and ∂b (Table 8.2). This approach is also used by Redpath in his report to the U.S. Geological Survey (URS 1986). The elevated levels of uncertainty in the slope value (± 30 -40%) in his study are largely associated with the assumption of a homogenous Q-structure.

Borehole	Depth Interval	Freq Range	Slope (m)	$\pm m$
JSR	20-30m	10-60Hz	-0.0014	0.0006
		70-100Hz	-0.0019	0.0014
	30-50m	10-120Hz	-0.0017	0.0006
	30-70m	10-120Hz	-0.0026	0.0008
	30-90m	10-120Hz	-0.0042	0.0014
AND	10-30m	10-75Hz	-0.0053	0.0031
		80-100Hz	-0.0289	0.0013
	10-45m	10-80Hz	-0.0103	0.0031
		10-85Hz	-0.0703	0.0421

Table 8.2. Results of slope error analysis ($\pm 1\sigma$).

In the upcoming figures presenting the results of the field tests and the calculated Q values, the vertical error bars in the $\ln(A1:A2)$ vs. f plots show the 5% error of the spectral amplitude ratio, based on the original amplitude calibration tests described in Chapter 7.1. Relative (%) errors were carried forward through the calculations using the established relationships given in Young (1962) for treatment of experimental error. The 5% error bars from the geophone recordings do not completely explain the scatter in the datapoints causing the error in slope, ∂m . In reality, variations in the soil over large source/receiver separations cause some scatter in the ratios to be inevitable, and this was observed to increase with increasing source/geophone separation. In addition, at some frequencies, signal interference caused one of the horizontal components to be 'noisy'. In these cases, the ratios are omitted from the slope fitting, as they represent points of low confidence. A method to improve the signal quality of future datasets would be to stack additional signals (>5) to compare with files containing fewer stacks to determine if the source of the noise was environmental. The noisy traces were

almost always the components out of line with the Minivib mass, where signal-to-noise ratio was low. In addition, the greatest number of noisy signals were those within the top 20m of the ground surface (subjected to surface effects: Minivib and urban noise, etc.) or those at a great distance ($>70\text{m}$) from the source where the signal-to-noise ratio is decreased. In these cases, the errors on the slopes were more elevated.

8.3 SPECTRAL RATIO - RESULTS

Nine datasets were collected for the spectral ratio processing at various depths in the two boreholes, as shown in Table 8.3.

	Dataset Number	Depth intervals and tool positions	Test Interv'l Name	Source Dir. (Vibe Pad)	Signal Taper	Window Length	# Stacks	Date Collected	Figure Num.
					(sec)	(sec)			
BH-GSC-JSR-01	1	30m (G1)-50m (G2)	JSR-2	90°	0.3	0.9	1	2-Nov-09	8.9
	2	30m (G1)-70m (G2)	JSR-3	90°	0.3	0.9	3 (10-20Hz), 2 (25-100Hz)		8.11
	3	30m (G1)-90m (G2)	JSR-4	90°	0.3	0.9	5 (10-45Hz), 3 (50-100Hz)		8.12
	4	20m (G2)-30m (G1)	JSR-1	45°	0.3	0.9	1	16-Nov-09	8.13
	5	30m (G2)-50m (G1)	JSR-2	45°	0.3	0.9	1		D-1
	6	30m (G2)-70m (G1)	JSR-3	45°	0.3	0.9	1		D-2
	7	30m (G2)-90m (G1)	JSR-4	45°	0.3	0.9	1		D-3
AR93-1	8	10m (G1)-30m (G2)	AND-1	90°	0.3	0.9	1	3-Nov-09	8.14
	9	10m (G1)-45m (G2)	AND-2	90°	0.3	0.9	1		8.15

Table 8.3. Parameters of the nine monofrequency data sets collected for spectral ratio calculation in boreholes BH-GSC-JSR-1 and BH-AR93-1. G1=Downhole tool 1 belonging to the GSC, and G2=downhole tool 2 belonging to Carleton University. Figures numbers of the interpreted results are also listed.

When the spectral amplitudes of the pulses decrease with frequency but the ratios between lower and upper frequencies vary linearly, we would have evidence that Q remains (nearly) constant while the attenuation increases with frequency. However, if the ratios vary non-linearly with frequency, we have evidence for a frequency-dependant Q . If the ratios remain almost constant, Q will be elevated.

Figure 8.9. (a) Composite plot of the upper (**30m**) and lower (**50m**) geophone spectral amplitudes from the monofrequency Minivib signals plotted at 5Hz intervals. The numbers above each peak indicate the spectral ratio of the lower to upper geophone. While variations in the peaks are due to the frequency filtering effects of the ground itself, the ratios at the center frequency remain fairly consistent. (b) The natural logarithm of the ratios from (a) are plotted against frequency and the slope is taken. The slope value is used in the calculation of Q (or damping). The slight decrease in the slope indicates there is very little amplitude loss due to intrinsic attenuation across the 10-100Hz frequency band ($Q=167$, or damping=0.30%).

The 'high-low' frequency response at 55Hz and 60Hz is unusual and seen in most experiments. This is observed to be related to the Minivib running near surface, but can also be influenced by near-by electrical sources. These data points were removed to reduce error in the slope fitting.

Notice that the spectral peaks do not have the same amplitude from frequency to frequency, but the ratios remain fairly consistent (0.45 ± 0.04). Although the ground force setting on the Minivib remained at 2000lb throughout the experiments, the ability of the pad to transmit low frequency (i.e. $< 20\text{Hz}$) SH signal is reduced. Within the Champlain Sea muds, the maximum amplitude of the SH wave is within the 25-75Hz range and is attenuated beyond this point. This effect is seen whether using an impulsive or swept sine source.

As such, a shortcoming of the broad spectrum spectral ratio approach is the reduction of amplitude content in the higher frequencies, resulting in a great deal of scatter on a $\ln(A_2:A_1)$ vs f plot. This is caused by taking the ratio of a very small number to a very small number, producing a very large or very small value. With the Minivib/monochromatic approach, the amplitude ratios of the signal at frequencies greater than 60Hz are reduced, but elevated enough to be measured, creating confidence in the results from a part of the frequency range which is generally not accessible (see Figure 9.2).

As a check on the consistency of the spectral ratios at the center frequency, ratios were computed for 2Hz on either side of the peak frequency. These should all be similar to the central frequency - if not, it is likely that noise interfered with the

signal, causing a distortion in the shape of the amplitude peak. This resulted in spectral ratios which had a greater range of ratios, such as those shown below in grey (Figure 8.10). These ratios were identified as 'low confidence' and excluded. In addition, this plot shows how the decreasing signal-to-noise ratio with increasing frequency causes the ratio values to be more 'scattered' in this plot, as shown in Figure 8.9a.

Figure 8.10. Expanded spectral ratio plot including ratios calculated for 2Hz on either side of the central frequency. This approach helps to identify signals which are affected by noise in the time domain, such as those in grey. Here, the shape of one of the component's frequency peaks is distorted, causing the ratio to rise or fall significantly. The unusual frequency response at 55 Hz and 60 Hz is associated with noise generated by power sources, such as urban transmission lines and the Minivib.

Figures 8.11 and 8.12 present the analyses of the monofrequency data from the JSR-3 (30m - 70m) and JSR-4 (30m - 90m) test intervals, plotted at 5Hz intervals. As with the 30m-50m interval, the Q values in the massive soils are very high. There is an increase in the RMS spectral amplitude in Figure 8.12 due to three signals being stacked at the time of data collection to increase the signal to noise ratio on the lower geophones.

Figure 8.11. Monofrequency spectral ratios from test interval JSR-3. Composite plot of the upper (30m) and lower (70m) spectral amplitudes from monofrequency Minivib signals plotted at 5Hz intervals. The slight decrease in the slope indicates there is very little amplitude loss due to intrinsic attenuation across the 10-100Hz frequency band ($Q=191$, or damping=0.26%).

Figure 8.12. Monofrequency spectral ratios from test interval JSR-4. Composite plot of the upper (30m) and lower (90m) spectral amplitudes from monofrequency Minivib signals plotted at 5Hz intervals. The slight decrease in the slope indicates there is very little amplitude loss due to intrinsic attenuation across the 10-100Hz frequency band ($Q=169$, or damping=0.30%).

The three lower test intervals in the JSR borehole indicate elevated Q 's and relatively little observable variation in Q over the 10-100Hz frequency bands tested.

This soil unit is relatively homogenous, and the increase in source-receiver spacing has accounted for the majority of the increased data point scatter in the spectral ratio plots. However, in the upper test interval, JSR-1 (20m – 30m), where significant reflections are seen on the high-resolution landstreamer reflection profile (see Figure 4.2) and soil properties are known to vary, the slope value is not constant across the frequency range tested.

Figure 8.13 presents the results from the monofrequency tests in test interval JSR-1 (20m – 30m). Across this depth interval damping is relatively low (0.4%, $Q \sim 116$) for wavelengths ($\lambda = v/f$) greater than 2.5m (10-65Hz), but at frequencies greater than 65Hz a change in damping behaviour is observed (damping of 0.6%, $Q \sim 87$). Wavelengths in this interval are calculated using the shear wave velocities from the downhole profile (shown in Figure 4.3). To rule out any software processing issues, Fourier transforms were carried out on the time series data with other software. Borehole geophysical logs, seismic reflection data, and core logs reveal that sediment textural changes occur in this soil interval at meter and sub-meter scales. At the deeper test intervals within the borehole, the soil has fewer visible textural variations and no prominent reflections, and the damping remains fairly steady (0.2% to 0.4%) across the 10-100Hz frequency band. These results suggest a stratigraphic influence may affect Q values in the upper soil column.

Figure 8.13. Spectral ratios in the JSR-1 test interval, where soil properties vary from those in the lower massive soils. At wavelengths smaller than 2.5m, the spectral ratios increase significantly, and damping appears to increase. Within this depth interval, borehole geophysical logs, seismic reflection data, and core logs reveal that sediment textural changes occur in this interval of overburden at meter and sub-meter scales.

Figures 8.14 and 8.15 present the results of the amplitude analyses for Anderson Rd in test intervals AND-1 (10m – 30m) and AND-2 (10m – 45m). At this location, road noise was a more significant factor than at the John Shaw Rd borehole. The borehole is located next to a busy rural road with irregular, but fast moving traffic. Minivib pulses were shot at breaks in the traffic, however, a greater degree of urban noise has likely caused a slight decrease in the signal-to-noise ratio.

The attenuation calculations for the frequency range of 10Hz-80Hz produces results with slightly more elevated damping values than those at BH-GSC-JSR-01, but with limited variation in the slope of the ratios. However, at frequencies greater than 80Hz, a significant drop in the ratios occurs when the amplitudes for the 10m geophones increases dramatically. There is no visible introduction of broadband surface noise into the traces so this does not appear to be due to an urban (road) noise effect. At 10m depth, the velocity of the shear waves has been measured at 122m/s, therefore, the wavelength at which this effect begins is approximately 1.5m. The tool is clamped within a zone where surface waves are observed (see note on Figure 8.5b). It is possible that the tool is not placed far enough away from the source to be in the far field regime, and we are detecting elevated amplitudes from surface waves. Alternatively, we may be detecting a resonance effect which is set up at these frequencies based on the near surface geology. However these large amplitudes are caused, the net observed effect is that these higher frequency waves are significantly attenuated ($Q=9-13$, damping=4-5%) by the time they reach the 30m and 45m depths. Amplitude decrease due to geometric spreading of surface waves occurs much more rapidly than body waves (at a rate of $1/R^2$ for surface

waves, vs. $1/R$ for body waves (Kramer 1996)). This is a significant contrast from the very high Q values of greater than 100 which we have observed at depth in the JSR borehole, and implies that the near surface is capable of significant attenuation of high frequencies (80-100Hz) at this location.

To confirm that this effect is restricted to the near surface, spectral ratios were computed from the 30m and 45m spectral amplitudes from the AND-1 and AND-2 datasets. Although the same number of stacks were used, the signal was generated at different times, so it was expected that there may be some additional data point scatter in the resulting $\ln(A_2:A_1)$ vs f plot. Figure 8.16 presents the results of this analysis, and indicates that the Q 's are again elevated (approximately 100) and that the intrinsic losses are low.

The results at this location indicate that dramatic variations in Q can exist using a surface seismic source, and may account for some of the low Q measurements reported in literature in soft soils using such investigation methods. An upper receiver placed in the near field where surface waves exist may lead to the conclusion that ground attenuation is much greater than should be expected for body waves. As an example, a survey carried out in fairly homogeneous San Francisco Bay muds reported a Q value of $13 \pm 35\%$ ($V_s = 290$ m/s) with a reference geophone placed at 6.6m depth and a lower phone at 24.4m (URS, 1986).

However, the attenuated high frequencies in the near surface which we observed in this study are much greater than those which would be considered damaging from an earthquake in the Ottawa area. Low frequency earthquake waves with

wavelengths much greater than 20m (i.e frequencies less than 10-15Hz) would not be significantly attenuated by a frequency-dependant Q value of 10 in the uppermost 10-15m of soil.

Figure 8.14. Spectral ratios and Q calculations from the AND-1 test interval. Significant amplitude increases recorded at 10m may be indicative of a near field effect or surface waves at frequencies greater than 80Hz. Significant damping of these high amplitudes are observed with Q values decreasing from 77 (10-75Hz) to 14 (80-100Hz).

Figure 8.15. Spectral ratios and Q calculations from the AND-2 test interval. Significant amplitude increases recorded at 10m may be indicative of a near field effect or surface waves at frequencies greater than 80Hz. Significant damping of these high amplitudes are observed with Q values decreasing from 63 (10-75Hz) to 9 (80-100Hz).

Figure 8.16. Spectral ratios and Q calculations from the depth interval between 30m and 45m. Data points are scattered as the waveforms were generated in separate datasets with different signal inputs.

8.4 DISPERSION

The dispersion datasets are analyzed to determine whether the approach described by Kjartansson (1979) for rock produces Q values which are similar to the results from the spectral ratio processing. Table 8.4 presents the parameters of the datasets used in the analysis.

BH	Dataset Number	Depth intervals	Test Interval Name	Freq. Range	Comp't Analysed	Source Dir.	Signal Taper	Num. Stacks	Figure Number
BH-GSC-JSR-01	1	30m (G1)-50m (G2)	JSR-2	10-120Hz	H1	90°	0.01	1	8.18
	2	30m (G1)-70m (G2)	JSR-3	10-120Hz	H1	90°	0.01	1	8.18
	3	30m (G1)-90m (G2)	JSR-4	10-120Hz	H1	90°	0.01	1	8.18
	4	20m (G2)-30m (G1)	JSR-1	10-120Hz	H2	45°	0.3	1	8.18
AR93-1	8	10m (G1)-30m (G2)	AND-1	10-90Hz	H1	90°	0.01	1	8.19
	9	10m (G1)-45m (G2)	AND-2	10-90Hz	H1	90°	0.01	1	8.19

Table 8.4. Parameters of the dispersion datasets collected in 2009.

The dispersion data are analyzed only in the time domain. The signals recorded at the upper and lower depth intervals were imported into Signal Analyzer one frequency at a time. The H1 components were chosen as they were in-line with the source direction and received the majority of the energy.

Figure 8.17 presents sample waveforms used to detect variation in velocity with frequency. The signals were collected in the JSR borehole with geophones clamped at 30m and 50m. The H1 components from each geophone are shown responding to a 20Hz signal. Note the variation in signal amplitude. The time of an early peak (T_1) from the upper geophone is chosen using picking software and subtracted from the arrival time of the same peak at the lower geophone (T_2). Note that the Δt between each waveform remains consistent, so it wasn't necessary to pick the first arrival. The travel time divided into the distance between the geophones gave the interval velocity between the two geophones. The results are tabulated in Table 8.5.

Figure 8.17. Sample waveforms from the dispersion experiments at the JSR borehole. Note the difference in amplitude scales. Similarities between the shapes of the waveforms made peak recognition straightforward. The difference in travel time divided into the distance between the geophones provides the interval shear wave velocity of the soil.

Freq	High Cut Filter	T₁	T₂	ΔT	$V_s=d/\Delta T$
Hz	Hz	sec	sec	sec	m/s
10	12-17	0.378	0.467	0.089	224.7
20	22-27	0.365	0.453	0.088	227.3
30	-	0.359	0.447	0.088	227.3
40	-	0.354	0.443	0.089	224.7
50	-	0.35	0.439	0.089	224.7
60	-	0.347	0.435	0.088	227.3
70	-	0.344	0.432	0.088	227.3
80	-	0.342	0.43	0.088	227.3
90	-	0.341	0.428	0.087	229.9
100	-	0.339	0.427	0.088	227.3
110	-	0.337	0.425	0.088	227.3
120	-	0.336	0.424	0.088	227.3

Table 8.5 Sample dispersion table from the JSR borehole with downhole tools clamped at 30m and 50m, $\therefore d=20m$. The resulting $\Delta T=0.088\pm0.001\text{sec}$, producing $V_s=226.9\pm1.5\text{m/s}$.

As a check, these Δt 's could be compared with the Δt 's measured between first arrivals during downhole VSP surveys using an impulsive source. In all cases, the Δt 's from the vibratory source and the impulsive source (either hammer or correlated Minivib) were within $\pm0.001\text{s}$ of one another (regardless of the wave peak chosen in the monochromatic experiment).

Figures 8.18 and 8.19 present the results of the velocity variation with frequency analyses using the John Shaw Rd and Anderson Rd data sets.

Figure 8.18. The results of the velocity variation tests in the John Shaw Rd borehole, showing that velocity variation is negligible within the 10-120Hz frequency band of the lower 3 test intervals, with standard deviations of less than 1.50 m/s. In the upper test interval (20-30m), the average velocity is 189.0 m/s with a standard deviation of 2.17m/s. The light blue data points at 10 and 60Hz were omitted from the linear regression because of poor signal quality.

Figure 8.19. Results of the velocity dispersion tests in BH-AR93-1, showing negligible change in velocity within the 10-90Hz frequency band, with standard deviations of less than 0.6m/s. Near surface interference of the instrument clamped at 10m made the travel time data very difficult to interpret at frequencies <20Hz and >90Hz. Data points omitted from the regression are shown as light green or red.

Least squares regressions were used to produce linear equations describing the observed variation of velocity with frequency. The greatest variation of velocity with frequency is within JSR-1 but this interval also had the greatest uncertainty with the largest standard deviation. Overall, the slope terms are extremely small, and any increase in velocity with frequency (within the 10-100Hz band) is observed to be negligible (i.e. within the error limits of the linear equations).

Borehole	Interval	Linear Equation	Fit Standard Error (Fit)
		(m/s)	(m/s)
JSR	1	$V_s = 186.86 + 0.0322f$	2.01
	2	$V_s = 225.45 + 0.0215f$	1.33
	3	$V_s = 249.06 + 0.0065f$	0.76
	4	$V_s = 267.71 + 0.0042f$	1.18
And	1*	$V_s = 157.92$	0.58
	2*	$V_s = 174.09$	0.46

Table 8.6. Linear equations (derived using least squares regression) representing the variation of velocity with frequency within the test intervals. *At Anderson Rd, the variation of velocity with frequency was so negligible that the slope became very small and negative (<-0.001), which is unrealistic. Therefore I neglect the slope term and account for the uncertainty in the velocity with the standard deviation.

Based on the equations from Table 8.6, the predicted values for velocity at 10Hz and 100Hz are calculated and shown in Table 8.7. Using the fit standard errors allows us to predict the widest range of velocities we would expect in this 10-100Hz range.

Borehole	Interval	Intercept	Slope	Fit Standard Error ($\pm 1\sigma$)	V_s (f=10Hz)	V_s (f=100Hz)
				(m/s)		
JSR	1	186.86	0.0322	2.01	187.2	190.1
	2	225.45	0.0215	1.33	225.7	227.6
	3	249.06	0.0065	0.76	249.1	249.7
	4	267.71	0.0042	1.18	267.8	268.1
And	1	157.92	0	0.58	157.9	157.9
	2	174.09	0	0.46	174.1	174.1

Table 8.7. Using the linear equations from Table 8.4, the velocities at 10Hz and 100Hz are calculated. Using the standard deviations, the lower most expected value (10Hz- 1σ) and the uppermost expected value (100Hz+ 1σ) can be computed.

Table 8.8 presents the results of Q calculations for each borehole's test intervals using Kjartansson's Equation (6.9). The velocity values at hypothetical points $f_2(100\text{Hz})$ and $f_1(10\text{Hz})$ are taken from Table 8.5. I used both the average velocity values, and the most extreme end points ($10\text{Hz}-1\sigma$, $100\text{Hz}+1\sigma$) to calculate Q and damping values over the order of magnitude I experimented within. These form a range of Q (or damping) values which may be expected based on the velocities measured. The constraints imposed when using Equation 6.9 are that Q is constant (or very weakly frequency dependant), and much greater than 10. From our spectral ratio experiments, we know these conditions are met, save in JSR-1 (65Hz-100Hz) and AND-1 and -2 (75-100Hz).

Borehole	Interval	Average Velocities V_s		Velocities with error $V_s \pm \sigma$	
		Q	D (%)	Q	D (%)
JSR	1	47.3	1.1%	19.7	2.5%
	2	85.4	0.6%	35.8	1.4%
	3	312.0	0.2%	86.4	0.6%
	4	519.0	0.1%	71.3	0.7%
And	1	99.4	0.5%	99.4	0.5%
	2	138.3	0.4%	138.3	0.4%

Table 8.8. Calculated values of Q and damping based on Kjartansson's (1979) approach to Q estimation from estimates of velocity dispersion.

In the following chapter, the results of the pulse broadening, spectral ratio, dispersion, and lab measurements are summarized, compared and discussed.

9.0 DISCUSSION

A compilation of the experimental Q values from this study is presented in Table 9.1. The *in situ* testing resulted in elevated Q values ($Q > 100$, $\xi < 0.5\%$) in the soft, homogenous soils at depths greater than 30m. At depths of less than 30m, it was observed that frequencies in the high end of the testing band ($> 70\text{Hz}$) were more attenuated in lower velocity soils with varying grainsize, moisture content, and stiffness properties.

Testing Approach												
BH	Geophone Position	Monofreq. Data Set #	Frequency Range	Pulse Broadening		Spectral Ratio		Dispersion		RC Lab testing		
				Q	D (%)	Q	D (%)	Q	D (%)	Q	D (%)	
BH-GSC-JSR-01												
	10-95m		10-70Hz	>>200	<0.25%							
	20-30m	4	10-100Hz					19.7 - 47.3	1.1% - 2.5%			
			10-65Hz			115 ± 50	0.43 ± 0.19					
	30-50m	1 5	70-100Hz			87 ± 65	0.58 ± 0.43					
			10-100Hz			167 ± 59	0.30 ± 0.11	35.8 - 85.4	0.6% - 1.4%			
	30-70m	2 6	10-100Hz			200 ± 95	0.25 ± 0.11					
			10-100Hz			191 ± 61	0.26 ± 0.08	86.4 - >200	0.2% - 0.6%			
	30-90m	3 7	10-100Hz			199 ± 86	0.25 ± 0.11					
			10-100Hz			169 ± 58	0.30 ± 0.10	71.3 - >200	0.1% - 0.7%			
	10m (Sample)					192 ± 83	0.26 ± 0.11					
	30m (Sample)										37	1.36%
	70m (Sample)										27	1.85%
										41	1.23%	
BH-AR93-1	5-60m		10-70Hz	>>200	<0.25%							
	10-30m	8	10-90Hz					99.4	0.50%			
			10-75Hz			75 ± 44	0.67 ± 0.39					
			80-100Hz			14 ± 1	3.6 ± 0.2					
	10-45m	9	10-90Hz					138.3	0.40%			
			10-75Hz			61 ± 18	0.81 ± 0.24					
			80-100Hz			9.0 ± 5.4	5.5 ± 3.3					
	30-45m			10-100Hz			106	0.47%				

Table 9.1. Resulting Q and damping values from BH-GSC-JSR-01 and BH-AR93-1.

9.1 *IN SITU* PROCESSING APPROACHES FOR VERY SMALL STRAIN

Pulse Broadening: The pulse broadening approach was useful to assess the frequency loss in a wavelet over increasing depth. With very little observed widening of the pulse width ($<0.5\text{ms}$) with a broadband signal, the frequency loss was approximately 2Hz over 85m of Champlain Sea muds. Therefore, I expected that we would not see significant amplitude attenuation or velocity dispersion in our very small strain downhole monofrequency tests. However, at such low attenuation values, any noise riding on the signal could become a greater factor than the widening of the pulse itself. Removal of tube wave frequencies is an important consideration with this approach. This, in addition to inconsistencies with a hammer source is likely why we obtained a slightly negative slope at the Anderson Rd borehole. Using the Minivib as a source increased our signal to noise ratio and provided a highly repeatable signal for interpretation.

Spectral Ratios: The spectral ratio approach with the monofrequency source was successful in transmitting high amplitude shear wave energy into the ground from frequencies between 10-100Hz. At frequencies below 60Hz, the method was especially effective, because the near surface naturally attenuates broadband signals above this frequency. This allowed me to observe amplitude attenuation at depths greater than 10m across a spectrum of 10-100Hz with higher confidence than with the traditional spectral ratio method. Having clearer resolution of the trends in the $\ln(A_2:A_1)$ vs f graph allowed me to identify where Q did not appear to be constant over our test frequency range, as indicated in Table 9.1. I was therefore able to

identify frequency-dependant amplitude attenuation at frequencies greater than 70Hz in materials with stratigraphic variation on the 1-3 meter scale.

Dispersion: The dispersion approach to Q estimation uses velocity variation with frequency to estimate Q. I applied this approach using a series of low taper monofrequency signals to determine whether a frequency dependence of Vs could be observed. Results of the data analysis showed that velocity variation with frequency was nearly negligible, and therefore, an elevated Q was computed using Equation 6.9. Interestingly, there is agreement between the results of the spectral ratio and dispersion measurements in the firmer clays at depths greater than 30m where the measured variation in velocity with frequency was almost negligible.

A drawback to this approach is that it neglects any frequency-dependant contribution from stratigraphy, which is a necessary consideration for the measurement of total Q in most soils. For that reason, the dispersion measurements may be able to represent an average Q in reasonably homogeneous soils (such as the deeper intervals of Champlain Sea muds tested), but may not be very representative of conditions in a low Q (or highly dispersive) material. More research is required.

Damping values calculated using all three the approaches described above were low ($<0.5\%$) at low strain levels ($<10^{-5}\%$). These results are in agreement with Idriss' 1990 published damping curve values at low strains, represented in Figure 9.1. Repeat tests with the Minivib mass oriented at varying angles from magnetic north resulted in Q values which fell within each other's error limits.

Figure 9.1. Published damping curve for clay (after Idriss 1990) corresponds well with this study's *in situ* test results.

As the Champlain Sea deposits are fairly homogenous and the Q values were elevated in the most massive portion of the soil column (depths > 30m), it is difficult to assign a dominant soil property which affected the Q (or damping) calculations. Certainly strain level was the external controlling factor, and increasing stiffness with increasing depth also had an effect, but it did appear to be negligible in the low-strain conditions. In interval JSR-1, the most noticeable soil property was the increase in porosity upward from 45 to 60%, and the transition from a marine clay to a varved glacio-lacustrine deposit from 20m upwards. At 70Hz, the damping increased from 0.43 ± 0.19 % to 0.58 ± 0.43 %, indicating that increased damping was occurring at seismic wavelengths of 3m or less. This is likely attributable to a stratigraphic effect, where the amplitudes of the waves were decreased through

interaction with (or reflection from) layers on the order of 3m thickness or less. As identified by Spencer et al. (1986), cyclic stratification is known to cause frequency-dependant attenuation, which hides the effects of intrinsic Q .

Finally, the significant attenuation ($Q=9$ to 14) observed in the Anderson Rd borehole at frequencies greater than 65Hz (when the upper geophone was placed in the near surface) is attributed to attenuation of surface waves *and* body waves, as opposed to body waves only. This may help to explain why early seismic convention led to the belief that a low V_s also meant that Q_s would be low (see Table 6.1). Research published in the last decade is starting to reflect the observation that Q can be quite high in soils of relatively low V_s (Langston et al. 2005; Rix et al. 2000).

9.1.1 TESTING CONSIDERATIONS - MONOFREQUENCY APPROACH

A number of factors must be considered in the application of the monofrequency technique. The first being that a signal time window must be selected which takes into account the surrounding stratigraphy. A longer time signal is desirable for improved resolution in the frequency domain, but cannot be compromised by strong interlayer reflections, which will increase or decrease the amplitude of the spectral peak at the signal's frequency. If reflections strongly affect the amplitudes, the result will be a plot of spectral ratio vs. f with significant data scatter. Downhole shear wave velocity logs and geological information are strongly recommended tools in the assessment of window length and tool separation. This technique is well suited for thick packages of massive muds as reflections are minimized which allows for longer signal windows.

Secondly, the signal taper should be fairly long to result in a sharp peak in the frequency domain. A sharp taper is preferable for the dispersion experiments which will create a strong pulse arrival in the time domain which is easily recognized.

Geophone position is critical for the success of a survey. Geophones placed too near the surface may record significant surface wave noise which will interfere with amplitude ratios. A recommended depth of placement of the upper geophone is several meters below the depth of significant vibrations and surface waves (i.e. generally >10m for Minivib/traffic noise, but this will also depend on soil type and sources). Distance between geophones will depend on the soil layering, but it was generally observed that the reliability of the spectral ratio slope decreased with distance from the source and with increasing distance between the geophones. In our case, sixty meters of separation, even in a low-loss soil produced more data point scatter in the spectral ratio plot than 20m or 40m. In a different medium with significant layering of different materials, this distance would certainly need to be reduced. It was also observed that the 'goodness of fit' of the regression line through the spectral ratio plot decreased with increasing frequency. Higher frequencies tended to cause more scattering as they interfered with small scale heterogeneities in the soil. Geophone location should also depend on the depths of significant reflectors. Amplitude ratios will be compromised if geophones are placed too near to a significant reflector.

An advantage of a frequency domain method for Q calculation is the ability to compute the root mean square (RMS) of the horizontal spectral components in each

tool. The reasons for this are to account for any misalignment of the compasses which orient the geophone block to magnetic north, and also the effect of SH signal rotation. As these effects aren't completely understood, the analysis of a single component for spectral amplitude is not recommended. Two orthogonal components can be summed as vector quantities in the frequency domain but not in the time domain, making a frequency domain processing technique a preferred choice.

9.2 LAB VS. *IN SITU* DATA

As discussed in Chapter 4, the resonant column (RC) test results concur that damping in these soils is fairly low, and if sample disturbance has played a role, damping is quite likely even lower than measured. Other RC tests in the soft silty soils of the Fraser Delta have yielded damping ratios as low as 0.75% (Stewart 1992), and RC tests in Champlain Sea soils at much higher strains (10^{-3}) have shown values of $\sim 1.75\%$ (Law et al. 1985). Based on these measurements, Idriss' 1990 damping curve for clay (Figure 9.1), and the results from the field testing, it seems reasonable to expect that damping should be less than 1% at low strains, but this will need to be confirmed with future testing of undisturbed block samples.

Strain has a very large role to play in the comparison of lab vs. *in situ* Q measurements. Shear waves generated by seismic sources (e.g. hammer, Minivib) have been estimated to produce disturbances on the order of $10^{-6}\%$, perhaps up to $10^{-5}\%$ depending on distance from the source. Weak motion earthquakes can produce disturbances up to $10^{-4}\%$. Lab data curves plotting damping vs. shear strain show that damping can change considerably over this range, up to 70% in some cases (Campbell 2009). In addition, results from Kim & Novak (1981) indicate that under lab conditions at very low strains, samples under constant confining pressure show velocities that continue to rise beyond the 1000 minute mark. This may mean that lab conditions overpredict damping values of Q_{in} , depending on the soil and the time required for consolidation. High porosity soils will likely continue to consolidate well beyond 1000 minutes. These factors suggest

that mismatches may occur between field, lab, and earthquake data when computing attenuation (Q_{in}) or damping terms, even under the best of specimen conditions.

Results from the non resonant (NR) testing show more rapid variation in damping and velocity versus frequency within narrower frequency windows than were observed in the *in situ* amplitude or dispersion data sets. As the NR testing method is particularly sensitive to sample inhomogeneities, these results are likely influenced by sample disturbance leading to poor low strain transfer functions. The best agreement between lab and *in situ* tests was found in the sample collected within 10m of the ground surface. However, as the velocities and damping values were generally in the range of the downhole measured values, and the results are encouraging for this sensitive soil. Future research with this method for varying strain and frequency tests would require carefully treated samples.

9.3 INTRINSIC VS. EFFECTIVE Q

The effect of elevated Q's from seismic measurements has been noted in recent papers. Campbell (2009) discusses the discrepancy between Q's measured from earthquake data versus Q's measured in the field, and even in the lab. The model proposed by Dainty (1981) explains effective attenuation, $1/Q_{ef}$ as the sum of two terms: $1/Q_{in} + 1/Q_{sc}$. Q_{in} , the intrinsic term (or anelasticity term) is frequency independent, and Q_{sc} , the scattering attenuation term, is frequency-dependent. Scattering would occur largely at reflective boundaries within the material and is frequency-dependant due the thickness of the layering in the particular soil.

Campbell theorizes that seismic surveys measure the Q_{in} term, and not Q_{ef} because of the relatively simple seismic pulses which are generated at surface.

In earthquake data, however, the complexity and size of the wavefront involves both terms, and a scattering (or frequency-dependent) term (related to stratigraphic effects) would tend to lower the overall Q_{ef} (relationship is $1/Q$) or increase the damping. Campbell also theorizes that lab (RC) tests also measure Q_{in} when carried out at very small strains and low ($<100\text{Hz}$) frequencies, and produces damping values lower than we would expect to see in a large soil mass from earthquake data. High frequency lab tests, (i.e. kHz range) would more likely estimate a Q_{ef} because scattering would occur in the microfractures of the specimen.

The elevated Q 's from the amplitude and dispersion experiments reflect this observation – that Q_{in} is the result of these downhole interval tests. However, as the soil mass is quite homogenous, I conclude that the within these zones of very elevated Q_{in} , the contribution from scatter, Q_{sc} , may not be significant in the 10-70Hz range.

10.0 CONCLUSIONS

In Chapter 3, four goals were identified for the damping studies carried out in this research. These included: (1) whether *in situ* Q is constant or varies with frequency and soil properties, (2) whether a monofrequency *in situ* approach reduces uncertainty in Q estimation versus traditional broadband techniques, (3) whether velocity dispersion is occurring *in situ* within the seismic exploration band (10-100Hz), and (4) how field measurements of Q relate to lab measurements of damping ratio. Each item is addressed in the summary below.

(1) Variation of Q in Homogenous Muds: It appears that the very high *in situ* Q_{in} 's measured in the homogenous muds are largely a product of low strain motion acting on porous, saturated, fine grained homogenous soils (i.e. low friction between grains). At depths greater than 30m, the downhole testing produced elevated Q 's (100+). Within the intervals tested at both the John Shaw Rd and Anderson Rd boreholes, the soils are known to be relatively massive, saturated, and fine grained (98% of grain sizes < 0.004mm (silt)) with porosities of approximately 0.5. An interval- Q between the geophones placed 15m to 60m apart in both boreholes indicated that the relative loss (ratio) between the two amplitudes was fairly consistent across the frequency band of 10 to 100Hz. This led to very high Q 's with the spectral ratio approach which appeared to be only weakly frequency dependent, if at all. These results were in reasonable agreement with time domain pulse broadening measurements, which revealed very little loss of shear wave frequency content over the entire borehole. In reality, the interval Q measured is most likely a Q_{in} , a non-frequency-dependant term which makes up part of the effective Q (Q_{ef}). A

scattering term, Q_{sc} , is the second component of Q_{ef} , and is typically not measured when using surface shear wave techniques. This theory may, in part, explain why seismic methods predict Q 's which are elevated when compared to the Q 's predicted from the analysis of earthquake data. In addition, marine clays have been theorized to be low-loss materials if elevated porosity in the soil prevents significant particle friction during induced motion. This approach predicts that a marine silty-clay of high porosity would typically be a low-loss material.

The 20- 30m range (JSR-1) within a zone of low velocity and high porosity soils, did reveal a small decrease in Q values at frequencies greater than 70Hz. Comparison with borehole geological/geophysical logs indicates this effect is likely a result of seismic wave interaction with stratigraphic heterogeneities.

(2) Monofrequency approach reduces uncertainty in Q : Field research into a new approach for spectral ratio calculation has produced a technique using monochromatic signals which greatly reduces the scatter encountered in the traditional broadband approach. This technique allows the interpreter to observe the relative attenuation of a single frequency signal between two geophones, thereby allowing better conclusions to be drawn about the frequency dependence of the Q factor, and about the overall shape of the amplitude reduction with frequency over the frequency band tested. Figure 10.1 presents a comparison of the spectral ratio technique using broadband and monofrequency signals in BH-GSC-JSR-01. In the traditional spectral ratio approach, taking ratios in frequency bands where little energy exists produces erratic or scattered values, as very small numbers are being

divided by very small numbers. With the monofrequency approach, discrete frequency bands are transmitted into the soil, permitting us to observe the relative amplitude decay between two downhole tools, even though the measured amplitudes are decreasing with increasing frequency. The resolution of the slope is clearer across the entire frequency band, and the approach is able to provide information where broadband signals lose most of their amplitude content.

Figure 10.1. Improved frequency testing range with the monofrequency spectral ratio approach. The top panel shows the spectral amplitudes of horizontal shear wave (SH) energy reaching the 30m and 50m geophones. The lower panel shows the natural logarithm of the spectral ratios for the broadband and monochromatic approaches. The monochromatic approach produces less scatter in the data points due to the discrete energy pulses at narrow frequency ranges. The uncertainty in slope is significantly reduced, and observations on the change of relative amplitude reduction vs. frequency can be made more confidently.

(3) Results – Dispersion: In the intervals tested, the shear wave velocity variation across the 10-120Hz range was quite small (i.e. $\sim \pm 2\text{m/s}$). Given these measurements, Kjartansson's approach of Q estimation using velocity dispersion generally agrees with values obtained from the spectral ratio approach in low-loss soft soils for estimates of Q_{in} . The best agreement between the spectral ratio and dispersion approaches occurred in the deeper intervals, where the soil was most homogeneous and the variation of the velocities was lowest. In the shallower, lower velocity soils ($z < 30\text{m}$), the mismatch is representative of the travel times having a larger standard deviation. Although the amplitude approach tended to produce higher Q's (or lower damping) than the dispersion approach (based on the error ranges used for the upper and lower velocities ($V_2(f)$ and $V_1(f)$)), the similarity between the measurements suggests that this approach may be a valid one for these low loss soils, *for the frequency independent portion of the Q measurement – Q_{in} .*

(4) Field versus Lab Testing: During the lab component of the research, three specimens from the JSR borehole were tested in the resonant column (RC) apparatus at the University of Waterloo. A newly developed technique called non-resonant (NR) testing was also carried out, which evaluates the frequency dependence of damping and shear modulus at low strains. The RC damping ratios ranged from 1.23-1.85% (Q of 27 – 41) at strains of $< 10^{-5}\%$. However, the velocities were lower than those measured in the field, suggesting that the specimens may have been disturbed, mimicking the effect of a higher strain. Alternatively, differences in strains between $10^{-6}\%$ and $10^{-5}\%$ can produce differences in damping response, as the lab results were collected in the viscoelastic non-linear range. All

these explanations may contribute to the discrepancies seen between the lab and field results. The variation in the NR results with frequency was more significant than what was observed in the field, suggesting sample disturbance may have affected the results. For future testing of Champlain Sea muds in the lab, it is recommended that block samples be used.

Summary: The advantages of any field approach are, of course, the measurement of bulk soil properties in a natural state of consolidation. Once removed from the ground, a sample may no longer represent the general characteristics of the material on a large scale, and if disturbed, lab test results will be affected. The challenges of a field approach are the heterogeneities of the soil, the limitations of the source, and the analysis of records which are often contaminated with noise/interference. In this research, uncertainty caused by these effects was reduced by the use of a monofrequency source, the repeated testing of both boreholes, switching the position of the top and bottom tools, changing the orientation of the Minivib mass, and varying data collection parameters.

A primary advantage of lab testing for geotechnical earthquake engineers is the broad variety of strain conditions which can be recreated with the testing apparatus. Lab approaches have been successful in determining low to high strain soil properties in the earthquake range of frequencies. However, the testing of fine grained soils (clays and silts) is challenging both in terms of sample extraction and testing. For Champlain Sea silts, this can be significant if the soil sensitivity is greater than 20, as some sample disturbance may be unavoidable.

The advantage of *in situ* testing is that different intervals can be bracketed to test for varying levels of Q_{in} . In a massive fine grained soil such as those tested here, the Q_{in} 's measured may represent the low end members of the actual damping which could occur during weak earthquake motions, providing conservative damping estimates.

For all these reasons, the best approach is a combined one. All techniques have their limitations, but each one improves our understanding of soil conditions. Lab measurements can predict soil behaviors over a variety of strain conditions, while *in situ* testing can bracket soil intervals and test the variation in Q (Q_{in}) over the entire borehole. A spectral soil to rock (SSR) approach using earthquake data will provide an idea of attenuation caused by a complex waveform, and the level of scattering (Q_{sc}) which may act to reduce the overall intrinsic Q measured with surface seismic and lab techniques.

It is reasonable to expect that Q 's (or damping) measured with different testing methods (lab, field) at different scales will be different. Applications for each of these approaches will also be different. A geotechnical engineer may only be interested in Q_{in} for low strain vibrating equipment. In this case, downhole testing methods would provide a realistic Q for the case where waveforms generated at surface may be relatively simple. In earthquake engineering, a downhole seismic Q_{in} should be compared with a lab Q_{in} and an earthquake Q_{ef} to get a sense of scale, stratigraphic effects (scattering) and ranges of possible Q or damping.

11.0 FUTURE RESEARCH

Carrying on from these results, areas of future research could include:

1. Non-resonant testing in the lab: Continued lab testing with the NR method.

The potential to understand the dependence of damping and shear wave velocity on frequency at a variety of strains would be very useful to the geotechnical community. The samples must be collected with a stainless steel thin-walled wide diameter Shelby tube, or better, as block samples to avoid soil disturbance and produce credible results for the geotechnical community.

2. Large strain testing in the lab: using wide diameter Shelby tube/block samples, continue with larger strain testing (triaxial compression) to form damping response and modulus reduction curves from 10^{-4} to 10^{-1} .
3. Monofrequency testing in situ: Additional work on the monofrequency approach at boundaries between reflectors to estimate Q_{in} .
4. Q_{sv} : Study the contribution of Q_{sv} to total attenuation by moving the Minivib to greater distances from the borehole and using a shear vertical signal.
5. Earthquake data analysis: Additional research on Q estimation from earthquakes at JSR seismic stations would provide interesting comparison between simple and complex wave fronts – is this a reasonable way to estimate bulk scattering effects i.e $Q_{in} - Q_{ef} = Q_{sc}$?

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APPENDIX A

SPATIAL CORRELATION OF SHEAR WAVE VELOCITY DEPTH FUNCTIONS WITHIN THE OTTAWA, ON AREA

Introduction

Earthquake-induced horizontal shearing forces at ground surface are based, in part, on the geotechnical and geophysical properties of the near surface at a particular building site. Observations from large earthquakes world-wide have concluded that the most damaged structures tended to be located in areas where soils of low shear wave velocity (and thus stiffness) were present in the near surface (Anderson et al. 1986, 1996). Current 2005 National Building Code of Canada (2005 NBCC) guidelines recommend measurements of average shear wave velocity to a depth of 30 m (V_{s30}) as a primary parameter to estimate seismic shaking levels. A 2005 NBCC seismic site category (A through E) can then be given to a site based on its V_{s30} value (NRC 2005). Mapping of shear wave velocity in the near surface (soil and rock) is therefore of interest in hazard mapping programs.

Almost all natural soils are heterogeneous in nature and are highly variable in their properties at some scale. In the field of geotechnical engineering, high factors of safety have provided a method of handling uncertainty in soil property variation. However, numerous basic geostatistical tools exist to investigate the inherent lateral and vertical spatial variability of soil properties due to variation in deposition and stress history. These have been applied to study spatial correlation of shear wave velocities in soft soils from seismic cone penetrometers (SCPT) and spectral analysis of surface wave (SASW) surveys (Thompson et al. 2007), tip resistance, q , from cone penetrometer (CPT) tests (Fenton 1999, Elkateb et al. 2003b), and undrained shear strength (Soulie et al. 1990). The results of the study by Elkateb were then used to assess the effect of soil variability on earthquake induced ground response. When combined with probabilistic (stochastic) simulation, these approaches have proven effective in risk analysis.

In the Ottawa area, a seismic site category map has been produced using a combination of 686 shear wave velocity measurement sites across the city and

stratigraphic information derived from a database of 20,000 water wells and engineering boreholes (Bélanger 2008) within the study area. Of the 686 shear wave sites, 158 sites were identified as having sufficient seismic reflection data points in the thick soft soil overlying bedrock to clearly define the vertical (1-D) variation of velocity and develop average shear wave velocity- depth ($V_{s_{av}}\text{-depth}$) equations in these locations. This appendix will describe the geostatistical methods used to laterally interpolate the parameters of the 158 $V_{s_{av}}\text{-depth}$ equations to untested locations, which allowed us to make improved $V_{s_{30}}$ estimates in locations where only the thickness of the sediment was known.

Ottawa Area Geohazard Studies

In Ottawa, ON, a $V_{s_{30}}$ 'microzonation' or seismic site category map was developed as part of a demonstration study jointly carried out by the Geological Survey of Canada (GSC) and Carleton University (Hunter et al. 2009). Shear wave velocity data were collected at 686 surface seismic sites and 16 borehole locations (downhole shear wave logs). In areas of complex bedrock topography, 25km of high-resolution seismic 'landstreamer' data was also collected (Figure A1). This resulted in a database composed of thousands of shear wave velocity data points which defined the variation of shear wave velocity with depth within the primary material types (bedrock, till, soft clayey silt) directly below each geophysical site.

Figure A1. Location of seismic test sites within the city of Ottawa, collected between 2005 and 2008.

Velocity-depth equations in soft soils

As part of this demonstration project, vertical variation of shear wave velocity in the soft post-glacial soils was investigated, as this material is known to have a significant impact on the level of shaking felt at a site during an earthquake. At 158 sites where the soil is thickest, average shear wave velocity is well defined by site-specific linear average velocity-depth ($V_{s_{av}}\text{-depth}$) equations (Figure A2). In addition, using all the city-wide velocity data points measured in the soil, a city-wide $V_{s_{av}}\text{-depth}$ equation was also developed which could be applied anywhere in the city to predict $V_{s_{av}}$ in the soil column, given a thickness. Details of the development of these equations can be found in Hunter et al. (2010).

Figure A2. Locations of borehole shear wave surveys and reflection sites with velocity-depth equations developed for post-glacial soils. Equations fall within areas of thick soft-soil deposits where reflections are generally the most prevalent features on the seismic records.

Most commonly, bedrock and till were present in the top 30m of the surface. V_{s30} was then estimated by calculating the travel-time weighted shear wave velocity through all three units. V_s in bedrock and till were reasonably given one interval velocity ($\pm 1\sigma$) as interpreted from seismic refraction data collected during this program. However, the significant positive velocity gradient with depth in the soft soil required that depth functions be defined to accurately predict V_{sav} with increasing depth.

Paleozoic/Precambrian bedrock in the Ottawa area is overlain by a thin layer of till, averaging 3.85m in thickness. Overlying the till is a thick glaciomarine-marine sequence which was deposited within the Champlain Sea approximately 11,000 yrs BP. The relatively homogenous glaciomarine clayey-silt is locally known as 'Leda Clay', and although it averages 26m in thickness at our test sites, we have measured

thicknesses of up to 150m within the city. The city-wide V_{sav} -depth equation is given by

$$V_{sav} = 123.86 + 0.88z \pm 20.30 \text{ m/s} \quad (10\text{m} < z < 100\text{m}) \quad (\text{A1})$$

where V_{sav} is the average velocity down to depth z within the soft post-glacial material $\pm 1\sigma$. The site-specific V_{sav} -depth equations are of the same form, but have different slope, intercept, and error parameters (Hunter et al. 2010).

The next step involved developing a way to use the known parameters at the 158 seismic sites to predict a unique set of new parameters for locations where boreholes existed, based on spatial location. The linear form the equations indicates that calculations of V_{sav} in soft soil across the city is controlled by the thickness distribution of z . Having glaciomarine soil thicknesses from the local database of 20,000 boreholes meant that we could examine the distribution of z across the city to determine how extensive our search range(s) should be for interpolation of seismic parameters to borehole locations. Borehole coverage is shown in Figure A3.

Figure A3. Map showing the locations of ~20,000 water well and engineering boreholes in the Ottawa area (after Bélanger 2008).

One approach was to simply assign to a borehole location the parameters of the nearest geophysical site to it within a reasonable search radius. Any sites outside of this search area would receive the city-wide velocity depth parameters from equation A1. However, this would be based only on spatial location and not take into account any trends in the bedrock or surficial geology. It seemed more accurate to consider an approach where the parameters assigned to a borehole location could depend on all the sites within a chosen search radius, and where the shape of the search area may be influenced by geological features.

To use this approach, however, required consideration of several factors: How big of a search radius to use around each borehole? What kind of distance-weighting function should be used? Can slopes and intercepts be averaged for a series of equations? How to best incorporate geological trends? Answering these questions required knowledge of geological conditions in Ottawa, combined with an examination of the thickness dataset to better understand trends which exist within the map area.

Determinate trends in soil thickness

Ottawa sits in a topographic depression called the Ottawa Embayment. The Embayment is linked to the ancient St Lawrence rift system, and lies within the Ottawa-Bonnechere Graben, which is approximately 55km in width, stretching between Ottawa and Montreal (Figure A4). The Graben is dissected by prominent normal faults oriented NW-SE which affect the bedrock topography of the Ottawa area (Rimando 2005).

Figure A4. Regional map of the Ottawa-Bonnechere Graben (grey area in upper map). Enlarged area shows City of Ottawa boundary with major fault structures. Regional structures after Wilson (1946). Terranes and their boundaries after Davidson (1986). Figure after Rimando (2005).

These formations are cut by several generations of brittle faults whose relative ages have been inferred from field studies of key outcrops within the city of Ottawa. From these studies, three periods of deformation (D1, D2, D3) and faulting of Phanerozoic age (542Ma – present) have been defined. Paleo-stress field calculations carried out for these three periods were made through studies of the relative ages of local faults and kinematic analyses of their surface features. From these determinations, relationships between periods of faulting and tectonic events occurring along the evolving eastern margin of North America were made (Rimando 2005).

These three periods of deformation created a successive series of faults whose predominant orientations can be seen in the bedrock geology map of Ottawa (Bélanger 2008). Of interest to microzonation activities is the bedrock topography and presence of fault controlled basins or ridges which have led to bedrock depressions infilled with soils of late- and post-glacial age. In Ottawa, these features have the greatest control over the distribution of low shear wave velocities. Superimposing a map of fault boundaries over a map of Holocene sediment thickness indicates deep basins (>80m) are present where normal large faults exist (Gloucester, Hazeldean, etc.).

Surficial features are also visible in the Holocene thickness map, although their regional effect on V_{s30} is not as significant. The Rideau River bed shows a thickness range of up to 15-20m which lowers the V_{s30} values along the river valley due to deposits of soft soil. Glacial sediments deposited as eskers which are buried or exposed raise shear wave velocities, which have been measured between 200m/s to 900m/s (average 580m/s).

A 3D map showing 'inverted' soft-soil thickness from the borehole database was used to better identify trends (Figure A5). Thickest soil accumulations are evident in areas of significant faulting.

Figure A5. 3-D image of the Holocene soil thickness from borehole database logs. On right are marked the locations of features running through the dataset. The inferred fault-derived features are typically 5 times deeper than the features caused by surficial phenomenon (rivers, glacial deposits, etc).

Based on knowledge of the predominant faulting orientations in the city, combined with the determinate trends observed in Figure A3, the city was divided into 8 zones (Figure A6, Table A1) which were examined separately to take into account the varying strike of the controlling feature in that region. This was important because sites which fell on the flanks of a fault where bedrock was shallowing (or deepening) over short distances (order of several hundred meters) showed that changes in V_{sav} -depth parameters could be expected, and therefore, should be honoured in spatial analysis of soil thickness (Figure A7). These trend areas encompassed the thick pockets of Leda Clay in the Carp, Constance, Orleans, Ramsayville and Rang St. Thomas areas. Only 8 V_{sav} -depth functions were associated with seismic sites outside these areas, and consequently, almost all of the borehole locations in the Central and Rideau areas received the city-wide V_{sav} -depth equation to calculate soft soil variation with depth. In the Central area, thickness was more or less randomly distributed, and within the Rideau River areas, moderate thicknesses of Leda Clay followed the trend of the river bed which was up to 1km wide.

Figure A6. Figure (a) shows the areas of the city where determinate trends are based on tectonic and surficial features. Figure (b) shows the interpreted fault outlines from the regional bedrock geology map (Belanger 2008) superimposed over the soft soil thickness map. Major fault boundaries align with deep basins filled with glaciomarine sediments.

Area	Predominant Feature	Strike	Controlling Feature
Carp	Fault	110-	Bedrock -
Constance	Fault	120-	Bedrock -
Central	none	-	-
Ramsayville	Fault	105-	Bedrock -
Orleans	Fault	070-	Bedrock -
Rang St Thomas	Fault?	010-	Bedrock -
Rideau N-S	River bed	approx	Surficial - river
Rideau E-W	River bed	045-	Surficial - river

Table A1 – Predominant trends identified in the Ottawa area which impact the thickness of Holocene soils across the region.

Figure A7. V_{sav} from downhole shear wave surveys plotted with V_{sav} from nearest surface seismic sites within a 2km radius of the logged borehole. Figure (a) shows shear wave velocities versus depth in an area where there is little variation in bedrock topography. Figure (b) shows how shear wave velocities in soft soil can vary in an area of significant bedrock fluctuation (here, flanks of a fault). Figure (b) shows how bedrock topography plays a role in distribution of shear wave velocities.

Working within these trend boundaries, interpolated V_{sav} -depth equations unique to each borehole location were developed which would be based on the distance to, and distribution of, available geophysical sites around it, as well as incorporating some element of the geological trend within which it fell (Figure A8). The task remained to characterize the spatial variability of the soil thickness within each trend area.

Figure A8. Creating a new velocity-depth function at each borehole was based on the parameters of the geophysical sites found within its search area.

Characterization of Soil Thickness

As identified by Elkateb et al. (2003b), the main elements of spatial variability related to this study of lateral thickness variability are: (1) statistical characteristics (mean, standard deviation, distribution), (2) spatial correlation structure describing changes from one location to another, and (3) the limit of spatial continuity beyond which the data are no longer correlated. Considering all these elements, a radius of influence around each borehole was chosen by examining semivariograms of glaciomarine soil thickness within each area. Similar approaches have been taken by Elkateb et al. (2003b), Thompson et al. (2007), and Soulie et al. (1990).

Statistical Characteristics

Throughout the remainder of the text, the results in the Carp area will be used to exemplify the studies carried out in each area. Figure A9 shows the results of the initial statistical thickness calculations for the Carp area. The block contains 4384 boreholes from the database, indicating a mean thickness of 10.12m and a standard

deviation of 11.79m. The distribution of soft soil thickness in this area was non-normal, which has implications which will be discussed in the upcoming section. Although the mean indicates a fairly shallow average thickness for the soft soil, these values are derived from water wells, and are therefore preferentially drilled where groundwater can be found in near surface bedrock aquifers.

Figure A9. Basic statistical characterization of the soil thickness data in the Carp fault trend area.

A necessary condition for the semivariogram modeling is stationarity, which requires that the geostatistical values of mean and variance do not change from location to location. We know, however, from observation that a strong WNW trend exists in the thickness data. To use thickness as a random variable and meet the stationarity condition, the data must be detrended. The steps followed for detrending of the data are outlined in the following section.

Geostatistical Characteristics of the Data

Choosing a radius of influence around each borehole was accomplished by examining semivariograms of glaciomarine soil thickness from each trend area. The semivariogram model is a mathematical representation of spatial variability which is used to predict the value of a variable at locations where data do not exist. The underlying principle is that two observations closer together are more similar than two observations further apart (Clark 1979). The basic equation of the semivariogram is

$$\gamma(h) = \frac{1}{2n} \sum ((g(x) - g(x+h)) + \dots)^2 \quad (A2)$$

where n is the number of sample pairs, h is the distance between a pair of samples, and $g(x)$ is the value of the sample at a particular position. Pairs can be located in any direction: up or down, left or right, or at an angle from one another. In order to determine how values change with distance in a dataset, the more directions and pairs which are sampled, the better. These values can be plotted on a graph with the distance between pairs plotted on the x axis, and the value of the semivariogram on the y axis. When samples are located close together, the difference in values between each location is generally small, but as the points get further apart, the difference grows until, at some distance, the values are independent of one another and the sill is reached. It is at this point that the radius of influence (or range) can be measured from the x axis.

A semivariogram represented in this way is called a 'cloud' semivariogram and is a representation of all the pairs of points measured in the dataset (Figure A10). In practice, even a modest sized dataset has a very large number of data pair combinations – n factorial(!). Pairs can become so numerous that the plot is unintelligible and relationships cannot be made. To overcome this, the pairs can be grouped into distance intervals (or lags) and the values within each interval are averaged (Figure A11). The result is an 'ordinary' semivariogram (Figure A12) which is generally much simpler to model.

Figure A10. Sample cloud semivariogram from the Ottawa dataset. Each dot on the graph represents half the difference between two points, separated by a certain distance. In most cases, a data set contains too much data to be represented meaningfully on a single semivariogram, so an ordinary semivariogram is often used to model the data.

Figure A11. Ordinary semivariograms and lag distance. (a,b) Lag distances are specified and the variances between pairs are averaged within each lag. Figures (a) and (b) show how data points are averaged over an entire 360° and the meaning of a lag distance. Figure (c) shows how looking at semivariogram with a window surrounding an azimuth can help to detect anisotropy within a dataset.

Figure A12 shows the basic elements of the ordinary semivariogram. When a semivariogram is extrapolated back to zero distance, it may not approach zero

variance. This is due to randomness or errors in the data at distances closer than the sample spacing, and is called the nugget (Houlding 2000). The nugget is an upward shift, and when added with the partial sill creates the sill. Variance is the mean squared deviation of each value from the mean value of the sample points. In a tightly constrained dataset, this value is small. The model curve is the best fit mathematical function to the dataset and the most common fit types are linear, spherical, and exponential, although many other types are possible (logarithmic, wave, power, quadratic, nugget, etc). Fitting the right model to the data is often a matter of trial and error, and often requires a combination of two (or more) model types (i.e. nugget and spherical are used in Figure A12) (Clark 1979).

Figure A12. Elements of a semivariogram. The semivariogram shows the detrended soil thickness data from the Carp area.

Semivariograms have a few assumptions which must be satisfied: the first is that the samples are normally distributed, and the second requires that any spatial trend present in the data is removed prior to analysis. The semivariogram relationship is also finite, meaning that at a great distance apart, pairs may begin to exhibit a recurrence in variance – even after an apparent sill has been reached (Houlding

2000). Therefore, in reality, a semivariogram's information can only be extended so far before the information becomes meaningless. In practice, it is rare for the lag distance to pass about half the total sampled extent (Clark 1979).

Due to the presence of bedrock and surficial determinate trends, thickness values will change more quickly in one direction than another. However, for a model to be fit to a semivariogram and the radius of influence (range) determined, the data must be isotropic, or equal in all directions. Therefore the trend must be removed before the range of the semivariogram can be determined.

Figure A13. Cloud semivariogram showing the effect of the determinate trend (fault) (a) and the residual semivariogram after the trend is removed using a 2nd order polynomial.

Figure A13 presents post-glacial thickness from the Carp area. The cloud semivariogram on the left is interpreted as having a significant trend in the values oriented in the NW-SE direction, shown by the parabolic upturn in the graph. To

produce a meaningful semivariogram, the trend must be removed and the graph must be recalculated using the remaining 'residual' points. To proceed with geostatistical modelling, the trend must first be identified in the sample data. The cloud semivariogram on the right, produced with ESRI Geostatistical Analyst software, shows the data with the NW-SE trend removed and the sill of the semivariogram levelling out.

Removing the trend in the data first requires modelling it with either a constant, first, second, or third order polynomial. The constant surface would remove a flat, horizontal surface from the dataset, and a first order polynomial would remove a flat but inclined surface from the dataset. The Ramsayville area is an example of a first order polynomial trend (Figure A14). Second and third order polynomial surfaces would remove more complex trends which rise and fall along a particular axis. The Carp area is an example of an area which was detrended with a second order polynomial. Applying a constant surface or first order polynomial to a trend with ridges and bumps would not effectively remove the trend and leave a significant portion of it behind. The Geostatistical package allows a user to model different polynomial functions to the data, determine which one will best recreate the trend, and then remove it. Models can be evaluated by comparing which polynomial fit produced the lowest root-mean square error. Often, the higher order polynomials with more curves will produce a more realistic fit for complex geological conditions, but also produce edge effects from the increasing exponents (Figure A15). A trade off exists between a better fit and edge effects. Once the trend has been identified, modelled, and removed, the semivariogram can then be plotted without the trend and the minimum radius of influence (or range) of the residuals can be determined. The semivariogram characteristics of the detrended soil thickness data, along with the order of the polynomial used for trend removal can be found in Table A2.

Figure A14. 3-D representation of the linear trend in the Ramsayville area. This type of trend is best removed with a first order polynomial.

Figure A15 (a) 2nd and (b) 3rd order polynomials fit to the data to model the depth of the Holocene soil along the Carp fault prior to removing the trend. While the 3rd order polynomial modestly improves the definition of the feature, the edge effects (red grid cells indicated by arrows) are more prominent around the area circumference.

Area	Boreholes (Count)	Mean (m)	Standard Deviation (m)	Polynomial Order for Trend removal	Model Type	Lag Width (m)	# Lags (m)	Sill (m ²)	Nugget Effect	Horizontal Range (m)
Carp	4384	0	10.25	2	Spherical	500	50	86	1	4700
Constance	1876	0	11.53	2	Spherical	500	25	136	15	3500
Central	11709	3.99	6.16	No detrend	Spherical	480	25	38	11	3000
Ramsayville	961	0	10.02	1	Spherical	400	25	72	11	2300
Orleans	2018	18.93	17.3	No detrend	Spherical	500	50	250	50	1500
Rang St Thomas	315	0	10.14	2	Spherical	500	50	110	12	1500
Rideau N-S	2678	0	7.1	2	Spherical	100	50	48	20	1600
Rideau E-W	315	0	6.58	2	Exponential	140	25	52	19	1500

Table A2. Statistical properties of detrended thickness data for different trend areas within Ottawa.

The Central and Orleans areas were not detrended. The nature of the Central area was its lack of large scale determinate trends and therefore it was not detrended. The Orleans block was quite complex in that it contained several steep fault scarps (as determined by GSC landstreamer profiling) infilled with soft soil and therefore did not follow a simple pattern of spatial variation with distance when detrended. As a result, the thickness data in this area produced a semivariogram which rose and fell quite sharply and didn't reach a sill. This also caused such a large nugget effect. (Where the nugget variance is small relative to the sill variance, high correlations exist over short distances.) No type of detrending would be able to remove these features without zooming in on each one at a much smaller scale. As we are interested in the range of the smallest feature for these purposes, the first sill reached was chosen as the horizontal range.

Interpolating Velocity-depth Equation Parameters

The next step involved calculating a new set of linear parameters (slope, intercept, error) for borehole locations by interpolating the parameters of the V_{sav} -depth equations which fell within the search radii defined for the trend area. One approach was to calculate the distance to all the geophysical sites within a borehole's search radius and assign a weight factor to each site based on a $1/r^n$ approach. This weight factor would be assigned to all the reflections which produced that geophysical site's linear equation, and the same process would be repeated for the next geophysical site falling within the borehole's search radius, until all the reflections from all the sites in that borehole's search radius had been given weights. A single least-squares regression was fit through all the weighted reflection points, with the nearest reflections to the borehole having the highest weights. The parameters of the regression line were assigned to the borehole, and the new V_{sav} -depth equation for that borehole was determined. This approach would require stepping through all the boreholes in the database in an automated fashion, and was termed the 'weighted reflection' approach.

A second approach was considered which involved creating three separate grids of the slope, intercept, and error parameters for all the geophysical sites within each trend area, and intersecting the borehole co-ordinates through the three grids to extract the parameters (slope, intercept, error) needed for the new velocity depth equations at each borehole location. This was termed the 'gridded parameter' approach. A deterministic gridding method was desired for this interpolation, because it would preserve the parameters at each geophysical site.

First, however, it needed to be proven that the second approach of interpolating slopes, intercepts, and errors would produce comparable results to the weighted-velocity approach. The 'gridded parameter' method and the 'weighted reflection' method were computed for 18 boreholes (12% of the geophysical site data) and the resulting equations from each method are shown in Table A3. As the greatest depth of interest for the maps was 30m, 30 was substituted into the equations for the depth parameter, z . Regardless of whether a $1/r$ or $1/r^2$ weighting factor was used, it was determined that the average velocities were consistent within each other's errors 100% of the time. The $1/r^2$ showed a smaller difference between the two methods (average of 1.1% difference with $w=1/r^2$, vs. 2.5% with $1/r$). Once established that the 'gridded' method produced comparable velocities to the 'weighted reflection' method, final grids were produced of slope, intercept, and error within each trend area using the determined search radius for that area.

						at 30m		
Water Well #	Trend Area	$1/r^n$	# sites veraged	Gridded Parameters	Weighted Reflections	$V_{\text{Save parameters}}$	$V_{\text{Sweighted refl.}}$	difference (%)
15000	Carp	2	10	$1.07z + 88.19$	$0.99z + 89.2 + /-$	120.2	118.9	1.2
15314	Carp	2	12	$1.25z + 92.65 + /-$	$0.94z + 98.31$	130.1	126.5	2.9
15287	Carp	2	12	$1.38z + 100.07$	$1.01z + 108.03$	141.4	140.1	1.0
15200	Carp	2	3	$1.21z + 126.97$	$0.92z + 136.38$	163.2	163.9	0.4
15197	Carp	2	4	$1.38z + 147.55$	$1.10z + 131.79$	163.0	164.7	1.0
15102	Carp	2	11	$1.48z + 112.12$	$1.54z + 112.87$	156.5	159.0	1.6
15062	Carp	2	8	$1.57z + 108.24$	$1.75z + 106.07$	155.3	158.5	2.0
15136	Carp	2	10	$1.49z + 109.27$	$1.66z + 108.10$	153.9	157.9	2.5
UGE0	Centra	2	5	$2.09z + 105.51$	$2.15z + 103.21$	168.2	167.7	0.3
UGE0	Centra	2	2	$1.63z + 128.97$	$1.26z + 136.86$	177.8	174.6	1.8
15230	Const	2	5	$1.90z + 108.62$	$1.71z + 114.50$	165.6	165.8	0.1
15286	Const	2	2	$1.28z + 140.91$	$1.14z + 144.81$	179.3	179.0	0.2
15014	Orlea	2	5	$1.32z + 127.40$	$1.42z + 123.54$	167.0	166.1	0.5
15014	Orlea	2	6	$1.19z + 114.24$	$1.41z + 110.60$	149.9	152.9	1.9
15261	Orlea	2	4	$1.19z + 133.71$	$0.80z + 145.64$	169.4	169.6	0.1
15015	Rams	2	3	$1.40z + 87.95$	$1.28z + 92.41$	129.9	130.8	0.7
15169	Rams	2	2	$1.70z + 90.88 + /-$	$1.62z + 92.59$	141.8	141.1	0.5
15015	Rams	2	4	$1.25z + 92.72 + /-$	$1.22z + 92.83$	130.2	129.4	0.6

Ave. 1.1

Table A3. 18 sites evenly distributed across the city were chosen to compare the two approaches for computing a linear velocity-depth function: weighted reflections vs. gridded parameters. The results show that both methods produce velocities at 30m depth within an average of 1.1% of each other.

During the gridding, diagnostic indicators were produced to evaluate 'goodness of fit' of the data to the interpolated surface. The cross-validation technique involves removing a data point from the dataset, gridding without it, and comparing the interpolated result in its place against the measured value which was removed. This is repeated for the whole dataset, removing one point at a time, and an average root mean square (RMS) value is produced for the entire gridded surface. The model with the lowest RMS is generally the one to keep. Table A4 presents the results of various trials with different interpolation techniques and parameters within the

Carp area. Power values of one and two were selected for the $1/r^n$ IDW weighting calculation gridding, and Kriging was also attempted using an optimized number of lags and lag distance.

Starting with an IDW gridding approach for the slope, intercept, and error grids, a choice of power value, n , had to be made for the weighting factor, $1/r^n$. Although initially starting at $n=2$, it was possible to optimize the power value by having Geostatistical Analyst determine the lowest RMS value for the data. This value was very close to one, making the optimum weighting function $1/r$. Kriging was also attempted with the datasets. Lag distance and number of lags were chosen based on rules of thumb for the technique and adjusted until optimum values were chosen. As a starting point, the number of lags multiplied by the lag distance should be approximately equal to half the distance between the two furthest points, and there should be at least 30 data pairs for any one semivariogram point. For each grid, the RMS error was noted. The parameters chosen for each trend area were those which exhibited the smallest RMS error for all three grids.

Area	Parameter	Gridding Variables						
		Grid Type	IDW	Kriging		Major Axis	Minor Axis	RMS
			Power	Lag size	# lags	(m)	(m)	
Carp	slope	IDW	2			4700	4700	0.282
	slope	IDW	1			4700	4700	0.2718
	slope	OK		2000	20	4700	4700	0.2713
	intercept	IDW	2			4700	4700	19.18
	intercept	IDW	1			4700	4700	17.78
	intercept	OK		2000	20	4700	4700	18.03
	fse	IDW	2			4700	4700	2.553
	fse	IDW	1			4700	4700	2.43
	fse	OK		2000	20	4700	4700	2.435

Table A4. Sample table from Carp area showing the errors generated from various interpolation techniques for the slope, intercept and error grids. IDW has the lowest RMS error, and was used for the gridding. Fse=fit standard error, OK=ordinary Kriging, Opt=optimised power value.

Once the IDW approach was selected, a further refinement was performed. An alternative to the circular search neighbourhood is an elliptical neighbourhood using an IDW approach (as seen in Figure A8). ArcMap's Geostatistical Analyst is capable of searching within an ellipse which would allow for geological trends in the data to be honoured by aligning the major axis of the ellipse along the strike of a trend. In this case, the search radius would become the minor axis, and the major axis would be the minor axis multiplied by a factor greater than one. The optimized factor to calculate the major axis length was 2 for Carp and Ramsayville, and 3 for Orleans where very deep narrow fault-controlled basins are known to exist (Table A5).

	RMS Errors		
Carp	Slope	Intercept	Error
Circular	0.282	19.180	2.553
Elliptical	0.274	17.820	2.410
Orleans	Slope	Intercept	Error
Circular	0.300	18.070	3.885
Elliptical	0.293	16.760	3.720
Ramsay	Slope	Intercept	Error
Circular	0.237	11.570	3.857
Elliptical	0.235	9.990	3.370

Table A5. RMS error comparison for the circular versus elliptical search neighbourhood axes. The elliptical method showed slightly lower RMS error values for grids when the optimized major:minor axes were chosen.

Using an extraction technique, the three grids in each trend area were intersected at each of the borehole locations. These values were transferred to a spreadsheet and became the new velocity-depth parameters in the post-glacial materials with a unique function at each borehole location. In Constance, Central, Rideau N-S and NE-SW, and Rang St. Thomas, too few velocity-depth equations existed to produce grids of slope, intercept and error. In these areas, a circular search neighbourhood (radius in Table A2) was used with a $1/r^2$ weighting function to weight the reflections. Where a borehole did not contain a geophysical site within its search

neighbourhood, the city-wide velocity-depth function was used. The majority of the city's boreholes were too distant from the 158 seismic sites with V_{sav} -depth functions to receive their own unique V_{sav} -depth function. Of the total number of boreholes within the city limits, 27% had a unique velocity-depth function, and 73% received the city-wide velocity depth function.

Conclusions

A study of shear wave velocity variation in the city of Ottawa was undertaken by examining the distribution of soft soil thickness across the city, and the variation of shear wave velocity vs. depth (V_{sav} -depth) equation parameters, as determined from seismic surveys.

Based on the analysis of shear wave velocity data collected across the city, one city-wide V_{sav} -depth equation (Eq. A1) and 158 site-specific V_{sav} -depth equations were developed. In these equations, V_{sav} is a dependant variable of thickness, and therefore the distribution of thickness across the city affects the calculation of V_{s30} . Based on examination of the soil thickness map and knowledge of geologic features, the city was divided into eight areas of differing soil thickness trends. Using geostatistical approaches, semivariograms were developed for each region which defined the spatial range of thickness. These ranges were then used to examine the lateral variation in average shear wave velocity.

While seismic surveys defined the vertical variation in V_{sav} , it was shown that the lateral variation in soft soil velocity can be predicted by creating grids of the V_{sav} -depth equation parameters. When compared to a more direct method of 'weighting reflection' velocities, grids of slope, intercept and error of the linear equation proved to be a useful representation of the variation of V_{sav} -depth equation parameters within a trend area. Where seismic measurements were not made, V_{sav} depth equations could be predicted by intersecting a series of borehole locations through the parameter grids of slope, intercept, and error where the thickness of the soil layer at each borehole was known. V_{sav} down to the base of the soil layer could then be estimated at the new location.

As only 27% of the boreholes within the city fell inside the range of a seismic site(s), the majority of the V_{sav} estimates in the soft soil were made using the city-wide V_{sav} -depth equation. This resulted in larger velocity errors at these locations. However, the standard deviations in the glacial till and bedrock were much larger than those of the soft soil, resulting in larger error ranges when these material types were present in the top 30m.

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APPENDIX B - GEOTECHNICAL RC LAB DATA

The following appendix contains the test results in graphical form for the lab resonant column and non-resonant column, described in Section 3. Each test contains the data recorded prior to testing, and any data collected post test.

System Constants	
Dist. From Center to Accelerometer, R_{acc} (cm)	6.8
Amplification Channel 1	100.00
Amplification Channel 2	3.16
Sensitivity (mV/g)	493.8
Mass polar moment of inertia of the top cap I_0 (kg. m ²)	2.30E-4
Mass polar moment of inertia of the driving plate I_0 (kg. m ²)	6.91E-3
I_0 Total(kg. m ²)	7.14E-3

Sample 1 – 10m

Sample Details	
Diameter of the Sample (cm)	7.47
Height of sample (cm)	14.40
Sample Mass (kg)	0.99
Sample Volume (m ³)	6.30E-04
Sample Density ρ (kg/m ³)	1585.39
Area polar moment of inertia of sample J_p (m ²)	3.05E-06
Mass polar moment of inertia of the sample I (kg. m ²)	6.97E-04

Sample Consolidation at 60kPa

Resonant Frequency f_R

Damping (%)

Sample 2 – 30m

Sample Details	
Diameter of the Sample (cm)	7.19
Height of sample (cm)	15.86
Sample Mass (kg)	1.16
Sample Volume (m ³)	6.45E-04
Sample Density ρ (kg/m ³)	1790.07
Area polar moment of inertia of sample J_p (m ²)	2.64E-06
Mass polar moment of inertia of the sample I_p (kg. m ²)	7.49E-04

Sample Consolidation at 235 kPa

Resonant Frequency f_r

Damping (%)

Sample 3 – 70m

Sample Details	
Diameter of the Sample (cm)	7.31
Height of sample (cm)	17.20
Sample Mass (kg)	1.31
Sample Volume (m ³)	7.21E-04
Sample Density ρ (kg/m ³)	1815.48
Area polar moment of inertia of sample J_p (m ²)	2.79E-06
Mass polar moment of inertia of the sample I (kg. m ²)	8.73E-04

APPENDIX C - FORMULAE FOR NON-RESONANT COLUMN TESTING

Solving for Damping and Vs using Non Resonance (NR) Method

Detailed derivation of the formulae for the NR method can be found in the following texts: Lai et al. 2001, Khan 2007, Khan et al. 2008.

However, a sample of the simplified MathCAD code used to solve for the damping and velocity values with the NR dataset collected for this study can be found below. The three input files for each sample test derived from the experimental data are also presented in this appendix.

An initial input file containing four columns is read into the MathCAD program.

Column <1>: Frequency

Column <2>: Peak of the Input Sine Wave Voltage (Corrected)

Column <3>: Peak of the Output Sine Wave Voltage (Corrected)

Column <4>: Phase lag between the two peaks in radians

Let the transfer function, $TR = \text{Column 2} / \text{Column 3}$

Knowing other parameters:

- **System constants**: mass polar moments of inertia (I_o), magnetic force factor of the coils (B), resistance of the coils (R), radius of the accelerometers (r_a), radius of the magnets (top plate) (r_m), system sensitivity (S), and
- **Sample Parameters**: the area polar moment (J_p), mass polar moment of inertia (I), mass (m), geometry (Length= L , radius= r), and density (ρ) of the soil sample

The complex shear modulus ($G_s^* = G_{\text{real}} + G_{\text{imaginary}}$) can be solved for iteratively as a function of frequency ($\omega = 2\pi f$) using the equation

$$\frac{T_{o(\omega)} e^{-i\theta(\omega)}}{\varphi(\omega)} = \frac{I\omega^2}{\sqrt{\frac{\rho L^2 \omega^2}{G_s^*(\omega)}} \tan \sqrt{\frac{\rho L^2 \omega^2}{G_s^*(\omega)}}} - I_o \omega^2$$

$$\text{where } \frac{T_{o(\omega)}}{\varphi(\omega)} = \frac{\frac{B r_m}{R}}{TR \left[\frac{g}{S \omega^2 r_a} \right]}$$

Once $G_s^*(f)$ is known, $V_s = \sqrt{\frac{G_{\text{real}}}{\rho}}$ and $\xi = \frac{G_{\text{imaginary}}}{G_{\text{real}}}$ can be calculated as a function of f .

Lai and Khan have created a code which will compute graphs of V_s and ξ versus frequency in MathCAD given the four experimental columns 1-4. The following tables show the experimental results from the NR testing, and the results are shown in the text in Figures 5.8-5.10.

Freq: 10m sample	V1 corrected	V2 corrected	Phase (rad)	TR
10	7.9130	1.3887	-0.0088	5.6981
15	8.5495	4.5553	-0.0381	1.8768
20	4.7867	7.6796	-0.0908	0.6233
21	3.8991	8.9989	-0.1033	0.4333
22	2.7079	9.5627	-0.1471	0.2832
22	3.3403	9.7809	-0.1307	0.3415
23	2.2484	10.7875	-0.2014	0.2084
24	1.1362	11.5746	-0.3937	0.0982
25	0.4481	12.6959	-1.8698	0.0353
26	1.4338	13.6487	-2.8421	0.1050
27	2.6079	14.3039	-2.9828	0.1823
28	4.0227	15.8668	-3.0348	0.2535
33	11.3466	21.9023	-3.1027	0.5181
38	19.8752	29.0209	-3.1171	0.6849
38	19.8664	28.9589	-3.1172	0.6860
43	29.6811	37.1460	-3.1277	0.7990
48	41.2108	46.8564	-3.1328	0.8795
53	53.6038	57.2540	-3.1361	0.9362
58	65.4588	66.8060	-3.1386	0.9798
63	82.0991	80.9001	-3.1406	1.0148
68	96.6601	92.7305	-3.1410	1.0424
73	114.5137	106.9480	3.1399	1.0707
78	132.5317	124.4070	3.1240	1.0653
83	153.0672	139.6610	3.1364	1.0960
88	173.4177	156.7910	3.1359	1.1060
93	199.5170	177.4490	3.1344	1.1244

Freq: 30m sample	V1 corrected	V2 corrected	Phase (rad)	TR
15	19.9603	2.6971	-0.0258	7.4006
20	23.7171	4.9901	-0.0371	4.7529
24	36.3662	7.1151	-0.0455	5.1111
25	35.0697	7.7096	-0.0407	4.5488
30	31.1295	11.0680	-0.0716	2.8126
35	24.0333	15.0208	-0.1403	1.6000
36	23.2238	15.8114	-0.1739	1.4688
37	20.7509	16.7980	-0.2292	1.2353
38	18.2811	17.7878	-0.3337	1.0277
39	17.2945	18.7776	-0.5870	0.9210
40	16.3047	19.5682	-1.4114	0.8332
41	17.2945	20.5548	-2.4085	0.8414
42	18.7523	21.5414	-2.7297	0.8705
43	21.2473	22.7273	-2.8510	0.9349
44	23.7171	23.7171	-2.9112	1.0000
45	26.1868	24.7037	-2.9447	1.0600
50	41.0147	30.6330	-2.9949	1.3389
55	63.2456	37.0619	-2.9707	1.7065
60	94.8683	43.9873	-2.8287	2.1567
65	121.5580	51.3870	-2.3103	2.3655
70	132.4362	60.2730	-1.8735	2.1973
80	120.5460	78.0766	-1.5192	1.5439
90	83.0098	99.8015	-0.9707	0.8317

Freq: 70m sample	V1 corrected	V2 corrected	Phase (rad)	TR
16	59.7987	6.9159	-0.0531	8.6465
26	44.4616	18.2811	-0.0848	2.4321
36	52.3673	34.5953	-0.0867	1.5137
44	47.4342	51.3870	-0.0829	0.9231
45	43.4813	55.3399	-0.1007	0.7857
46	42.5010	57.3005	-0.1201	0.7417
56	28.6566	83.9901	-0.2032	0.3412
66	9.8821	116.6248	-0.7712	0.0847
67	7.9057	120.5460	-0.9928	0.0656
68	6.9159	124.5305	-1.2719	0.0555
69	6.7198	128.4517	-1.6043	0.0523
70	7.3112	132.4362	-1.8961	0.0552
71	8.1018	134.3968	-2.1839	0.0603
72	9.6829	140.3419	-2.3900	0.0690
82	32.6031	181.2000	-2.9498	0.1799
92	61.7593	228.1000	-3.0438	0.2708
102	95.8486	281.2000	-3.0825	0.3409
112	134.3968	337.5000	-3.1133	0.3982
122	173.9253	400.0000	-3.1229	0.4348

APPENDIX D – ADDITIONAL DOWNHOLE SHEAR WAVE TESTING

Figure D-1. Shear wave testing results for the upper 100 ft of the well. The data points represent the shear wave velocity (V_s) in ft/s. The line of best fit shows a general increase in V_s with depth, indicating increasing stiffness of the formation.

Figure D-1. Monofrequency spectral ratios from test interval JSR-2. Composite plot of the upper (30m) and lower (50m) spectral amplitudes from monofrequency Minivib signals plotted at 10Hz intervals. The Minivib source was oriented at 45° from magnetic north. In this dataset, the slope is increasing slightly instead of decreasing as is expected. This is interpreted to indicate that the amplitude loss due to intrinsic attenuation across the 10-100Hz frequency band is extremely low and beyond the detection limit of the dataset ($Q > 200$, or damping $< 0.25\%$).

Figure D-2. Monofrequency spectral ratios from test interval JSR-3. Composite plot of the upper (30m) and lower (70m) spectral amplitudes from monofrequency Minivib signals plotted at 10Hz intervals. The Minivib source was oriented at 45° from magnetic north. The slight decrease in the slope indicates there is very little amplitude loss due to intrinsic attenuation across the 20-90Hz frequency band ($Q=199 \pm 86$, or damping= $0.25 \pm 0.11\%$).

Figure D-3. Monofrequency spectral ratios from test interval JSR-4. Composite plot of the upper (30m) and lower (90m) spectral amplitudes from monofrequency Minivib signals plotted at 10Hz intervals. The Minivib source was oriented at 45° from magnetic north. The slight decrease in the slope indicates there is very little amplitude loss due to intrinsic attenuation across the 10-80Hz frequency band ($Q=192\pm83$, or damping= $0.26\pm0.11\%$).