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THE DEVELOPMENT AND EVALUATION
OF A
CROSSHOLE SEISMIC SYSTEM
FOR
CRYSSTALLINE ROCK ENVIRONMENTS

BY

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the requirements for the Degree of
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Abstract

Crosshole seismic techniques appear to be useful in the evaluation of the mechanical properties of inhomogeneous crystalline rock masses, particularly when near surface weathered zones prohibit the use of high resolution surface seismic methods. A project has been undertaken to develop a crosshole seismic system specifically designed to meet the criteria of the Canadian Nuclear Fuel Waste Management Program. Because the rock masses to be studied are at depth, the system was required to operate in environments with high hydrostatic pressures of up to 15 MPa, and was required to be non-destructive, because of the high cost of the boreholes.

The system that has been built utilizes piezoelectric ceramics in a wall-clamped accelerometer geometry as both receiving and transmitting transducers. Although the peak power of the transmitter is limited by the design criteria, the highly controllable nature of the piezoelectric ceramic allows the use of a continuous transmitted signal. The impulse response of the rock mass is then estimated using cross-correlation techniques. In this manner significant noise reduction is obtained, and the detection of low level signals is enhanced. The received signals are digitized and processed in the field using a microcomputer system.

Tests have been conducted which demonstrate that crosshole seismology is capable of evaluating the
mechanical properties of rock masses. P-wave transmissions on paths of up to 300 m and S-wave transmissions on paths of up to 50 m have been recorded, and show that both velocity and amplitude information in the data reflect geological and geophysical effects observed using other methods. The system has also been used to study tube wave phemenena, which can provide information on the shear modulus of the rock in the absence of direct shear arrivals. Although problems exist in accurately determining the position of the probes and in calibrating the amplitude response of the system, valuable information can be collected and interpreted with the present system. Future developments in instrument design and in methods of tomographic reconstruction hold great promise in several fields of application, including mining engineering and rock mass assessment.
Acknowledgements

The work that formed the basis of this thesis was part of a larger project, and was only possible because of the contributions made by the other members of the project team. The supervisor of the project and of this thesis, Dr. G. F. West, provided invaluable contributions to all aspects of the work. His part in the design and construction of the electronics was especially critical to the system development.

Dr. Joe Wong was a co-worker on the project from start to finish. His insights into the interpretation problems were most appreciated. Joe’s practical approach to making the best use of available (and not always fully functional) equipment resulted in important field data that would not otherwise have been collected. B. Polzer also assisted in the field work.

The Geophysics Technical Support Staff contributed to the success of the project. A. Wieckowski (electronics), M. Bloore (computing), and J. Marrs (logistics and bench assembly) all gave freely of their time. The Physics Department machine shop was responsible for the manufacture of the hardware. The advice and skilled work provided by W. Culbert, A. French, R. Carder, D. Curling, and J. Friel is gratefully acknowledged.

The completion of the manuscript was accomplished with assistance from many quarters. Critical readings of the
first draft by Dr. West and N. Bregman were very helpful. Many of the figures were prepared in a professional manner by K. Khan, J. Kostilek, and R. Cunha. Plotting software developed by D. Boerner was useful. I would like to thank Dr. C. H. Chapman and Dr. D. York for their time in evaluating the final draft.

The Faculty, Staff and Students of the University of Toronto Department of Physics should be commended for their work in making the department a excellent home for physics research. A project like the one described would be difficult without the facilities and cooperation that were provided.

The contributions made by my wife, Maria, are difficult to enumerate and impossible to properly acknowledge. Her patience and support during the course of the work were unending. She also spent many hours word processing and proof reading the manuscript.

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# TABLE OF CONTENTS

<table>
<thead>
<tr>
<th>Chapter Title</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>ABSTRACT</td>
<td>i</td>
</tr>
<tr>
<td>ACKNOWLEDGEMENTS</td>
<td>iii</td>
</tr>
<tr>
<td>TABLE OF CONTENTS</td>
<td>v</td>
</tr>
</tbody>
</table>

## CHAPTER ONE: INTRODUCTION

1.1 Introduction  
1.2 Crosshole Seismology  

## CHAPTER TWO: SYSTEM DESIGN AND FABRICATION

2.1 Design Rationale  
2.2 Transmitter  
2.3 Receiver  
2.4 Clamping Mechanism  
2.5 Fabrication Details  

## CHAPTER THREE: SYSTEM EVALUATION

3.1 Test Sites  
3.2 System Performance  
3.3 Travel Time Analysis  
3.4 Tube Wave Analysis  
3.5 Amplitude Variations  

## CHAPTER FOUR: CONCLUSIONS AND RECOMMENDATIONS

4.1 Conclusions  
4.2 Alternative Applications  
4.3 Recommendations  

REFERENCES  

APPENDIX A - ROCK PHYSICS  
APPENDIX B - ELECTRONICS  
APPENDIX C - DASP - DATA ACQUISITION AND SIGNAL PROCESSING
Chapter One: Introduction

1.1 Introduction

In the normal practice of exploration seismology a source of elastic wave energy is placed at one point on the earth's surface and the resulting seismic waves are monitored at other surface locations. As the waves from the source propagate downwards, subsurface structures cause some of the energy to be redirected upwards by reflection, refraction and diffraction. Reflection seismology is the central tool in the search for petroleum in sedimentary environments. In other applications, however, where the targets are smaller, velocities higher, and the geological environment more complex, the method has been much less successful. An important factor is often the presence of a surface overburden layer that is dispersive and attenuative, and that can be unpredictably variable in such important parameters as composition, thickness, and degree of water saturation. Because the seismic energy in a standard survey technique must pass through this zone twice, high frequencies are either attenuated or scattered, and high resolution is impossible. The surface layer problem is even serious in environments with outcropping crystalline rock, where highly fractured zones to depths of fifty metres or more are common.

One objective of the Canadian Nuclear Fuel Waste Management Program (CNFWMP) is to find a suitable site for a high level radioactive waste repository. A detailed
hydrogeological model of the site will be an essential part of the suitability studies, since the mass transport of radioactive materials by the ground water is thought to present the greatest risk to vault integrity. The water flow in crystalline rock masses is controlled by the major faults and fracture systems, which must be located and characterized as part of the evaluation. At present, a number of large crystalline rock bodies are being explored in order to refine both the hydrogeological modeling and the technology for collecting the data required by the models.

The mechanical discontinuities that act as channels for the water flow also affect the propagation of seismic waves. The effects include attenuation, reduction in wave velocity, coherent reflection, and scattering. As a result, these features are a good target for seismic exploration, but will require much higher resolution than is available from standard surface seismic techniques.

A method that has been suggested for collecting high resolution seismic data is to place the seismic instrumentation under the surface layer by lowering the source into one borehole and the receiver into another. High frequencies are no longer scattered or absorbed in the overburden, and amplitude studies are more reliable, since the coupling of the receiver to the rock is more consistent and predictable. This also makes possible many different source-receiver geometries (Figure 1.1), and allows detailed probing of the rock between the holes.
Figure 1.1 Possible geometries for crosshole seismic investigations.
A project to develop a hole-to-hole seismic system for high resolution investigation of crystalline rock masses was undertaken at the University of Toronto Geophysics Laboratory by three workers, G.F. West, J. Wong, and P. Hurley, with the support of Atomic Energy of Canada Limited (AECL) and the Canadian Department of Energy, Mines and Resources. The author was responsible for the design and construction of the transducer sections of the probes, which is the subject for the second chapter of this thesis. This work is discussed with reference to the design rationale for the entire system, and to those details of the complete system that are relevant.

An important part of the development was an analysis of the data to verify the usefulness of the system for investigating rock masses. This was done by comparing results obtained with the system to observations available from other sources. In some cases, comparison with simple theory was possible. This analysis is presented in Chapter Three. Some of the data are also used to evaluate selected aspects of the system performance.

The remaining portion of this chapter provides a review of previous work in crosshole seismology. In the closing chapter, some possible applications of the University of Toronto system outside of the radioactive waste program are given. Appendix A is a detailed review of theoretical and laboratory work on the mechanical properties of crystalline rock.
1.2 Crosshole Seismology

Crosshole seismology was first used to study attenuation, as a means of eliminating the geometrical effects of the near surface, and to avoid any near surface inhomogeneities. Ricker (1953) and McDonal et al. (1958) used explosive sources (typically 0.45 kg dynamite) with wall clamped geophones for receivers. Very clear seismograms were obtained by Ricker (1953) over distances of about 500 metres in a shale for which Q was measured to be about 20 (high loss). McDonal et al. (1958) stress that the measured wavelet changed with successive shots, an indication that borehole damage was occurring. Error in fixing the exact time of the blast (shot time) was not mentioned in either study, probably because high resolution travel times are not required in attenuation measurements.

One of the first reports of crosshole measurements for exploration is by Bois et al. (1972). The experiment was conducted in a sedimentary environment in the frequency range 0 to 160 Hz with wall clamped geophones. Useable seismograms were recorded over straight line paths of up to 900 m between holes of up to 655 m apart. Sources were primacord charges of 0.2 to 0.3 kg. Resolution of the shot time was reported to be +/- 0.1 ms. The measurement of the arrival time of the direct compressional wave was complicated by interference from refracted arrivals caused by large velocity changes (2.5 to 4.4 km/sec) in the target area, with the result that the accuracy was reduced
to $\pm 5$ ms. A major structural feature was identified from the data, as were the major lithological units, using an iterative ray-tracing method of reconstructing the velocity field. The authors concluded that their system was incapable of resolving small velocity inhomogeneities, probably due to the difficult environment tested, and the lack of high frequency sensitivity that was intrinsic to the source and receivers used.

Some systems have been developed specifically for engineering purposes, where the holes are usually shallow. Further, the materials to be tested normally have very high loss; the ranges quoted must be weighed with this in mind. Ballard (1976) presents a system in which pipe instead of the normal cable is used to lower the instruments into the hole. The transmitter consists of a pipe with a mechanism at its lower end which is clamped tightly into the borehole. The pipe, otherwise isolated from the rock, is then connected to a vibrator on the surface. Vibrations are transferred through the pipe to the clamping point where a geophone is located to mark the "shot time". The receiver geophone is also lowered by pipe and clamped in the desired position. Compressional and shear travel times at distances up to 50 metres and at depths up to 30 metres were measured with the aid of an enhancement (signal averaging or stacking) seismograph. The vibrator was controlled with a tone burst generator, with frequencies selected between 100 and 250 Hz. The use of such a system is clearly limited to shallow depths.
Auld (1977) describes another mechanical system designed for engineering applications. The source consists of a plate which is locked into the hole hydraulically, and a guided "hammer" which may be dropped from 0.3 metres above the plate. The receiver is a standard geophone clamped with hydraulic pressure. Since the hammer blows were untimed the authors recommended the use of three holes, in line, permitting the velocity to be calculated from the difference in travel times. Satisfactory results were obtained in holes about 5 metres apart to depths of 100 metres. A serious disability of this and the previous system was inability to clearly identify the shear wave arrival, which is necessary for the determination of the elastic moduli.

A source of elastic energy that has received much attention is the marine seismic sparker. It is a relatively simple instrument consisting of a DC power supply (typically 5 kV), a bank of capacitors (typically between 10 to 100 μF.) and a pair of electrodes across which the charged capacitors are discharged. The resulting high energy spark explosively creates a gas bubble and seismic energy travels through the water and into the rock. McCann et al. (1975) were among the first to use such a source downhole. Seismograms were measured over paths up to 140 metres through a high loss material using a hydrophone (sensitive to changes in water pressure) as receiver. The sparker had about 1000 joules of energy stored in the capacitors. The system was capable of
detecting refractions and reflections as well as the direct
P-wave arrivals; the S arrivals were not visible in the
records shown and are not mentioned.

Fowler (1977) developed a system very similar to that
of McCann et al. A string of 12 hydrophones was employed
to increase survey efficiency. A sparker was built having
the same basic specifications as McCann et al’s, but with
additional options. The spark gap could be bridged by a
wire which would "explode" during the discharge, it could
be simply left exposed to the borehole fluid, or, as in the
McCann et al. case, it could be enclosed and immersed in a
electrolyte. In comparison experiments the exploding wire
gave the best broadband signal (up to 5 kHz.), but was
difficult to "reload" with fresh wire. Serious
reverberations occurred with the enclosed electrolyte.
Better engineering could probably have cured some of the
problems. Except for the lack of a clear S arrival, tests
of this system (path lengths of up to 45 metres and depths
to 70 metres) were successful.

Two more sparker systems are described by Carabelli
(1970) and Hall et al. (1979). Carabelli discusses the
effect of varying the capacitance and the voltage, in order
to adjust the frequency content. This allowed seismograms
to be collected with energy in bands up to 5 kHz across 50
metres of hard rock. Hall et al. combined a very simple
electrode construction with equally simple, easily made
hydrophones, to create a system with high timing resolution
(+/- 10 μs), high frequency content (1 to 7 kHz), and good
range (up to 100 metres in crystalline rock). The sensitivity of these systems to the S phase is not discussed.

To expand the range of investigation, Gustavsson et al (1982) has returned to the use of an explosive source. The explosives are specially prepared to give high timing accuracy (better than 0.1 msec) and low, non-destructive power. The charges used were between 0.03 and 0.1 kg., and were recorded with wall clamped geophones in crystalline rock to distances of up to 650 metres and at depths of up to 450 metres without signal processing. In addition S wave energy seems to be identifiable in some of the records shown.

Another type of source has been used by Thill (1978). Piezo-ceramics are the active elements in hydrophones and can be driven with a high voltage pulse to produce seismic waves. Many sonar signal generators and sonic logging sources are of this type. Thill combined a piezoelectric source with a hydrophone to do rock damage assessment in mine workings. The problem of dry holes was solved by encasing the instruments in a pliant oil-filled container. By pressurizing the oil, coupling to the rock was possible. The operating range of the system was claimed to be 10's of metres at 20 kHz.

Aki et al (1982) used a piezoelectric source with a specially made hydrophone receiver to investigate a possible geothermal reservoir in crystalline rock. Seismic records were obtained over a few tens of metres at
frequencies as high as 12 kHz, which show recognizable shear arrivals in most cases. An extension of the experiment used custom designed detonators as a source, with three component, wall-clamped geophones as receivers. This combination also gave good results, with good S identification, at distances of tens of metres, and at frequencies into the kilohertz range. There was no estimate of the damage caused by these charges.
Chapter Two: System Design and Fabrication

2.1 Design Rationale

As can be seen from the review in Chapter One, the demands of in-hole operation combined with the specific needs of various applications has forced the use of many different approaches to what is conceptually the same problem, that of producing seismic energy in one borehole, and detecting it in another. Full definition of the requirements of a particular system is therefore an essential first step in the design process.

The standard cored borehole used in the Canadian Nuclear Fuel Waste Management Program is HQ size, with a diameter of 75.7 mm, a typical size for operations requiring good core recovery. In order to avoid problems of jamming against small rock chips or irregularities in the borehole wall instruments must be limited in diameter to about 65 mm. The depth of investigation required in these holes could be up to 2 km, since preliminary plans call for a vault depth of about 1 km. This requires that the sondes be able to withstand hydrostatic pressures of up to 20 MPa and be fully operational with 2 km of cable separating the power supplies and control electronics from the active parts of the probes.

Stringent restrictions imposed by the waste management program are on the peak transmitter power, which limits the useful range of the system, and the number of available boreholes, which increases the need for an extended range
of operation. Each borehole drilled increases the risk of establishing a direct path for the transport of dissolved waste to the surface. Blast damage makes measurements by other geophysical and hydrogeological instruments unreliable, which could result in the need for even more holes. It was therefore necessary that the crosshole system use a low peak power source to minimize borehole damage, and that transmission distance be maximized within this constraint.

The rock type to be studied is important to the system design. Crystalline rock masses typically have velocities in the range 5.0 to 7.0 km/sec, densities between 2.5 and 3 \((\text{kg/m}^3 \times 10^3)\), and very low seismic losses. For a simple model of a spherical wave, the amplitude of the particle motion is of the form

\[ A = A_0 \frac{2\pi}{\alpha} \frac{1}{r} \]

2.1.1

\(Q\), the quality factor, is expected to be more than 100, and possibly as high as 500 in sound crystalline rock. In sedimentary or soil engineering applications the \(Q\) is often below 25, and long range operation requires the use of low frequencies. As can be seen from Table 2.1, frequencies up to 5 kHz are feasible in crystalline rock for ranges of up to 1 km, and are desirable for high resolution in space (short wavelength) and time.

In order to do a complete and detailed interpretation of how the mechanical properties of a rock mass affect the seismic transmission, the full seismic waveform must be
## Total Amplitude Loss Factor

**P velocity = 1500 m/sec**

<table>
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<tr>
<th>freq. (Hz)</th>
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<th>10.0</th>
<th>100.0</th>
<th>500.0</th>
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**P velocity = 4500 m/sec**

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**Quality factor Q = 5. (soil and alluvium)**
**Quality factor Q = 50. (sedimentary)**

Table 2.1: Loss factors in materials with different seismic Q values as a function of frequency and distance.
## Total Amplitude Loss Factor

**P velocity = 6000. m/sec**  
**quality factor Q=100. (crystalline)**

<table>
<thead>
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<th>100.0</th>
<th>500.0</th>
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**P velocity = 6000. m/sec**  
**quality factor Q=500. (good crystalline)**

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recorded. This is also necessary because measurements of the shear wave travel time are essential for accurately estimating parameters such as the elastic moduli and effective crack density. Seismic amplitude variations should also be very diagnostic of fracture zones. Full waveform recording will also allow the identification of events other than the first arrival, such as refractions, reflections and guided waves, which, provided that the spatial sampling is dense, will be of great importance in unravelling complex situations. For seismic travel times of up to 0.25 seconds digitized at a rate of 10 kHz, each data set will contain 2500 points. Since a full coverage survey would have more than 1000 of these records a substantial data acquisition and mass storage capability is required for the system.

The prototype system presented in this thesis was designed to meet the above criteria. The design was based on the use of high frequency piezoelectric transducers in a wall-clamped accelerometer design. The characteristics of the transducers were exploited by using a continuous signal with low peak power, extracting the impulse response of the earth by numerical cross-correlation of the transmitted signal with the received waveform. The system consisted of two sondes, a transmitter and a receiver, which are controlled by digital signals from the surface. Both sondes contain appropriate electronic circuitry for decoding the digital signals and operating the wall clamping mechanism. The receiver contained signal
Figure 2.1. Photo of main components, including two probes, two surface control boxes, and the microcomputer with plotter.
preamplifiers, and the transmitter includes a high-voltage driving circuit. The surface electronics consist of two sets of probe control electronics, a signal generator, an analog to digital converter and a microprocessor which controls the data acquisition and storage on floppy disks. The main parts of the system are shown in Figure 2.1.

The rest of this chapter is concerned with the development of the mechanical components required for the system. The electronics development is reviewed in Appendix B. The signal processing required by the design is outlined in Appendix C, as are some aspects of the data acquisition system.

2.2 Transmitter

The choice of a seismic source to meet these criteria was the key part in the development of the system presented. While an explosive source is the obvious choice for long distance transmission, difficulties involving shot time measurement, frequency content, amplitude repeatability, and borehole disruption (both mechanical and chemical) combine to make such use unattractive. Gustavsson et al (1982) have shown that custom-made explosives improve timing and amplitude repeatability and reduce borehole damage, giving good seismograms at distances of up to 650 m. The blasts did however cause changes in point resistivity logs that were run before and after the experiment. Care was also taken to avoid setting blasts in fracture zones identified with the standard
borehole logs, even though transmitter positions within these low velocity zones would be important to the interpretation. Furthermore, explosive sources are also inefficient for the collection of large data sets, since the transmitter must be repeatedly removed from the hole for reloading.

Sparkers sources have been successfully used in shallow holes. The frequency content and power can be selected by adjusting the voltage and capacitances used. The use of long cables for deep hole work can seriously limit total power, however. The resistance of the spark gap, modeled by Caulfield (1962) to be between 1 and 10 ohms in salt water, is in series with the cable resistance. For short cables the resistance is small. For example Hall et al (1979) used about 150 m. of RG 58 A/U coax with a resistance of 5 ohms. Long cables will have much larger dc resistances, however, (2000 m. of RG 58 A/U is 70 ohms) and introduce much larger losses, unless very large conductors are employed. Such cables are expensive and may still cause difficulties through inductive and transmission line effects.

An alternative to a custom cable is to move the capacitor bank downhole. A steady current could be used for charging, and the discharge could be triggered in a number of ways. The design of this circuit would be difficult, however, since the capacitor must be physically large while the available cross-sectional area within the hole is small. The temperature characteristics of
electrolytic capacitors, and the effect of this on the repeatability of the spark, would require careful study, especially in high heat-flow environments. Finally, the pulse repetition rate would be limited by the power restrictions of the cable.

The detectability of a signal is related to the ratio of the transmitted energy to the "energy" of the background noise, which could be of either electrical or natural seismic origins. With an impulsive source, the total signal energy is transmitted in one pulse, which is limited in magnitude by the strength of the borehole wall. The signal-to-noise ratio can be increased by adding seismograms for a fixed geometry to one another, since the transmitted signals will add coherently while the noise will cancel. With real impulsive sources, the advantages of stacking can be limited by the repeatability of the source, which may be poor. This is particularly true of explosive sources where inexact determinations of the shot time combined with small positioning differences between shots will cause a spreading out of the first arrival, and poor time resolution. Although sparkers are more suitable than explosives for stacking, stacking in itself is inefficient if a large amount of noise reduction is necessary, since the reduction depends upon the square root of the number of seismograms stacked. The relatively low pulse repetition rate provided by sparkers would then be a limiting factor.

The ultimate extension of stacking is to use a fully
continuous source, and to extract the impulse response of the earth by cross-correlation of the received and transmitted signal. This technique is extensively used in reflection seismology, where the source is usually a vibrator and the signal is a sinusoid with a time varying frequency, often called a chirp. The main concept in such work is a non-destructive controlled source, which meets the design criteria of this project ideally.

A controlled source must be capable of continuous transmission of a complicated signal. Piezoelectric material has been used extensively to transform electrical energy into elastic energy for such high power applications as ultrasonic cleaning and sonar scanning, as well as for high resolution applications such as biomedical ultrasonic imaging. It is capable of faithful reproduction of signals with a broad frequency content as required by the design criteria. Furthermore, a piezoelectric transmitter in an accelerometer geometry and with wall clamping will radiate shear energy more efficiently than could an impulsive pressure source. Finally, this type of source will provide less excitation of tube waves, which are strongly excited by simple pressure sources, and which can act as small secondary sources as they move along the borehole.

A piezoelectric ceramic transducer is a high strength ceramic material which can be forced to have piezoelectric properties (poled) through exposure to a strong electric field while cooling through its ferroelectric Curie point. Once poled, the material may be used to convert mechanical
\[ \frac{V}{F} = \frac{F g_{33}}{L W} \quad \text{"P"} \]

\[ \frac{V}{W} = \frac{F}{L W} g_{15} \quad \text{"SH"} \]

\[ \frac{V}{L} = \frac{F}{L W} g_{15} \quad \text{"SV"} \]

\[ \Delta T = V d_{33} \]
\[ \text{or} \]
\[ \ddot{T} = \ddot{V} d_{33} \]

\( d_{ij} \) = Charge Coefficients

\( g_{ij} \) = Voltage Coefficients

\[ K_{ij} = \frac{\text{Input Energy (Ele or Mech)}}{\text{Output Energy (Mech or Ele)}} \]

\( E_{ii} = \frac{\text{Stress}}{\text{Strain}} \quad i = 1, \text{Stress } \perp \hat{p} \)

\( i = 3, \text{Stress } \parallel \hat{p} \)

\( i = 5, \text{Shear Stress} \)

\( E_{ii} \) = Electrical Circuit used.

\text{Figure 2.2 Fundamental Relationships for piezoelectric ceramic elements.}
Table 2.2 Physical Properties of the piezoelectric ceramics used in the probes.

<table>
<thead>
<tr>
<th>PROPERTY</th>
<th>Ai4R</th>
<th>Ai5R</th>
<th>UNIT</th>
</tr>
</thead>
<tbody>
<tr>
<td>Relative Dielectric Constant ($k_{33}$)</td>
<td>1400</td>
<td>2400</td>
<td></td>
</tr>
<tr>
<td>Dissipation Factor (Loss Tangent)</td>
<td>0.003</td>
<td>0.015</td>
<td></td>
</tr>
<tr>
<td>Coupling Coefficient ($k_p$)</td>
<td>0.60</td>
<td>0.62</td>
<td></td>
</tr>
<tr>
<td>Frequency Constant</td>
<td>213</td>
<td>198</td>
<td>kHz cm</td>
</tr>
<tr>
<td>Density</td>
<td>7.6</td>
<td>7.6</td>
<td>$10^3$ kg m$^{-3}$</td>
</tr>
<tr>
<td>Curie Temperature</td>
<td>325</td>
<td>310</td>
<td>°C</td>
</tr>
<tr>
<td>Q ($Q_m$) (Radial Mode)</td>
<td>550</td>
<td>80</td>
<td></td>
</tr>
<tr>
<td>$g_{33}$</td>
<td>18.7</td>
<td>25.0</td>
<td>$10^{-3}$ v m N$^{-1}$</td>
</tr>
<tr>
<td>$g_{31}$</td>
<td>9.0</td>
<td>10.4</td>
<td></td>
</tr>
<tr>
<td>$g_{15}$</td>
<td>22.0</td>
<td>36.0</td>
<td></td>
</tr>
<tr>
<td>$d_{33}$</td>
<td>290</td>
<td>385</td>
<td>$10^{-12}$ m v$^{-1}$</td>
</tr>
<tr>
<td>$d_{31}$</td>
<td>140</td>
<td>175</td>
<td></td>
</tr>
<tr>
<td>$d_{15}$</td>
<td>380</td>
<td>500</td>
<td></td>
</tr>
<tr>
<td>$E_{33}$ (Open Circuit)</td>
<td>7.4</td>
<td>5.3</td>
<td>$10^{10}$ N m$^{-2}$</td>
</tr>
<tr>
<td>$E_{11}$ (Open Circuit)</td>
<td>8.5</td>
<td>7.0</td>
<td></td>
</tr>
<tr>
<td>$E_{55}$ (Open Circuit)</td>
<td>3.5</td>
<td>2.5</td>
<td></td>
</tr>
</tbody>
</table>
energy to electrical and vice-versa. A block of piezoelectric ceramic may be modeled conceptually as two surfaces of opposite electric poles connected by a spring. Any change in the strain (i.e. in relative displacement of the spring) will cause a change in the electric field associated with the poles. Similarly, a change in the external electric field will produce a stress in the spring. In general, the equations that completely describe the behaviour of piezoelectrics combine both electrical and mechanical terms, as shown by Mason (1948). In practice, at frequencies far from natural resonances, a piezoelectric ceramic block will behave like a capacitor electrically, and like a stiff spring mechanically. Figure 2.2 shows some of the standard relationships for piezoelectric ceramics.

In the interest of obtaining the maximum possible response from a driven piezoelectric, resonant modes are often used. This requires a lumped mechanical system with characteristic dimensions that are comparable to the wavelength required. Since the wavelength of a 3 kHz elastic wave in a typical piezoelectric ceramic is about 1 metre, it is clear that such a device would be impossible in a 0.075 metre borehole. While it is possible to reduce the effective elastic modulus of the ceramic through the use of flexural modes of vibration, the result is necessarily a source with low mechanical impedance, and would transmit inefficiently into crystalline rocks. It would also be possible to build a probe that resonated in a
Figure 2.3 Simple accelerometer design using piezoelectric ceramics.
Figure 2.4 In-hole geometry of piezoelectric ceramic accelerometer.
direction parallel to the hole axis, but this configuration would provide poor radiation of compressional energy perpendicular to the hole. Resonant methods also limit the bandwidth of the transmitted signal.

The ceramic can be used in a non-resonant design, and the high mechanical impedance of the transducers exploited by coupling directly to the borehole wall. This type of design will also facilitate the use of broadband signals, and will allow the adjustment of the frequency band to maximize either transmission distance or time resolution for different environments or applications.

One way to do this is shown schematically in Figure 2.3. A piezoelectric element is sandwiched between the borewall and a mass. From the relationships in Figure 2.2, a sinusoidal driving voltage \( V_0 e^{i\omega t} \) will produce a strain in the element of magnitude

\[
\Delta T = d_{33} V_0 \quad 2.2.1
\]

If the wall and the mass are rigid, the instantaneous displacement of the mass is given by

\[
\vec{u}_r = \Delta T \hat{r} = d_{33} V_0 e^{i\omega t} \hat{r} \quad 2.2.2
\]

The magnitude of the reaction on the wall from this motion will be

\[
F = |\vec{u}_r| = md_{33} \omega^2 V_0 \quad 2.2.3
\]

In order to build the transmitter, piezoelectric and stainless steel elements were arranged as shown...
schematically in Figure 2.4. This mechanical system is comprised of a coupling plate, a transducer and a backing mass. The piezoelectric ceramics used in the probes are manufactured by Almax Industries of Lindsay, Ontario. The specifications of the materials used are given in Table 2.2. The transmitter was built using Ai-4R material. In order to get the maximum possible displacement from the voltage available, thin disks were assembled using conductive epoxy, in the manner indicated in Figure 2.5. The disks were 22 mm in diameter, and 3.2 mm thick. Four stacks of 8 disks were used, each assumed to operate as a single block with thickness 25.6 mm. The anvil (coupling plate) was 305 mm long, and the effective mass about 2.5 kg. The mass is at best an estimate, for reasons discussed below.

The model assumes that the force on the borehole will be equal to that due to the inertia of the backing mass and is an approximate low frequency limit. At higher frequencies the compliance of the components will cause a departure from equation 2.2.3. A rough estimate for the frequency at which this occurs can be made by considering the simple mechanical resonances of the system.

The geometrical resonances defined by the boundary conditions and the wavelength are all outside the required operating range. Typical acoustic impedances for stainless steel, ceramic, granite and water are 40, 22, 15, and 1.5 \((\text{kg m}^{-2} \text{sec})^{-1} \times 10^6\) respectively. The \(\lambda/2\) free resonance (lower impedance materials on both sides) for the backing
Figure 2.5 Arrangement of piezoelectric ceramic elements in transmitter.
mass is approximately 200 kHz. The ceramic may resonate in either a \( \frac{\lambda}{2} \) or \( \frac{\lambda}{4} \) mode depending on the quality of the coupling to the mass. The open-circuit elastic modulus and dimensions of the elements place these at 85 and 42.5 kHz, well above the required band of 0.5 to 5 kHz.

The lowest resonance is that caused by the compliance of the ceramic resonating with the mass. Treating the system as a mass and spring, the spring constant is given by the Young’s modulus, the thickness and the surface area of the ceramic,

\[
k = \frac{EA}{T}
\]  \hspace{1cm} 2.2.4

The resonance of the mass-spring system is given by

\[
w = \sqrt{\frac{K}{m}} = \sqrt{\frac{EA}{mT}}
\]  \hspace{1cm} 2.2.5

Substitution of the values above and in Table 2.2 give a resonance of 6.7 kHz. The result assumes that the backing mass is rigid. The elongated shape, however, suggests that flexing may reduce the mass that reacts directly to the strain in the ceramic. This would increase the resonance frequency. A xylophone key with dimensions and material similar to the backing mass would have a resonant frequency of about 1 kHz (Josephs, 1967), indicating that this effect could be strong. The flexural modes are damped by the hardware used in the probe fabrication, and by the materials used to seal out the borehole fluid. These components also marginally increase the mass-spring resonance, as will be discussed below.
A transmitter of this type will produce a radial point force on the borehole wall. The far field radiation pattern for a point force in an infinite elastic medium has been calculated by White (1965), and is shown in Figure 2.6a. The pattern is characterized by a two lobe P-energy pattern, and a S-energy torus, coaxial with the force. Greenfield (1978) has shown that this result is suitable for wavelengths long compared to the borehole diameter. As the wavelength is decreased, the symmetry of the pattern is disturbed, with more energy beamed in the direction of the force. This is shown in Figure 2.6b as a comparison of the maximum P-energy along the force axis, and in the radiation pattern of the S-energy (the component tangential to the hole) in Figure 2.6c. For NQ holes in crystalline rocks with $V_p$ in the range 4.0 to 6.0 km/s, a frequency band of 0.5 to 5 kHz has a range of $\lambda_p/2a$ ($\lambda_p$ is the compressional wavelength) of 10 to 200. At the highest frequencies in slow crystalline rock, only slight distortion from the low frequency limit of the pattern is expected.

2.3 Receiver

The conversion of body wave energy into water pressure that is needed for hydrophone-type detectors depends upon the distortion of the cross-sectional area of the borehole. Any compressional wave incident on the hole will produce this effect. Shear motion perpendicular to the hole axis will change primarily the shape of the hole but not the area, and thus produces no pressure in the water. A
a. Radiation pattern of a point force in an infinite elastic medium. (after White (1965))

b. P-wave displacement amplitude versus $\lambda_0/2a$ in forward and backward directions for a point force in a hole of radius $a$, with a wavelength $\lambda_0$. Amplitudes normalized to a. (after Greenfield (1978))

c. Radiation pattern for the tangential shear component of the s-wave in the plane perpendicular to the borehole at the point of the force, as a function of the P-wavelength and the borehole diameter. Result is symmetric about the x-axis. (after Greenfield (1978))

Figure 2.6 Radiation patterns associated with downhole seismic sources.
vertically travelling shear wave is therefore undetected by a hydrophone in a vertical hole. White (1965) has shown that non-vertical shear waves produce some pressure changes in a borehole, but only a portion of the available energy is converted. All of these waves cause particle motion in the borehole wall, however, and could be detected with a motion sensitive receiver. The detection of borehole motion also enhances the detectability of compressional waves, provided the mechanical impedances are properly matched.

While wall clamped geophones are commercially available, all such instruments are designed to operate in large diameter holes for vertical seismic profiling in the frequency band below 500 Hz. Since the response and dimensions of these instruments did not comply with the needs of this project it was necessary to design and construct a wall-clamped, motion-sensitive, receiver.

The receiver design parallels that of the transmitter. In the configuration shown in Figure 2.3, small motions of the borehole wall will be transferred to the mass by the ceramic. If the motion is of the form \( \vec{u}_r = \vec{u}_0 e^{i\omega t} \), the force across the ceramic produced by the inertia of the mass will be given by

\[
\vec{F} = m \ddot{\vec{u}} = m \omega^2 \vec{u}_0 e^{i\omega t}
\]

2.3.1

From the relation in Figure 2.2, this force results in a voltage of magnitude

\[
U = \frac{1}{LW} g_{33} m \omega^2 |\vec{u}_0|
\]

2.3.2
The receiver discussed so far has been limited to the detection of radial motion, which is only one component of the three that are available. It is possible to detect the vertical and tangential motions of the borehole wall with the mechanical system in Figure 2.4 if shear sensitive piezoelectric ceramics are used. Returning to the spring and electric charge analogy, an external electric field perpendicular to the spring will cause shearing forces on the charges. Similarly, if a shear strain is applied to the spring, the electric field perpendicular to the spring will be changed.

Shear sensitive piezoelectric elements were made from normally polarized ones by removing the electrodes and then coating the sides parallel to the poling direction with conductive paint. The piezoelectric relationships for this arrangement appear in Figure 2.2. In the geometry shown in Figure 2.4, vertical or tangential motion ($\mathbf{U}_V$ or $\mathbf{U}_T$) of the wall will result in a inertial shear force on the ceramic. If the normal vector to the painted surfaces is parallel to the motion, the magnitude of voltage produced is

$$V = \frac{q_{15} \omega^2 |\mathbf{U}_V|}{L} \quad \text{or} \quad V = \frac{q_{15} \omega^2 |\mathbf{U}_T|}{W}$$

2.3.3

where $|\mathbf{U}_V|$ and $|\mathbf{U}_T|$ are the magnitudes of the vertical and tangential particle displacements.

The receiver ceramics were Ai-5R, and were shaped as blocks with a length of 21.5 mm., width of 21.5 mm., and thickness of 25.4 mm. These were arranged in pairs, with 6 blocks in total for the three orientations. The other
mechanical components are identical to the transmitter. In the low frequency limit the elastic parameters are not affected by the placement of the voltage-sensing electrodes. The ceramics may therefore be considered identical to each other in the estimation of the frequency response for each channel.

The high frequency limit for the radial component can be estimated from equation 2.2.5 to be 7.7 kHz, and is subject to the discussion of Section 2.2. The spring constant controlling the shear response can be calculated from equation 2.2.4 by replacing $E_{33}$ with $E_{15}$, which is the modulus defined for this geometry. The calculation of the resonance is more difficult since the effective mass is hard to define. Using 2.5 kg gives a frequency of 5.3 kHz. The vertical channel may well see a greater effective mass, however, since the entire instrument mass is geometrically available as reaction mass, and flexural effects will be minimized. This would lead to enhanced sensitivity to low frequencies of vertical motion. The effective mass for the tangential component may be reduced by torsional flexibility in the backing mass, reducing the low frequency sensitivity. The sensitivity of the shear channels could also be reduced by slippage between the components, a problem that is minimized in the compressional channel with the prestress imposed by the assembly hardware and the external confining pressure.
2.4 Clamping Mechanism

Efficient operation of both the receiver and the transmitter depends strongly on the quality of the mechanical contact between the rock and the transducer. Ideally, the face of the transducer should be bonded directly to the rock surface. In practice bonding is impossible, and irregularities in the borehole wall combine with localized small scale fracturing to reduce the quality of the contact. The response of the system is changed by the introduction of a soft spring (compliance of water and asperities), thus reducing the sensitivity to wall motion. A fully decoupled instrument acts like a high mechanical impedance hydrophone, with low sensitivity to changes in borehole fluid pressure.

The transducer's anvil serves to eliminate part of the coupling problem. The anvil is a stainless steel plate (Figure 2.4) that fits the curvature of an NQ borehole on one side, and has a flat surface in contact with the ceramic on the other. It provides a large surface for the distribution of stress over many of the small irregularities on the wall surface to better approximate a perfect smooth contact. The large surface area also acts to trap pore fluid in small depressions in the wall, which then provides at least partial coupling to this portion of the surface.

An important problem is how to decenter the instrument in the borehole with sufficient force to guarantee that the
Figure 2.7a Schematic diagram of the clamping mechanism magnetic drive.
anvil is in contact with the wall. In standard borehole instrumentation it is common to use a simple bow spring to provide a continuous force for the whole time that the instrument is downhole. The difficulty in using this method is that the large surface area of the anvil provides a good opportunity for catching irregularities in the hole, and jamming the instrument when moving between positions. This method also results in high wear on the anvil, which is precision machined with O-ring grooves for sealing.

Another standard tool for clamping is the caliper arm. This device consists of a strong arm on a pivot, actuated by a worm gear. A rotating high pressure seal is required to keep the borehole fluid from damaging the motor that drives the gear. A seal capable of withstanding the expected 20 MPa pressures for periods of many hours is not commercially available in the dimensions that are allowed, and the construction of this device was thought to be beyond the capabilities of the available facilities.

In the clamping device used, a magnetic coupling transfers mechanical power produced by an electric motor in a pressure-tight stainless steel tube through the wall to components exposed to the fluid. A 4-pole cylindrical magnet attached to the motor drives a similar magnet, through a stainless steel plate of 3 mm thickness. Because this is a low torque connection, it is necessary to run the motor at moderately high speed, and to use a speed reducing gear box to increase the torque. The device is shown schematically in Figure 2.7a.
Figure 2.7b Schematic diagram of cart, slot and bronze cover assembly of the clamping mechanism.
The low speed axle of the gear box is used to wind (and unwind) a high strength nylon line. The line pulls a small cart along a slot in the backing mass which has variable depth. The cart eventually becomes wedged between the mass and the borehole wall (Figure 2.7b). The motion of the cart is reversible, through the use of a second line, wound in the opposite sense, and a pulley block at the bottom of the backing mass. The slot is covered with a phosphor bronze cover, which is fastened securely to the pulley block at the bottom, and by means of a light spring at the top. This cover protects the nylon line that drives the cart, and provides a smooth surface for the cart to run on. The cart has two wheels that run on the mass, and one that runs on the bronze cover. In the case of failure of the clamping mechanism or jamming, it was planned that sufficient force to break the nylon line could be applied by pulling on the probe cable, thus releasing the cart. The force required would be far less than the breaking stress of the cable. The bronze cover and the cart are considered expendable in these conditions.

In crude tests in the laboratory, the clamping mechanism was capable of lifting the entire instrument, which has a mass of about 15 kg. In a vertical borehole, the instrument is supported by the cable and by buoyancy, so that most of the force is available for clamping.
2.5 Fabrication Details

Since the work reported here was essentially a feasibility test, the instrument system had to be designed around existing cables. The cable used for the transmitter was a standard six conductor armoured logging cable produced by Vector Cable Company that was 305 m long. The availability of only 6 conductors required that the probe functions (transmit, clamp, unclamp, etc) be selectable from the surface. This is discussed in Appendix B. The receiver cable was a twenty-four conductor cable produced by Mark Products which was 650 m long. Each channel used a different pair of conductors which eliminated a need for multiplexing circuitry. Control of the probe was identical to the transmitter. The receiver cable was on loan for this project from the Geological Survey of Canada, through the cooperation of J. A. Hunter.

The probes are constructed entirely of stainless steel, and are mechanically identical. Figure 2.8 is a schematic diagram of an assembled probe. The uppermost section is a coupling between the probe and the cable, which is attached directly to a 63.5 mm diameter tube, housing the downhole electronics. The circuit elements are assembled on aluminum rails with a connector at the bottom to facilitate insertion into a partially assembled probe. The lower end of this 760 mm length tube is threaded on to the gearbox holder of the clamping mechanism. This segment contains the motor and is designed to allow the electrical
Figure 2.8 Schematic diagram of a probe for crosshole seismic investigations.
leads to pass the gearbox by means of four narrow pipes that serve as the structural support of the probe through this section. The gearbox has a reduction factor of 24, and was custom built of brass and stainless steel. All openings of the probes are protected with double 'O ring' seals. The design of these seals places the groove on an inner massive block, with the other surface on a lighter, outer member. In this configuration, the confining pressure will act to improve the quality of the seal.

The lower end of the gearbox holder is attached to the transducer assembly with a standard flange, and is sealed with a viton gasket. The flange is necessary to allow threading of electrical leads through the gearbox holder and into the transducer assembly. The connection also serves to mechanically decouple the transducers from the rest of the probe by means of the gasket material. This seal failed in initial tests when the very soft gasket material was pushed into the central portion of the flange. However, the flange was sucessfully modified by adding an internal brass ring to support the gasket.

The transducer assembly is milled from a solid bar of stainless steel. The ceramic elements are housed within the anvil which is completely hollow. There is also room in the anvil for the electrical leads necessary for transducer operation, and for leads required by the operation of any similar, independent probes hung below. The outer surface of the anvil was turned to the radius of an NQ hole before separation from the bar. The rim of the
anvil facing the backing mass is grooved for an 'O ring', which is the main seal.

The backing mass is the structural connection to the rest of the probe. It is mostly solid and has a matching 'O ring' groove cut on the side facing the anvil. The sealing between the two pieces is completed with a PUC (plastic) spacer that is smoothly polished to mate with the O-rings. This material was chosen as a compromise between the very stiff material that is desirable for sealing, and the very soft material needed to permit free relative motion between the anvil and the backing mass. The maximum diameter of the whole assembly is 61 mm. The three pieces are held together by 4 stainless steel bolts for instrument integrity and 28 nylon bolts which provide uniform deformation of the O-ring for sealing at low confining pressures. High ambient confining pressures will increase the O-ring deformation and improve the quality of the seal.

The sealing arrangements will have some effect on the sensitivity of the transducers. By acting as a spring parallel to the ceramics, the spacer will reduce the sensitivity (or power) of the probes, and increase the primary resonance frequency. The spring constant may be calculated from the estimated effective area (the cross-section of the O-rings), thickness, and Young's modulus, as in Equation 2.2.4. The resulting value is $5.5 \times 10^8$ Pa·m compared to the value of $5.8 \times 10^9$ for the receiving ceramics, and $4.4 \times 10^9$ for the transmitter. The O-rings are softer than the PUC, further reducing the
coupling. The bolts (size 6-32) will also act as a parallel spring. The spring constant calculated for the steel bolts is $1.7 \times 10^8$ and for the nylon $1.5 \times 10^7$. It is clear that the ceramics are much stiffer than the sum of the other components, and therefore will support most of the stress produced by relative motions between the anvil and the backing mass. The spacer and the bolts will have the advantageous effect of damping any transverse modes of oscillation in the backing mass, especially when the anvil is pressed against the borehole wall.
Chapter Three: System Evaluation

3.1 Test Sites

In this chapter, some of the data collected with the cross hole seismic system are examined. Interpretations of the results are presented and discussed in terms of measurements made with other methods wherever possible. The purpose of the chapter is to evaluate the usefulness of the system for rock property exploration, and not to offer detailed and definitive analysis of a particular site or data set.

Tests were conducted at two locations. The site used for preliminary tests was the borehole geophysics test facility of the Geological Survey of Canada, near Ottawa Ontario. There are four 75.7 mm boreholes available with cores. A preliminary study of the cores has been made by Bernius (1981a,b). Hole BC-81-1 is 305 m deep, penetrating 12 m of lower Paleozoic dolomitic shale (Oxford Formation), 9 m of sandstone and sandy shale (March Formation), 42 m of sandstone (Nepean Formation), and 240 m of a mixture of late Precambrian (Haydrinian) granitic and gneissic rock. Holes BC-81-2, -3, and -4 are spaced at 10, 30, and 100 m from -1 respectively, and are 120 m deep. While the holes are described by Bernius as vertical, the exact orientation has not been determined. Figure 3.1 is a plan view of the site, and Figure 3.2 is a preliminary section in the plane of the holes. At the time of the testing, standard borehole logging had not been carried out.
Figure 3.1. Site map for Borehole Geophysics Test Facility near Bell's Corners Ontario. (after Bernius (1981b))
Figure 3.2. Preliminary geological section of Borehole Geophysics Test Facility as inferred from the core log. There is no control on the dips of the sedimentary bedding or the fractures because the holes are vertical and in-line. (after Bernius (1981b))
The second site was a granite batholith located in south-west Manitoba at the Whiteshell Nuclear Research Establishment, selected by AECL as the location for an Underground Research Laboratory (URL). The rock mass is made up of a 300 m section of slightly altered pink granite, over fresh grey granite. The contact between these units is marked with a zone of heavy fracturing. This site is the subject of intense investigation.

Two pairs of boreholes on this site were used for crosshole seismic experiments carried out by J. Wong in October 1982. Experiment 1 (E1) was conducted in holes M1a and M1b. These hole are 13.3 m apart at the surface. Drill hole M1a is 326 m deep and has casing in the top 150 m. Drill hole M1b is 150 m deep, and is uncased. The holes used in Experiment 2 (E2), URL6 and M2a, were 175 m apart. M2a is identical to M1a. URL6 is 400 m deep, and is uncased. All M-series holes are percussion drilled to diameters that vary continuously from 165 to 152.2 mm URL6 is a NQ size (75.7 mm) diamond drill hole with full core recovery. A detailed analysis of the core is available (Ejeckham, 1982). Television logs (Dugal, 1982) of the holes are also available.

In a typical experiment, the trailer-mounted transmitter winch is placed near one borehole with the transmitter control electronics. A truck containing the receiver winch, receiver control electronics and signal acquisition system is parked near another hole. The control modules are connected together with a twisted pair
cable to provide a timing link between the transmitter and the signal acquisition. Tripod and pulley assemblies are assembled over the holes and used to lower the sondes. A clamp is used to eliminate the tension in the exposed portion of the receiver cable, reducing noise caused by wind motion.

Data collection begins when the sondes are clamped into position and the transmitter started. Data from the receiver is digitized and stacked into the memory of the microcomputer. The stacked data may then be correlated with the transmitted waveform to check on data quality. The final step is the transfer of the data from the memory on to the floppy disks. During the correlation and transfer of data, either one or both of the probes may be moved to the next desired position. For the collection of three channels of data, and correlation of one, the procedure takes approximately 6 minutes. This is reduced to 3 to 4 minutes if the correlation is held for a later time. A typical day would see the collection of about 300 uncorrelated data sets (100 positions), with occasional correlation to check data quality.

3.2 System Performance

The signal selected for use with the system was the pseudo-random binary sequence (PRBS). A PRBS is a sequence of binary pulses of variable duration, arranged randomly within a fixed length of time. They are easily generated and may be chained together for a continuous periodic
Figure 3.3. Computer simulated data sets. Raw data (a) is in the upper left, correlograms (b) in the upper right, and Fourier power spectra below. Note the changes in plot gain from trace to trace.
transmission. Geophysical application of the PRBS has been discussed by Foster and Sloan (1972), Goupillaud (1976), Cunningham (1979) and Duncan et al (1980). The properties of the PRBS are discussed in Appendix C.

A portion of a PRBS is shown in the top trace of Figure 3.3a. The autocorrelation of a PRBS produces a triangle with the base length defined by twice the fundamental clock period used in the creation of the sequence, with the apex at the zero lag position. The triangle is repeated in time with a period equal to the sequence length, which must be long enough to contain all arrivals. The cross-correlation of two sequences, identical except for a time delay in one, results in a shifted triangle, with the lag to the apex equal to the delay. This is shown in the top trace of Figure 3.3b, where a delay of 7.5 ms was introduced into one sequence before correlation. Electronic limitations (see Appendix B) forced the use of a sequence with the frequency response indicated in the top trace of Figure 3.3c. This is the expected power spectrum for a PRBS with a fundamental clock period of 0.3 ms, and it has a distinct notch at 3.3 kHz. A sequence length of 1023 clock pulses (~ 0.3 s total length) was commonly used, and a digitizing interval of 0.075 ms was standard.

The remaining traces in Figure 3.3 show the possible effects of the system and earth response on the appearance of the signal, through the use of a four pole recursive butterworth filter. The signal in the second trace
simulates the response of the analog electronics. A 500 Hz low cut filter is necessary to reduce 60 Hz cable pickup and low frequency seismic noise caused by wind. A high cut filter (set at 6.6kHz in the Figure) prevents aliasing.

The response of the transducers is modeled in the third trace. The dependence will enhance the high frequencies and in the absence of dispersion, attenuation and scattering, the correlogram should appear similar to the third trace of Figure 3.3, where the low cut corner has been moved up to 3.3 kHz. The fourth trace shows the effect of attenuation of high frequency by moving the high cut corner to 3.3 kHz. This effect may be expected in fractured regions, caused either by higher rates of attenuation or by scattering at short wavelengths. Scattering, expected in fracture zones, will also affect the recorded data by increasing the coda (pulse length), as will dispersion.

Some examples of data collected in situ are presented in Figure 3.4. As in Figure 3.3, the raw data is in 3.4a and the correlograms in 3.4b. The fourier power spectra in 3.4c are calculated for only the portion of the total correlation that has apparent seismic energy, so that the noise present at times before and after the arrival does not dominate the spectra. The top trace is data collected with a probe separation of about 20 m. in near-surface granite. Fifty sequences were averaged (a fifty-fold stack) for this record, which shows clear pulsations with random spacing in the raw data. With a
Figure 3.4. Data collected in the field. Raw data (a) is in the upper left, correlograms (b) in the upper right, and Fourier power spectra below. Note the changes in plot gain from trace to trace.
signal to noise ratio of this magnitude a pulse waveform and a stacking seismograph would be adequate for signal recovery. The correlogram has a clear P arrival at 3.1 ms, which is very similar in appearance to the third trace of Figure 3.3, that is the high frequency parts of the system passband are strongly emphasized. Since selective wave attenuation of low frequencies by natural mechanisms is unlikely, the low amplitude of the low frequency parts must be due to the instrument response. The coda of the shear arrival is spread out slightly, which may be indicative of different transmission properties for the shear, or may be due to low amplitude secondary events associated with tube waves in the transmitter hole. The signal to noise ratio (s/n) in the correlogram is about 100.

The remaining records in Figure 3.4 are of decreasing signal amplitude. The raw data in the second trace, recorded on a 180 m path, was averaged over 50 sequences. While the transmitted signal may be visible over the background noise in this record, the measurement of a clear first break with an impulsive system would require a long averaging time. The initial cycles of the direct arrival are similar to the fourth trace of Figure 3.3, indicating strong attenuation of the high frequencies. The character of the arrival is most probably due to the bimodal distribution in the power spectrum, combined with the phase shifting produced by the filter (Fig. 3.3) and by filtering and possible dispersion in the rock (Fig. 3.4). In the correlogram, the s/n is better than 10, with a first
arrival at 28.5 ms.

The final trace was recorded with the probes spanning a section for which the borehole television logs indicated heavy fracturing. Despite doubling the averaging time (100-fold stack), the uncorrelated data shows only a trace of the desired signal. After correlation, the s/n is about 5, with a first break at 29 ms. The power spectrum is intermediate between those above, with a moderate enhancement of the high frequencies. This indicates a relatively broadband attenuation along the path, which allows the natural response of the transducers to shape the spectrum. It should be noted that these records have had no processing other than correlation. Digital filtering and other noise reduction methods could further improve the s/n in all of these cases.

The records of Figure 3.4 show an undesirable feature common to all data collected. In the bottom two traces, a low amplitude event appears approximately 20 ms before the first arrival. This event is a 'ghost' of the first arrival, which actually occurs late in the record. The ghosts in lower traces are related to the previous correlation triangle in the periodic chain, and are not seismic signals or a predicted artifact of the PRBS correlation process. In cases where the complete set of arrivals is longer in duration than 20 ms, interference of ghosts of the late arrivals with the first break can degrade travel time resolution. The origin of this event has not been identified, despite extensive computer
modeling. Deconvolution could remove the problem but was not warranted. Future experiments with the analog electronics, particularly the high-voltage drive circuit where a non-linear response from saturated transformers is possible, may reveal the source.

In the computer graphics software that has been developed, an interactive plotting routine provides for the selection of the first break from a trace displayed on a high resolution graphics terminal. The first break is often of small amplitude, but is usually identifiable by its high frequency content. The travel time is calculated by adding one clock period to this time, to account for the width of the correlation triangle.

Using the first break as a reference, the software then allows the selection of a portion of the waveform for amplitude analysis. In standard processing, a window of 4.8 ms (64 data points) after the first arrival was used for Fourier transformation. The resulting power spectrum was divided into two sections, and integrated, producing numbers representing the low and high frequency amplitudes.

Seismograms measured on different occasions with the same geometry provided a test of the repeatability. The data used for this test is shown in Figure 3.5. One set of data was collected sequentially while the other is a collection of seismograms which were recorded on different days, totally out of sequence. Table 3.1 shows that for this small sample, the repeatability of the travel time selection is within 1 digitizing interval (+/- 0.075 ms).
Figure 3.5. The data on the left were collected by fixing the receiver position and moving the transmitter. The data on the right were extracted from a number of fans collected with the receiver moving and the transmitter fixed. Data acquisition and plotting constants are identical for each data set. The orientation of the transducers may differ in the two cases.
Table 3.1 Travel time and amplitude values extracted from the data sets of Figure 3.5. The left column is for the left hand data set which was used as the standard for the relative changes of the third column.

<table>
<thead>
<tr>
<th>Time (ms)</th>
<th>Data Set 1</th>
<th>Data Set 2</th>
<th>Change Between Measurements</th>
</tr>
</thead>
<tbody>
<tr>
<td>30.45</td>
<td>32.25</td>
<td>32.175</td>
<td>- .075</td>
</tr>
<tr>
<td>29.775</td>
<td>29.85</td>
<td>29.175</td>
<td>- .075</td>
</tr>
<tr>
<td>28.575</td>
<td>28.500</td>
<td>28.20</td>
<td>+ .075</td>
</tr>
<tr>
<td>28.20</td>
<td></td>
<td></td>
<td>0.0</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Low Frequency</th>
<th>Data Set 1</th>
<th>Data Set 2</th>
<th>Change Between Measurements</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.67</td>
<td>1.51</td>
<td></td>
<td>+ .16</td>
</tr>
<tr>
<td>1.51</td>
<td></td>
<td></td>
<td>0.0</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>High Frequency</th>
<th>Data Set 1</th>
<th>Data Set 2</th>
<th>Change Between Measurements</th>
</tr>
</thead>
<tbody>
<tr>
<td>2.87</td>
<td>3.65</td>
<td>3.64</td>
<td>+ .99</td>
</tr>
<tr>
<td>4.46</td>
<td>3.24</td>
<td>3.24</td>
<td>+ .37</td>
</tr>
</tbody>
</table>
The digitizing interval provides the main limit on the timing accuracy for very short travel paths. More complicated time picking schemes, including filtering, interpolation and deconvolution could improve the timing resolution if necessary.

At the present time, there is no way to control or measure the direction in which the transducer anvil is pointing. Because the receiver is sensitive (in theory) to the motion of a single point on the borehole wall, the position of this point relative to the incoming wave must be known. Similarly, the position of the point displacement that caused the wave is required, so that the polarization of the wave can be estimated. The absence of this information makes a quantitative study of the instrument response and sensitivity impossible. The repeatability of the clamping could also have a large effect because of the importance of the mechanical coupling. Textural changes in the hole, for example grooves and ridges from the drill, or small rock chips only partially attached to the wall, could make even minor differences in the positioning of the probes very important. Variable clamping could also affect the frequency response, by reducing the mechanical impedance 'seen' by the transducer assembly.

These effects are evident in the data through the changes in the appearance of the waveforms (i.e. 270,350). With data set 1 as the reference, the average difference in amplitude numbers between the two data sets is 28% for the
low frequency, and 66% for the high, as calculated with a 64 point window (4.8 ms). While the repeatability of the amplitude is not very good, the anomalies produced by the main target of study, fractures, can exceed an order of magnitude in size. Strong correlations between amplitude anomalies and known fracture zones, discussed below, support the usefulness of amplitude data.

An experiment to test the effect of the designed clamping mechanism was not possible, because the hole diameters at one test site were too large and because of difficulties with the data acquisition system during initial tests in standard NQ holes. A test was conducted in the large diameter holes using a bow spring arrangement on the receiver in order to estimate the importance of clamping.

The top set of seismograms in Figure 3.6 are from the radial channel, with the transmitter and receiver both unclamped. While unclamped the probes are almost certainly in contact with the wall over some part of their length, due to inclination of the hole. The lower set of seismograms was recorded with the receiver clamped with the bow spring and the transmitter unclamped. The average gain in the primary arrival produced by the clamp across all frequencies is 2.2, as calculated from the amplitude numbers produced with a 64 point window. Gain in the later arrivals is also enhanced by at least this amount. The travel times repeat to within 1 digitizing interval, as above. The effect of the clamp is not uniform in this example, with a variation in gain from 4 to 0.75.
Figure 3.6. The data on the left were collected with both instruments unclamped. A bowspring arrangement was used to clamp the receiver for the data on the right. The transmitter was stationary for the duration of the experiment.
This may be due to the roughness of the borehole wall, and to small particles of rock grinding between the anvil and the wall during movement due to the permanent clamping from the bow spring. The increase of signal level obtained with clamping (up to a factor of 6 if both instruments were to be clamped) is highly desirable, since it will add directly to the usable range of the system.

3.3 Travel Time Analysis

A transmitter that is easily and accurately controlled with an electrical signal removes much of the timing inaccuracy found in seismic systems that use explosive or mechanical sources. The piezoelectric transmitter outlined above has low peak power however, and is only useful because the normally attenuative overburden is avoided in the crosshole geometry. Unfortunately, the use of boreholes introduces other inaccuracies to the measurements.

The most serious of these is the poor knowledge of the location of the probes. In most cases, precise borehole orientation surveys are not available. Where surveys have been conducted, the holes often show considerable departure from the vertical, with the bottom of the hole displaced horizontally tens of metres from the top. For accurate velocity determinations, the distance between the probes must be known to better than 1%. The absolute location of the probes must also be known for the accurate positioning of anomalous regions. For example, a deviation of 1 degree
over 500 m of one hole will change the distance to a vertical hole by 8.7 m. With a hole separation of 100 m the systematic error in the velocity determination caused by the variation in path length would exceed 5%.

A fan of seismograms is collected by fixing either of the instruments at one depth, and moving the other. Fan shooting is important since relative changes in travel time are a function of the experiment geometry, allowing verification of the relative probe positions in favourable geologies. It also enables fast collection of data, since only one probe is unclamped, moved, and clamped for each seismogram. By changing the depth of the apex of the fan, particular portions of the rock body may be selected for detailed study.

In a rock mass with constant velocity \( c \), the travel time is controlled by the length of the travel path. For a fan geometry, it is given by

\[
c^2 t^2 = D^2 + (Z_R - Z_T)^2
\]

where \( Z_R, T \) are the receiver and transmitter depths, and \( D \) is the hole separation. This assumes that the holes are parallel. Rearrangement gives

\[
t^2 = \frac{Z_R}{c^2} - \frac{2Z_T}{c^2} \frac{Z_R}{c^2} + \left( \frac{D^2}{c^2} + \frac{Z_T^2}{c^2} \right)
\]

This is the equation of a hyperbola in \( t \) and \( Z_R \). In a geological situation where the assumption of uniform velocity is valid, the hyperbola of first arrivals can
Figure 3.7. A seismic fan collected with the transmitter fixed in hole M1a, and the receiver traversed in hole M1b. Amplitude changes from trace to trace are not easily interpreted since all seismograms are scaled to have the same peak-to-peak maximum amplitude. The tangential shear channel is displayed. All channels had similar travel time and amplitude characteristics for the first arrivals.
provide a measure of the velocity that is dependent only on the relative time between seismograms. This is done by using the travel time at the apex as the reference, and dropping all but the $Z^2$ term of equation 3.2.2. If the absolute travel time is known, the depth of the transmitter and the separation between holes may also be determined.

A seismic fan from E1 is shown in Figure 3.7. The transmitter was fixed at a depth of 100 m in M1a, and the receiver moved in 1.13 m steps in M1b. The first arrivals at depths between 67.8 m and 137.9 m are clear, and seem to lie on a smooth hyperbola. A least-squares linear regression of the square of the measured time on to the measured depth, the square of the measured depth, and a constant (ie Equation 3.2.2) was done to verify the experiment geometry. It was accomplished using the standard relation

$$B = (X^TX)^{-1}X^TY$$

for the vector of dependent variables $Y$, the matrix of independent variables $X$, and the coefficients $B$. The routine used (NAG G02CJF) is part of a commercially available library produced by The Numerical Algorithms Group (1982).

The confidence interval for the coefficients was estimated from the variance-covariance matrix provided by the subroutine, and adjusted for the finite degrees of freedom in the data. Searle (1971) gives the interval for
Figure 3.8. Results of the least-squares fit to the P-wave arrivals of Figure 3.7. The crosses are centred at the measured travel times. The line is the best fit.
the $i^{th}$ regression coefficient as

$$t_f, \alpha / 2 \sqrt{\sigma_{ii}}$$

for $\sigma_{ii}$ the diagonal element of the variance-covariance matrix, and $t_f, \alpha / 2$ an appropriate factor from a $t$-distribution with $f$ degrees of freedom, and $100(1-\alpha)\%$ confidence levels. A factor of 1.5 was used in all calculations, representing an 80% confidence level for 4 degrees of freedom. In most cases more than 4 degrees of freedom were used; the correct factor for each case would reduce the confidence interval indicated only slightly. The interval for each coefficient was then used in the standard error propagation rules to produce error estimates for the calculation of the velocity, receiver depth, and hole separation.

The result of the regression is shown in Figure 3.8. Combining the regression coefficients with Equation 3.2.2 gives a velocity of $5.56 \pm 0.04 \text{ km/s}$, a transmitter depth of $100.83 \pm 0.06 \text{ m}$ and a hole separation of $16.31 \pm 0.02 \text{ m}$. The curve fits the data to within the measurement errors in most places. The difference in hole separation represents a deflection in one hole of less than 2 degrees over 100 m.

The second arrival in the seismograms of Figure 3.7 can easily be identified as shear by the distinctly different curvature of the hyperbola. The shear arrival time is identifiable for receiver depths from 72 to 145 m, with a maximum path length of 47 m. The regression of the travel times (Fig. 3.9) gives a velocity of
Figure 3.9. Results of the least-squares fit to the S-wave arrivals of Figure 3.7. The crosses are centred at the measured travel times. The line is the best fit.
3.24 $\pm$ 0.03 km/s, a transmitter depth of 100.38 $\pm$ 0.1 m and a hole separation of 17.88 $\pm$ 0.02 m. The discrepancy in the hole separation may be due to the scatter in the shear times on the ends of the hyperbola, combined with the low velocity zone in the centre of the fan (discussed below). The points on the ends of the hyperbola contribute significantly to the velocity, since they contribute most to the slope of the asymptotes, which is the reciprocal of the velocity squared. A slight over-estimate of the velocity from this source would inflate the predicted hole separation, as is evident from Equation 3.2.2.

Another fan of data is shown in Figure 3.10. This data is from E2, with the transmitter fixed at a depth of 200 m in URL6, and the receiver raised in steps of 5 m in M2a. The P-wave arrivals are again distinct and approximately hyperbolic, with a maximum transmission distance of 230 m. The shear arrival does not appear in the any of the traces. Since the S-wave was the largest amplitude signal at the smallest separations of E1, much higher attenuation of the shear wave is indicated. The velocity calculated from the hyperbolic fit to the P-wave arrival times (Fig. 3.11) is 5.90 $\pm$ 0.06 km/s, with a hole separation of 177.0 $\pm$ 0.2 m and a transmitter depth of 210.13 $\pm$ 0.03 m. The measured values are 175 m and 200 m, respectively. The elevation difference between the reference points on the well collars is not known, but is less than 10 meters. The difference between the depth
Figure 3.10. A seismic fan collected with the transmitter fixed in hole URL6, and the receiver traversed in hole M2a. The plotting scale is constant from trace to trace so that amplitude changes are preserved.
Figure 3.11. Results of the least-squares fit to the P-wave arrivals of Figure 3.10. The crosses are centred at the measured travel times. The line is the best fit.
predicted from the regression and that measured on the cable is probably caused again by a non-uniform velocity distribution.

The velocities predicted from the hyperbola are produced by the arms, and particularly the lower arms, of the data curves. The coefficient of $Z^2_R$ represents the slope of the asymptotes to the hyperbola, and is used directly in the calculation of the velocity. The regression process gives linear weight to the squared data ($t^2$ and $Z^2_R$) which then dominate the choice of the $Z^2_R$ coefficient. The other coefficients are sensitive to the geometrical parameters, but are of reduced reliability unless the velocity in the central portion of the fan is the same as that on the arms. More detailed inspection of the travel times indicate that both the E1 and E2 fans have velocity anomalies near the apex, and have an apparent increase in velocity with depth.

If verification of the borehole positions is necessary, high density data in a region of nearly constant velocity is required. The curve fitting must be done with care to avoid bias towards the greatest times and depths. Fehler (1978) has done this by linearizing Equation 3.2.2, and using an iterative fitting procedure. In the remaining discussion, the measured probe positions will be used because of the obvious problems in the present analysis. This is supported by the relatively good agreement of the predicted geometries to the measured ones.

Once the position of the transmitter and receiver have
Figure 3.12. Average velocities for E1. The velocities are calculated for each transmitter-receiver pair using the straight-ray path length and the measured travel time. The velocity ratio is also plotted on the same scale, but is unitless.
been established, the analysis of the travel times is possible. The simplest way to analyse the times picked from a fan is to remove the geometric factor by calculating a velocity for each trace, using the straight line distance between the probes. This velocity is an average of the velocities of all materials traversed by the ray.

The average P-wave and S-wave velocities for E1 are plotted in Figure 3.12 as a function of receiver depth. They show a clear low velocity zone in the central portion of the fan. Within the zone the velocities are approximately constant, with a 20% drop for the P-wave and 23% for the S-wave referenced to the highest velocities. The higher velocities calculated from the regression coefficients are due to the lengthened travel times (low velocity) in the central portion of the fan. This caused a flattening of the hyperbola, with an increase in the slope of the asymptotes. The inflated velocity would then contribute to an increase in predicted hole separation when the regression coefficients are recombined as indicated above.

The configuration of the velocity anomaly cannot be quantified with one fan, although the sharp changes in the average P-wave velocity indicate that the zone may extend from 90 to 106 m. The position of the apex of the fan (transmitter at 100 m) in the anomalous zone means that the velocities calculated for receiver positions outside of the zone are lower than the actual value. For the lowermost ray-path, the difference would be about 3.5%, neglecting
refraction. In order to define the boundaries of the zone fully, transmitter positions above and below would be necessary, so that the effects of refraction could be predicted using the correct velocity contrast.

Low velocities can be caused by both lithological and mechanical changes in the rock mass. In the core log of URL6, 200 m away from M1a, the lithology is very regular from the surface to 200 m depth. It is therefore unlikely that a 15 m thick change in lithology exists between M1a and M1b. Fracturing is thus the most probable source of the low velocities. The theoretical developments of O'Connell and Budiansky (1974) would allow an estimate of the crack density provided the P-wave and S-wave velocities in both the fractured and unfractured rock were available. While the velocities in the unfractured rock are not available, the highest velocities in the fan represent the least fractured rock, and can be used to check the compatibility of the field results and the theory.

Figure 3.13 is a reproduction of figure 9 of O'Connell and Budiansky (1974). Using values of

\[ \bar{V}_p = 4.58 \quad \text{(receiver at 101.7)} \]
\[ \bar{V}_s = 2.48 \]
\[ V_p = 5.50 \quad \text{(receiver at 133.3)} \]
\[ V_s = 3.17 \]

Figure 3.13 gives a crack density parameter, \( \hat{E} \), of 0.4 and crack aspect ratio parameter, \( \omega \), of 5.5. These values may in turn be used to calculate the crack porosity, \( \eta \), from
Figure 3.13. Theoretical curves relating P-wave and S-wave velocities and crack density for varying saturation. See text for description of the variables. (after O'Connell and Budiansky (1974))
the relations given by O'Connell and Budiansky (1974)

\[
\omega = \frac{\widetilde{K}}{c} a
\]

\[
= \frac{4\pi \epsilon}{3} a
\]

The compressibility of the rock was calculated from the relationships of Table A1 using the density from a standard gamma-gamma borehole log, $2.7 \times 10^3$ kg/m, and the velocities $U$ and $V$ above. The result was $4.55 \times 10^{10}$ Pa. The compressibility of the water was taken to be $2.25 \times 10^9$ Pa. From these values the crack porosity in the anomalous zone is 1.5%. The use of velocities higher than 5.50 and 3.17 km/s to represent the uncracked rock would lead to higher porosity estimates.

A crack porosity of 1.5% is very high for normal granite. Nur and Simmons (1969b) give experimental results on samples from 3 granites for which the maximum is 0.45%. At high crack concentrations, the self-consistent energy approach used in the development of Figure 3.13 becomes unreliable and the value of 0.4 is near the limit for which O'Connell and Budiansky (1974) claim accurate results. Still, the lack of quantitative theory does not detract from the conclusion that the anomaly is caused by pervasive fracturing in the zone, increasing the bulk crack porosity. The estimate of the crack porosity of 1.5% is not considered reliable in itself, due to the many approximations made in the derivation of the necessary
Figure 3.14. Average velocities for E2. The velocities are calculated for each transmitter-receiver pair using the straight-ray path length and the measured travel time.
Figure 3.15. A simulation of the refraction problem in crosshole seismology using values typical for the data presented. With holes 175 m apart, the first arrival time in rock with a P-velocity of 6.0 km/s is 29.17 ms. A direct arrival on a 175 m path in rock with a velocity of 4.8 km/sec will arrive at 36.46 ms. A arrival with a 10 m path in the slow rock and a 175 m path in the fast rock (longer in time then the true refraction) will arrive at 31.25 ms. Thus using only the first arrival times and the straight-ray path length the minimum velocity will be 5.6 km/sec. The real 20% anomaly is reduced to 6.7%.
input for the theoretical relationships, and because of the limitations of the theory in highly fractured environments.

The above results cannot be compared to measurements made using other techniques at the present time. Because the holes were percussion drilled, no core is available for counting fractures. The casing in M1b makes the direct examination of the borehole wall with a television log or acoustic televiwer impossible. Part of M1a has been inspected by television, but not in the zone of interest.

The average P-wave velocities for E2 fan of Figure 3.10 are plotted in Figure 3.14. There are a number of small anomalies in the data, particularly at 150, 205 and 320 m. The velocities are lower than nearby velocities by between 2% and 3%, well within the measurement accuracy for closely spaced data. The location of the low velocity zones agrees well with the location of amplitude anomalies, and with fracture zones discovered by core and television logs, as discussed in Section 3.4. With large hole separations, the detection of thin elongated features will be complicated by refraction, which will reduce the magnitude of the anomaly. Figure 3.15 shows graphically how this occurs. Prediction of crack density parameters from the velocities in Figure 3.14 is pointless without corrections for refraction.

Another way to assemble crosshole data is to traverse the receiver and transmitter together to maintain a constant path direction. In this geometry each seismogram measures the travel time through a different volume of rock
Figure 3.16. Data collected with the probes traversed together. The transmitter was in hole BC-81-1, and the receiver in BC-81-3, 30 m apart. The measurements were made at 1 m intervals from a depth of 113 m to 20 m.
and is less directly dependent on the geometry since the path length remains unchanged. The presentation is particularly appropriate to the study of relative changes in the rock properties when the geology is uniform or horizontally layered. The initial interpretation scheme assumes that there is little or no refraction taking place, and that the boreholes are co-planar and parallel. While these assumptions are not valid for detailed quantitative analysis, they are sufficient for data evaluation. Butler and Curro (1981) discuss the detailed analysis of such records.

The seismograms in Figure 3.16 are from Bell's Corners and were collected with the transmitter in BC-81-1, and the receiver in BC-81-3, 30 m away. The instruments were at equal depths and were moved up the holes in 1 metre steps. Problems (since rectified) in both the wiring and the logic of the data acquisition electronics during these early tests introduced noise to the correlation process, and effectively removed all amplitude information from the data. Although the noise level in the seismograms is high, the time of arrival of the first seismic energy remains interpretable. Traces without data are the result of instrument malfunction.

The flat lying sedimentary section at Bell's Corners allows direct comparison of velocity structure measured on a traverse with the crosshole seismic system to the geological structure determined from core analysis (Fig. 3.2). The data shows clear changes of the first arrival
Figure 3.17. Theoretical and experimental variations of P-wave velocity with porosity for sandstone. (after Wyllie et al (1956))
time with depth. In the upper 40 meters of the record, assuming a 30 m path, the velocities range from 3.7 to 5.3 km/s. These are consistent with values reported in the literature (Fig. 3.17) for sandstones. This section of the record has significant character, with noticeable fluctuations in the travel time. At a depth 65 m the times begin to decrease systematically corresponding to average velocity changes from 4.3 km/s at 64 m to 5.6 km/s at 78 m. Below 78 m the travel times slowly decrease, to a maximum velocity of 6.3 km/s at the bottom.

This data may be compared directly with the core log and preliminary section. The upper section lies entirely in the Nepean Sandstone. This formation is highly variable, and includes thin bands of conglomerate and turbidite (Bernius, 1981a). The fluctuations in the velocity are most probably due to variations in porosity, a phenomena examined by Wyllie et al (1956), and others. Sandstone was modeled as a system of quartz particles in a fluid matrix and a net compressibility calculated as an average of the two materials. The compressibility was then used to predict the velocity as a function of porosity (Fig. 3.17). If increased porosity were the only cause of reduced velocity, Wyllie's results would indicate a variation from 1% at the highest velocities to 6% at the lowest. The porosity of the core has not been measured, but these values seem low. These rocks are well lithified, and solid bonding between the grains which could produced velocities higher than that expected for packed quartz
grains only.

The core log indicates that a 17 m transition from sedimentary into paleo-weathered crystalline rock begins at about 65 m, which corresponds to the zone of increasing velocity. A simplistic interpretation is that, as the depth increases, an increasing proportion of the travel path is in the faster crystalline rock, instead of the sandstone or the relatively slow alteration materials. The fractures that are indicated in the logs may also act to reduce the velocity, but resolution on the scale of a few centimeters would be necessary to separate lithological and mechanical effects.

The crystalline rocks in the lower part of the section would be expected to produce the uniform high velocity observed. This is due to the more consistent porosity and lithology that is characteristic of igneous rock. Fractures are reported in the core log, but do not appear as distinct velocity anomalies in this data set. Although fracturing can produce local velocity changes of 30% or greater (see appendix A), the time that the wave spends in the fracture zone will be short, producing small changes in the total travel time. Moreover, dipping fractures would affect many adjacent traces, and therefore cannot be resolved with this geometry. The gradual velocity increase in the bottom 40 m may be due to lithologic changes, but may also be due to a drift in the distance between the holes.
3.4 Tube Wave Analysis

Tube waves are compressional waves propagating mainly in the borehole fluid. Their velocity is reduced from the bulk fluid value by the yielding of the borehole wall. White (1965) has used the steady state mechanical properties of thick-walled pipes to estimate the tube wave velocity as

\[ u_T = u_W (1 + \frac{K_W}{M})^{-1/2} \] 3.3.1

where \( u_W, \rho_W, K_W \) are the velocity, density and compressibility of the borehole fluid respectively. \( M \) is the effective rigidity of the pipe and is calculated from the elastic properties of the pipe material, \( E \) and \( \sigma \), and from the inner radius \( b \), and outer radius \( a \).

\[ M = \frac{E(a^2 + b^2)}{2[1 + \sigma(1 - \sigma)(a^2 - b^2) - 2\sigma b^2]} \] 3.3.2

If the outer radius is very large this reduces to

\[ M = \frac{E}{2(1 + \sigma)} = \mu \] 3.3.3

the shear modulus of the pipe material. This result is a low frequency, long wavelength approximation. As was seen in the previous section, the present system has not detected direct shear arrivals at distances of greater than 50 m in sound granite. White's result (Eqn. 3.3.1), holds some possibility for measuring the insitu shear modulus,
Figure 3.18. A partial fan of data collected in the E2 holes. The left data set is from the vertical shear channel, the centre from the radial compressional channel, and the right from the tangential shear channel. Note that the sixth trace from the top of the right hand data set is from the radial channel. Automatic scaling in the plotting routine has removed much of the amplitude information.
provided tube waves are present.

The distortion of the borehole wall due to a tube wave is similar to the Rayleigh and Stoneley waves of planar geometry. Rayleigh waves are plane-polarized with particle motion that is elliptical on the boundary. By comparison, tube waves have only vertical and radial components of motion at the borehole wall.

A strong tube wave event is shown in a set of seismograms in Figure 3.18. All three channels are displayed, and show the polarization characteristics of the tube wave clearly. Channel one is sensitive to vertical borehole motion, channel two to radial motion, and channel three to tangential motion. The tube wave originates at a depth of 300 m in the receiver hole, and is seen to be propagating in both directions, with large amplitudes in channel one and moderate amplitudes in channel two. The wave is undetected in channel three. These results provide good evidence that the probe does have selective sensitivity to the different components of motion, and that it is operating in an accelerometer mode. If the probe were acting as a hydrophone (i.e., pressure sensitive) the radial component would have a large response, and the two shear channels would have small responses, caused by internal conversions in the transducer.

Tube waves are observed in most seismic experiments carried out with the receiver placed in a water-filled borehole. In many instances, the events are considered to be noise. The high amplitude tube waves dominate the
record at times when low amplitude events such as reflections and shear wave arrivals are expected. This is particularly true in vertical seismic profiling work (i.e. Hardage, 1981).

Tube waves can be produced in different ways. Hardage (1981) has observed tube waves caused when a body wave passes the borehole at the location of a large impedance change. Tube waves were also created by a change in casing thickness, and by the bottom of a hole. This mechanism was also discussed by White (1965) in a derivation of tube wave reflection coefficients. Kitsunezaki (1971) has theorized that tube waves can be generated by an injection of pore fluid into the hole from a permeable formation stressed by an incident body wave. Huang and Hunter (1982) have developed a scheme for inferring fracture permeability from the amplitude of the tube wave generated by the mechanism suggested by Kitsunezaki (1971). Paillet and White (1982) discuss the identification of fractures from the tube wave amplitudes observed in acoustic logs with full-waveform recording.

There are four tube wave events in Figure 3.19 (seismic fan E2). The strongest source is located at a receiver depth of 300 m and it sends strong tube waves in both directions in the hole. This depth matches with a highly fractured zone in the television log (Fig. 3.20). Another downward travelling tube wave originates at about 155 m, and may be caused by the bottom of the casing, or by the fracturing which is indicated in the log. The upward
Figure 3.19. This is the radial channel of the same seismic fan as shown in Figure 3.7. Four tube wave events are present, which can best be identified by comparison to Figure 3.21.
Figure 3.20. Fractures detected by television logs. (after Dugal 1982)
arm of this event appears to be blocked by the casing which is present above. The fourth event originates in a fracture zone at 315 m depth. The upward propagation from this source is faint, and disappears above 300 m. This may be due to attenuation or scattering caused by the fractures between 300 and 315 m.

The measurement of velocities for the tube waves is simplified by the unattenuated and undispersed nature of the waveform. The starting time of the wave is not accurately known, but only the relative travel time between traces is important. The velocity can be obtained best by identifying the arrival time of a particular phase in the waveforms, rather than the actual onset time.

Times chosen in this manner for the tube waves of E2 are shown in Figure 3.21. The measurement error in the relative travel times is less than $\pm 0.035$ ms ($\pm 1/2$ digitizing interval). The error in the relative depth measurements is $\pm 0.01$ m. A typical error for the tube wave velocity between any two adjacent depths is $\pm 1\%$. The main source of error is timing, which could be improved by using either interpolation or a smaller digitization interval.

An alternative way to increase the reliability of the tube wave velocity determination is to use a number of data points in a least squares regression. This provides an average velocity over a particular interval for which confidence limits can be calculated. This was done for the data of Figure 3.21 in a manner identical to that used in
Figure 3.21. Travel times selected for the events of Figure 3.19 are indicated by crosses. The lines represent the least-squares fit to the data. Event 1t is divided into two parts as discussed in the text.
### Table 3.2

**Tube Wave Velocities from Regression Analysis**

<table>
<thead>
<tr>
<th>Tube Wave</th>
<th>Velocity (km/sec)</th>
<th>80% Confidence Interval</th>
</tr>
</thead>
<tbody>
<tr>
<td>tt</td>
<td>1.372</td>
<td>.002</td>
</tr>
<tr>
<td>ut</td>
<td>1.387</td>
<td>.003</td>
</tr>
<tr>
<td>lt</td>
<td>1.379</td>
<td>.006</td>
</tr>
<tr>
<td>bt</td>
<td>1.392</td>
<td>.006</td>
</tr>
<tr>
<td>lt1</td>
<td>1.36</td>
<td>.02</td>
</tr>
<tr>
<td>lt2</td>
<td>1.39</td>
<td>.01</td>
</tr>
</tbody>
</table>
the previous section. In this case the receiver depth was regressed onto the travel time and the intercept, i.e.

\[ Z_R = c t + Z_{RO} \] 3.3.4

The velocities obtained in this manner are listed in Table 3.2.

The combination of Equations 3.3.1 and 3.3.3 allows the calculation of the shear modulus from the tube wave velocity and the velocity and density of the borehole fluid.

\[ \frac{1}{\mu} = \frac{1}{\rho_w} \left[ \frac{1}{V_T^2} - \frac{1}{V_W^2} \right] \] 3.3.5

Experimental verification of this result has been reported by Riggs (1955) and White (1965), from measurements made in sedimentary rock. Kitsunezaki (1971) on the other hand, reports that shear velocities calculated by combining known formation densities with the tube wave velocities are smaller than measured shear velocities by a consistent factor of 0.53. His result was obtained with a variety of materials, including soils. Although direct measurements of the shear velocity is not possible with the data from E2, comparison with the results of the direct measurements from E1 above will certainly allow the detection of a factor of 0.5 given the uniformity of the lithology and the similarity of the P-wave velocities.

Equation 3.3.5 requires other parameters besides the tube wave velocity. Gamma-gamma logs show a uniform density of 2.72 +/- .02 \times 10^3 \text{ kg/m}^3 through this region.
### TABLE 3.3

**VARIATION OF TUBE WAVE VELOCITY WITH FLUID TEMPERATURE AND SALINITY**

\[ V = \left( \frac{1}{\frac{1}{V_w^2} + \frac{\rho_w}{\mu}} \right)^{\frac{1}{2}} \]

For:
- \( \mu = 2.43 \times 10^9 \text{ Pa} \)
- \( V_s = 3.00 \times 10^3 \text{ m/s} \)
- \( \rho_k = 2.7 \times 10^3 \text{ kg/m}^3 \)

**FLUID: DISTILLED WATER**

<table>
<thead>
<tr>
<th>T (°C)</th>
<th>( \rho_w (\text{kg/m}^3) )</th>
<th>( V_w (\text{m/sec}) )</th>
<th>( V_t (\text{m/sec}) )</th>
</tr>
</thead>
<tbody>
<tr>
<td>0</td>
<td>1.0000</td>
<td>1404</td>
<td>1350</td>
</tr>
<tr>
<td>10</td>
<td>0.9997</td>
<td>1446</td>
<td>1388</td>
</tr>
<tr>
<td>20</td>
<td>0.9982</td>
<td>1482</td>
<td>1419</td>
</tr>
</tbody>
</table>

**FLUID: SEA WATER (3.5% SALINITY)**

<table>
<thead>
<tr>
<th>T (°C)</th>
<th>( \rho_w (\text{kg/m}^3) )</th>
<th>( V_w (\text{m/sec}) )</th>
<th>( V_t (\text{m/sec}) )</th>
</tr>
</thead>
<tbody>
<tr>
<td>0</td>
<td>1.028</td>
<td>1449</td>
<td>1390</td>
</tr>
<tr>
<td>10</td>
<td>1.027</td>
<td>1490</td>
<td>1425</td>
</tr>
<tr>
<td>20</td>
<td>1.025</td>
<td>1522</td>
<td>1454</td>
</tr>
</tbody>
</table>
TABLE 3.4

VARIATION OF SHEAR MODULUS WITH TUBE WAVE VELOCITY

Using:

\[
\frac{1}{\mu} = \frac{1}{\rho_w} \left[ \frac{1}{V_T^2} - \frac{1}{V_W^2} \right]
\]

For:

\[
V_W = 1.445 \times 10^3 \text{ m/s} \quad \rho_W = 2.7 \times 10^3 \text{ kg/m} \quad \rho_W = 0.9997 \times 10^3 \text{ kg/m}
\]

<table>
<thead>
<tr>
<th>(V_T) (km/s)</th>
<th>(\mu) (km/s)</th>
<th>(V_0) (km/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.330</td>
<td>1.148</td>
<td>2.062</td>
</tr>
<tr>
<td>1.350</td>
<td>1.419</td>
<td>2.293</td>
</tr>
<tr>
<td>1.370</td>
<td>1.833</td>
<td>2.606</td>
</tr>
<tr>
<td>1.390</td>
<td>2.543</td>
<td>3.069</td>
</tr>
<tr>
<td>1.410</td>
<td>4.042</td>
<td>3.869</td>
</tr>
</tbody>
</table>
The tube wave velocities are close to the fluid velocity because of the high strength of crystalline rock. The results are therefore very sensitive to the velocity and density of the borehole fluid. An indication of this is given in Table 3.3, where tube wave velocities are calculated using a constant shear modulus and borehole fluid parameters for different temperatures and salinities. Shear parameters for the range of tube velocities of Table 3.3, calculated with assumed water temperature of 10°C are given in Table 3.4. Temperature logs in URL 6 (Soonawala, 1982) and M2a (Tomsens, 1983) indicate temperatures of 8.0 to 9.5°C between 100 m and 350 m depths. A fluid resistivity log for URL 6 (Soonawala, 1982) indicates fluid resistivities below 7.0 X 10^4 ohm·cm, which corresponds to a salinity of less than 0.5%. The volume of other dissolved solids in the borehole fluid is unknown but is likely much less than 1%.

The results in Table 3.5 have been calculated for distilled water at the extremes of the allowed temperature range. Referring to the direct shear velocity measurements of the previous section, it is clear that the values from the tube waves span a reasonable range for the true shear velocity. This is in agreement with White (1965) and Riggs (1955) and disagrees with Kitsunezaki (1971).

The variation in fluid parameters are expected to be small between closely spaced measurements, assuming that the hole is in thermal and chemical equilibrium. However, the results of Table 3.2 indicate real fluctuations in the
TABLE 3.5

SHEAR PARAMETERS FROM TUBE VELOCITIES

\[
\frac{1}{\mu} = \frac{1}{\rho_w} \left[ \frac{1}{U_{T}} - \frac{1}{U_{W}} \right]
\]

for distilled water at 10 C: \(\rho = 999.7 \text{ kg m}^3\), \(U = 1446 \text{ m s}^{-1}\)

for distilled water at 8 C: \(\rho = 999.9 \text{ kg m}^3\), \(U = 1438 \text{ m s}^{-1}\)

\begin{tabular}{cccc}
10°C & & & 8°C & & \\
\hline
\(U_T\) & \(\mu\) & \(U_S\) & \(\mu\) & \(U_S\) \\
km/s & Pa & km/s & Pa & km/s \\
1.372 (tt) & 1.887 & 2.65 & 2.099 & 2.79 \\
1.387 (ut) & 2.406 & 2.99 & 2.761 & 3.17 \\
1.379 (lt) & 2.100 & 2.79 & 2.366 & 2.96 \\
1.392 (bt) & 2.643 & 3.13 & 3.078 & 3.38 \\
1.36 (lt1) & 1.60 & 2.44 & 1.75 & 2.53 \\
1.39 (lt2) & 2.54 & 3.07 & 2.94 & 3.30 \\
\hline
\end{tabular}

shear velocity observed directly in data set E1 is 2.5 to 3.0 km/s, increasing with depth.
tube wave velocity on a scale much smaller than can be explained by the fluid variations indicated by the resistivity and temperature logs. Event bt (Figure 3.21) originates at the bottom of the fracture zone, and gives a shear velocity of 3.38 km/s. The velocity for all of the points of lt give a velocity of 2.96 km/s, slightly lower than for bt. The upper portion of lt (lt1), which lies entirely in the fracture zone, gives a shear velocity of 2.53 km/s (at 8°C). The lower portion, travelling in the same volume of rock as bt, gives a value of 3.30 km/s. The temperature log Tomsons (1983) indicates slightly elevated temperatures between 315 and 322 m, with water flow that is correlated with slightly saline water in URL 6 (Soonawala 1982). These factors would act to increase the tube velocity, and hence the predicted shear velocity, in the region of lt1. Despite this, the fracture zone at 315 m appears to produce at least a 23% decrease in shear velocity. Since the tube wave samples only the rock close to the borehole, this result can only be extended to the rock mass if independent evidence of a continuous anomalous zone exists.

Another type of tube wave event is evident in the lower portion of E1 (Figure 3.7). The events between 35 and 45 ms are similar in appearance to the direct arrivals but occur much too late in the record to be body waves direct from the transmitter. Wong (1982) has suggested that these could be the result of tube wave propagation in the transmitter hole, converted at fractures or zones of
Figure 3.22. Travel times selected for the events of Figure 3.7 are indicated by crosses. The lines represent the least-squares fit to the data.
high impedance variations to body waves.

In order to test this hypothesis, times were chosen for two T-wave to P-wave events, labelled on Figure 3.22 as t1p, and t2p. T-wave to S-wave arrivals are also apparent, but are difficult to time because of interference from other arrivals. Scatter in the data, combined with the increased number of variables, made simple regression of the T-P times unreliable. By subtracting from the times a value corresponding to an assumed tube wave velocity and path length, the problem reduces to that of Section 3.2. A tube velocity of 1.38 km/s for t1p was consistent with a P-wave velocity of 5.9 km/s and a T-wave to P-wave conversion depth of 146 m. The same velocity for t2p corresponds to a p velocity of 6.0 km/s, and a "source" depth of 149 m. Since the transmitter hole is reported to be 150 m deep, t2p may be due to a conversion at the bottom of the hole. The source of t1p is not at the bottom, indicating that fractures may be responsible.

Although two tube waves can be identified in the E1 record the velocities are lower than expected. Event pt has a velocity of 1.36 km/s, and event tst (a T-wave to S-wave to T-wave arrival) has a velocity of 1.32 km/s. The direct shear velocity is about 3.1 km/s in the vicinity of these events, which would normally give a tube wave velocity of about 1.39 km/s. The velocity reduction could be due to the casing. The effect of casing on tube wave velocity has been discussed by White (1965) and Henriet et al (1983), but only for the case where the rigidity of the
pipe exceeds the shear modulus of the hole. Although the necessary parameters for the calculation of the casing rigidity are not available, standard casing for holes of this size would have a rigidity lower than the 29 GPa shear modulus calculated from the formation density and shear velocity.

3.5 Amplitude Variations

The discussion of amplitude repeatability in Section 3.2 illustrates the difficulties involved in extracting meaningful values for the attenuation of the body waves through the rock mass. Much of the data that has been collected shows distinct amplitude anomalies, however. As will be seen below, amplitude measurements promise to be useful for locating zones of high attenuation. Since variations in crystalline lithology are unlikely to cause energy loss in seismic waves, such zones may be unambiguously interpreted as zones of anomalous porosity. Attenuation mechanisms for crystalline rock are reviewed in Appendix A.

The data set of E2 (Figure 3.10) is a good example of the type of amplitude variations observed. When the travel times were selected, the amplitudes were measured in the manner described previously. The results for the two frequency bands were summed to give the total amplitudes. These were multiplied by the geometrical length of the particular ray path to eliminate the effects of spherical spreading, and are plotted in Figure 3.23.
Figure 3.23 A graphical representation of the data available for the seismic fans of E2. The average P-wave velocity and the average attenuation value are plotted in the central portion as a function of depth. Velocities are in km/s, and amplitudes as described in the text. The straight-ray paths are drawn over these, and the fracture logs (television and core combined) included for comparison.
There are four zones of low amplitude in the data set. The smallest (Zone B) spans two traces at 200 and 205 m, where mild attenuation is indicated, matching a low velocity zone (fig. 3.10). Zone A is located between 140 m and 175 m, and has very low signal levels. It is associated with the largest velocity anomaly. The amplitudes above this zone are lower than those for equal path lengths (i.e. 115 m compared to 285 m) in the lower section of the fan. The seismograms collected between 245 and 265 m (Zone C) show high attenuation of the high frequencies with the low frequencies relatively unaffected, and minor, if any, velocity reduction. All seismograms below 295 m (Zone D) show signal loss, with high losses indicated at depths near 300 and 320 m. Again, low velocities are found at the same locations as the low amplitudes.

The geometrical ray paths for this data are plotted in Figure 3.23 with the television logs for both boreholes. The zones of low amplitude translate directly into zones of high fracture frequencies in the log for M2a, with the exception of Zone B. The zone of highest signal loss, Zone A, matches the zone of highest fracture frequency, with Zones D and C corresponding to lower frequencies, respectively.

Finding the connections between the fractures located by television logs in different holes is necessary for the type of hydrogeological assessment planned for radioactive waste disposal sites. A standard method is to isolate
particular fractures and then to pump traceable fluids along them. For environments with large hole separations, and multiple fracture zones with low permeability, this can be a slow process. The importance of crosshole seismology is its ability to sample the rock mass away from the borehole. This allows the extrapolation of data collected with high resolution borehole logging techniques into the rock mass. A qualitative examination of the data presented shows how this can be done.

The first conclusion that can be drawn from the seismograms (Figure 3.10) is that the fracture zone indicated at 265 m in URL 6 (Figure 3.22) is not directly connected to the zone at 250 m of M2a. The high amplitudes between 270 and 290 m would not be possible if this connection were made. The attenuation at 250 m could only be caused by a fracture zone local to M2a, or one that was parallel to the ray paths ending in the fracture zone. Based upon this, a tentative connection between the fractures at 205 m in URL 6 can be made to those at 250 m in M2a. The high amplitudes below 270 m may indicate a thinning of the zone towards URL 6.

The T-wave to P-wave event that appears below 320 m (Fig. 3.10) is similar to those discussed above. It is a compressional wave arrival from the fractures at 270 m depth in URL 6, and has an amplitude and frequency content comparable to the normal direct arrivals. This indicates that no regions of high attenuation were intersected and puts a lower limit to fracture zone 4 at 320 m. All direct
arrivals from the transmitter depth of 200 m are attenuated, however, which requires a fracture system extending away from M2a between 300 and 320 m. A reasonable case therefore exists for the connection of the fractures at 270 m in URL 6, to those at 310 m in M2a.

The zones above 175 m cannot be defined with this data set. It is clear that some connection is possible, since amplitudes are suppressed above 150 m. Another fan with the transmitter located at 145 m would be necessary to check if the fractures at 150 m in URL 6 are connected to those at 150 m in M2a. This depth may also mark the beginning of more pervasive facturing which produces the velocity gradient at shallow depths noted in the direct arrivals. The fan of data collected for E2 represents half a day of work for a two man crew.

Assuming that the radiation pattern and frequency response of the instrument are well calibrated, there are three main factors which restrict the use of the amplitude data for quantitative determination of rock quality. The most serious of these is the effect of substantial fractures in the borehole wall on signal amplitudes. The repeatability of the mechanical coupling of the instruments will likely be greatly reduced by the increased roughness of the borehole in these regions, emphasizing features which may be local to the hole. In badly fractured zones, small rock chips and fault gouge between the anvils and the borehole wall could dominate attenuation effects of the rock mass.
Quantitative interpretation will also require careful attention to the various wave phenomena that are present in the data. This includes refractive changes of the path length in the attenuating medium caused by the velocity variations in the fractured zones. This is a possible cause of the high amplitudes at the 290 m region of Figure 3.10, which exist despite the fractures indicated at 210 m in URL 6, and at 250 m in M2a. The seismograms between 245 and 265 m all show enhanced low frequencies and attenuated high frequencies, which is strong evidence that scattering and wave-guide or resonance phenomena are taking place. Such effects are expected in situations where the wavelength used is comparable to the target under study. The dominant wave lengths are from 1 to 10 m in the data presented.

The final difficulty is that of modelling the loss mechanisms. Present theoretical and laboratory results (Appendix A) indicate that frictional sliding and viscous water flow will be the dominant mechanisms. The proportion of each that is present in a particular environment is unpredictable, and may depend on the frequency of the wave. Quantitative inversion of amplitude data for crack density parameters will require more theoretical work.
Chapter Four: Conclusions

4.1 Conclusions

A crosshole seismic system has been designed, built and successfully tested. An important feature of the system is its continuous coded waveform transmitter, which allows the calculation of the impulse response of the rock mass by cross-correlation of the received signal with a copy of the transmitted waveform. The source has a steady acoustic power output of the order of a watt which is appropriate for electrodynamic transducers, and a low peak power which prevents damage to the borehole. Piezoelectric elements are employed in an accelerometer geometry and are conveniently driven by a pseudo-random binary sequence as the coded signal. The receiver uses ceramic piezoelectric sensitive to both shear and compressional motions to give three component sensitivity.

Clear compressional wave arrivals have been detected with instrument separations of up to 230 m in unaltered, little fractured granite. Shear wave arrival amplitudes are larger than the P-wave at short distances, but appear to suffer higher attenuation. The maximum distance for shear detection has been 50 m. The presence of tube waves in the records holds some hope for extracting shear parameters in experiments with large hole separations. However the calculation is strongly dependent on borehole fluid parameters which are not always known accurately.

The task of finding and characterizing zones of
anomalous mechanical properties can be accomplished with the system. Although velocity anomalies alone are not enough to distinguish a fracture zone and a change in lithology, the high resolution makes possible accurate comparisons with rock mechanics theory. The presence of high attenuation or results from standard borehole logging can be used to separate lithological changes from porosity changes if necessary. Quantitative interpretation of any crosshole data must include corrections for refractive effects, especially when the vertical extent of the variations are small compared to the hole separation.

A number of deficiencies have so far prevented a full evaluation of system performance. The most serious of these is lack of orientation sensing or control, which prohibits detailed analysis of the radiation patterns of the transducers. The effects of clamping were not fully tested, initially because of instrumentation difficulties and later because of the incompatibility of borehole size and instrument design. The system does transmit and receive seismic energy in the 0.5 to 5 kHz band, and certainly has some selective sensitivity to different components of motion of the borehole wall as shown in tube wave arrivals. Travel time resolution is presently +/- 0.075 ms, and improvement is possible through higher digitizing rates and improved signal processing. The reliability of amplitude measurements will be unknown until uncertainties introduced by the clamping and orientation problems are resolved.
4.2 Alternative Applications

Although the system presented should be of particular use to the problems of the radioactive waste disposal program, there are other applications for high resolution crosshole seismic data. After an ore body is discovered using geophysical reconnaissance methods, extremely high resolution definition of the body is required for planning the mine workings. This is presently done with extensive (and expensive) drilling. In situations where the target body is expected to have a seismic signature, a crosshole system could be used for this purpose. Examples of common mine targets that would normally produce seismic anomalies include massive sulphides (P-velocity of 4 km s\(^{-1}\) (Waboso and Meren, 1978) in crystalline host) and coal (P-velocity from 0.9 (Sontang and Wolfe, 1983) to 2.5 km s\(^{-1}\) (Mosher and Mason, 1983) in sedimentary host).

The engineering of any excavation requires data on the mechanical properties of the host material. While the high operating frequencies of the system make foundation and soil evaluation unfeasible, tunnelling and mining applications are possible. At the present time the data required is often collected using hydraulic presses for quasi-static determinations of the stress-strain relationship. This method has an inherent shortcoming in that only a small volume of rock is evaluated.

As can be seen in Table A1, the P-wave velocity, S-wave velocity and density completely define the elastic properties of a perfectly elastic material. Using the
Figure 4.1. Young's modulus and shear modulus calculated from the average velocities in Figure 3.12, and the density measured with a gamma-gamma borehole logging device.
density measured by a standard gamma-gamma density log, the velocities of Figure 3.12 were used to calculate the dynamic Young's (E) and shear (G) moduli for E1 (Figure 4.1). An anomaly in E is very pronounced, with a maximum difference of 33%. These values are representative of the entire rock mass between the holes, and not the small region around the hole sampled by acoustic logging or quasi-static devices like a Goodman press. Although comparisons of quasi-static measurements to dynamic measurements in the laboratory do not show complete agreement (Simmons and Brace, 1965), empirical evidence could allow the combination of a small number of quasi-static measurements with a crosshole survey to give complete coverage of an excavation site.

4.3 Recommendations

Although the present system is adequate as is for high resolution investigation of crystalline rock velocities, there are changes that should be made to improve the performance. A new high voltage circuit has been designed and bench tested by G. F. West. This circuit can provide nearly twice the driving voltage presently available and also allows the use of higher clock frequencies. By doubling the frequency, the first null in the PRBS power spectrum will move out of the operating band. Combined with the increased voltage, this should significantly increase total signal power, and thereby the maximum range of the system. The circuit should be incorporated into the
transmitter probe as soon as possible.

Amplitude or attenuation studies appear to hold great promise for detecting fracture systems in crystalline rock. Unfortunately such measurements are difficult with the system. The main difficulty is the inability to remove the effects of the instrument radiation pattern, since there is no way to control or monitor probe orientation. Direct measurement of the azimuthal orientation could also be done using magnetometers, provided no strong anomalies in rock magnetic susceptibility are present and that the magnetic field sensors are far enough from the clamping mechanism drive. The inclination could be measured with a simple pendulum-type dip meter, although high sensitivity is difficult in the small cross-section available. If sufficiently accurate these measurements could be used to calculate the borehole location, removing a major source of error. Direct control of the probe orientation will require increased mechanical complexity of the probes which may not be desirable or necessary, since proper calibration of the radiation pattern will allow amplitude corrections from the orientation measurements.

The repeatability of the clamping is also very important for amplitude studies. Good clamping can increase signal amplitudes by a factor of two or more, but consistency is required. The clamping mechanism discussed above must be fully tested and improved if necessary. The importance of this factor would allow for increased complexity in the mechanical design of the probe. Either a
caliper arm or hydraulic ram would provide higher clamping forces, but require high cost engineering to fit into a 65 mm probe.

Finally, the sophistication of the interpretation can be much increased. While the geometry of anomalous zones can be estimated from single sets of seismograms, the real strength of the crosshole technique is that it allows full probing of the entire rock mass between the holes. Full data sets (ie many ray paths from different probe positions) will allow the complete two dimensional reconstruction of the plane between the holes. Reconstruction techniques for geophysical applications have been discussed by Bois et al (1972), Lytle and Dine (1980), Lakshmanan et al (1982), Gustavsson et al (1982), Wong, Hurley and West (1983) and others. Most of the techniques presently in use do not adequately address problems caused by refraction and incomplete data sets (ie no vertical ray paths). Nevertheless, comparisons between known geology and the tomographic reconstructions have been qualitatively successful as reported in the literature. At present, reconstructions provide a powerful tool for reducing large crosshole data sets to a form that is manageable for interpretation purposes. It is hoped that future developments in seismic tomography will enable the production of a quantitative cross-section of the elastic properties of a rock mass spanned by a pair of boreholes.
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Appendix A: Rock Physics

A.1 Introduction

For the purposes of this thesis, the term crystalline rock refers to all igneous and metamorphic varieties with low porosity (typically <1%, insitu, at depth). Even though the porosity is low, formations of crystalline rock usually contain a range of physical discontinuities, such as intergranular cracks and voids, joints and joint systems, and even major faults and fractures. Nevertheless, crystalline rock is usually assumed to be perfectly elastic macroscopically, particularly at the low strain amplitudes associated with exploration seismology. The assumption has been tested on the laboratory scale (i.e. Simmons and Brace, 1965) by comparing measurements of various elastic parameters, through the relationships shown in Table A1, and has been found to be valid at moderate confining pressures (> 100 kPa). At lower pressures the effects of open cracks cause irregularities in the stress-strain relationship, and there is some departure from perfect elasticity. The presence of cracks and the fluid that fills them in most environments can also cause attenuation and dispersion of seismic energy, although seismic losses in crystalline rock are normally much lower than losses in other rock types.

Mechanical properties can be measured directly in the laboratory on small samples. Static or quasi-static measurements of the stress-strain relationship in various
<table>
<thead>
<tr>
<th>TABLE A1  ELASTIC RELATIONSHIPS AND DEFINITIONS</th>
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\[
\begin{align*}
\rho &= \text{Density} \\
V_p &= \text{Compressional Velocity} \\
V_s &= \text{Shear Velocity} \\
\beta &= \text{Adiabatic Compressibility} \\
E &= \text{Young's Modulus} \\
G &= \text{Shear Modulus} \\
K &= \text{Bulk Modulus} \\
\lambda, \mu &= \text{Lame Constants} \\
\sigma &= \text{Poisson Ratio} \\
\chi &= \text{Axial Modulus}
\end{align*}
\]

Seismic Velocities In Terms Of Elastic Moduli

\[
\begin{align*}
V_p &= \frac{\lambda + 2\mu}{\rho} \frac{1}{2} \\
V_p &= \frac{E (1-\sigma)}{(1-2\sigma)(1+\sigma)} \frac{1}{2} \\
V_p &= \frac{K + 4/3 G}{\rho} \frac{1}{2} \\
V_s &= \frac{\mu}{\rho} = \frac{G}{\rho} \frac{1}{2} \\
V_s &= \frac{E}{\rho} \frac{1}{2} \frac{1}{2(1+\sigma)}
\end{align*}
\]

Elastic Moduli in Terms Of Seismic Velocities

\[
\begin{align*}
\sigma &= \frac{1}{2} \frac{(V_p/V_s)^2 - 2}{(V_p/V_s)^2 - 1} \\
G &= \rho V_s^2 \\
\beta &= \frac{1}{\rho(V_p^2 - 4/3 V_s^2)} \\
E &= \frac{2G(1 + \sigma)}{} \\
K &= \frac{1}{\beta} \\
\mu &= G \\
\lambda &= \frac{2G\sigma}{1-2\sigma} \\
\chi &= K + 4/3 G \\
E &= \frac{\mu(3\lambda + 2\mu)}{\lambda + \mu} \\
\sigma &= \frac{\lambda}{2\lambda + \mu} \\
\sigma &= \frac{3K - 2G}{6K + 2G}
\end{align*}
\]
geometries allow direct calculation of the elastic moduli, and are reported extensively in the engineering literature. Dynamic measurements using resonant bar techniques have also been used successfully to measure a number of elastic parameters, but are difficult to extend to in situ conditions of pressure, temperature and saturation, because of the interaction between the confining medium and the resonator. The method best suited to exploring the effects of environment on the sample is also the most applicable to high-frequency seismology. In a direct parallel with seismic measurements, ultra-sonic transmitters and receivers are used to study wave propagation through the sample, and measure the wave velocity and attenuation from which the elastic moduli may be calculated.

A.2 Velocity Measurements in the Laboratory

The presence of physical discontinuities, for example micro-cracks, pore spaces, and joints, have a large effect on seismic wave propagation in laboratory measurements. Hughes and Cross (1951), who were among the first to use the pulse method, showed this by measuring P-wave velocity as a function of pressure in a sandstone (Figure A1). At low pressure the elastic moduli of the bulk rock is reduced by the distribution of fine pore spaces (gaps) between the mineral grains, resulting in a lower P-wave velocity. As the pressure is increased, the spaces are forced to close, and the elastic moduli rise towards those of the homogeneous mineral grains.
Figure A1. Velocity versus pressure for a dry (x) and saturated sandstone (o). The pore pressure equals the confining pressure for the saturated case. (after Hughes and Cross (1951))

Figure A2. Velocity versus pressure for a granite sample. (after Nur and Simmons (1969a))
Birch (1960, 1961) presented a systematic study of P-wave velocity for many types of crystalline rock. In some cases the velocity changed by as much as 50% over the pressure range 0-1 GPa, while variations between different rock types (at high pressures where the crack effect is small) were about 15%. Identical behaviour is reported by Simmons (1964) in a study of S-wave velocity for the same samples used by Birch.

The effects of cracks are also strongly dependent on the amount of porewater present. The problem has been addressed by many authors, including Todd and Simmons (1972) and Spencer and Nur (1976). Nur and Simmons (1969a,b) have shown that the reduction in P-wave velocity at low pressures is much less significant for saturated samples, probably due to the decreased compressibility of the void caused by the presence of the water (Figure A2). If the pore fluid is allowed to leave the rock as the confining pressure is increased (i.e. the static pore pressure=0), the velocity increases to the same high pressure maximum as in the dry case. If the pore fluid pressure is made to increase with the confining pressure, the voids do not close, and the P-wave velocity remains almost constant (as was also observed by Hughes and Cross, 1951). The S-wave velocity is only slightly changed by the presence of fluid, since fluids cannot support shear stresses.

The effects of temperature on velocities has received only a small amount of attention. At the low temperatures
and pressures expected in the shallow crust, changes in the crack structure and in the fluid saturants as a function of temperature can be expected to have the largest contribution. King and Paulsson (1981), in an experiment on a small block of granite in a mine working, measured an increase in velocity with increased temperature. The maximum temperature used was 75 °C. The increase was attributed to the closing of micro-fractures due to thermal expansion. At higher temperatures, increasing temperature resulted in velocity decreases, possibly the result of the either vapourization of insitu water, or thermally induced fracturing. Spencer and Nur (1976) show that at temperatures up to 400 °C, reductions in the velocities could be related to changes in the compressibility of the pore fluid, with some possible changes in the shear modulus of the rock above 300°C. They also showed that permanent fracturing occurred in the samples with each exposure to successively higher temperatures. Experiments of this type are important to the radioactive waste disposal program, since it is expected that the waste will release considerable thermal energy into the rock.

A.3 Theoretical Causes of Velocity Variations

As a result of the laboratory observations, theoretical work on the mechanical properties of crystalline rock has concentrated on the effects of porosity on an otherwise homogeneous rock matrix. Cracks in particular, even when they represent only a fraction of
the total porosity, are most important. Walsh (1965) compared theoretically the effects of isolated spheroids, ellipsoids, and penny-shaped cracks on the static compressibility and found that the cracks caused as large an increase in compressibility as did spheroids with equal cross-section. Crack-like shapes are therefore favoured in much of the more recent work, and crack porosity, instead of total porosity, is the principle parameter of the models.

The dynamic deformation problem is of greatest interest to seismic interpretation and has received attention from many authors, including Kuster and Toksoz (1974a,b), O'Connell and Budiansky (1974), Toksoz et al (1976), Budiansky and O'Connell (1976), Bruner (1976), and Chatterjee and Mal (1978).

Kuster and Toksoz (1974) have made use of scattering theory with a long wavelength assumption and with a low density of inclusions (saturated cracks) so that multiple scattering could be ignored. These calculations showed that the presence of dry cracks could cause a significant reduction in the P and S velocities, which could persist for S even in the case of total saturation. Toksoz et al (1976) provide some experimental confirmation of the model (Figure A3), but with heavy emphasis on the sedimentary environment.

O'Connell and Budiansky (1974) were among the first to attempt to include crack interaction effects in the wave propagation problem. Using the long wavelength
Figure A3. Comparison of theoretical velocities (Kuster and Tokoz model) with experimental determinations (Nur and Simmons (1969a)). Differential pressure is the difference between the confining and pore pressure. The pore pressure was 0.1 MPa. (after Tokoz et al (1976))

Figure A4. Comparison of the Nur and Simmons data with the O'Connell and Budiansky theory. Some details of the theory are discussed in Section 3.3. (after O'Connell and Budiansky (1976))
approximation, the "self consistent energy" method was developed, in which the effect of each crack is calculated as if it were imbedded in a homogeneous material with the effective elastic parameters of the cracked body. Budiansky and O'Connell (1976) have compared velocities predicted by this work to laboratory measured values, with some success (Figure A4). Although four measured values are necessary, the P and S velocities in the cracked rock and the P and S velocities in the uncracked rock matrix, the theory has been used to predict in situ crack parameters (Wright and Langley, 1980); unfortunately, these predictions have not been independently verified.

Chatterjee and Mal (1978) have expanded on the work of Kuster and Toksoz (1974a) by considering a multiple scattering formulation. The calculations were made using spherical and cylindrical inclusions, because of the complexity of the problem, although they are not a good representation of cracks. Parallel calculations were made using the self consistent method for comparison. Results show that discrepancies occur in the second order coefficients of the crack concentration, an indication that the self consistent method may be inaccurate at high crack porosities. Further difficulties with the self consistent method are illustrated by Bruner (1976), who used a variation of the method to arrive at results substantially different from those of Budiansky and O'Connell (1976) at high crack porosities.

Direct experimental verification of the different
models has been attempted by Hadley (1976) with a sample of Westerly granite. Porosity measurements were made by direct observation with a scanning electron microscope, providing information on both the volume and geometry of the voids. Velocity measurements were taken from the literature and comparisons made with both the Budiansky and O'Connell model, which includes crack interaction, and the Kuster and Toksoz model, which does not. The velocities predicted from the porosity measurements fit the data equally and moderately well for both models. Since this sample did not have a low crack density it seems that the self-consistent method may not represent crack interaction as well as intended, or the crack interaction effects may not be important for modeling the crack densities found in the sample tested.

The effects of large-scale features not present in hand samples, such as faults, fractures and joints, have been poorly studied. This is probably due to the fact that most of the simplifying assumptions made in the micro-crack analysis (i.e., the long-wavelength approximation and the random orientation assumption) are not appropriate for large features. Laboratory samples are normally chosen to exclude such features because of handling difficulties, and because of obvious problems associated with the wavelengths and scales of investigation involved.

Stesky (1979) has simulated jointing by cutting core samples perpendicular to the core axis (thus perpendicular to the direction of propagation) and re-assembling.
Observations indicate that jointing may contribute to the total crack density, but has no other effect. In rocks with high initial crack density the creation of up to four "joints" had only a very small effect on P-wave velocity as a function of pressure, while samples with low initial crack density showed large effects at low pressures. The results were compared by Stesky to a theoretical model proposed by Walsh and Grosenbaugh (1979), with qualitative success. The model treats cracks as planes of indefinite dimensions, held open by a distribution of asperities. At present the theory has only been developed for the case of static compressibility, with the exception of the extension made by Stesky for the comparison of P-wave velocity measurements.

A.4 Attenuation

The fact that seismic waves are attenuated as they propagate through the earth can be seen in any set of seismograms collected as a function of distance from the source. The attenuation loss is expected in even a homogeneous perfectly elastic material, because of geometrical spreading of the wave front. Real earth materials are neither homogeneous nor perfectly elastic, making possible many mechanisms for energy loss from the initial wavefront. These may be divided into two broad categories: elastic mechanisms, which include geometrical spreading, scattering, reflection, and other elastic wave phenomena, and inelastic mechanisms, often called
attenuation mechanisms, through which energy from the wave is dissipated in the rock. Some of the parameters used to describe attenuation and their interrelationships are given in Table A2.

Most experimental work in rock has shown that attenuation is approximately proportional to frequency, or equivalently, that Q is independent of frequency. Knopoff (1964) and Bradley and Fort (1966) provide tabulations of early work. There is great variability in the techniques used for the measurements however, and many important details such as previous lab history and moisture content of the samples are not available. As a result, the absolute values of the data are questionable. Toksoz and Johnston (1981) provide a "state of the art" review of attenuation research, through a comprehensive collection of reprints, and cover the complete range from theory to observations in the laboratory and the field. Measured values for Q in rock range from below 10 in a saturated sandstone (Bradley and Fort, 1966) to 900 for an evacuated limestone (Pandit, 1971) and over 2000 in extremely dry crystalline rock (Tittman, 1981).

Detailed theoretical models have been developed for a variety of attenuation mechanisms. In dry rocks, the most important mechanisms include intrinsic inelasticity in the constituent minerals, relaxation of grain boundaries, and friction on grain boundaries. In the near surface environments of exploration geophysics, and at low strain amplitudes, losses due to friction at grain boundaries
Table A2.

\[ Q = \frac{E}{\Delta E} \]

\( Q = \text{ratio of total energy to energy lost per cycle.} \)

\( \alpha = \text{coefficient of exponential decay in a simple attenuating plane wave model.} \)

\[ A(x,t) = A_0 e^{-\alpha x} e^{-i(kx-\omega t)} \]

Where \( \alpha = \text{coefficient of exponential decay along the wave vector.} \)

Generally \[ \frac{1}{Q} = \frac{2\alpha C}{\omega} \]

Where \( C \) is the velocity of the wave used.

Clearly if the attenuation of the wave is proportional to frequency, \( Q \) must be independent of frequency.

High \( Q \) means small energy loss (\( Q = 500 \) is high in near surface rock)

Low \( Q \) means high energy loss (\( Q = 50 \) is low)
dominate by orders of magnitude. Theoretical developments of this mechanism are given by Walsh (1969), Gordon and Davis (1968) and others. The importance of friction is supported by Pandit (1971) and Tittman (1977), who show that when friction points are locked by complete vacuum drying of a sample, the $Q$ increased dramatically.

The presence of even minor amounts of fluid in the rock has a large effect on the attenuation. As indicated in the paragraph above, the affinity of quartz (and perhaps other minerals) for water means that just the moisture present in laboratory air can lubricate surfaces within the rock enough to allow slippage, and thus frictional losses. Larger volumes of fluid can also have an effect. A number of mechanisms have been suggested for this case, including inertial flow of the fluid relative to the rock matrix, called Biot flow, shear losses near the crack or grain boundary surfaces (Walsh, 1969), and various inter-crack flow, or "squirting" mechanisms (O'Connell and Budiansky, 1977). In the case of partial saturation, even greater fluid flows are allowed, as considered by Mavko and Nur (1979), and Dutta and Ode (1979a,b). Johnston et al (1979) have attempted to evaluate the relative contributions of some of these mechanisms for a sample of sandstone, and some of their results are shown in Figure A5. As indicated by Johnston et al (1979), the complexity of the problem renders such work mostly qualitative.

Features of the rock environment, such as pressure and temperature, will affect the attenuation measured in
Figure A5. Theoretical attenuation coefficients for different attenuation mechanisms as a function of frequency. (after Johnston et al (1979))
crystalline rock. The effects are expected to be similar to those on velocity, since the presence of cracks is the controlling factor in both cases. Pressure has been shown by Gordon and Davis (1968), Toksoz et al. (1979), and others, to decrease the attenuation by as much as a factor of two over a range of 0-400 MPa, which indicates a much larger dependence than for velocity. Temperature dependence has been poorly studied. Tittmann (1977) showed that temperature had a very small effect on a very dry olivine-basalt, while Gordon and Davis (1968) have shown a large effect in a quartzite. The increase in attenuation in the quartzite was probably caused by induced thermal fracturing which would have no effect in the very dry rock where all cracks, old and new, were locked by the lack of lubrication.

Separating the attenuation in a seismic record into compressional component and a shear component is an attractive idea. It is very difficult in the laboratory, since even the slightest amount of scattering or dispersion will both remove energy from the direct arrivals and cause interference. These effects complicate the measurement of direct arrival energy, particularly for the shear wave. Laboratory measurements of attenuation in the literature are most often made with a resonant technique, some of which are torsional, some longitudinal, and they are not directly comparable to the seismic case. Although some attempts have been made to measure separate P-wave and S-wave Q's by pulse methods (Toksoz et al. 1979), coverage
of the problem is incomplete.

Attenuation measurements in situ are rare. In the work that has been done (Ricker, 1953, and McDonal et al, 1958), great care was taken to insure that elastic losses were minimized and predictable, by choosing large volumes of homogeneous rock for study. This sort of requirement is impractical for an exploration tool, however. In the crystalline rock environment insitu, fracture zones and other major discontinuities can be expected to cause reflections and scattering, as well as inelastic attenuation. In situ measurements may therefore contain both inelastic and elastic scattering components. This will make quantitative interpretation more difficult. At the same time, signal amplitudes will be very sensitive to the presence of such features, and thus will provide excellent anomaly detection capability. This is in contrast to velocity measurements, where scattering effects do not interfere so seriously, making quantitative interpretation possible, although absolute changes may be small.

A.5 Crystalline Rock Measurements In Situ

The study of crystalline rock in situ can be broken into four main subsections according to techniques used. Standard methods of reflection seismology have been used in a few cases, with some success. Borehole logging methods have received more attention, particularly for mining and engineering applications. Borehole to surface or surface
to borehole work has been considered by some authors, but few results have been reported. Finally, direct transmission methods have been used in various configurations including crosshole, inter-mine shaft, and across partially isolated blocks.

Many of the difficulties associated with normal reflection seismology in crystalline rock are described by Hajnal and Stauffer (1975). The main problem encountered was that of transmitting sufficient energy into and through a near surface low-velocity layer (estimated thickness was 40 metres), in order to "see" isolated, irregular surfaces with plane wave reflection coefficients of about 0.1. Typical velocities in a meta-volcanic unit were measured by refraction to be about 6.6 km/sec with a surface layer of about 5.8 km/sec. The low velocity was attributed to fracturing and partial water saturation.

A sophisticated reflection survey of a granite batholith has been reported by Mair and Green (1981). A commercially available high-resolution, vibrator-type (MINI-SOSIE) system was used, and a number of data processing and correction methods applied. The resulting sections showed a good reflector at depths of up to 600 metres, which correlated well with a 50 metre thick fracture zone that was intersected by boreholes in the survey region. Care was necessary in processing the data to remove the effects of a 20 metre thick clay overburden.

Refraction surveys are possible (Lam and Wright (1980), Wright et al 1980) and useful for mapping the
bottom of the low velocity surface layer and for finding a representative average velocity for the rock mass under the middle of the profile. However, the method has limited use for probing the rock mass at depth, because there are no strong velocity horizons in the target area to produce headwaves.

The use of single boreholes for detailed seismic analysis is prevalent in the petroleum industry. Acoustic and full-wave logging tools provide important information for evaluating the porosity, permeability and other physical parameters. While the large differences in magnitude of these parameters in crystalline compared to sedimentary rock make direct application of such methods difficult, the use of modified systems is under study.

Geyer and Myung (1971) point out that the full-wave logging systems provide the capability for measuring P-wave velocity, S-wave velocity, and attenuation. From these the elastic moduli can be calculated directly. A comparison of moduli derived in this manner with those measured using static methods in the laboratory (for core taken from the tested holes) showed that the dynamic moduli for a number of crystalline rock types were systematically higher than the static moduli. This is in agreement with the results of Simmons and Brace (1965) (see Section A.2) since the hydrostatic pressure in the hole was higher than that in the laboratory (atmospheric). Similar comparisons for dolomite and anhydrite did not show this effect, probably because of the absence of brittle micro-fractures. The
full-wave logs also showed excellent identification of fracture zones indicated in the core log, both in the calculated moduli, and in the loss of signal amplitude that was evident in the figures presented.

A logging system specifically designed for use in crystalline rock masses has been reported by King et al (1974). It is small diameter (45 mm) has a small transmitter-receiver spacing (305 mm), and operates at high frequencies (>20 kHz). This geometry allows the isolated study of small sections of the borehole wall. Both transducers are forced into contact with the wall hydraulically, which removes the need for borehole fluid for acoustic coupling to the rock. The use of pusher rods makes possible the study of holes of any orientation, with any degree of saturation. The deepest (along hole) reported use is 65 metres (King et al, 1978). It is reported that lithological identification is possible in the absence of fractures, and that fracture zones were detectable, particularly if core was available to identify lithology.

Simmons and Nur (1968) have attempted to compare P-wave velocity as a function of depth in a borehole to the laboratory behaviour of P-wave velocity as a function of pressure. The sonic logs indicated high (about 6.0 km/sec) velocities from the surface to depths as great as 3 km in a body of granite. Over the lithostatic pressure range associated with this depth range (0-100 MPa), laboratory measurements have all shown a large increase in
velocity. This discrepancy may be due in part to the total saturation of the in situ rocks and to substantial horizontal stress at all depths. Fracturing induced in the samples during the coring operation could be a contributing factor. It is interesting to note that local velocity variations in the logs were as great as 15%; this could be an indication of zones of fracturing.

A similar experiment was carried out in an intensely fractured quartz-diorite body close to the San Andreas Fault by Stierm and Krorvach (1979). The results, completely different from those of Simmons and Nur (1968), show in situ velocities much lower than laboratory values at all depths. In a hole that had artesian water flow, the measured velocities ranged from 4 km/sec in intact sections to 2 km/sec in an area identified as a shear zone in the core log. The laboratory velocities were about 6 km/sec. A thorough analysis lead to the conclusion that dilatent macro-cracks were the source of the velocity reduction. The dilatancy was thought to be due to the high in situ stress field, with contributions from high pore pressures and gouge filling.

Pratt et al. (1977) have conducted an in situ experiment by cutting slots in an granite outcrop in order to isolate five sides of a 3 metre cube. Flatjacks were inserted into the slots to allow stressing of the block, which had been chosen to include three joints with a simple geometry. The results agreed with behaviour seen in the laboratory. Increasing stress increased the compressional
velocity (P-wave velocity range was 4.2 to 4.9 km/sec) as did total saturation. It was noted that transmission losses on ray paths perpendicular to the joints made detection impossible without sufficient applied stress to force joint closure. Velocities at the edge of the block were lower than those measured through the centre, and laboratory measurements on samples at room pressure were lower still. This was thought to be the result of induced cracking caused by the release of in situ stress.

A detailed analysis of a fracture system has been attempted by Aki et al. (1982) as part of a hot dry rock geothermal study. The target was a fracture system induced by pumping water into a borehole at high pressure. Using an audio-frequency crosshole transmission method, it was possible to identify a major lithological contact and a major vertical fracture system. Some aspects of the fine structure (on a scale of a few metres) of these features were resolvable through a detailed analysis of the received waveforms. Zones of small cracks (<3m in extent) were identified in both the velocity and attenuation data. It was noted that signal amplitudes were strongly frequency dependent in some regions, an indication that scattering was a major loss mechanism. The experiment used a range of wavelengths from a few tens of centimetres to about 10 metres.

Kaneko et al. (1979) monitored compressional wave travel times and signal amplitudes between nearby mine shafts during a mining operation that was expected to cause
breakage in the test area. Over a six month period the travel times increased by about 3% while signal amplitudes decreased by more than 15%. This agrees qualitatively with laboratory experiments where propagation properties were measured while core samples were strained to breakage. The result supports the view that attenuation measurements are more sensitive to the presence of cracks than are velocity measurements.
Appendix B: Electronics

The design of the electronics was controlled by the desire to operate the transmitter on a 2 km length of standard six conductor logging cable. Because of the need to operate the probes in a variety of modes, including clamping mode, unclamping mode and in either a receive or transmit mode, plus provision for future sensors, conductors could not be dedicated to single operations. The cable length also would have made direct operation of the various features difficult, due to losses along the cable. Instead, the probes were designed to operate on a DC power supply, according to a set of digital commands transmitted on one line, and controlled where necessary by a single analog "signal down". This arrangement requires two power lines, a control line, a "signal up" line, a "signal down" line, and a reference potential (ground). This is shown in Figure B1. Many of the details of the electronic design were worked out by A. Wieckowski.

For the transmitter, the "signal down" line carries either the clamping motor speed control signal or the signal to be transmitted. The "signal up" line carries a monitor signal which produces an audible indication of the load on the motor during clamping operations, the signal waveform during transmission and (in the future) orientation sensor outputs when the sensors are operated. The "probe control" line carries a digital signal which causes the probe to switch from one mode of operation to
Figure B1.

SURFACE ELECTRONICS

- PRBS FROM RECEIVER
- MOTOR CONTROL
- MOTOR MONITOR
- POWER SUPPLY
- SIGNAL DOWN SELECTOR
- FUNCTION ENCODER
- SIGNAL UP SELECTOR

PROBE

- POWER SUPPLY
- SIGNAL DOWN SELECTOR
- FUNCTION DECODER
- SIGNAL UP SELECTOR
- MOTOR
- MOTOR MONITOR
- HIGH VOLTAGE DRIVER

CABLE

TRANSMITTER ELECTRONICS

BLOCK DIAGRAM

EXPANSION CAPABILITY FOR ORIENTATION SENSORS, ORIENTATION CONTROL, ADDITIONAL PROBES, ETC.
another. The design provides for expansion to allow for multiple probes, probe orientation detection and control, or other functions.

The cable available for use with the receiver probe had 24 conductors. This allowed the independent transmission of each of the three channels of data, thus avoiding the need for multiplexing the transducer outputs onto the "signal up" line. In initial tests, a direct connection from the transducer preamplifiers to the surface with a common ground connection was found to suffer from high noise levels due to pick-up of spurious signals. As a result, the final design used 2 conductors for each channel, in a balanced, transformer-coupled configuration. This is shown in Figure B2. The probe control features of the receiver are identical to those of the transmitter.

The high voltage drive circuit is an essential part of the transmitter. It is required to drive a large capacitance (approximately 40 nF in the final design) with a high voltage, essentially square wave. In the circuit used, a DC supply voltage of 30 to 40 volts is switched in polarity across the input to a step-up transformer according to the the PRBS waveform control signal on the "signal down" line. This was accomplished using SCR circuitry and produced a driving signal of about 1 kilovolt peak to peak.

The transmitter drive circuit was not fully satisfactory for a number of reasons. The most serious problem was that the output transformer required a
resistive load in parallel with the transducers to improve the frequency response. The current demand from the SCR’s was increased by the load, with the result that the shortest clock period attainable was 300 μs. A 300 μs base period places the first null in the transmitted signal (see Fig. C4) at 3.3 kHz, or almost in the middle of the desired operating band. The circuit was used because no better one was available at the time.

To obtain maximum voltage output, the receiver transducer elements were in the form of blocks and had low capacitance (∼200 pf). This made special demands on the signal conditioning circuitry. It was essential to minimize the capacitance of the parallel connecting leads so the first stage of the amplifiers had to be placed immediately adjacent to the transducers. This stage was also required to be very low noise, to allow the detection of the smallest possible signals. The downhole electronics also included a set of line driving amplifiers. The transformers used in the coupling to the surface provided a voltage gain of 3 from their winding ratios, and acted effectively as low cut filters, because of their inherent frequency response.

The uphole electronics for the receiver consists of an analog section and a digital section. The analog section is comprised of coupling transformers to complete the balanced transmission lines, fixed gain high cut filters, variable gain broadband amplifiers, and fixed gain variable high cut filters. The total gain available along the lines
is selectable in steps from 30 to 6,000. The low cut
filtering has a corner frequency of about 0.3 kHz, and the
high cut is selectable in steps from 3.3 to 10 kHz, to
prevent aliasing at various digitizing rates.

The digital section is the heart of the signal
processing system. The PRBS is generated with the same
clock that controls the analog-to-digital converter to
ensure high timing accuracy. It is generated in the manner
proposed by Becciolini and Marclay (1974), using a shift
register to implement the recursive algorithm given in
Appendix C. The signals are sampled simultaneously on all
three channels, by means of separate track and hold
circuits, and the samples are transferred to the computer
sequentially via an analog multiplexer. At the end of each
PRBS sequence the electronics generates a signal to the
computer to reset the stacking operation. Overhead in the
computer software limits the minimum time between samples
to 20 μs which gives a maximum Nyquist frequency for 3
channel operation of 8 kHz. When sampling only one
channel, the Nyquist frequency may be increased to 24 kHz.

Both sets of upheole electronics contain identical
circuitry for probe control. Besides this, the transmitter
contains only a power supply and a repeater for sending the
PRBS from the receiver down to the probe. All of the
electronics are supplied from the surface electronic
modules, which are operated from standard 115 v, 60 Hz
power. The total power consumption of the system,
including the computer, is less than 1000 watts, and is
easily supplied by a portable motor generator.
Appendix C: DASP—Digital Acquisition and Signal Processing

The design of the crosshole seismic system presented here is based on the use of cross-correlation of the received waveform with a copy of the transmitted signal. This process will provide high rejection of noise, enabling the detection of very small signals. The waveform used for this purpose was a maximal length Pseudo-Random Binary Sequence (PRBS). A maximal length PRBS is a sequence of randomly distributed binary pulses, and is uniquely described (Foster and Sloan, 1972) by:

- the frequency of a basic clock \( F_c \),
- a length parameter \( n \),
- and coefficients \( C_1, \ldots, C_n \) in the recursion relation
  \[
  Q_j = C_1 Q_{j-1} + C_2 Q_{j-2} + \ldots + C_n Q_{j-n},
  \]
  \( (Q_i = 0 \text{ or } 1, \ C_i = 0 \text{ or } 1, \text{ with modulo } 2 \text{ addition}) \)

provided that the polynomial

\[
1 + C_1 x + C_2 x^2 + \ldots + C_n x^n
\]

is primitive and irreducible and that \( C_n \) is 1. Sequences generated by the recursion repeat exactly every \( 2^n - 1 \) clock periods. An example of a sequence for \( n=5 \) is shown in Figure C.1, where the sequence has been shifted to be double sided, instead of positive only as indicated by the recursion relation.

PRBS's have been used in many applications, particularly for digital communications, and have received
BASIC CLOCK PULSE SEQUENCE

PSEUDO-RANDOM BINARY SEQUENCE (PRBS)

REPEATS AFTER 31 CLOCK PULSES
some attention in the geophysical literature. Goupillaud (1976) and Cunningham (1979) have discussed the use of PRBS coded seismic vibrator signals for multiplexing vibrators and for reducing side lobe amplitudes, respectively. Duncan et al (1980) report on the development of a wide band electromagnetic sounding system which uses a PRBS source.

The cross-correlation process for two PRBS signals is illustrated in Figure C.2. By considering the correlation process as a chain of shift-multiply-add operations, it is clear that the sum calculated for most lags will be of a series of $\pm 1$'s, with a result of $-1$. At one lag, however, the two sequences will exactly align, and the result will equal the number of clock pulses, $2^n - 1$. Thus for an infinite bandwidth signal, the correlation will consist of a comb of triangles with a width of $1/F_c$, height (normalized to the signal) of $2^n - 1$, and a repetition rate of $(2^n - 1)/F_c$ (Figure C.3). The region between the triangles is made up of a single side lobe of magnitude $-1$. For the case of $F_c$ and $n$ large, the sequence is a very good simulation of white noise, and the cross correlation of a reference sequence with one that has been processed through a linear system will be an accurate representation of the impulse response of the system.

Foster and Sloan (1972) have shown that the use of a PRBS with correlation for measuring the impulse response of a linear system will result in a increase in the ratio of signal power to noise power by a factor of $2^{n-2}$, where $n$ is
Figure C2. (after Duncan et al. (1980))
Figure C3. (after Duncan et al (1980))

AUTO-CORRELOGRAM OF PRBS (arbitrary units)