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Date 20 Nov 1995
Abstract

In the early 1980's a group of Russian researchers reported finding a new seismoelectric phenomenon associated with sulphide minerals. They asserted that strong bursts of broadband electromagnetic emissions appeared when a seismic wave passed through a sulphide orebody. This discovery has aroused considerable commercial interest because of the economic importance of mining sulphide minerals for zinc, copper, lead, gold, and other metals. However, there have been no reported follow-up studies of their findings, and no evaluation of the exploration potential of this phenomenon.

This thesis reports a number of field experiments designed to examine the Russian claims and explore the properties of the phenomenon. These field trials have confirmed the existence of this relatively unknown seismoelectric phenomenon and have substantially increased our knowledge of its characteristics. In the tests 0.2-0.5 kg explosive charges were detonated to provide a strong source of seismic energy. When the seismic disturbance passed through a zone of sulphide mineralization, high frequency electromagnetic emissions were produced.

The electromagnetic emissions appear in the form of brief pulses, 2 to 5 microseconds in duration. Typical peak amplitudes, measured 80 to 120 m from the zones of sulphide mineralization, are 10 mV/m and 2 nT. The spectrum of the electromagnetic pulses spans a very wide range of frequencies, from 1 kHz to 3 MHz. Fourier analysis of the digital records from one of our sites shows a peak in the emission spectrum, at 1.1-1.4 MHz. This spectral peak was consistently reproduced from various portions of the sulphide orebody.

New measurement and interpretation techniques were developed to study the electromagnetic signals. These techniques were applied to data from each field trial to demonstrate the potential of seismoelectric techniques for exploration. The results show that the high frequency seismoelectric phenomenon can be used to locate massive sulphides in underground mines.

A physical framework to describe the field measurements has been developed. It is proposed that the seismic wave induces electrification via crack formation in the orebody. The electrified surfaces of the crack recombine rapidly in a gas discharge to produce the observed range of electromagnetic frequencies.
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CHAPTER 1

INTRODUCTION

1.1 Preface

This thesis is an account of my investigation of an unusual and potentially useful phenomenon that can be induced in natural sulphides. The aspect of the phenomenon that has stirred the most interest is the ability to convert seismic energy into electrical energy. There are many other physical processes that do this, such as piezoelectricity and electrokinetic effects. However, these processes produce an electrical signal (current, charge or voltage) that is linearly related to the stress applied to the material, whereas the phenomenon under study does not. An example of this non-linear response is shown in Figure 1.1. It is immediately obvious that the electrical response in Figure 1.1 differs from the seismic response. Furthermore, the spectrum of the electrical response spans a range of 1 to 2000 kHz, which is three orders of magnitude greater than the seismic disturbance that produced it. The non-linear nature of this conversion is uncommon, but not unknown, as similar conversions are observed in laboratory rock failure experiments (Nitsan, 1977; Yamada, et al, 1989).

The discovery and initial study of high frequency seismoelectric effects from sulphides was by a group of Soviet geophysicists led by G.A. Sobolev (1980). This group consisted of G.A. Sobolev, V.M. Demin, Y. Y. Maybuk, and V. F. Los; most references in this thesis to Sobolev et al. refer to the efforts of this group. Apart from the extensive efforts of the above-mentioned Soviet scientists the only systematic research of which I am aware on this phenomenon has been conducted by the Geophysical Instrumentation Group, headed by R.D. Russell, at the University of British Columbia.

The phenomenon was discovered in the late 1970's by Sobolev and his group whilst investigating the piezoelectric effects of rocks in situ. Piezoelectric signals look similar
Figure 1.1 The response of an electric field antenna (above) and a geophone to a nearby explosion (0.5 kg of pentolite). This record is from an underground experiment in the Lynx Mine, Vancouver Is., BC.
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to geophone signals, that is, they contain approximately the same frequency content (nearly all of the energy is below 2 kHz). Field trials conducted near sulphide minerals revealed another type of signal. The new type of signal was a relatively high amplitude pulse (compared to piezoelectric signals) of short duration (e.g. Figure 1.1). Pulse bandwidths were typically in the MHz range (Sobolev et al., 1980). Sobolev et al. initially labeled this phenomenon PRRER, pulsed radio range electromagnetic radiation (1982), but have subsequently renamed it as RPE, the radio pulsed effect (private communication with M. Maxwell during a visit to Moscow, 1992). In deference to the discoverers, the high frequency seismoelectric phenomenon associated with sulphides will be called RPE throughout this thesis.

An attractive property of RPE is that it appears to be distinctly associated with sulphide minerals. Base metals, such as tin, lead, zinc, nickel and often copper, come from the mining of sulphide orebodies. Also, significant amounts of gold and silver are mined from sulphide ores. Sulphides are metallic salts of sulphur, for example: cinnibar, HgS; pyrite, FeS; sphalerite, Fe/ZnS; calcopyrite, CuS; galena, PbS. Consequently, any phenomenon associated with sulphide minerals is of great interest to the mining industry because of the economic significance of sulphides.

A number of geophysical techniques have been developed to detect sulphide orebodies. Among the most popular are aeromagnetic, induced polarization, and various EM methods. However, many orebodies do not respond well to these established methods, or the response is masked by the physical properties of the host rock. New methods that can reliably detect these orebodies are eagerly sought by the mining industry. Another problem with many established methods is that they cannot be used in operating mines because the mine infrastructure produces too much interference. This restricts the mine operator's ability to reduce costs: core-drilling costs far more than most geophysical surveys.
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The problems with current methods of exploration have provided an impetus to study the possible use of RPE for exploration. Seismoelectric effects have been studied sporadically in the West since 1936 (Thompson, 1936; Martner and Sparks, 1959, Butler et al., 1994), but not used for exploration. In contrast, seismoelectric exploration methods were actively pursued by Soviet scientists from the 1950's (Volarovich et al., 1959, Volarovich and Sobolev, 1969, Neyeshtadt et al., 1972) up until the break-up of the Soviet Union in the 1990's. Most of these efforts were in the area of piezoelectric phenomena. As of 1992 (information from M. Maxwell and R. D. Russell after a visit to Moscow and St. Petersburg), the same scientists were still researching and applying seismoelectric methods. Figure 1.2 illustrates schematically the seismoelectric method.

Figure 1.2  Schematic of an underground seismoelectric method of exploration. An explosion generates a seismic wave which propagates outward. As the seismic wave passes through regions of sulphides EM emissions are detected by an array of antennas.
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of exploration. An explosive charge (or an alternative seismic source) produces a seismic pulse, which propagates outward. Antennas receive electrical disturbances created by the pulse, and transmit the signal to a recording device. The delay the between blast and the reception of an electrical disturbance provides information about the location of the target. Nearly all of this delay is due to the time taken for the seismic energy to reach the region of conversion. Therefore, if the seismic velocity is known then the distance between the seismic source and the target can be calculated.

In 1990, as part of the Geophysical Instrumentation Group at the University of British Columbia, I was given the task of investigating the properties of RPE and its possible use as an exploration tool. The principal hypothesis to be tested in my research is that RPE can be used to explore for massive sulphides. Two major tasks needed to be undertaken: to show that a phenomenon fitting the description of RPE exists, and to devise and evaluate an exploration method based upon RPE. Note that in 1990 there was still doubt about the existence of RPE because of the lack of independent confirmation. The Instrumentation Group at U.B.C. decided that field tests near massive sulphides would give the most decisive results. These field trials are the cornerstone of this thesis.

Is the hypothesis true or false? It is true. Positive results from field tests in four underground locations show that RPE does exist, and that it can be used as an exploration tool. Details of the field trials and examination of the results are found in Chapter 4. Each field trial aimed to show that pulsed EM signals were stimulated by the passage of seismic waves through the orebody, and characterize these signals. In addition, various aspects of the exploration method, such as range, new instruments, charge size and charge placement, were tested in each trial.

Arrivals from the records obtained from these field trials needed to be picked (event selection) and interpreted. Despite many similarities, adaptation of simple seismic techniques proved to be inadequate because of fundamental differences between RPE
and seismic methods. In addition, the complex structure of most mines presents difficulties in displaying the data. Methods and rules for picking, displaying and interpreting were devised and are detailed in Chapter 3. In that chapter I describe how to recognize an RPE signal, and describe methods used to analyze the dataset or invert for structural information. These techniques form a basic tool-set for the processing aspects of an exploration method.

The practical side of an exploration method requires the collection of data. One aim of the U.B.C. research program was to provide a complete description of the equipment and specifications necessary for success in seismoelectric work. Special care is needed because the electromagnetic signals are small, millivolts and nanotesla, and are often measured in the presence of strong electrical interference. Furthermore, the wide bandwidth of RPE signals presents challenges in the design of low noise sensors to acquire these signals. Electric and magnetic field data were recorded in our field trials, as well as some geophone data. All of this data was recorded in digital format, a first for both U.B.C. and our Russian counterparts. Chapter 2 details the instrumentation and methods used for data acquisition. In addition, improvements in various aspects of the instruments and method are suggested.

A realistic model of the processes that produce RPE is needed to fully exploit and understand RPE. Sobolev et al. (1982a) have provided an explanation, but their model cannot explain some aspects of the phenomenon. RPE is difficult to explain because the EM fields are large, and the pulses are very short in duration. I provide an alternative physical model of RPE in Chapter 5 that is based upon triboelectric effects. This model of RPE explains many effects seen in the field and laboratory, and explains why sulphides in particular should exhibit this phenomenon.

1.2 Background

RPE differs from other seismoelectric phenomena in two key areas: it is not repeatable,
and a threshold of seismic power is necessary to initiate the process (Sobolev et al., 1983). The first claim needs some qualification; the signals are not exactly repeatable in arrival time or amplitude, but on repeat experiments RPE signals will generally appear at approximately the same time with similar amplitudes. The implication of the second claim is that nothing short of an explosive source will excite the mechanism. RPE is clearly non-linear in its response to the seismic input.

Sobolev et al. (1982) reported that other physical effects are associated with RPE. Ultrasonic transducers placed in the sulphide orebody measured vibrations coincident with RPE emissions. Also, both light and X-rays were registered on photographic film inserted into the orebody, but none from film put in the host rock. A container made of wood prevented light from reaching the film in the tests for X-rays. In another paper, Sobolev et al. (1984b) describe an experiment where a gamma ray spectrometer detected gamma rays, with energies in the 100 to 700 keV range, that were coincident with the seismic pulse.

U.S. and world-wide patents on an exploration method based upon RPE have been granted to Sobolev et al. (1986). Two interesting details come from these patents: the power spectrum of the signal depends upon the minerals comprising the orebody, and better signals are obtained if two successive shots are used instead of one shot. In the patent it is mentioned that the first shot "charges up" the orebody and second shot, smaller than the first, is used to extract the signals. The charge is monitored by an electrometer, which is presumably connected to the orebody under study, and the second blast is initiated when the electrometer has reached a peak value. The phenomenon of seismically induced static charge is described in a short paper by Sobolev et al. (1982b), in which a build-up of charge appears on the surface of the orebody, with residual decay times of seconds. This effect appears to be quite important in their implementation of a seismoelectric exploration method utilizing RPE.
These developments came to the attention of R.D. Russell when he met Sobolev at an IASPEI meeting in London, Ontario, during 1981. Sobolev offered to demonstrate their equipment and techniques, and in 1983 both Sobolev and Demin arrived in Vancouver to do two underground tests (Sobolev et al., 1984a). The two sites were the Giant-Yellowknife Mine in the North-West Territories, and the Sullivan Mine at Kimberly, British Columbia. At Sullivan the tests were for RPE only, and at the Giant-Yellowknife both piezoelectric and RPE type responses were sought. Several geophysicists from U.B.C., M. Maxwell, B. Narod, and P. Whaite, also participated in these experiments.

To enhance portability, all of the Soviet equipment was battery powered. Principal instruments were a Hitachi reel-to-reel VCR for recording the data, and induction coil magnetic antennas. A current-carrying wire wrapped around the explosive provided a time break for the data. Data from the VCR was divided into high frequency data on the video channel, and low frequency plus voice data on the audio channels. The tapes were analyzed by replaying the signals to a triggered oscilloscope, and photographs of the oscilloscope screen were used to gather timing information.

These collaborative tests showed encouraging results, but were not completely convincing. At this stage, only a few years after discovery, the Soviets were probably still perfecting the method. Also, the methods of interpretation and presentation were very different from those routinely used by Western nations. In hindsight, the style of presentation was appropriate given that the survey lines ran parallel to the ore-zones and the small size of the data-set (see Chapters 3 and 4). The Soviets were already convinced that the blips on the oscilloscope traces were not artifacts, as they were familiar with this type of signal and immediately recognized RPE, but many Western observers were not prepared to share their confidence until presented with more convincing data.

In 1986, B. Narod and M. Maxwell duplicated the tests at the Sullivan Mine using locally
available equipment (see Chapter 4 for details of the Sullivan site). A Racal tape recorder allowed the recording of four channels at a bandwidth of 300 kHz. Magnetic field sensors consisted of a UTEM coil borrowed from Cominco, and a custom designed coil made by B. Narod, called UBC I. Blast time-breaks were produced by the same method used by the Soviets, a current carrying wire wrapped around the explosive.

Bursts of short duration EM pulses, similar to those obtained by the Soviets, were produced during these tests. The signal to noise ratio was low, around 2 to 4. Furthermore, the best receiver turned out to be the geophone: the UTEM coil behaved erratically, and UBC sensor had a higher noise level. Despite these instrument faults and quirks, the records showed bursts of signals at approximately the same times found by Sobolev and Demin. The experiment was deemed to be successful in duplicating the work of the Soviet scientists. However, a profusion of signals immediately after the blast did cast some doubt upon the source of the signals. EM pulses were expected to start 5 to 6 milliseconds after the blast because the main orebody was about 30 meters from the shots, but the signals often started earlier. The cause was narrowed to two possibilities: small amounts of sulphides near the shotpoint, or interference from the wire-break circuit. Sulphides near the shotpoint were considered to be the dominant mechanism because of the common times of strong signals seen by both field trials, an unlikely event for wire-break interference.

Laboratory tests on sulphide samples by Sobolev et al. (1982) reveal that RPE is not easy to produce on a small scale. The simplest test is to uniaxially compress a sample and examine the response. This type of test did not show the sulphide samples to be particularly active. In fact, samples containing quartz gave the greatest signals. Sobolev et al. found that if a low frequency AC voltage was applied to the sample then the sulphides produced much stronger signals; the responses from other types of rock were unchanged. It would appear that the applied voltage serves the same role as the reported charge build-up on the orebody (Sobolev et al., 1982b). However, a later sample
testing method (1992, private communication between M. Maxwell and R. D. Russell with G. A. Sobolev) involved pressing the sample onto a knife edge to fracture the sample while a current is passed through it. The laboratory methods of Sobolev et al. indicates that they consider both fracture and low frequency electric fields vital for the production of RPE.

The importance of fracture and low frequency electric fields is reflected in their explanation of RPE. In 1982 (Sobolev et al., 1982a) they proposed that cracks generate the RPE pulse, and the pulses are amplified and shaped further by natural semiconductor circuits within the orebody, which are powered by nearby piezoelectric minerals such as quartz or sphalerite. I find this explanation inadequate as it requires piezoelectric materials, and seems too complex for a natural phenomenon. Their explanation is not mentioned in later publications, and I suspect it has fallen out of favor.

In 1992 R.D. Russell and M. Maxwell visited Moscow and Leningrad (now St. Petersburg) to see what avenues the ex-Soviet scientists were pursuing in the area of seismoelectricity. Other than a short paper by Demin and Sobolev in 1988, there had been no recent reports. M. Maxwell and R. D. Russell discovered that the field of piezoelectricity had moved from research and development to application and field surveys (under the auspices of a company called Rugidfizika). Demin and Maybuk were continuing the study of RPE (in Moscow), but were at a less advanced stage than Rugidfizika in application. However, they were also busy producing case histories. In fact, the ex-Soviets had quietly collected an impressive set of case histories, none of which have appeared in Western literature (or probably in Soviet literature). With the break-up of the Soviet Union into various countries and the resulting economic chaos there is some uncertainty as to continuation of these programs.

By 1990 there was a strong desire for independent confirmation of the validity of Sobolev's work. A number of mining companies had heard or read of the Soviet work
and wanted to know if such a phenomenon was real. If so, would it be feasible to use it for mineral exploration? As part of a broad program to discover and develop seismoelectric phenomena for geophysical exploration, the Instrumentation Group of U.B.C. was given the resources and funds to provide this information. My thesis is the embodiment of the research effort in RPE at U.B.C during 1990 to 1995.
CHAPTER 2

FIELD INSTRUMENTATION AND METHODS

2.1 Overview of Equipment and Procedure

Most of my field studies were carried out in an underground environment. An advantage to underground work is that the EM noise background can be very low, due to the shielding of the overlying rock. A great disadvantage is the limited choice in shot and instrument placement. Furthermore, the underground environment is harsh on equipment because of the moisture, mud, dust, and, in many sulphide mines, the acidity. Unless otherwise stated, this section of the thesis implicitly assumes that the described equipment and methods are for underground work, as this aspect is most relevant in understanding the case histories in Chapter 4.

The basic aim of our field trials was to measure the electromagnetic response of the earth to a seismic stimulus. Collecting this data involves setting off an explosion to produce the seismic disturbance, whilst recording the response from various electric and magnetic field antennas placed near the area of interest. This procedure is broadly similar to seismic methods. However, the placement of sensors is not as crucial because electromagnetic disturbances travel at least four orders of magnitude faster than seismic waves, and all of the EM sensors receive the signal within microseconds of each other. Hence, the time lag seen on the records is principally due the propagation time for the seismic wave to reach the target. Unlike seismic methods, multichannel recording does not give further information about target location, but offers redundancy and better discrimination.

Stimulating RPE requires a fairly powerful source, which is provided by an explosive. To obtain good signal reception we want to be near the shotpoint, but not so close as to endanger the crew or equipment. We have generally kept our sensors and ourselves
between 50 and 200 meters from the blast. This distance is sufficiently close to receive signals reliably. In an underground mine there is little choice in sites to place shots, equipment, and people, so shot placement is done with a lot of thought. Areas where there is clearly unstable rock formations, or where damage to mine infrastructure might occur are avoided. Ventilation is an important consideration because the explosion produces dust and fumes, which may contain harmful gases such as carbon-monoxide. Equipment and crew are best deployed upwind of the blast, preferably in an alcove or short drift/crosscut, and out of the direct path of the blast airwave and other mine traffic.

Each explosive charge is placed in a short (1 to 3m) drill hole. Sobolev and Demin (private communication with R.D. Russell) often place their charges on the tunnel surface because it is more convenient than drilling numerous small holes. Surface charges do not efficiently produce seismic energy, therefore, a larger charge is necessary. A typical charge size for our work ranges from 0.2 to 0.5 kg, for the Russians 2 to 5 kg is more typical. I feel that the extra damage done to mine infrastructure by this technique more than offsets the time saved by not drilling shot holes. One effective way to increase productivity is to fire the shots in salvoes. Because it is not safe to approach the area of an explosion for a period of ten to fifteen minutes after the blast (gases and spalling/falling rock) firing a sequence of four shots at two minute intervals saves approximately 30 to 40 minutes that would otherwise be spent waiting for the dust to settle.

The requirement for broad band, low-noise EM measurements necessitated custom-built sensors and pre-amplifiers. Sensor bandwidths were typically 1 kHz to 100 kHz, with some allowing measurement up to 5 MHz. The upper frequency bands are not typically used in geophysics, therefore, most commercially available sensors lack the ability to make broad band measurements in this range. Another factor is that most geophysical work is done on the surface and EM measurement is limited by ambient noise. The underground environment is generally quiet at these frequencies, with most transients
occurring in sync with the mine grid (e.g., a transient every half cycle, or 8.33 ms). Thus, the ability to detect signals is often limited by the intrinsic noise properties of the sensor, which can be optimized by design and component selection.

Figure 2.1 Schematic of the instrumentation used in seismoelectric research at UBC.

The EM sensors and geophones were remotely set, and linked to a central instrumentation site via shielded cable; an arrangement similar to many seismic surveys. A flow diagram of the signal path is shown in Figure 2.1. This arrangement assures flexibility in sensor placement, as each has a separate cable, and reduces the possibility of EM interference from the computer/digitizer system. All of the EM sensors have integral pre-amplifiers so that the cables do not alter the signals significantly.

Signal conditioning electronics, the computer based digitizer, test equipment and batteries are located in one area, the instrumentation site. This centralized arrangement
allows one person to monitor and adjust the incoming signals, check sensors and batteries for proper operation, and to save the digitized data to a portable computer. Batteries power all of the equipment in order to enhance portability without compromising the EM environment. DC power avoids the problems of switching transients often found in AC equipment and in mains inverters. The batteries also supply power to the EM sensors through extra wires in the cables.

2.2 The Seismic Source

Various types of seismic sources have been tested and only one stands out: pentolite (PETN/TNT mixture) boosters. This type of explosive in sizes 0.18 to 0.5 kg has worked very well in the field trials described in Chapter 4. During my first field trial (Sullivan Mine) I tested a number of small sources, including a sledge hammer, shotgun blanks, and detonators, and observed no electromagnetic response. Blasting agents or emulsion-type explosives (ammonium-nitrate/fuel oil mixtures) of 0.1 to 0.3 kg sizes were also tried in this test (and in sizes up to 10 kg in later surface trials in Australia) with little success. This was probably due to poor coupling to the host rock rather than a lack of energy. The pentolite boosters were very strongly recommended by Grant Scott, a miner/powderman on loan to us from the Sullivan Mine workforce. After hearing our needs Scott suggested that the boosters were the correct explosive for the task, and he insisted on gathering some for our experiment. Scott was vindicated as the pentolite explosives succeeded where the other sources failed.

Pentolite differs from most commonly available explosives in that it has a very high velocity of detonation (7500 m/s) and it is relatively dense (1600 kg/m$^3$). Normally, pentolite is used to boost the explosive process in large charges made of ammonium-nitrate/fuel oil mixtures (Tour, 1992). High velocity explosives, such as pentolite, are efficient in producing seismic pulses in underground rock because the characteristic impedance of these explosives and the impedance of underground rock are well matched (Nicholls, 1962; see Appendix A.1).
Figure 2.2 Example of the seismic pulse (above) produced by a small explosive in underground rock, and the amplitude spectrum of the seismic disturbance (below).
Chapter 2: Field Instrumentation and Methods

The results from my underground field-work indicate that approximately 0.5 kg of pentolite explosive is needed to reliably induce RPE in ore-zones 75 m from the shotpoint. This combination of charge size and range corresponds to a peak pressure of about 150 kPa at the orebody (see Appendix A.2 and A.3). Therefore, the size of explosive charge should be adjusted so that a seismic stress of approximately 100 to 200 kPa is delivered to the target (Appendix A.2 gives the scaling laws).

An example of the seismic wavelet and spectrum obtained by my methodology is shown in Figure 2.2. The pulse is very broad-band for a high energy seismic pulse, and contains substantial energy from 200 to 1000 Hz (a 1000 Hz low-pass filter was used in acquiring this data).

2.3 Obtaining a Shot Moment

The shot moment or time-break is an electrical signal indicating that the explosion has begun. This is a very important signal as it is the reference for comparing events from other shots. In this case, a simple concept does not translate into an easily realizable one. It is relatively easy to devise a method to obtain a shot moment signal; the problem is in devising one that does not pollute the EM environment during measurement.

Initially, two solutions were pursued to remedy the problem of electrical noise. One method was based upon a modified blasting box that quickly disconnects the current source (in less than 0.01 ms) once the detonation has been initiated. This eliminated the possibility of further wire contacts inducing false signals, however, the current pulse into the detonator produces a large EM transient. Another method is to wrap a loop of wire around the explosive, put a small current (0.2 mA) through the loop, and monitor the resistance change by measuring the voltage across the loop as it is broken; we call this method a wire-break. Because the wire-break is independent of the type of detonator used a seismic detonator is not necessary. The ability to use ordinary electrical detonators and fused caps (detonators initiated by a flame at the end of a slow-burning
fuse) is a desirable feature as it allows the use of detonators from the mine magazine. Unfortunately, neither method met our needs as both created too much electrical noise. A further disadvantage for the wire-break is that the additional conductor attached to the explosive package may compromise safety.

My solution to the problem of electrically isolating the blast from the antennas and other electrical equipment was to use a fiber optic cable and fuse blasting caps. The idea behind the fiber optic time-break is that the intense light emitted by high explosives can be used to obtain an accurate time break. In practice, a small length of unsheathed optical fiber is attached to the detonator or explosive. Some of the light from the explosion is collected by the fiber optic strand, and is transmitted through the length of the fiber optic cable to an optical receiver, which translates the light into an electrical signal (Kepic et al., in press; details are in Appendix B.1). After the blast, the burnt end of the fiber optic cable is trimmed away and the cable is reused.

The fiber optic time break has eliminated the strong EM activity, 0 to 3 ms after the blast, that frequently occurred in previous experiments (see Appendix B.2, and the Sullivan Mine and Mobrun Mine field trials). This blast-related activity masked early signals, and placed some doubt upon the source of later arrivals. The fiber optic trigger is also very accurate. A side-by-side comparison of the fiber optic and wire-break methods on a detonator showed that the fiber optic trigger occurred first, and the wire-break signal followed 10-20 μs later. This demonstrates the great accuracy of the fiber optic time-break because the wire-break is known to be a very accurate method of determining time of detonation (Burrows, 1936).

2.4 Magnetic Field Sensors

A magnetic field transducer (or magnetometer) produces an electrical signal that is simply related to the magnetic field. There are many ways to achieve this result, but the simplest and most effective method for the frequency range of 1 kHz to 5 MHz is to use
wire coiled onto a ferrite rod. Faraday's law of induction describes the coil's behavior in a magnetic field

\[ V = -NA \frac{dB}{dt} \]  

where \( V \) is the voltage across the coil. The voltage is proportional to the rate of change in the magnetic flux enclosed by the coil, which is equal to number of turns \( (N) \times \) cross-sectional area \( (A) \times \) magnetic flux density parallel to the coil axis \( (B) \).

A simple electrical circuit representation of the coil is an inductor in series with a resistor. The inductor represents the self-inductance \( (L) \) of the coil and the resistor represents the resistance of the wire comprising the coil. The effects of coil resistance can be ignored for most purposes because it does not play a significant role until very low frequencies (sub 10 Hz), where the impedance of the inductor is less than that of the resistor. A shunt capacitor needs to be added to complete the electrical representation of the coil. This capacitor approximates the behavior of the distributed capacitance between coil windings. The effect of the magnetic field may be represented as either a voltage source in series with the inductor using equation 2.1, or as a current source in parallel with the inductor. The current source has a current \( (i) \) of

\[ i = \frac{NAB}{L} \]  

Note that the current is proportional to the magnitude of the magnetic field \( (H) \).

Shunting the coil with a resistor and measuring the voltage across the resistor provides a passive means to measure the current produced by the coil. This type of magnetic sensor is electrically equivalent to a damped parallel-resonant circuit, and I have labeled this arrangement as the "damped resonator" arrangement to distinguish it from the more common inductive pick-up, which is an undamped coil. The transfer function of the damped resonator design is obtained by considering the voltage \( (V) \) across the parallel combination of current source \( (i) \), inductance \( (L) \), resistance \( (R) \), and capacitance \( (C) \). Conservation of current gives:
\[
\frac{NAB}{L} = V \left( \frac{1}{Z_L} + \frac{1}{Z_r} + \frac{1}{Z_c} \right) \tag{2.3}
\]

In the frequency domain the transfer function is
\[
\frac{V(\omega)}{B(\omega)} = NA \frac{j\omega}{\left(1 + j\omega \frac{L}{R} - \omega^2 LC\right)} \tag{2.4}
\]

If the coil is more than critically damped,
\[
R \leq R_c \quad \text{where} \quad R_c = \frac{1}{2} \sqrt{\frac{L}{C}} \tag{2.5}
\]

then the transfer function has low and high-pass corner frequencies \((f_L, f_H)\) respectively. When the coil is under-damped, \(R > R_c\), the response is peaked at the resonance frequency \((f_o)\).

\[
f_H = \frac{1}{2\pi RC} \quad f_L = \frac{R}{2\pi L} \quad \text{and} \quad f_o = \frac{1}{2\pi \sqrt{LC}} \tag{2.6}
\]

For a well damped coil, \(R \ll R_c\), the damped resonator produces a voltage proportional to the magnetic field between frequencies \(f_L\) and \(f_H\). This is only strictly true near \(f_o\), but if the resonance is well over-damped then we can neglect the roll-off near \(f_L\) and \(f_H\) for most pass-band frequencies.

The principal benefits of the damped resonator design are that the frequency response is flat over a range of useful frequencies, and this response extends beyond the coil's resonance. Induction coil sensors generally measure frequencies below the coil's self-resonance, or at resonance, which limits their use as broadband sensors. The pass-band responses of the magnetic sensors used in my research were chosen to be between 1 kHz and 300-3000 kHz, as this region encompasses most of the observable RPE spectrum, and it avoids the large amounts of power-line noise in the 10 to 1000 Hz band. In addition, these sensors were required to detect very small signals in a low noise environment, therefore, considerable effort was made in optimizing both the sensor noise level and the frequency response (Appendix B.3).
The measured properties of the UBC I, UBC IV and UBC V sensors, which were used in my field trials, are given in table 2.1 (see Appendix B.4 for details). Overall sensitivities were adjusted in the UBC IV and V designs based upon the experience gained from two previous field trials with UBC I, which was found to be set too low. A sensitivity of about 0.1 V/nT was found to be suitable for the measurement of RPE signals 50 to 200 meters from the source.

2.5 Electric Field Sensors

Unlike the magnetic sensors, there were several techniques used to measure the electric field. The three main techniques were: the dipole, measurement of the potential between a pair of stainless steel stakes driven into the earth some distance apart; the long wire antenna, measurement of the potential difference between an insulated wire overlying the earth and a ground stake; and the parallel plate dipole, the measurement of the potential between two freely suspended, parallel, electrode plates. Each has its particular advantages and characteristics, which will be discussed in the following sections.

Grounded Dipole

Grounded dipole refers to the measurement of the potential between two points in the earth. This is achieved with a differential pre-amplifier connected between two stakes set into the earth. The stakes are used to ensure that a low resistance electrical contact is made with the earth; the resistance between stakes (2-10 meters apart) is usually of the order of 10 kΩ. In many geophysical applications porous pot electrodes are used, but these are primarily used to remove low frequency effects (Telford et al., 1986) that are not important in my studies. Each pre-amplifier is housed in a metal die-cast box to protect the electronics from the elements and has banana jacks to connect wires to the electrode stakes.

The pre-amplifiers used in my work are based upon the three op-amp instrumentation
amplifier design (see appendix C for schematic), which provides a high input impedance and good common-mode signal rejection (Horowitz and Hill, 1989). In addition, the use of a differential amplifier prevents ground loops. I constructed two models of amplifier, the T-box (named after the pattern of electrical tape affixed on the housing) and high bandwidth pre-amplifier (HBW), for both my use and for general use for the research group. Specifications for these amplifiers are tabulated in table 2.2. The T-box type of pre-amplifier is principally used by the U.B.C. research group in tests for low frequency seismoelectric phenomena, and the frequency response, 0.1 Hz to 30 kHz, extends lower than is necessary for RPE work. High bandwidth pre-amplifiers have either a 1 kHz or 0.1 Hz low-cut.

![Equivalent circuit of a small dipole antenna.](image)

Principal weaknesses of the grounded dipole are a susceptibility to magnetic fields because of the loop formed by the wires to the stakes (Wu and Thiel, 1989), and an uncertain frequency response beyond 100 kHz. Keeping the leads to the stakes short and straight reduces the former problem. The latter problem is due to stray input capacitance to the pre-amplifier, and the input capacitance of the amplifier. The equivalent circuit of a short dipole antenna (small compared to EM wavelengths) is shown in Figure 2.3 (Casey and Bansal, 1991). The input resistance and capacitance of the pre-amplifier, $R_i$, and $C_i$, respectively, are critical in determining the frequency response and noise characteristics of the system.
and $C_i$ are included in the equivalent circuit (Figure 2.3) because of their influence on the sensor response. A low-pass filter is formed by the source resistance $R_s$, and input capacitance $C_i$ with a corner frequency of

$$f_L = \frac{1}{2\pi R_s C_i}$$

(2.7)

The source resistance, $R_s$, is the resistance between the two ground stakes. Typical values of $R_s$ and $C_i$ are 10 kΩ and 50 pF respectively, which results in a 300 kHz low-pass response. This effect precludes the use of grounded dipoles for wide-band measurements. Note that at higher frequencies the dielectric currents of the earth predominate, and provide the voltage across the wires. At very high frequencies the frequency response is flat (if limitations on amplifier bandwidth are neglected) because the capacitors (Figure 2.12) are the only significant circuit elements at high frequencies. However, the input voltage is lowered by the voltage dividing action of the stray capacitance verses the source capacitance $C_a$ at high frequencies.

**Long Wire Antenna**

The long wire antenna (LWA) proved to be very useful in my early experiments because of its large effective height (effective height equals the open circuit voltage produced by an antenna divided by the electric field parallel to it) and ease of implementation. The long wire antenna is a long straight insulated wire placed near the ground with one end connected to an differential amplifier. The other input to the amplifier is connected to a ground stake at one end of the wire.

The long wire antenna arrangement was introduced to the U.B.C. instrumentation group by A. Boyle in 1990, then at C.R.A. Group Special Equipment, Melbourne, Australia. A debate about what the long-wire was actually measuring arose between us. Further controversy was prompted by a series of letters between Wait (1989) and Wu and Thiel (1989b) over a paper by the latter (1989a), which was about the improved magnetic field immunity of the long wire antenna over grounded dipoles. R.D. Russell and I
believe we have settled the question of what is measured (1991 IEEE/URSI meeting in London, Ontario; see Appendix B.5). The long wire antenna can be modeled as a summing amplifier with each segment of wire acting as a local capacitive pick-up. This action of distributed measurement contrasts greatly with the measurement made with a grounded dipole, which infers electric field behavior from the measurement of the electric potential between two points.

The long wire antenna was originally used as a long straight insulated wire on the ground with one end connected to a charge amplifier. The "ground reference" of the charge amplifier circuit was connected to a stake at one end of the wire. A problem with using the long wire antenna with a charge amplifier is that the capacitance of the wire must be known, or measured, to accurately determine the average field strength. However, a much greater problem is that if more than one antenna is set out then multiple ground points due the grounded stake of each antenna alters the electric field. Both problems can be circumvented by the use of a high input impedance differential amplifier. The transfer function of this arrangement (in the frequency domain) is

\[
V = \frac{j\omega RC_w}{1 + j\omega RC_w} \bar{V}
\]

(2.8)

where \( C_w \) is the capacitance of the long wire, and \( R \) is the input impedance of the amplifier (treatment of the problem is similar to the charge amplifier arrangement except that the current collected by the long wire flows through the input impedance of the amplifier, \( R \), to the ground stake; see Appendix B.5 for details). At frequencies above the RC pole the response is equal to the mean potential difference under the wire, \( \bar{V} \), referenced to the ground stake. Equation 2.8 implicitly assumes that the capacitance per unit length of the wire is constant, otherwise the potentials are weighted by the capacity per unit length (Appendix B.5). The pole is typically below 100 Hz (\( R > 5 \, \text{M}\Omega \) and \( C > 1 \, \text{nF} \) for 50 metres of wire) and is not of consequence in my measurements. I do not use the long wire antenna for high bandwidth measurements because it suffers the same uncertainty in high-end frequency response as its grounded counterpart.
Parallel Plate Dipole

The parallel plate dipole is indispensable for measuring vertical fields and high-bandwidth signals. It has a flat response from 1 kHz to 4.5 MHz (upper end limited by the pre-amplifier; Figure 2.4) and it is very portable (Baum, 1980). This is a non-contact antenna consisting of a very high input impedance amplifier connected to a pair of freely suspended conductors. The electrical equivalent circuit is the same as the grounded dipole (Figure 2.3), but the source resistance term $R_s$ can be neglected as it is near-infinite.

Figure 2.4 Test of the impulse response of the UBC IV magnetic antenna and parallel plate dipole antenna. The plot is an FFT of a 20 Msample/s time-domain record.
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The transfer function of the parallel plate dipole system in the frequency domain (Matsui, 1991) is

\[ V = \frac{j\omega RC_s}{1 + j\omega R(C_s + C_i)} E l \]  

(2.9)

where \( l \) is the distance between plates, \( E \) the electric field strength, \( R \) the input impedance of the pre-amplifier, \( C_i \) the input capacitance of the preamplifier, and \( C_s \) the capacitance of the antenna (i.e. the amount of dielectric coupling).

The input resistance and the total capacitance of the system forms a high-pass filter. Beyond this RC pole the output of the parallel plate dipole is flat until the pre-amplifier starts to roll-off (at about 5 MHz for the high bandwidth pre-amps). The corner frequency of the high-pass filter is about 100 to 200 Hz in my implementations of the parallel plate dipole (\( R=100 \text{ M}\Omega \) and \( C=20 \text{ pF} \)). In addition, it can be seen from 2.9 that the effective height of the antenna is strongly influenced by the input capacitance \( C_i \).

The high bandwidth series of pre-amplifiers (specifications are in the dipole section) were designed with this application in mind, and have only 8 pF input capacitance. This is necessary as most practical dipoles for in-mine use have only 5 to 15 pF of capacitance. Hence, the effective height of the antenna is usually only half of it's physical size.

To obtain a good signal-to-noise ratio the plates should be placed far apart and made as large as possible (Appendix B.6). For reasons of portability my implementation of the antenna is limited to a maximum dimension of about 2 meters. Very large plates are clumsy so I settled on plates sized 0.4 m x 0.4 m. This design resulted in an effective antenna height of about 1.2 m and an antenna capacitance of 12 pF.

An unexpected advantage of the parallel plate dipole in underground use is the enhancement of the effective height of the antenna by the dielectric properties of the earth. I initially found that the electric field strengths measured by the parallel plate dipole were consistently higher than the grounded dipole measurements, by a factor of 10 over the 1 to 30 kHz band. An explanation for this discrepancy is that in
underground use the antenna is in a air-space formed by the tunnel, and that the continuity of \( D = \varepsilon E \), combined with the high dielectric constant of the surrounding rock, increases the electric field in the tunnel. The tunnel geometry lowers this enhancement by a factor of two (for a cylindrical void, Kittel, 1986) when measuring across the tunnel. Note that there is no increase of the electric field along the tunnel direction.

2.6 Signal Transmission and Cables

There are two principal concerns about cabling: EM interference, and capacitive loading of the pre-amplifier. Transmitting broad-band signals is difficult because most cables perform differently for audio verses radio frequencies (Ott, 1988). For low frequency measurements shielded twisted pair cables with differential transmission of the signal has worked well. The shield is connected to ground at only one end to prevent ground loops. Differential transmission and reception removes common-mode noise such as inductive pick-up. Ott (1988) recommends this type of system over coaxial cable for audio frequencies. Unfortunately, the twisted pair cable used in my field trials does not perform well beyond 50 kHz. This may be due to imperfections in the shield of the cable (i.e. breaks) and asymmetry in the twisting of the wire pairs. The shield is made of aluminum foil and has 100% coverage, but after considerable field use the shield is broken and segmented because of cable bending and unbending during use. An uninsulated drain wire prevents the shield from becoming disconnected, but cannot prevent uneven coverage. In hindsight, a combination of braided and foil shield would be better for field use.

At high frequencies coaxial cable works very well. Below 10 kHz, coaxial cable works poorly because the resistance in the shield inhibits the mutual inductance between centre conductor and shield that normally cancels inductively coupled signals (Ott, 1988). In addition most coaxial cables are only shielded by braided wire. This type of covering results in a 80 to 90% physical coverage, which reduces shield effectiveness. As
with the twisted pair, a better solution is the use of braid and foil shield, which I have used with the parallel plate dipole antenna.

When the wavelength of the signal becomes comparable to the length of the cable then the cable behaves like a transmission line (Ott, 1988). To prevent reflections the cable should be terminated with a resistance equal to the characteristic impedance. For the twisted shielded pair it is about 100 Ω, for standard instrumentation coaxial cable it is 50 Ω. Terminating the cable places an extra load on the preamplifier, which must deliver sufficient current to produce the required voltage. For a 5 volt signal this can translate into 100 mA of current, which is beyond the capabilities of most pre-amplifiers. In addition, at low frequencies the cable acts as a large capacitor (10 nF for 100 m of shielded twisted pair). Charging and discharging this capacitance can require up to 100 mA of current for fast slewing signals. These capacitive loads can alter the frequency response of the pre-amplifiers. To overcome the loading problems of cables a buffer should be placed after the pre-amplifier.

2.7 Amplification and Filtering

When a signal is digitized it is very important to maximize the dynamic range; otherwise, the smaller signals will "staircase" because of the limited representation of the waveform. The role of an amplifier is to adjust the signal amplitude so that good use of the available dynamic range is made, and to avoid corruption from external noise sources. In the U.B.C. system (see fig 2.1), fixed gain pre-amplifiers are used to boost the signals to levels where the cables will not unduly affect the signal-to-noise ratio, then the signal is further amplified and filtered before digitization by a bank of Tektronix AM502 amplifiers. Because the EM noise varies so greatly from site-to-site the last factor of 10-100 of gain is achieved with post-amplifiers, which are adjustable in a 1-2-5-10 type gain sequence. The AM502 is a plug-in module, four of which share a special housing (model TM515 that supplies the power and provides a mounting structure for portable use. An extra custom built module was used to supply DC power directly from batteries;
To prevent aliased records, the signals are filtered below the limit set by the Nyquist frequency of the digitizer. A multi-pole anti-alias filter was avoided because these filters tend to distort the pulse shape, and simple one or two pole filters were used. The pre-amplifiers are of fixed band-width for simplicity and ruggedness, and the post-amplifier, a Tektronix AM502, shapes the final pass-band characteristics. The AM502 has both low-pass filters (6 dB/octave with 1 Mhz, 300 kHz, 100 kHz, 30 kHz, 10 kHz, 3 kHz, 1 kHz, 300 Hz settings) and high-pass filters (12 dB/Octave with 0.1 Hz, 1Hz, 10 Hz, 100 Hz, 1 kHz, 10 kHz settings). I generally set the pass-band to 1 kHz to 30 kHz when using the RCE digitizer at 125 ksamples/s. For broadband measurements the AM502 units were bypassed because of the 1 MHz maximum bandwidth of the AM502.

The system bandwidth should be maximized because the electromagnetic signals from RPE consist of very brief (<5 μs) pulses, and low-pass filtering of the signal below it's natural bandwidth will lower the signal-to-noise of the signal. The amplitude of a pulse is proportional to the bandwidth, but the RMS amplitude of Guassian or white noise (the principal type of intrinsic noise in my system) is proportional to the square root of the bandwidth. Hence, the signal-to-noise ratio of a band-width limited RPE pulse falls with the square root of bandwidth, and it is optimal to preserve the natural bandwidth of the pulse (see Appendix B.7 for details).

An insidious problem with many low noise pre-amplifiers is the demodulation of AM radio signals (Horowitz and Hill, 1989). This problem can prevent successful surface work unless remedied. Demodulated AM radio signals are mostly music and voice, therefore, most of the interference is filtered by the 1 kHz high-pass on the post-amplifier. However, a sufficient amount of energy remains to give less than optimal signal-to-noise. Most low-noise amplifiers designed for audio frequencies suffer from this behavior to some degree. Unfortunately, many popular low-noise op-amps (such as the OP-27) suffer more than we can typically tolerate (Appendix B.8). To avoid AM
demodulation an amplifier should be chosen for high slew-rate performance, or a full power bandwidth of approximately 1 MHz or better (the two are essentially the same specification).

2.8 Analog to Digital Conversion

RCE Digitizer

Most of my data has come from electrical signals that are converted from their analog form to a digital representation by a computer based digitizer made by RCE Electronics (Santa Barbara, California). The actual digitizer is a printed circuit board that fits inside IBM compatible personal computers. In this case, the computer is a portable made by Epson. The advantage of this system is that the portable computer (and digitizer) can be powered by a 12 V battery, removing the dependence on mains electricity. This digitizer system was acquired by R.D. Russell in the late 1980's to observe relatively low frequency seismoelectric phenomena, such as piezoelectricity, and some aspects of the RPE phenomenon. The RCE-based system is not ideal for studies of RPE because of it's limited speed and buffer depth, but it was available in early 1991 when affordable high speed digitizers were still on the drawing board and the research group had many cables, amplifiers and filters for the unit. Therefore, it was a logical choice at the time. In addition, the unit's portability is a great asset in the underground environment.

The basic operation of the digitizer is that the analog signal is fed to an analog-to-digital converter and the resulting digital representation is stored in a memory buffer on the board. Analog signals are wired to the board via an interface box and ribbon cable to a special receptacle on the board. We replaced the standard cable with a twisted pair type to minimize cross-talk on the ribbon cable. Software for the host computer is provided so that the board may be operated in manner similar to a digital oscilloscope.

Key features of the RCE digitizer are: the ability to digitize 1 to 16 independent electrical inputs (or channels); convert the analog signal to a 12 bit representation (including sign);
a fairly large on-board buffer of 64 ksamples; a very flexible triggering system; and a maximum sampling rate of 1 Msample/s. In addition, the digitizer is relatively easy to configure via custom programs on the host computer. Because the RCE board has only one analog-to-digital converter there is a trade-off in sampling speed and in the number of channels used. The board samples each electrical channel in a sequential order (e.g. 1-2-3-4-1-2-etc. for four channels), so the maximum speed for a given number of channels is equal to 1 Msample/s divided by the number of channels (which has to be a power of two because of the board's architecture).

In general, I operate the board with 8 channels at a rate of 125 ksamples/s. This setting gives a 65 ms time window before the on-board buffer is exhausted, which is appropriate for underground tests as by this time the seismic wave has traveled about 250 meters and is too weak to induce a response. The resulting Nyquist frequency is about 62 kHz and I generally set a 30 kHz high-cut on the (Tektronix) amplifiers that feed the signals into the board. With eight channels I am able to dedicate one to the trigger signal, have a three component array of magnetic and electric sensors, plus a spare channel for a geophone.

**Gage Digitizer**

In late 1992 the geophysical instrumentation group at U.B.C. purchased a high speed digitizer from Gage Electronics (Montreal), a model CS220-1M, for my RPE studies. The Gage digitizer is similar to the RCE board in that it needs an IBM compatible host, and has a large on-board buffer. Principal differences are in speed, 40 Msamples/s vs. 1 Msample/s, an 8 bit A/D, and a maximum of two channels. This digitizer needs a large buffer (1 Msample) because at high sample rates the buffer memory is quickly consumed. I have used the board in the two channel configuration at a sample rate of 10 Msamples/s. With the 1 Msample buffer this gives a 51 ms time window and a 5 MHz Nyquist frequency. Another very useful feature is the ability to amplify or attenuate the signal before digitization, giving signal input (or dynamic) ranges of +/-5 V down to +/-
200 mV. With the extra gain adjustment the use of external post amplifiers, such as the Tektronix AM502, are unnecessary.

The 8 bit representation does not leave much room for error in setting the gain of amplification before digitization; a factor of 10 too little will result in a staircased reconstruction of the signal, too much gain results in clipping and a total loss of signal structure beyond the dynamic range of the system. Unfortunately, there is a certain inherent variability in amplitude of the RPE response. In my only field test of this digitizer I became experienced enough to guess within a factor of two the right gain, and after a few shots I had the system at an optimum setting: just enough gain for the largest signals to push the dynamic range of the system.
### Table 2.1 Characteristics of the magnetic sensors used in the field trials.

<table>
<thead>
<tr>
<th></th>
<th>UBC I</th>
<th>UBC IV</th>
<th>UBC V</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pass-Band (kHz)</td>
<td>0.7 - 300</td>
<td>1.5 - 3500</td>
<td>4 - 950</td>
</tr>
<tr>
<td>Sensitivity (mV/nT)</td>
<td>1.67</td>
<td>100</td>
<td>100</td>
</tr>
<tr>
<td>Noise (fT/Hz(^{1/2}))</td>
<td>360</td>
<td>12</td>
<td>8</td>
</tr>
<tr>
<td>Noise p-p in pT</td>
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<td>225</td>
<td>60</td>
</tr>
<tr>
<td>Dynamic Range nT</td>
<td>6000</td>
<td>100</td>
<td>100</td>
</tr>
<tr>
<td>Sensor Size</td>
<td>40 cm x 5 cm Dia.</td>
<td>80 cm x 5 cm Dia.</td>
<td>80 cm x 5 cm Dia.</td>
</tr>
</tbody>
</table>

### Table 2.2 Pre-amplifier characteristics. These pre-amplifiers were used to amplify and buffer the measurements of electric potential.

<table>
<thead>
<tr>
<th></th>
<th>T-Box</th>
<th>HBW</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pass-Band (kHz)</td>
<td>0.1 - 30</td>
<td>1 - 5000</td>
</tr>
<tr>
<td>Input Impedance (MΩ)</td>
<td>4</td>
<td>100</td>
</tr>
<tr>
<td>Gain</td>
<td>28</td>
<td>25</td>
</tr>
<tr>
<td>Noise (nV/Hz(^{1/2}))</td>
<td>90</td>
<td>8</td>
</tr>
<tr>
<td>Noise p-p in μV</td>
<td>160</td>
<td>160</td>
</tr>
</tbody>
</table>
CHAPTER 3

INTERPRETATION AND PROCESSING METHODS

3.1 Introduction

Prior to this thesis the only information available on the interpretation of RPE data was a brief outline from the patent of Sobolev et al. (1986). Most of their published work provides details about the EM signature and associated phenomena of RPE. Some useful hints about locating targets with seismoelectric methods can be found in the monograph about piezoelectric methods by Sobolev and Demin (1980); it contains the boundary delineation method (described section 3.5). This chapter adds substantially to the current pool of knowledge by giving details about a suite of techniques used to analyze and interpret data from the field trials described in Chapter 4.

The basic steps that I use to analyze the results of a survey are: (1) plot each shot record; (2) pick the RPE events within each record; (3) construct a histogram of the number of events verses arrival time, and obtain a statistical measure of the survey’s confidence in locating sulphides; (4) plot the event data from the whole survey in a scatter-plot format (arrival time vs. offset/shotpoint location) and look for patterns; (5) try to delineate geological features or provide a tomographic image from the pattern of events. This methodology has evolved over time with the advances in data collection methods, and with greater understanding of the RPE phenomenon. Analysis of the earliest field trial, at the Sullivan Mine, was fairly crude and principally involved steps 1, 2, and 4 with very little interpretation. In contrast, data collected from the Lynx Mine was processed and analyzed using all of the above mentioned steps because of the amount and quality of data obtained.
3.2 Spectrograms

With the arrival of the Gage digitizer a problem arose, the problem of plotting and analyzing signal traces with over 500,000 points. There are too many points to plot as an amplitude vs. time trace; even plotting at 100 points per mm would result in a trace 5 m long. The principal purpose of plotting the shot record is to present the data so that the RPE signals and their arrival times are clearly observable. Segmenting the data (e.g. 2000 segments of 250 points) and plotting the peak-to-peak value of each segment provides a possible solution, but discarding most of the data in this way negates the benefits of a high-bandwidth digitizer. One of the principal benefits of the Gage digitizer is the ability to capture the full power spectrum of RPE signals (1-5000 kHz). A solution to obtaining both high resolution time and frequency information is to plot a spectrogram of the data.

A spectrogram plots the evolution of the power spectrum with time (Cohen, 1995). It is an amplitude vs. frequency vs. time plot. The amplitude is typically represented by a colour or is contoured, and the other two quantities form an orthogonal co-ordinate system. An example is shown in Figure 3.1. There are many sophisticated ways to process the data so that the frequency/amplitude estimation is optimal (Cohen, 1995), but I use the most basic form: a moving window or segment with a windowed periodogram estimate of the spectrum. The trace is divided into 512 segments of 1024 points, and a spectral estimate is made on each of the segments. Each segment represents a 0.1 ms time window, which gives precise timing of an event (the P-wave travels about 0.6 meters in this time). Rather than computing a 1024 point FFT, I produce 8 overlapping segments of 256 points and compute eight 256 point FFT's. From the 256 point FFT's an average power spectrum is constructed. Figure 3.2 illustrates the way these sub-segments are formed from the 1024 point segments of data. The start of the data is zero-padded because of the overlapping of sub-segments. This scheme of an average spectral estimate sacrifices resolution in frequency for a reduced spectral
Figure 3.1  An example of a spectrogram from a high frequency record. This example is from the Lynx mine, and is the spectrogram of the time series plotted in Figure 1.1.
variance (Press et al., 1989). A Parzen window is used for the FFT calculations (Press et al., 1989).

I used the Press et al. (Numerical recipes, 1989) spectral estimate scheme to reduce the graininess of the noise. RPE pulses do not benefit significantly from increased frequency resolution, but the contrast between signal and noise can improve markedly with an averaged spectral estimate. In addition, the smoother estimate allows the removal of continuous (in time) bands of noise in the spectrogram by subtracting a reference spectrum. The reference spectrum is gathered from a noise recording (no shot) gathered in the field. It can also be obtained from sections of the shot record where there is no apparent signal, but I used the former method because it works well and it is a little easier to implement. There is still some variance in the amplitude of the spectral estimate so I normally subtract a little more than necessary (about 5% more than the reference) to completely eliminate the bands. A cost of removing the bands by over-subtracting is that some "hole burning " in the pulse spectrum can occur for weak signals.

![Time Series Diagram](image)

**Figure 3.2** An illustration of the segmentation scheme used in dividing the time-series data into overlapping estimates of the amplitude spectrum vs. time.
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Figure 3.3 A comparison between a spectrogram and an amplitude vs. time trace. The signal is clearly observable in the spectrogram, but the trace shows no evidence of a signal.
Spectrograms have only been applied to data from one field trial (see Chapter 4, Lynx III), but the high bandwidth acquisition system provided astounding improvements over the previous system as judged by side-by-side comparison with the new system. The ability to distinguish signals from noise improved dramatically; an example is shown in Figure 3.3. A large part of this improvement is due to the ability to see concentrations of signal energy in certain frequency bands without the prior knowledge of which frequency bands will be most useful. There are other benefits of using time-frequency analysis. For example, narrow-band signals usually need to be removed in order to see low amplitude arrivals on amplitude-vs-time traces, but in a spectrogram plot the narrow-band signals are visually distinguishable from the broad-band pulses of RPE (horizontal lines versus vertical lines). As mentioned previously, these narrow band signals can be removed effectively by subtracting a noise template.

If the spectral peaks of RPE signals are distinctly related to mineral composition (Sobolev et al, 1986a) then it may be possible to note compositional differences within the orebody from changes in the spectrum as the seismic wave travels through the orebody. Many sulphide orebodies have stringers or fringe zones that are different in composition from the interior of the orebody. This concept has not been tested, but it would be of immense value for interpretation purposes.

3.3 Picking Events

Each shot from a field trial has a record that contains the response of various sensors to effects of the blast. Typically 10-20 ms of pre-blast and 40-50 ms post-blast data is collected on each record. A record contains the response from at least two to six EM sensors. Each of these records is plotted and events that might be due to RPE are picked from the EM sensor traces on the plot.

The EM signature of RPE is a brief pulse, or group of pulses of anomalous amplitude (Sobolev et al., 1982a). Other than this general description there are few useful
characteristics to help in the search for RPE signals within shot records. Electrical discharges and power switching transients share similar characteristics, and are often indistinguishable from RPE. However, signals from the latter are often related to periodic signals, such as the power grid, and may be identified by their periodic occurrence. This was the case with records obtained from the Sullivan Mine, and some from the Lynx Mine, which contained transients in sync with the 120 V, 60 Hz, mains power. In general, any short duration pulse found in the data following the blast is considered to be possibly due to RPE, unless there is an obvious alternative explanation. An occasional exception to this rule is the case where the pre-blast data contains significant amounts of "RPE-like" signals; in this case the record is discarded because a natural or man-made disturbance is in progress and the post-blast signals are probably due to the same process.

Once a signal is identified what are the important characteristics? For example, does a bigger signal deserve more attention than a smaller one? Because the mechanism responsible for RPE is not certain, any interpretation based upon the peak amplitude of the signal is a gamble. Furthermore, peak amplitudes are poorly reproduced in repeat shots. Therefore, I tend to place little importance upon the peak amplitude of the signal, and give each identified signal an equal weighting. The arrival time is an important characteristic because it gives the distance the seismic wave has traveled from the shotpoint. Furthermore, the arrival time is generally reproducible, within the limits set by the duration of maximum forcing of the seismic wave. The power spectrum of the signal may be another useful characteristic (Sobolev et al., 1986). If the type of mineral strongly influences the RPE power spectrum then the various mineralogical differences within the orebody would produce different classes of signal.

As mentioned previously, the basic process of picking events is to look for anomalous pulses amongst power-grid-related spikes and other interference. Spectrograms, high bandwidth recording, and fiber-optic blast sensing have made this job much easier than
it was originally. However, defining an event is not always easy. RPE signals will often appear as bursts of spikes/pulses that continue for many milliseconds. Sobolev et al. (1986) claim that this is an indication of the dimensions of the orebody (multiple signals as the seismic wave sweeps through). Therefore, it is important for interpretation purposes to represent the duration of this activity. Should a burst of pulses be weighted more heavily than a solitary pulse, and how should the weighting be done? If each resolved pulse is given equal weight then one or two shot records can dominate further interpretation because one burst may consist of hundreds of pulses. In addition, the pulses within a burst may not be readily resolved. After considering these issues, I decided to define an event as any pulse or group of pulses within 0.3 ms to 1 ms of the time of the largest pulse within the group. For example, a long burst of pulses is represented as a sequence of events spaced equally apart in time (say 0.3 ms). The time interval of 0.3-1 ms is somewhat arbitrary; however, this time window reflects the accuracy in reproducing arrival times in repeat experiments.

This arbitrariness can be remedied by defining a start and end time to an event, but this solution requires substantial changes in some of the methods and algorithms to be described later in this chapter. These algorithms were designed for single events because this was the predominant type of signal seen in records produced by the lower bandwidth RCE digitizer, which has been the workhorse of my field trials. Spectrograms of data acquired by the (newer) Gage digitizer show periods of activity in nearly every record, and an RPE event should be defined with a start and end time to gain maximum benefit from this type of high bandwidth system.

3.4 Presentation of Raw Survey Data

It is standard practice by many geophysicists to present seismic data by plotting many amplitude-vs-time traces next to each other (i.e. the wiggle-trace format and its variants). Trace position on the plot may depend upon shot-to-receiver distance (a gather), or shot to receiver midpoint position on the survey line. This type of presentation has been
enormously successful since the 1950's in interpreting 2-D surveys to pick geological features. Ground penetrating radar data is also presented in the same manner.

Given the success of "wiggle-trace" system of presentation, and the similarity between seismoelectric methods and seismic methods, it would appear to be a good candidate for displaying field data. However, I find that "wiggle-trace" plots are inappropriate for presenting RPE survey results. Firstly, the raw or processed amplitudes on the seismic and radar traces are representative of a physical measurement of a well defined property. For example, in the seismic reflection method amplitude changes represent an abrupt change of material properties where the amplitude is proportional to the contrast in impedance. As previously mentioned (see section 3.2, Picking Events), we have no real basis with which to judge RPE amplitudes. "Wiggle-trace" plots are designed to draw attention to amplitude anomalies between traces, a property that RPE presentation could do without given the great variation in amplitude of RPE arrivals. In short, I find that "wiggle-trace" plots are ineffective because they tend to emphasize unknown properties in RPE data.

I prefer to use a symbol to represent an anomalous pulse, which I refer to as an event. With this scheme each event has equal importance, and it is the relative timing of events that dominates the presentation. Data from the earlier field trials (Sullivan and Mobrun, see Chapter 4) was corrupted by blast EM, and interpreting the data by analyzing "wiggle-trace" plots was almost impossible. It proved to be easier to pick the RPE events from each shot record and then interpret from a scatter-plot (offset/shotpoint vs. time). This technique works well with data from shot locations that are fired more than once (to improve the chances of receiving signals) because RPE data cannot be usefully stacked due to amplitude variation and jitter of the narrow pulses. Examples of both the wiggle-type and scatter-plot type of representation can be seen in Figure 3.4.
Figure 3.4 Examples of scatter (above) and wiggle-trace plots (below). These are from the same data-set, an experiment at the Mobrun Mine. Grey areas indicate where I picked signals. Note that the interpretations are slightly different in the placement of the second and third grey areas. The scatterplot includes information from repeat shots.
3.5 Statistical Analysis

Statistical analysis of the data-set has been of immense value in recovering useful information from surveys plagued with excessive EM noise and interference. It can also serve as a quality control measure for a survey.

When picking pre-blast events digitized from the lower bandwidth RCE system I found the presence of randomly occurring spikes that are indistinguishable from RPE; both types of events appear as the impulse response of the system. A lot of man-made noise has this character, and so do nearby electrical discharges (e.g. thunderstorms). This random background can obscure seismically induced phenomena. For example, both surface field trials in Queensland, Australia (see Chapter 4, Century), were hampered by large amounts of atmospheric noise (spherics) and an unresponsive orebody. Statistical analysis was the only usable tool in these cases.

If the background of noise spikes displays no discernible pattern or predictability in arrival time then the noise can be treated as a Poisson process (Montgomery and Hines, 1980; Barlow, 1994). The Poisson probability distribution is

\[ P(x) = \frac{\alpha^x e^{-\alpha}}{x!} \quad \text{and} \quad \alpha = \lambda t \quad (3.1) \]

where \( \lambda \) is the rate of events per unit time, \( t \) is the time-interval in which events are counted, and \( x \) is the number of events in the time interval \( t \). Mean and variance of the Poisson distribution are the same, \( \alpha \). As \( \alpha \), the expected number of events, increases the Poisson distribution approaches the normal distribution. With this description of the noise characteristics a model of the distribution of events within shot records can be constructed, and compared with the actual distribution produced by the field data. If there is considerable disparity between the two then this is an indication that the dataset contains events that are not randomly generated.
Once the events have been picked from the various shot records a histogram of the number of picked events per time interval after the blast is plotted. Note that it is important to pick both pre-blast and post-blast events for this purpose. The picked events of a survey are grouped according to arrival time after the blast and counted. Figure 3.5 displays an example from the Lynx Mine. I usually bin the data into 5 ms time intervals. Bins with negative interval limits represent pre-blast events. If the histogram is relatively flat then it is an indication that RPE has not been induced, or the response is not recognizable. However, if the histogram shows large peaks after the blast then it is probable that there is an orebody nearby. The position of the peak in the histogram indicates the approximate distance of the target from the shotpoints. This is very useful information when no other obvious patterns emerge from the data; it indicates the presence and approximate distance of an orebody.

Often it is intuitively obvious, as in fig 3.5, that RPE responses have been produced, but sometimes the peak is small or the noise is very high, and a procedure to numerically evaluate the likelihood that a peak in the histogram is due to a fluctuation in the noise...
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is needed. Hypothesis testing (Hines and Montgomery, 1980) provides such a procedure. It works by constructing a hypothesis about the data-set, then tests are performed on the data to see if this hypothesis can be rejected with reasonable confidence. The hypothesis to be tested is that any peak in the histogram is a natural fluctuation of the random process generating the background noise. If the hypothesis is true then pre-blast and post-blast bins should contain approximately the same count within statistical uncertainties. A nearby orebody would raise the count of the post-blast bin beyond reasonable expectations, and numerical tests would show the hypothesis to be false.

There are two types of error that may occur when hypothesis testing: type I and type II (Montgomery and Hines, 1980). A type I error occurs when the hypothesis is in fact true, but the test rejects the hypothesis. Type II errors are the converse: the hypothesis is false and the test accepts it. The practical consequence of a type I error is that a false target is identified. A type II error would result in a real target being overlooked. Both errors can be costly, and need to be small for the tests to have any significance. Type I errors can be controlled by increasing the sampling (i.e. shoot more). The other type of error depends upon both target contrast and sample size, therefore, a survey is designed to detect targets that provide a sufficient response. Another property of this type of testing is that a "strong" conclusion about the existence of an anomaly can be made; however, when the test supports the hypothesis then it is "weak" conclusion, and we can only conclude that we have failed to identify a target.

The T-test is a hypothesis testing method which can be used to compare means from two different populations, which are normally distributed. In my case I want to compare the mean rate of events before and after the blast at certain time intervals. The form of the t-statistic is

\[ t = \frac{\bar{x}_1 - \bar{x}_2}{S_p \sqrt{\frac{1}{n_1} + \frac{1}{n_2}}} \]

where

\[ S_p^2 = \frac{(n_1 - 1)S_1^2 + (n_2 - 1)S_2^2}{n_1 + n_2 - 2} \]  

(3.2)
where the subscripts denote the population (pre- and post-blast), \( \bar{x} \) the mean, \( S \) the variance, and \( n \) the number of samples from each population. Evaluating equation 3.2 gives a number expressing the amount of deviance between the means of the two populations. The t-statistic is analogous to the distance from the expected mean in units of standard deviation. This number is used with Student's t-distribution to calculate the probability that the hypothesis is false (tabulated in most statistics texts).

The use of a normal approximation to a Poisson process is reasonable if the number of events in each bin is greater than or equal to 5 (Barlow, 1994), and is accurate for background counts greater than 10 events per bin. The normal distribution allows the possibility of negative numbers in the count number, which is not possible, but as the expected number of counts becomes greater the portion of the normal distribution giving negative counts becomes vanishingly small.

The situation of comparing events from different bins is analogous if we substitute the following in to equation 3.2:

\[
\bar{x}_i = q_i \\
S^2_i = q_i \\
n_i = m
\]

where \( q_i \) is the number of events in bin \( i \), \( m \) is the number of records, and \( n_i = m \).

Note that for Poisson distributions the variance is equal to the mean. This substitution results in the following expression for the t-statistic

\[
t = \frac{mq_1 - mq_2}{\sqrt{mq_1 + mq_2}}
\]

If a new variable \( T \) is defined as the total number of events appearing in a certain bin position (e.g. 10-15 ms after the blast) summed over all the records then 3.3 becomes

\[
t = \frac{T_1 - T_2}{\sqrt{T_1 + T_2}}
\]

and we now have a method to compare pre-blast and post-blast bin counts. To use the
tables in various books we need to use the right curve, which depends on the number of degrees of freedom. This number is the number of random variables minus two, \( T_1 + T_2 - 2 \) (Hines and Montgomery, 1980). With very large degrees of freedom the t-distribution is nearly identical to the normal curve and the t-statistic is the distance from the mean in standard deviations (Hines and Montgomery, 1980).

An improvement in our analysis can be made if the pre-blast bins are grouped together and equation 3.4 is weighted for the greater pre-blast time interval.

\[
t_\star = \frac{k_2 T_1 - k_1 T_2}{\sqrt{k_2^2 T_1 + k_1^2 T_2}}
\]  

(3.5)

This formula is not strictly a t-statistic (Hines and Montgomery, 1980), but it is sufficiently close for most purposes. The time interval covered by the group of bins for each population is denoted by \( k \). For example, a count of events 0 to 10 ms before the blast \( (k_1 = 10 \text{ ms}) \) could be compared to the number of events 20 to 25 ms after the blast \( (k_2 = 5 \text{ ms}) \).

Equation 3.4 and 3.5 are valid for finding the probability that a particular bin will reach its value from a natural fluctuation; that is, they have significance only for the time interval tested. This is not equivalent to scanning across the histogram and testing until a significantly deviant peak is found (Barlow, 1989). The correct interpretation of a peak (with \( j \) counts) when scanning a histogram (of \( n \) post blast bins) for significant peaks is

\[
Prob(\text{histogram has a bin } \geq j) = 1 - (1 - Prob(\text{bin } \geq j))^n
\]

(3.6)

Equation 3.6 can be translated as follows: the probability that a histogram has a peak greater than \( j \) is equal to unity minus the probability that none of the peaks will exceed \( j \). Thus, the probability that a histogram contains a noise fluctuation greater than or equal to \( j \) is higher than that for a single time interval. The value of \( Prob(\text{bin } \geq j) \) is found from cumulative probability tables or computed once the t-statistic is calculated by equation 3.4 or 3.5 is obtained.
In practice, I construct a histogram from events picked from the records, and visually inspect for peaks. If a post-blast bin appears to be significantly greater than pre-blast bins then equation 3.4 or 3.5 is used with the event counts from the deviant and pre-blast bins to obtain a t-statistic \( t \). The t-statistic is used to provide a probability that the bins under comparison share the same mean (Hines and Montgomery, 1980). If I am not expecting to see a peak in a particular bin position then equation 3.6 is used to lower the significance of the peak, which increases the probability that the two samples share the same mean. This probability allows me to test the hypothesis that the peak can be attributed to the background of random pulses. If the hypothesis is rejected, then I conclude that the peak is due to RPE because this is the only phenomenon that I know of that can produce such a peak.

3.6 Boundary Delineation

Boundary delineation is an interpretation method based upon the assumption that the received signals come from edges or boundaries of an orebody. There is some experimental evidence (Mobrun Mine experiment, chapter 4) that RPE is more easily produced from the ore-zone periphery. With this assumption and some simple geometric concepts the boundaries of the orebody can be mapped.

The method is based upon the premise that when an EM signal is received the source of the signal is located somewhere on the seismic wavefront. In the underground context, this wavefront is approximated by a sphere centered upon the shotpoint. Note that if the medium is very inhomogeneous then the wavefront has a complicated structure, and the simple geometry would be a poor approximation. Luckily, many underground environments are relatively homogeneous (a 5% variation in local seismic velocities is the most I have measured) and detailed seismic modeling is unnecessary.

Boundary delineation in its present form is a two dimensional graphical method of interpretation. Each event has an arrival time \( t_i \) and a shotpoint location \( (x,y) \). From
Figure 3.6 A graphic illustration of the boundary delineation method. In this method the seismic wavefronts at the time of signal reception are reconstructed to form a picture of the interface.
our knowledge of the local seismic velocity ($v_p$) a circular arc of radius $r = v_p t_i$ is drawn around $(x,y)$ to show the wavefront position at time $t_i$. An example of this step is illustrated in Figure 3.6. In general, I use a scatter-plot to try to pick reflector-like grouping of events (i.e. a sequence of events slowly changing in arrival time from one shotpoint to the next) and then I use this list of events to plot the circular arcs. Plotting all of the survey's events at once will often produce a very confusing picture. After the circular arcs are drawn then common tangents are drawn between adjacent arcs to form a profile of the boundary (Figure 3.6).

There are some problems with the 2-D algorithm (a 3-D algorithm is possible, but it is very difficult to interpret and present 3-D results). The greatest problem is the choice of plane to work with. For example, the plane might be horizontal and include the shotpoints, or vertical and aligned with a regional geological structure. The choice of working plane will influence the type of picture seen by the interpreter, and needs to be chosen (or guessed) with care. Another difficulty is that there are two lines produced by the method: one on either side of the line of shotpoints (Figure 3.6). Drill data or other geological constraints may be used to reject one of the boundaries, but if this data is unavailable then both boundaries have to be considered.

3.7 Tomographic Reconstruction

The principal idea behind tomographic reconstruction is that a region of geology that responds to one shot will also respond to other shots. Arrival time information from shots at various positions allows the triangulation of the source position (Sobolev and Demin, 1980). The method of triangulation is perfect if the source of the signal can be reliably triggered and identified. Unfortunately, RPE is less than perfect in both reliability and uniqueness of signature. To overcome the imperfect behavior of RPE, I have developed a algorithm that attempts to reconstruct the structural form of RPE-producing regions. This tomographic reconstruction algorithm appears to be fairly
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robust, and has worked well in at least one instance when applied to "real" data (see Chapter 4, Lynx Mine III).

The algorithm for reconstruction is relatively simple. Firstly, the volume is divided into volume elements via regular grids. Each volume element is a cube 4 to 10 meters a side and has a number associated with it, which is initially set to zero. It is assumed that the seismic velocities are relatively homogenous and isotropic so that the seismic wavefront can be accurately represented by a spherical shells. A list of event arrival times and the co-ordinates of the shotpoint that generated each event is prepared. The algorithm scans the array of volume elements for volume elements that could be responsible for a particular event on the list. That is, the (Cartesian) distance between a volume element and the shotpoint responsible for the event is compared to the distance of the event (i.e. arrival time x acoustic velocity). If there is a match then the number associated with the volume element is incremented. Therefore, the number associated with each volume element corresponds to the number of circular arcs passing though that volume element (see Figure 3.7 for an example), which I shall call the count. If a region is actively producing RPE signals then the volume elements of this region will have a large number of counts. This is the basis of interpretation for my tomographic reconstruction algorithm.

This tomographic algorithm is not strictly an inversion process because it does not try to find a particular solution. There is no minimization of a norm. Instead, all possibilities are represented with the expectation that the real structure will dominate the picture if there is sufficient data.

There are two main departures from ideal conditions for the reconstruction algorithm in practical field data collection. One is that we are often imaging a three dimensional structure with a limited amount of freedom in shot placement. To locate a point source it takes at least four shot locations (Sobolev and Demin, 1980). The four points cannot be co-planar, otherwise, an image of the source will appear on the other side of the plane.
In underground experiments this means collecting data from different elevations, which I have been unable to do because of the potential difficulty in co-ordinating shot timing and data acquisition between two mine levels. Constraints from regional geology and drill data can be used to reject these image sources. The worst case scenario for tomographic imaging is a survey where the shotpoints are in a line. Because there is only one degree of freedom in shot placement (along the tunnel) the tomographic image will be a series of toroids and rings sharing the same axis as the tunnel.

![Diagram](image.png)

**Figure 3.7** A demonstration of tomographic reconstruction. In this example there are four targets and four shotpoints. Note how the poor resolution of the targets outside the region encompassed by the shotpoints.

The other significant problem is that most shot records contain more than one signal. Trying to match signals across records is impossible as they are usually indistinguishable (apart from amplitude, in which no two are alike). This can result in non-uniqueness in the problem of locating the source. An example is shown in Figure 3.7, where there is more than the three shotpoints necessary to locate a source in 2-D, but there are four indistinguishable signals from each location. The net result is that one or two other
possibilities arise for the source locations. Source location degeneracy is removed by having more shotpoints locations than is required for one signal.

Given the above mentioned problems and properties of RPE what is the best survey geometry? The ideal survey would be to surround the target of interest with a number of shotpoints on at least two different elevations. This is impractical for a lot of underground work (if we confine ourselves to the tunnels and cross-cuts, but not if we use drill-holes). For most surveys the minimum is to have at least two roughly orthogonal shot-lines, and to try and shoot from as many directions as the various drifts, cross-cuts, and shafts allow. At least two degrees of freedom in shot placement is necessary for tomographic reconstruction to work.

Figure 3.8  A map of the synthetic survey example. The Ore-zones are the shaded areas, and the shotpoints of the survey are marked by stars.
A synthetic survey demonstrates the capabilities of this algorithm. In Figure 3.8 we have a two dimensional problem in locating a long thin lens that is split by a fault. This is not an easy test for the algorithm, but the orebody's shape and the survey geometry represents a fairly realistic challenge. I generated 145 events from the 30 shotpoints by assuming an even (but random within) distribution of RPE events from the orebody. Each part of the orebody reacts in the same manner, and host rock does not react at all. In the mock data no events were generated more than 100 meters from the shot (the effective range of the charges), and a higher probability was given for generating RPE at distances of 50 meters or less (a factor of two). I usually shoot two shots per shotpoint in a survey. Applying this rule to the mock dataset gives a productivity of two or three events per shot; a fairly typical number for my surveys. Tomographic reconstruction produces a fairly good image from this dataset (Figure 3.9). Placement of the anomaly is accurate, but it deteriorates outside the area bounded by the shotpoints. This a consequence of fewer shots contributing constructively to the count in the outer regions of the survey. The small lens is poorly imaged because of the lack of shots to the north, but it is clearly detected nonetheless. Spatial resolution of the tomogram is somewhat poor because of the 10 meter shotpoint spacing and the comparatively narrow features to be imaged. In practice, 5 meter spacing would be preferable, and used if circumstances (time/cost) permitted.

A practical disadvantage of the reconstruction algorithm is that it does not provide quantitative information about material properties. This can be attributed to our lack of knowledge about the RPE mechanism. However, the count from my algorithm appears to give a good indication of the presence of sulphides, which is a very desirable result. Seismic and electrical methods (at present) do not give a quantitative value on the amount of sulphide. The presence and amount is inferred from the physical parameters (e.g. p-wave velocity or conductivity), often with poor accuracy.
Figure 3.9

The tomographic reconstruction of the synthetic orebody. Red indicates regions where there might be ore-bearing zones.
CHAPTER 4

FIELD TRIALS

4.1 Field Program Objectives

The principal aim of the field program was to confirm the existence of the phenomenon described by Sobolev et al. (1980, 1982). In essence, verify that pulses of EM energy are produced from sulphide minerals when seismically excited. Published material from Sobolev's group is rich with the descriptions of the character of the emissions (1982) and related properties (1984), but they give few details about the experiments. Why are they so sure that the signals come from the sulphides? Alternative explanations, such as triboelectric effects from the blast, would appear to be equally viable. Furthermore, what are the implications of the inability to exactly replicate the seismoelectric response (Sobolev et al., 1982), and how serious is the problem of replicating results? It is difficult to understand some of their claims without some direct experience with the measurement of RPE.

Another purpose of the field program was to investigate the possible use of this phenomenon as an exploration tool. Discussions between the Russians, ourselves (R.D. Russell and M. Maxwell of the U.B.C. Instrumentation Group), and Bob Smith, the head geophysicist of C.R.A. (an Australian mining company) led us to believe that Demin and Maybuk were using RPE for exploration. In fact, they were offering their services to Western companies. Some mining companies are known to be interested in RPE, but do not want to commit themselves to a method that is neither understood or proven. The techniques used by the Russians to acquire and interpret RPE is uncertain despite information from patents granted to Sobolev et al. (1986). Furthermore, no information on the effectiveness of their techniques exists. In short, there was an aura of mystery about the whole business of RPE-based exploration.
To summarize, the aims of the field program were to assuage any doubts about the existence of the RPE phenomenon and to evaluate its potential for mineral exploration. Whilst achieving these broad aims I intended to learn more about the phenomenon and develop a system that could reliably measure the EM attributes of RPE in the field. The measurement and analysis tools developed would also provide a basis to evaluate RPE as an exploration method.

4.2 Sullivan Mine

Two previous seismoelectric experiments were conducted at this site, in 1983 (Sobolev et al., 1984a) and 1986. Both involved members from the U.B.C. research group, but not myself. Due to the encouraging results from these two experiments, and a well defined geology, the site was chosen to test various seismic sources and new instrumentation. The objectives were to clearly observe seismoelectric emissions, if any exist, using our most recent instrumentation, and to determine the best source for accomplishing this.

Experiment Details

The Sullivan mine is a lead-zinc sulphide mine located in Kimberly, B.C., operated by Cominco Ltd. It’s orebody is a 160 million ton, gently dipping, iron-lead-zinc sulphide lens which lies conformably in Protozoic clastic meta-sedimentary rock (Hamilton, 1982). Footwall rocks are graded quartzwake and mudstone beds 10 to 30 cm thick. Primary mineral constituents are quartz, sericite, biotite and some pyrrhotite laminations up to 1 mm thick occurring 5 to 15 m below the "main band" of the ore body. The main band is 3 to 24 m thick and consists of a succession of fine grained pyrrhotite, sphalerite and galena beds. Typical grades of the main band are about 10% Pb and 15% Zn, with 25% Fe occurring mainly as pyrrhotite. The main band is overlain by a succession about 20 m thick comprising 35% sulphide-rich layers in four bands intercalated with three sulphide poor interbeds of mudstone and quartzwake. The hanging walls are graded quartzwake and mudstone.
Our experiments were conducted over a three day period, 16-18 April, 1991, at level 2700, in the eastern part of the mine (Figure 4.1). Participants in these experiments were R. D. Russell, B. B. Narod, and myself from U.B.C, and a blaster/loader (Grant Scott) loaned to us from the mine workforce. Shot points were placed in opening 2713 at 10 m intervals. The orebody is approximately 30 m above opening 2713 and dips 20 degrees north-west. Instrumentation and the portable computer containing the digitizing card were placed on a powder-car in opening 2711 near the triple junction with 2713, and the opening to the hoist. The nearest shot point was 40 m distant, and the furthermost was 140 m distant. Measurements of the P-wave velocity were on average 5700 m/s for this area, which is similar to the 5800 m/s measured in previous experiments at this site.

Figure 4.1 Map of the Sullivan Mine experiment. The orebody lies approximately 20-30 meters above the shotpoints, and dips north-east.
Numerous types of seismic sources were experimented with at shotpoint SP2 (Figure 4.1). Non-explosive sources were stacks of 20 blows from a sledge-hammer, and 1000 blows from a pneumatic rock drill. Explosive sources were: detonator, detonator with a small primer jacket, 1/2 and full stick of emulsion type explosive (<0.4 kg), pentolite primer (0.17 kg Trojan made by Austin Explosives), and 12 gauge blank shotgun shells fired from a tripod mounted "buffalo" gun. Little or no damage resulted from the use of these sources, and the same hole (SP 2) was used for several shots. Explosives were detonated with electric blasting caps from a special blasting circuit made at U.B.C. that disconnects the electrical circuit once detonation begins to prevent electrical interference. The delay between the current pulse and the blast was set to approximately 0.6 seconds by choice of a #25 blasting cap.

At least two magnetic sensors were deployed during each of the experiments, and an electric field sensor (Long Wire Antenna) was added on the last day. These sensors were placed 30 m northwest into opening 2711. Magnetic sensors consisted of three models, UBC 1, UBC 2 and some UBC 3 types. The UBC 3's proved to be too insensitive for this application, and UBC 2 was too narrow in bandwidth (1-10 kHz). Because UBC 1 is capable of a 300 kHz of bandwidth an AM demodulator was connected to this sensor (see Appendix B.8 for description of demodulator). Each of the sensor signals were transmitted to one of the Tektronix AM 502 amplifiers at the instrument site, and band-limited to 30 kHz before being recorded. The AM502's were recently purchased, and only four were available for this experiment.

This experiment tested the new amplifier system (AM 502) for the first time. Another first for RPE investigation was the use of a digitizer in the field. Previously, data were recorded on an analog magnetic tape system and then digitized later. The RCE digitizer card recorded 8 channels at 8 microseconds per sample. In general, a geophone close to the source would trigger the start of the acquisition process. Shots on the 18th of April used the EM emissions from the explosion itself to enable the trigger.
All of the equipment was battery powered with the exception of a portable Tektronix oscilloscope, which was only used for diagnosing problems with the equipment and examining noise from the sensors. An emergency shelter and lunch room 180 m northeast from the instrument site provided mains power for the oscilloscope and a 60 W safety light. Experiments with the mains on and off showed no discernible difference in extraneous noise.

Discussion of Results

Only the pentolite explosives produced any evidence of seismoelectric signals. Not a hint of signal was observed with the other sources, which contrasted greatly with the copious amounts of EM activity after the blast from pentolite sources. No distant geophone records are available from these tests, so there are no quantitative comparisons between the sources. A geophone placed near the shot produced no useful information because the ground motion was masked by air-wave effects. However, it was obvious that the pentolite explosives produced a seismic impulse with higher amplitude and frequency content than the other sources. The difference between the pentolite explosives and the emulsion type explosives (Forcite) could be felt through our feet: pentolite explosives gave a distinctive crack, whereas the other explosives produced a muffled thump (an effect noticeable a fraction of a second before the air-wave arrives).

A 100 m profile along 2713 between SP2 and SP12 (Figure 4.1) yielded 7 useful records from 6 shotpoint locations. All of these records show seismoelectric events, an example of one of these records is shown in Figure 4.2. Trains of EM pulses appear 4 to 10 ms after the explosion (Figure 4.2), which is consistent with excitation of the ore zones by the seismic wave. In some records the signals were clipped by the amplifiers because of the unexpectedly high amplitude of the phenomenon (>10 nT and >4 mV at the sensor). Signals from the AM demodulator (connected to UBC I) and the long wire antenna exhibited greatest amounts of signal. The former indicates significant amounts of high frequency energy not accessible to the lower bandwidth sensors, the latter demonstrates
the merits of measuring electric fields. As the shot point moved away from the sensors, smaller signal amplitudes were recorded.

Figure 4.2 A record from SP2, Sullivan Mine. The uppermost trace is from a geophone near the shot, and the others are from various EM sensors. It was expected that the signals would start at about 4 ms after the blast, the time to reach the main orebody. However, there is substantial amounts of signal earlier than 4 ms. These early signals may be due to small amounts of sulphide near the shotpoint, or emissions from the blast. This burst of early activity appears to cease after approximately 3 ms, and is characteristic of all of the records from the Sullivan Mine.

There is 3-5 ms of blast-associated EM in all of the Sullivan Mine records. The duration of the blast-associated EM masks any seismoelectric signals less than 20-25 m away from the shot point, and it makes it difficult to determine whether a signal is from the sulphide zones, or from some phenomena near the shot-hole cavity (O'Keefe and Thiel,
Two explosives were freely suspended 1-2 m from the tunnel ceiling and set off, to measure the intensity and duration of noise from the explosion plasma. Both shots gave less than 2 ms of EM impulses with much less amplitude than the shots in the drill-holes. This test indicates that the explosive plasma is not the source of the (seismoelectric) signals on other records.

A comparison between drill core results and reconstructed wavefronts, Sullivan Mine. The arcs in this diagram represent the seismic wavefront at times of significant EM activity. The burst of signal activity immediately after the blast was not included in this interpretation, as these emissions were thought to be possibly related to processes from the blast itself.

A comparison of signal arrival data and the drill data is shown in Figure 4.3. From Figure 4.3 it can be seen that most the arcs will intersect the orebody at some point. This shows that it is quite plausible that the orebody produces these signals. However, the elimination of events before 3 ms and the proximity of the main orebody makes it fairly easy for the data to fit this scenario. A previous seismoelectric experiments in this area (Sobolev et al., 1984a) obtained very similar arrival times, and also saw many early arrivals.
Conclusions

Narrow pulses of EM energy were observed at times after the explosion that are consistent with seismogenic effects occurring within the orebody. Effects from the blast cannot be unequivocally excluded as the source of the signals. However, the similarity of my results to the 1983 and 1986 experiments provides further support for the existence of RPE.

The instrumentation performed well in the underground environment with the demodulator and LWA performing very well in receiving these signals. A new type of explosive source, the pentolite primer, was discovered to be very effective in exciting the orebody without damaging the mine infrastructure.

4.3 Mobrun Mine

The principal aim of this experiment (Kepic et al., 1995) was to show that RPE originates from sulphide minerals. The Sullivan data did not prove this hypothesis because the data may have been contaminated with blast-associated effects. At the Mobrun Mine the plan was to move the shots away from the orebody and observe the resulting arrival times of the signals. If the signals arrived progressively later, and at times consistent with conversions within the orebody, then this would demonstrate that the orebody was responsible.

Experiment Details

The Mobrun Mine is operated by Audrey Resources, and is located near Rouyn-Noranda, Quebec. Rio Algom discovered the main lens in 1956 by means of a mobile EM road survey (Seigel, 1957). A recently discovered iron-copper-zinc orebody, the 1100 lens, was the object of our investigation (Figure 4.4). It is located 250 m south of the main lens, and is 360 m below the surface. The 1100 lens is hosted in a large pyroclastic unit bounded to the south by rhyolite and andesite flows, and by a thick rhyolite unit to
Figure 4.4 A map of the Mobrun Mine experiment. Map co-ordinates are in meters.
the north (Caumartin and Caille, 1990). It is a vertically-oriented massive sulfide body extending 300 m east-west laterally, and from 360 to at least 740 m below the surface, with a maximum of 50 m thickness. The orebody is split along its longitudinal axis forming two zinc-rich lateral fringes. The sulfides consist of fine to medium grained, partially granulated pyrite matrix, containing 5 to 15 percent sphalerite as irregular shards and clusters, and 1 to 5 percent disseminated, very fine grained chalcopyrite (Caumartin and Caille, 1990).

Our tests were performed in a ramp that descends south-east from level 6 to the top of the 1100 lens (Figure 4.4). Participants in the experiments were M. Maxwell, G. Mellema, and myself. One of the regular miners assisted with drilling of the shot holes and blasting.

The seismic source used in our tests was a 0.22 kg booster charge made of pentolite. Each charge was placed untamped into a 1.5-2 m drill hole, and detonated with a four-second electric blasting cap. The four-second delay allowed the sensors to fully recover from a momentary overload induced by large currents from the blasting box. Shotpoints were placed at 5 m intervals along the ramp, starting with shotpoint zero at the intersection of the ramp and the orebody, and finishing with shotpoint sixteen, 80 m further up the ramp and 50 m away from the nearest ore-zone (Figure 4.4). Shotpoints 12, 13, and 15 were not used because of the possibility of damage to mine infrastructure. After the initial traverse, the shot holes generally showed little damage, and were reused; some holes were used three or four times with no significant difference in seismic output.

An array of electromagnetic sensors and the recording site were located 100-130 m up the ramp from shotpoint zero. The electric field was measured by low noise (30 nV/√Hz) pre-amplifiers connected to either two stainless steel stakes set into the ground (grounded dipole), or one stake and a long insulated wire acting as a capacitive pickup (long wire antenna or LWA). Magnetic sensors consisted of an integrated preamplifier
fed by broadband coils wound on a ferrite rod (Appendix B.4). The bandwidths of the electric field sensors were 1 Hz to 30 kHz (long wire antenna) and 1 Hz to 10 kHz (grounded dipole), and of the magnetic sensors were 1-300 kHz and 2-7 kHz (the UBC I and UBC II models respectively). A bank of Tektronix AM502 amplifiers at the recording site amplified and filtered (generally with to a bandpass of 1 to 30 kHz) the signals from the sensors before they were recorded by the RCE data acquisition system at 62.5 kilosamples per second. To record the presence of frequencies beyond the Nyquist frequency of the recording system, a wide-band, precision rectifier (similar to an AM demodulator) was connected to the 1-300 kHz magnetic antenna (See Appendix B.8 for details). All of the equipment was battery powered to reduce electromagnetic interference and enhance portability. A geophone at the instrument site monitored the seismic field near the sensors, and was used to calculate P-wave velocities, which were found to be approximately 5400 ± 200 m/s.

Figure 4.5 A record of a shot at SP7, Mobrun Mine. The four lowermost traces are from EM sensors. These traces show signals attributable to the orebody, labeled as RPE, and the same type of near shot EM seen in the Sullivan experiment.
Discussion of Results

An example of the data acquired during a shot is shown in Figure 4.5. Most of the EM activity 0 to 3 ms after the blast is thought to be caused by small amounts of sulfides near the shot. Close examination of the early portions of the records shows that the "blast-associated EM" does not always start with the blast, and it is often delayed. I found that the delay could be attributed to the time for the seismic wave to reach a narrow zone of small pods of sulphide minerals. This zone cuts through the shot line at SP3 and SP9; records from these shotpoints tend to have blast EM starting at 0 ms. In addition, at SP3 and SP4 the smell of sulphur dioxide was distinctively present after the blast; indicating that our shots had been in contact with sulphides. I believe that relatively small amplitude emissions from the blast area were re-radiated by the blasting leads connected to the explosive. These wires pass near the EM sensors used in our experiments, thus, induced currents in the blasting leads may have communicated the near-blast emissions.

Pulses on the EM sensor traces in Figure 4.5 occurring later than 3 ms after the blast are examples of RPE. The LWA sensor trace illustrates best the unique properties of these signals: typically, a short duration pulse or a small group of pulses of one polarity. The earlier EM activity, by contrast, exhibits no preference in polarity and is more oscillatory in character. In addition, it is expected that the randomly polarized emissions from the shot area will be received most strongly by the sensor closest to the shotpoint, which is the LWA, because EM field amplitudes decay rapidly with distance from the source. It can be seen in Figure 4.5 that the LWA does in fact respond most strongly to the early, near-shot EM. These differing characteristics were used to pick the RPE signals. The appearance of RPE signals and the presence of early EM appear to be unrelated; sometimes there were no RPE signals and only early EM, and sometimes otherwise, but in most cases both appeared on our records. The LWA response in Figure 4.5 does put some doubt upon the second (and later) group of arrivals labeled as RPE because they
appear weakly on all antenna except the LWA, but the consistent negative polarity of the pulses suggests that these signals are strongly polarized and it may be that the LWA is favorably oriented. Note that the arrival of the seismic pulse at the instrument site (ISG trace in Figure 4.5), about 18 ms after the blast, is well after most of EM signals have been recorded, demonstrating that the EM signals are unrelated to any seismic-to-antenna coupling.

Sobolev et al. (1982) mention that a characteristic of RPE is the weak repeatability of size, shape, and arrival time of emissions, which they attribute to irreversible processes involved in the signal generation (e.g. the creation of cracks). To test this claim we refired many of our shotpoints. Interpretation was often difficult because it is hard to tell the difference between the RPE spike and the early noise from shots close to the orebody. Although the traces were not identical in appearance, I found that most repeat shots contained the same RPE signals with only slightly different arrival times (within ±0.5 ms) and different amplitudes than the original (Kepic et al., 1995). Some irreproducibility appears to be an inherent part of this seismic-to-electric conversion process because the seismic pulse shape was fairly consistent for each shotpoint. Despite the less-than-ideal reproducibility, our repeat shots show that the consistancy of signal arrival times is sufficient to identify targets.

The raw data displayed in Figure 4.6 supports the claim that the signals with RPE characteristics come from the orebody. In panel (a) the best looking traces were chosen from each shotpoint; best looking means that the first RPE signal dominates any early EM. This trace was generally from the HF dipole, which was furthest from the shots, but was not used throughout the tests so some shotpoints were not recorded with this sensor. Panel (b) is from the AM demodulator, which tends to emphasize oscillatory signals over pulses, but was used extensively during the tests and shows the later arrivals better than panel (a). Shotpoint to ore-zone distances, to an accuracy of about 5 m, were obtained from drawings provided by the mine. Distance estimates from
Figure 4.6  A plot of EM activity vs. offset, Mobrun Mine. In panel (a) traces were selected from the various antennas based upon the clarity of the first arrival. In panel (b) only the demod B trace was used. Grey areas indicate my picks.
Chapter 4: Field Trials

shotpoint to ore-zone, to an accuracy of about 5 m, were obtained from drawings provided by the mine. Given the distance to the nearest ore-zone and the P-wave velocity we can calculate the expected arrival times, which are marked by the intersection of the traces with a dotted line. Eight traces (SP 3 to 6, SP 9 to 11, SP 14, and possibly SP 1 and SP 2) contain a spiky signal that moves out with a slope consistent with conversions at the nearest portions of the orebody. This trend is very clear in panel (a), eliminating the possibility that EM emissions from the explosion are responsible: the high correlation with the expected arrival times makes an explanation based upon a random process implausible. Such quasi-random EM emissions have been reported by O'Keefe and Thiel (1991) in their observation of quarry blasts and they have proposed that spalling and stress releases from the shattered rock near the shotpoint are responsible. This type of process might explain the presence of the early or near-shot signals (see Figures 4.5 and 4.6), but these signals do not start at the same time as the blast and in some cases are noticeably late (SP14 for instance), unlike the emissions observed by O'Keefe and Thiel. The offset between the dotted line and the RPE arrivals is probably due to the time taken for the seismic energy to reach peak amplitude (approximately ms, see shotpoint geophone trace in Figure 4.5), or from errors in the estimates of arrival times from the mine drawings. This dataset provides firm evidence that the impulsive EM signals identified as RPE on our records arise from seismoelectric conversions by massive sulfides within the 1100 lens because no other nearby geological structures or known physical process could plausibly produce the same moveout.

The first arrivals appear to delineate the near edge of the orebody, as expected, but if the rate of signal generation is proportional to the density of sulfide minerals then we would expect sustained bursts of signals as the seismic wave sweeps through the orebody until there is insufficient energy to excite the RPE mechanism. A glance at Figure 4.6 shows that this is clearly not the case. The most responsive portions of the 1100 orebody appear to be the edges normal to the wavefront, or nearest to the shot. It may be that RPE is stimulated most efficiently at the interfaces between the host material and sulfide
minerals, or conductive orebodies such as the 1100 lens may attenuate signals emanating from interior points. Given that it is the edges of the orebody that respond best, what is the significance of the late RPE signals, which can be seen in Figure 4.6 on traces from SP2 to SP10? The plan view of the orebody in Figure 4.4 offers a clue: the western section of the orebody splits into two roughly parallel lenses. If the estimated travel times to the more distant lens are compared with the arrival times of these later signals a good match is obtained. In Figure 4.7 our identified RPE arrival times and the calculated travel times to the nearest portion of the two lenses, A and B (the B lens is north of A), from each shotpoint are plotted for comparison. Our picks from the data closely match the expected arrival times, too well to attribute to chance. The possibility that the second arrival pattern is due to a seismic reflection is unlikely as there is no obvious reflecting structure with the right geometry, and little indication of strong reflectors on our geophone records. Nor is it conceivable that S-wave conversion at the nearest ore-zone is responsible for the second group of arrivals because the ratio of first arrival time to second arrival time varies too greatly. The data shows two groups of RPE signals that appear to originate from seismoelectric conversion at the near edges of the A and B lens structures of the 1100 orebody.

The discussion of results so far has shown the good correlation between the signal arrival times with the known geology. Since we feel certain that the spiky signals are RPE, and that they originate from zones of economic interest, what could we have deduced about the orebody using a minimum of prior geological information?

A suitable method of interpreting the data is to look for boundaries by finding common tangents to the circular arcs in a section (described in Chapter 3, Boundary Delineation). Because boundary delineation is a 2-D method a plane orientation must be selected for the interpretation. For the interface to be mapped accurately it must lie near the section plane. Drill results, at 50 m spacing, indicates the orebody is approximately lens shaped, and when projected onto the surface strikes east-west along 10 000N (Figure 4.4); thus,
the 10 000N plane is a plane likely to be successful. Furthermore, the ramp intersects the orebody at elevation 4970 meters so a horizontal section in that plane would also be a good candidate. Figure 4.8 shows the resulting interpretations made from these crosssections. Shaded areas in Figure 4.8 mark where I have picked interfaces from the wavefront positions, which are shown as arcs in the diagram. The arcs were shortened for clarity, and others for which no common tangents could be found are plotted in the same direction as those that were successfully connected. In Figure 4.8a the interpreted interfaces correlate well with the estimated positions of the ridges associated with the A and B lens features. The first arrivals in Figure 4.8b correctly delineate the nearest interface, the top of the B lens, but the second interface has been mapped to a lower position (13-15 m). In fact, it is located at approximately the same elevation as the nearest interface, but lies in the 9 970N plane as is apparent in Figure 4.8a. Clearly, the selection of the 10 000 N plane (Figure 4.8b) was not a good choice for delineating the A lens.

![Graph](image_url)

Figure 4.7 A comparison between signal arrival times and estimated travel times to the nearest portions of the A and B lens features. The picks are from the traces displayed in Figure 4.6.
Figure 4.8 Application of the boundary delineation method to the Mobrun dataset. Arrivals from repeat shots were included in the analysis. Arrows in this diagram represent the seismic wavefront at times of significant EM activity. The method has successfully outlined two prominent interfaces associated with the A and B ore lenses.
Conclusions

The data obtained from the Mobrun Mine shows compelling evidence that RPE originates from sulphide minerals. It appears that the fringes of the orebody produced most of the signals. This may be due to the conductive inner portions of the orebody attenuating the signals, or favorable mineralogy within the fringes.

The inconsistent nature of RPE arrival times and amplitudes hindered interpretation, but it was found to be sufficiently consistent to determine the source of the signals. Including some simple geologic constraints enabled the boundary delineation interpretation method to map the nearest interface accurately, and to do fairly well on the more distant interface. These interfaces were 0 to 60 m distant from the shotpoint, thus demonstrating the ability of an RPE-based method to image a target to 60 m.

4.4 Lynx Mine I

The principal purpose of this experiment was to determine if the seismoelectric method could detect sulphide ore that was primarily sphalerite. Another key aim was to attempt to delineate the boundary of the orebody, or produce an image of the ore structure. Other aims were to extend shot-to-orebody and sensor-to-orebody distances, to evaluate new instruments, and to generally improve productivity.

Experiment Details

The Lynx mine, operated by Westmin Resources, is located in the centre of Vancouver Island, B.C. Typical ore grades from the Lynx mine are 7.8% Zn, 1.3% Pb, 1.2% Cu, 135 g/t Ag and 2.8 g/t Au. In decreasing order of abundance, the main sulphide minerals are pyrite, sphalerite, chalcopyrite, galena, tennantite and bornite (Pearson, 1993). The lynx orebody is a faulted and folded array of individual lenses along a 2700 m strike length. These ore lenses occur within a stratigraphic unit known as the Lynx-Myra-Price Horizon (Pearson, 1993). This horizon forms an asymmetrical anticline, with the ore on the
Figure 4.9  A map of the experiment on level 14, Lynx Mine. Map co-ordinates are in feet.
south side of the anticline called the S-zone, and the orebodies on the north side called
the G-zone. Our initial tests were conducted near lenses belonging to the S-zone of ore
lenses. S-zone lenses are several meters thick, and dip steeply at about 70-80 degrees. G-
zone lenses tend to be thicker, and dip at angle of approximately 40-50 degrees. In both
regions hanging wall rocks are andesite, and the footwall rocks are rhyolite.

The tests were on a portion of the S-zone on level 14 in the Lynx Mine in a passage
running roughly parallel with the orebody along 11100N between 2000E and 3000E
(Mine co-ordinates, Figure 4.9). Measurements were carried out during May 11 to 15,
and July 15 to 19, 1992. Participants in the May experiment were M. Maxwell, B. B.
Narod, and myself. In the July experiment R. D. Russell, M. Maxwell, K. E. Butler joined
me in the fieldwork.

Approximately 100 shots were fired, with most fired between 2100E and 2600E (Figure
4.9). Each shot consisted of a 0.22 or 0.45 kg primer charge (pentolite) placed in a 6 ft or 8
ft drill-hole. The May tests used the smaller charge size, larger charges were used in July
tests because I was concerned about the lack of observed signals from the May
experiments. The fiber optic trigger made its debut during the May experiments. Safety
fuse detonators were used with the fiber optic trigger to eliminate any wires leading into
the explosive.

These tests included some newly developed EM sensors (UBC V and the High
Bandwidth preamplifiers). The wider bandwidth of these sensors was not exploited, but
the lower noise proved to be very useful. Electric and magnetic field antennas were
placed at three locations and monitored EM in the 1 kHz to 60 kHz band. In general, two
orthogonal horizontal dipoles measured the electric field, and three orthogonal UBC V
magnetic sensors measured the magnetic field. In the May tests the UBC V sensors
exhibited positive feedback problems, and two were replaced with the older UBC I sensor
and a long wire antenna arrangement. Three component magnetic and electric field
data were acquired in the July tests with three improved UBC V antennas, two horizontal dipoles, and a small vertically oriented parallel plate dipole. A geophone was placed with the sensors to monitor ground motion, and obtain velocity information. P-wave velocities were measured to be about 5800 m/s.

All of the sensors were grouped together, but moved occasionally because of concerns about signal quality. In the May experiment the sensors were grouped near the instrumentation site at 2800 E (Figure 4.9), which is up to 300 m from the excited portions of the orebody. It was decided to move the sensors closer to 2100 E in the July tests, but half-way through the experiments we realized that the tunnel walls in this area were covered with a reinforcing mesh of steel. The sensors were moved further south to reduce the shielding effects from the nearby conductive structures.

Discussion of Results

I identified three types of seismoelectric response, which I have labeled as fuzz, spike, and burst types (Figure 4.10). Fuzz signals appear as an increase in random electromagnetic activity, with no preference in polarity or in frequency. A spike is a solitary pulse lasting less than 0.1 ms. Bursts look like a sequence of spikes on the high frequency antennas (1 kHz low-cut), but substantial amounts of lower frequency energy was evident on a dipole set with a 10 Hz low-cut. Of the three types of response only the spike signals can be clearly attributed to the orebody.

The fuzz signals were seen during our first test, when the antennas were placed at 2800E, but were not seen on any of the second test records with the antennas placed near 2100E. Also, only the shots sited from 2400E to 2600E induced these signals. The arrival times of the fuzz signals are before the seismic wave could have reached the main orebody; therefore, these signals appear to originate in the host rock near the shotpoints. It is possible that signals could be due to the presence of small amounts of sulphides in the host rock, or a number of nearby ore pods (a conclusion reached after discussions with
the Lynx Mine geologist, Mike Becher).

Figure 4.10 Examples of the three types of signal recorded from shots at level 14.

The burst signals have many characteristics described by Sobolev et al. (1982, 1984) and are clearly induced by the blast, but the signals come predominantly from shots in only one region, 2100E to 2200E. Records containing this type of signal usually have five or six bursts arriving from apparent distances of 10 meters to more than 200 meters from the shotpoint. If the bursts are due to sulphides then, according to Sobolev and Demin (1986), the number of distinct ore packets in the region is equal to the number of bursts, and the approximate size of the orebody can be deduced from the duration of the burst. Because many burst signals consistently arrive at times when the acoustic wave is more than 200 meters from the shotpoint we would expect that many of the other shotpoints would be able to induce these signals. Unfortunately, we were not able to detect enough of these signals at other shotpoints to identify the source. The burst type of signal may
contain valuable information about the geology, but I was unable to confirm which geological unit is responsible; a possible clue is that this type of signal was not seen in later tests on level 10 of the Lynx Mine.

The spike signals appear in records from all of the shotpoints, and many large signals arrive at times when the seismic wavefront has just entered the ore-zones. A problem with the spike signals is that they are very similar to man-made interference. Noise records (random records made in the absence of the seismic source) gathered in the mine also show this type of transient noise, and it was found that the arrival times of these spikes were random. To analyze the significance of the spike arrival times I constructed a histogram (Figure 4.11) that plots the number spikes in a 5 ms time interval at different delays from the blast. Two records contained anomalous amounts of noise (i.e. full of spikes) and were not included in this analysis.

Figure 4.11 Histogram of spike events, level 14, Lynx Mine. The high count 10 to 15 ms after the blast is consistent with the known location of the orebody.
Seismoelectric signals from the massive sulphides should typically arrive at times 5-20 ms after the blast because the seismic wave will be propagating with sufficient energy to excite RPE in the orebody at these times. There is a clear increase in the spike count in the histogram 10-15 ms after the blast (Figure 4.11). The 10-15 ms delay corresponds to a distance of 60-90 m from the shotpoints (5800 m/s acoustic velocity), which is consistent with the known orebody location relative to the shotpoints. To test the significance of the peaks in the histogram, a comparison between counts at negative time intervals, which are a measure of the background noise (man-made and natural), and peaks are made. A t-test performed on the count numbers indicates that the largest peak (10-15 ms) has a smaller than 1 in 10,000 chance of being due to random fluctuations in the background noise during the shooting. The chance that any of the count bins were to reach this count is less than 1 in 1000. Therefore, the peak is significant and cannot be attributed to a fluctuation in noise.

To locate a target properly, a number of shots are needed to triangulate the position of conversions. I was unable to do this with data from level 14 because the shotpoints are positioned in a line roughly parallel to the orebody (see Chapter 3, Tomographic Reconstruction). However, with the knowledge that the orebody runs roughly east to west, and is north of the shotpoints, the data can be plotted in a direction (the mapping of the arrival time to distance along a particular direction from each shotpoint) that will show trends in the data and indicate some of the orebody structure. The best direction for the level 14 data is due north because the earliest signals from the orebody will be due north of the shotpoint, later signals may come from other directions as the seismic wave spreads along the horizon. A plot of the spike data projected north (Figure 4.12) shows the correspondence between large spike signals and ore-zones running along 11200N. The signals that appear to come from distances further north are probably from regions higher or lower in elevation, or east and west of the easting shown. The purpose of the diagram is to show the significant increase in signals as the seismic wave strikes the ore-zones. An interpretation of the diagram is that there appears to be an
anomaly between 11150 and 11300N, 2100E to 2600E, and at an elevation similar to level 14 (10650). Also, there is some evidence of a smaller anomaly striking along 11025N near 2500E to perhaps 2800E. The lack of signals from shots west of 2100E is probably due to the small number of shots fired at these locations; therefore, little can be said about extent of the orebody west of 2100E.

Figure 4.12 Northward projection of arrival data, level 14 Lynx Mine. This plot highlights the sharp increase in signals once the seismic wave enters the ore-zones, which strike east-west on 11150 and 11200 N.

Conclusions

The sphalerite deposits at the Lynx Mine (level 14) are able to produce RPE. I was not able to accurately delineate the orebody due to the unfavorable shot-line geometry and a lack of data, but I could deduce the approximate distance of the orebody and, with some knowledge about the ore-horizon, gather information about the lateral extent of the orebody.
The significance of the different species of seismoelectric signal is ambiguous. However, the spike type of signals have the characteristics of RPE, and they appear to eminate from the orebody.

No clear difference was found in the response from the different charge sizes used, 0.22 or 0.45 kg. The large number of events 10-15 ms after the blast in the histogram indicates that the range of detection is at least 60-90 m. A new method of obtaining a time-break, the fiber optic trigger, was successfully implemented. No blast EM was observed in these experiments. I attribute this to the use of the fiber optic trigger and safety fuse detonators. The new magnetic sensors (UBC V) proved to be better at detecting small signals that were otherwise missed by the older generation of sensors.

4.5 Lynx Mine II

The principal aim was to produce a tomographic image of the orebody, and to experiment further in applying RPE for exploration.

**Experiment Details**

The orebody in this area belongs to the G-zone, and is near the apex of the anticline structure of the Lynx-Myra-Price Horizon (see Lynx I for some geological details, or Pearson, 1993). Work on level 10 was carried out on 19 to 21 July 1992 in a region near 5200E and 11500N in passages 10-581Dr, 10-501 Dr, 10-501X-C, and 10-531 Dr (Figure 4.13), during afternoon and night when the mine was quiescent. R.D. Russell, M. Maxwell, and K. Butler assisted me in these tests.

The shotpoints were arranged so that we would be able to triangulate the position of regions that consistently produced signals and produce an image of the ore-zone. Approximately 50 shots were fired in this survey from 18 locations using a 0.45 kg charge of pentolite explosive in a 8 ft drill hole for each shot. All of the shots were fired at least once with a fiber optic time break. To speed the shooting process the remote geophone
Figure 4.13  A map of experiments on level 10, Lynx Mine. Map co-ordinates are in feet.
was used to trigger the data acquisition system for many of the later shots, and the fiber optic was not used. This scheme worked well because of the excellent reproduction of the seismic wave obtained by using pentolite in a borehole.

The type of sensors used and the arrangement was identical to the July tests on level 14 (Lynx Mine I). Three UBC V magnetic sensors, two horizontal dipoles, and a vertical parallel plate dipole were placed at 5450 E and 11600N into blind cross-cut. A geophone was placed amongst the sensors for triggering and P-wave velocity measurement. P-wave velocities were calculated to be approximately 5500 m/s.

Discussion of Results

Analysis of the data from level 10 was much simpler than level 14 because there was only one variety of signal, the spike. An example of a typical shot record is shown in Figure 4.14. Additionally, we received twice as many signals per shot. A histogram of the number of spikes vs. delay (Figure 4.15) shows that most of the signals occur 5 to 10 ms after the blast, or approximately 30 to 65 meters from the shotpoints. It is very unlikely that this peak is due to count fluctuations (less than 1 in 8000). Reproducibility of data was also much better; for example, of four shots fired at SP6 three contained a signal at 10.6 ms.

The quality of data and the degree of freedom in shot placement allowed us to invert the data to produce a pseudo-tomographic image using the technique discussed in Chapter 3. Because all of the shots lie on a plane (level 10) anomalies above the shotpoints will have mirror images below the shotpoint plane; therefore we cannot ascertain whether an anomaly is above or below level 10 without using regional geology or constraints from drill data. A small number of parallel slices (tomograms) were generated and stacked to produce the final image because of uncertainty in source elevation and position due to the dipping nature of the orebody. This produces a smoother and more diffuse picture, but it suppresses the production of artifacts due to projecting the data.
Figure 4.14 An example of a record acquired from level 10, Lynx Mine. The topmost trace is from the fiber-optic blast sense circuit, and the fifth trace is from a geophone at the sensor site. Both electric and magnetic field antennas show RPE-like signals from 3 to 10 ms. The use of a fiber-optic blast sense and fuse blasting caps resulted in a clear recording of the onset of signal activity.

onto the wrong plane (a vertical distribution of shots would resolve this problem, see Chapter 3). Figure 4.16 is a plan view of a stack of 10 horizontal slices between 0 and 40 meters above level 10. There is clearly an anomaly striking east-west along 11430N. The anomaly shape and position is very close to the expected position of the ore-zones covered by the shot pattern. Extrapolating the available drill data puts the expected ore-zones along 11400N rather than along 11430, but this may be due to the dip of the
orebody (towards the north). Also, note that there are two weak anomalies, one along 11360 (5100E to 5100E) and another at about 11550N and 5150E. The first might be associated with the unmined ore horizon, and the second to pillars left from previous mining activity.

Conclusions

Because of the shotpoint positioning and good signal production on level 10 we were able to produce a fair image of the ore-zones in the area covered in the experiment. The principal anomaly in the tomogram is an elongated shape with the longitudinal axis striking along 11430 and extending between 5000E to 5200E. This image agrees rather well with the approximate position and shape of the orebody inferred from drill data.
Figure 4.16 An application of the tomographic reconstruction method, level 10, Lynx Mine. The data-set was approximately 100 events picked from 40 shot records. An outline of the approximate shape and location of the orebody is superimposed upon the image. The question marks indicate a lack of drill core data.
4.6 Lynx Mine III

The 1994 field trial at level 10 tested the concept of injecting a DC bias current into the orebody in the hope that it might produce more signals, or alter the response sufficiently to observe the role that tellurics play in RPE. Another important aspect of these tests was to use a very high bandwidth digitizer with wide-band sensors to learn more about RPE. The first objective was not achieved, but the second was spectacularly successful.

Experiment Details

Approximately 30 shots were fired between 22 and 24 November, 1994. A few early shots were fired with 2-3 small-diameter sticks (about 1" in diameter), but these shots lacked seismic energy, and did not stimulate signals from the orebody. Most shots were fired with 16 oz primer charges, which were squeezed with difficulty into the 2" holes we were using, but performed very well. Only two areas were used for firing. Firstly, we fired 17 shots in locations SP1 to SP4 (Figure 4.13), then we switched to locations SP10 and SP11 for shots 18 to 24, and the last two data collection shots (numbers 25 and 26) were back in the area SP1 to SP4.

Electric current was injected, via a custom borehole probe, 100 ft into a pre-existing borehole at 52E50 and 112N00. The current was approximately 5 to 10 mA at about 160 V. Originally we had planned to use a Phoenix IP transmitter, but resorted to wiring spare 12V cells and the batteries from an electric locomotive in series when it was determined that the IP transmitter produced large amounts of EM noise. Tests with the current alternately off and on were continued until shot 15. No current was injected into the orebody after shot number 15.

Our new (Gage) and old (RCE) digitizer systems were compared by using both simultaneously, although, each system had independent antennas. The Gage digitizer requires a standard computer system to act as a host because of power, bus, and monitor
requirements. A desktop (286 AT style) computer with a colour monitor was used as a host. The desktop computer was powered with a sinewave inverter (12 VDC to 120 VAC) to provide minimal electrical interference. The monitor was turned off during shooting to eliminate the possibility of line and video frequencies appearing on the sensors. The Gage digitizer was operated with two channels digitizing analog signals at a rate of 10 Msamples/s.

Both magnetic and an electric field antenna were used with each digitizer system. All of the EM sensors were placed in the area at 5200E and 11200N (Figure 4.13). The low bandwidth system (RCE digitizer) used two dipoles and three magnetic antennas (UBC V) with an operating bandwidth of 1 to 30 kHz (set by the AM502 amplifiers). In addition, the low bandwidth system received signals from a standard exploration geophone sited with the sensors, and from a geophone element mounted within the current injection probe. The Gage system used a parallel plate dipole with a high bandwidth preamplifier (0.1 to 5000 kHz) for measuring electric fields, and a UBC IV magnetic sensor (1.5 to 3000 kHz). Tektronix AM502 post-amplification was not used on these sensors because these amplifiers are limited to a bandwidth of 1 MHz. Coaxial cable was used for the high bandwidth sensors with no evidence of cross-talk or interference. The coaxial cable was constructed with both foil and braid shielding, and a 50 Ω impedance. Both cables were terminated by a 50 Ω terminator at the digitizer to reduce reflections.

Discussion of Results

The results from the current injection are inconclusive as there were no clear improvements in reproducibility or signal quality traceable to the current injection. Changing the current from shot-to-shot did not appear to directly alter the seismoelectric response. However, data from shots after the current injection tests were not as well reproduced. This may be due to the location of shots (SP10 and 11 mostly), or simply by chance. I feel that the current injection did not have much effect because the small
amounts of current injected could not significantly alter the conditions near the ore-
zone. Larger currents at the ore-brearing areas are needed to test the idea properly.

The Gage digitizer worked extremely well. We were able to see the EM signals more
clearly than in previous experiments; the best records were from the electric field
(parallel plate dipole) channel of the Gage digitizer. The full amplitude of the signals are
preserved by the greater bandwidth of the Gage digitizer (5 MHz vs. 60 kHz), so, rather
than just capturing only the very largest signals (usually one or two per record) we were
able to observe many more; for instance, most large transients are accompanied by many
smaller transients.

We have confirmed the very high frequency nature of the EM signals. Figure 4.17
displays two examples of the signals we were recording. The top example is a short
pulse with some oscillatory components (the bumps on top of the pulse), and the
example below is primarily oscillatory in nature. Both last for less than 5 μs, with rise-
times of less than 1 μs. The oscillatory signal appeared most often. One of the most
interesting features is that the frequency of the oscillations (approx. 1.3 MHz) was quite
reproducible from shot-to-shot, and within each record. An example of this consistency
can be seen in a spectrogram (Figure 4.18) of the electric field from a shot at SP4. This
behavior is consistent with a claim by Russian researchers that each type of ore/mineral
has distinctive spectral peaks (Sobolev et al., 1986). The mineral assemblages that make
up the ore in the level 10 area are fairly homogeneous, hence, the spectral response is
not expected to vary significantly. Unfortunately, they do not mention what the spectral
peaks are for different minerals.

The high-bandwidth tests provided have provide further insights on the phenomenon
of exhausting the orebody, and the problem of reproducing data in these tests. Both
aspects have been mentioned by our Russian forerunners (Sobolev et al. 1982). What I
mean by exhausting the orebody is that the orebody appears to produce fewer signals as
shooting progresses. For example, when we last conducted tests on level 10 (in 1992) we
Figure 4.17 Two examples of the type of pulse recorded by the high bandwidth system, level 10, Lynx Mine. Both pulses are short in duration, less than 5 μs. A common characteristic of pulses from the level 10 data-set is the consistency of energy in the 1.2-1.5 MHz band, which is evident in the oscillations in both pulses.
Figure 4.18
A spectrogram of the electric field disturbances created by a 0.5 kg shot, level 10, Lynx Mine. The spectrogram shows a peak in the 1.1 to 1.5 MHz bands. This signal came from a shotpoint in the same area that data for Figure 4.14 was collected.
found that after a particularly long session of shooting the ore-body was no longer responding - that is, we could not see any recognizable signals from the last eight shots. Again we have observed this phenomenon. Figure 4.19 illustrates this point. Note that of the seven shots that were sequentially fired from locations SP10 and 11 (which are less than 10 meters apart) there are very few remarkable high frequency responses from shots 21 and 22 (ignore the low frequency roll on shot 22). This behavior was also seen on a series of shots at locations SP1 to 4. It appears that the response of the orebody can be exhausted after several shots. After observing the gradual decrease in activity from our work in the area of SP1 to 4 (shots 4 to 15) we decided to immediately try shooting at

![Figure 4.19](image-url) Demonstration of an orebody tiring out, level 10, Lynx Mine. These are recordings of the horizontal E-field after a 0.5 kg shot. Shotpoints 10 and 11 are approximately 10 meters apart. After five shots in this area the orebody ceased to produce significant high frequency EM signals. However, the orebody responded to our shots after we returned to this area several hours later.
locations SP10 and 11 to see if we had exhausted the orebody, or whether it was no longer responding to shots from a particular direction. The activity on the earlier traces in Figure 4.19 demonstrate that it was the latter. Furthermore, the orebody appeared to recover after a short period of time. After shot 22 we exited the mine for a few hours and then returned to shoot both areas again. Both areas responded anew, with large numbers of signals.

![Figure 4.20](image)

**Figure 4.20** Four records of the electric field after a 0.5 kg shot at SP1. These shots were fired under nearly identical conditions. However, the traces appear to be dissimilar.

Perhaps it is only the large amplitude signals that disappear; these are the easiest to see on the voltage vs. time traces, but exceptional circumstances within the orebody may be needed to deliver these large signals. I can envision circumstances where this behavior might occur, such as the opening of cracks or joints by the action of the seismic wave.
Figure 4.21 Spectrograms of repeat shots at SP1. In contrast to Figure 4.20, the records are similar in appearance.
Chapter 4: Field Trials

The spectrograms, which are better at resolving small amplitude signals, appear to show less activity too, but the smaller signals are much more persistent. Therefore, it appears that the exhaustion of the orebody is principally a property of the large amplitude signals. This effect is less pronounced for the (more numerous) small amplitude signals.

The other, and possibly related, problem is that of reproducing our data. In the past, we have rarely been able produce identical copies of our data by shooting another shot in the same location. In general, the signals are fairly close in time of arrival, but in amplitude and shape they are very different. This property undermines the credibility of the seismoelectric method for sulphides. For an example look at Figure 4.20, which is composed of traces from shots at position SP1 over the two day period. The traces in Figure 4.20 appear to have little in common because the largest signals are scattered about in arrival time. However, the spectrograms of these traces (Figure 4.21) show a different story, the four panels are similar in appearance. The principal area that the spectrograms are different are in the location (in time) of large amplitude signals (look for the red-blue vertical lines), which are the only basis for comparisons in the voltage vs. time traces. The apparently fickle nature of the largest signals contrasts with the persistence of the more numerous smaller signals, especially those in the 1.1 to 1.4 MHz band. It seems that in the past we have been using the wrong yardstick for comparing records for reproducibility. The spectrogram provides a better measure to compare records.

Conclusions

The problem of reliably producing signals is better if the high amplitude signals are not given great importance. Analysis of the data with spectrograms shows that the lower amplitude signals are reproducible. The largest signals are either changed by the process of RPE, or require a lengthy time to recover.
The phenomenon of orebody exhaustion was demonstrated. However, the apparent effects of exhaustion of the orebody can be countered by alternating shot location or waiting for the orebody to recover. This indicates that angle of incidence and time after the last excitation are important in the RPE process.

Gathering higher frequency components beyond 100 kHz is a definite improvement. The benefits include better signal discrimination, possible identification of ore-type, and greater reliability in obtaining signals. With low noise, wide-band sensors I was able to produce a response from greater portions of orebody; instead of solitary spikes, periods of activity coincident with the passage of the seismic wave through the ore-zones were observed. The most useful sensor for broadband use was the parallel plate dipole. It gathered broad-band (<5 MHz) signals with good signal-to-noise, and proved to be convenient to use.

There is no firm evidence that injecting current into the orebody improves the seismoelectric method. I feel that idea has merit, but the amount of current injected was too small, and the probe too far from the orebody, to alter the conditions near the orebody sufficiently.

4.7 Century

In early 1992 the U.B.C. Geophysical Instrumentation Group was invited by C.R.A. to test a RPE-based exploration method upon a recently discovered sulphide orebody. The orebody had eluded detection by gravity, magnetic and TEM surveys (Thomas et al., 1992) because it is primarily composed of sphalerite, which has no distinctive geophysical trait. This experiment was our first attempt to perform RPE measurements on the surface. Previous trials were loosely based upon the Soviet style of underground work. The aim of the field trial was to detect the presence of a large deposit of sphalerite from surface measurements. If the orebody proved to be detectable then an attempt to image or delineate it was to be made.
Experiment Details

The Century deposit is a large (116 Mt) sulphide orebody located approximately 250 km north of Mt. Isa in Queensland, Australia. It consists of a smaller shallow southern block, which subcrops in the southwestern margin of the orebody, and a larger, deeper, northern block completely concealed beneath Cambrian limestone and recent alluvium (Thomas et al., 1992). Sulphide mineralization is hosted in dolomitic siltstones and carbonaceous shales in a sequence about 40 meters thick, and consists of four principal laterally continuous sub-divisions. Footwall rocks consist of shales and siltstones, and hanging wall rocks are dolomitic siltstones and shales. Most of the mineralization occurs as stratabound, banded sphalerite, galena, and pyrite within black carbonaceous shales. Typical grades are 10.3% zinc, 1.5% lead, and 35 g/tonne silver. The orebody did not produce an anomalous response to gravity, TEM, and magnetic surveys, and drilling was principally based upon geochemical anomalies found on the surface near where the orebody outcrops (Thompson et al., 1992).

The field trial was conducted during the period of September 14 to 23, 1992, by R.D. Russell, M. Maxwell, K.E. Butler, and myself. Experiments were performed in two areas over the southern portion of the orebody where it is relatively shallow. The first area was located south of an exploration shaft. The shotpoints formed a T shape with the arms extending east, west and south with the exploration shaft on the north-south line. The second area had a cross-shaped survey line that was centered on discovery hill. The two locations are referred to as the T and X surveys.

Shotholes were drilled to a depth of 6 meters into bedrock to give better seismic coupling. Two types of explosive were used, Anzomex (a pentolite explosive) and Powergel (an ammonium nitrate/fuel oil mix). Typically, 0.7 kg of Anzomex was used, with some charges up to 2 kg in size. Powergel was tested in sizes ranging from 1 to 6 kg with 0.5 kg of the pentolite explosive to ensure full detonation. Near the end of field trial some explosives were put into deeper drill-holes (up 60 m) near the X-survey in an
effort to get closer to the orebody. A total of 97 shots were recorded.

Two portable computers, each with a RCE digitizer, were available for this experiment. In general, one was used with one or two antennas to obtain a 500 or 250 kHz Nyquist frequency. The other computer was used in the same manner as in previous underground experiments: 8 channels with a Nyquist of 62 kHz with six channels used for EM signals and the other two for triggering and a geophone. Dipoles, long wire antennas and the ferrite-core magnetic sensors were used. In addition, a smaller version of the UBC V type magnetic antenna was constructed beforehand for use in boreholes. Most of the sensors were deployed within 50 meters of the shotpoint. A remote (<200 m away) dipole was often deployed in an effort to discriminate between locally generated signals and spherics.

The T site was surveyed first because of the proximity to the surface of high concentration ore. Several deep boreholes in the area allowed the lowering of the sensors to orebody depths and vertical component measurement of electric fields via the long wire antenna. Both high frequency (1 kHz to 30 kHz) and low frequency (10 Hz to 1 kHz, to look for possible piezoelectric responses) data were collected on the first two days. After this period only high frequency data was collected because of the absence of audio-frequency signals. Four days were spent in this area trying to obtain signals from a 40 to 120 m deep orebody.

The orebody is shallower at the X site, but the ore grades are poorer due to leaching. A greater variety of antennas and recording methods were used on this site because of the lack of success during the T survey. In general, there was a greater emphasis in obtaining signals with a bandwidth beyond 200 kHz. The spectral decomposition unit (Appendix B.9) was implemented in this survey to help discriminate between spherics and VLF noise, and broadband signals. Frequency bands from 50 kHz to 5 MHz were monitored by this unit, however, the use of 100 m of twisted pair cable between the dipole pre-amplifier and the instrument site limited the response to 50 kHz to 1.5 MHz.
Chapter 4: Field Trials

Discussion of Results

The results from both surveys were disappointing. No obvious seismoelectric signals appeared. All of the records are severely contaminated by a continuous background of VLF-type sources and spherics. The sinusoidal noise (VLF) could be removed with post-processing (narrow notch filters), but the spherics were similar in appearance to many of the signals seen in previous underground experiments and could not be removed easily.

Spherics are typically limited to 10 to 30 kHz frequencies (McCracken et al., 1984), which are determined by the atmospheric propagation modes. It was hoped that the higher frequency bands would show some seismoelectric responses. A relatively small number of broadband signals were detected in these higher bands on the spectral decomposition unit and on some high frequency electric field measurements, but we were not able to consistently reproduce any of these signals. The use of a remotely set dipole, separated from the other antennas by approximately 200 meters, did not provide discrimination between spherics and locally generated signals. In hindsight, the antenna should have been at least 0.5 to 1 km away to work effectively as a remote detector.

To answer the question of whether a response might have been induced a statistical analysis on the occurrence of spikes vs. time of blast was performed. The data from the X survey can be surmised as being of no significance: we failed to produce a recognizable response. Data from the T survey is more interesting. Figure 4.22 shows a histogram of spike arrival times from the T survey. Two peaks in the histogram are somewhat significant, one at 30 to 40 ms and the other at 90 to 100 ms. Data from previous seismic work gives an average P-wave velocity of about 2700 m/s for shallow depths; therefore, the anomalies come from apparent depths of approximately 100m and 300m. The first peak could conceivably come from the orebody below the experiments. However, it is unlikely that other peak comes from sources 300 m at depth because the seismic velocity used in this estimate is only valid for relatively shallow sources (less than 100 m), and is too low for a deeper source. A delay of 90 ms would should correspond to a depth of
more than 400 m. The seismic disturbance would be very weak at this depth, and very unlikely to induce RPE. If the late peak is from a seismoelectric conversion then it may be from a shallow source adjacent to the survey area.

Figure 4.22  Histogram of event arrival times, Century.

The mean noise level was estimated from a collection of noise records and pre-blast time intervals, giving a mean rate of 47 spikes per second (95% confidence interval on this estimate is 42-52). There is a less than 2% chance that a count fluctuation (due to the Poisson type process) would give a count above 28 at a particular time interval. The chance that such a peak would occur somewhere in the histogram is much higher at 25%. A figure of 25% would not be considered significant in many scientific tests, 1-5% is typically used. However, most of the bins are higher than the estimated noise level, which is an unlikely event in itself. In addition to forming an overall histogram, histograms of data from each arm of the T were compiled and compared. Each sub-group shared roughly the same pattern: two larger than normal peaks at about 10-30 ms and 90-100 ms. The sub-groups were too small in size to produce meaningful statistics, but the resemblance of each sub-group histogram to the survey histogram supports the
notion of a consistent response rather than a collection of random events.

If the estimate of the background noise is wrong (say it really was 51 instead of 47) then the peaks have no relevance and the histogram is well modeled by a random background of spherics. This is possible, but unlikely because a fair number of noise records were available to add to the pre-blast estimate. Of greater concern is the lack of identifiable signals; most of the signals look very similar to spherics recorded by Labson et al. (1985) and McCracken et al., (1984). This may be due to the relatively narrow bandwidth of the final processed records; most of the records were recorded with 1-30 kHz bandwidths, and then notch filtered at several frequencies in the 20-40 kHz range to remove dominating narrow band signals. That there is no apparent sub-species of signal attributable to these peaks in the histogram is perhaps not surprising given the similar bandwidths of spherics and the final records.

Conclusions

The Century orebody did not respond well to our attempts to induce RPE. However, the balance of arguments based upon statistical analysis of spike arrival time data from the T survey area indicates that a response was likely induced. I am unsure about the significance of the late arrival peak (90-100 ms) from this dataset. It is possible that the signals come from a relatively shallow area, but offset from the shotpoint by 100 to 150 m. There is also the possibility that this peak comes from a deeper target, or is simply not significant.

4.8 Other Field Tests

Immediately after the Century field trial another surface field trial was conducted at a recently discovered massive sulphide orebody (also in North Queensland, Australia) on a property owned by B.H.P. This test was very similar to the Century test in that the same instruments and procedures were used. Results from this trial were also similar: no obvious signals, but an indication that a significantly higher than expected count of
spikes occurred after the blast. Michael Maxwell (of the U.B.C Instrumentation Group) did the spike count and analysis of amplitude vs. time traces on this dataset. Spherics and VLF noise were again a hindrance.

A question that might be asked is do we see RPE signals everywhere? The previously described field trials have one thing in common: all conclude that RPE was induced. What about sites that do not contain sulphides? The Instrumentation Group at U.B.C. has conducted many other seismoelectric tests, searching for piezoelectric phenomena, in areas known to have very little sulphide mineralization. At two of these piezoelectric field trials, at the Paymaster site in Ontario (1991) and at Humboldt (1992) in Australia, 1-30 kHz records were made to check for RPE. No RPE-like signals were observed. I have not studied a barren site extensively, but there appears to be little indication of Nyquist-limited (typical recording bandwidths are in the 5 to 10 kHz range) pulse-like signals at the various piezoelectric field trials. In summary, there is no evidence of RPE at sites lacking sulphide minerals, which is consistent with the conclusions made by Sobolev et al. (1982a).
CHAPTER 5

A PHYSICAL MODEL FOR RPE

5.1 Introduction

Before I provide details about a particular mechanism to explain my observations, and those of Sobolev et al., I will discuss the important attributes of RPE so that it will be clear why particular candidate mechanisms are rejected. A number of mechanisms were proposed to me, and I searched extensively for potential candidates. Most candidate mechanisms failed to deliver half of the known properties of RPE and were quickly dismissed. A few of the rejected mechanisms will be analyzed to illustrate the issues involved in explaining the processes of RPE.

The rest of the chapter describes a mechanism based upon the formation of cracks within sulphide materials. My model addresses the issues of mechanical-to-electrical conversion and the production of the distinctive EM emissions of RPE. Physical parameters of this crack-based model are estimated, and compared with field measurements when possible. At present there is no accepted mechanism for RPE, and what follows is largely speculative.

5.2 Physical Attributes of RPE and Sulphides

The results from my field trials and those of Sobolev's research group provide ample demonstration that sulphides are an important, if not exclusive, class of materials to host RPE. In addition, the seismic flux must be greater than a certain threshold to create the observed electromagnetic signals. The field studies show that this threshold is of the order of 200 kPa. Below this level of stress very few signals appear. Furthermore, greater stresses do not increase the amplitude of the signals, but instead produce a greater number of signals (Sobolev et al., 1982a).
So why are sulphides so special? This is an important question to answer if insight into the process of RPE is to be gained. One distinctive property of sulphide minerals is that they are fairly reactive chemically compared to other rock types, and tend to oxidize or reduce in the presence of air and water (Sato and Mooney, 1960). These oxidation/reduction potentials will often produce telluric currents, which can be used to detect the nearby presence of an orebody (Telford et al., 1986). Another characteristic property of many sulphides is that electronic conduction is the dominant current flow mechanism. Most pure sulphide materials are semi-conductors, but with the addition of impurities conductivity increases greatly (Kittel, 1986). In general, host rocks consist of a silicate matrix, which is non-conducting, and the currents are carried by electrolytic ions in the pore fluids. This electronic conduction trait is exploited in the Induced Polarization method of exploration (Telford et al., 1986). Acoustic and mechanical properties of sulphide ores are not remarkably different from many other rocks (Carmichael, 1989), but I have noticed (from my laboratory tests) that sulphide samples are prone to fracture and spalling when uniaxial stresses greater than (approx.) 2 MPa are applied. Other than some shale samples, the host rocks (mostly igneous or metamorphic) were much stronger. All or some of these properties may be important, and there may be others; I have mentioned the properties that I feel are important.

Sobolev et al. (1982a) claim to have measured ultrasonic vibrations, coincident with the EM radiation, within the orebody under test. The presence of ultrasound is a key indication of micro-cracking (Yamada et al., 1989). Further evidence for the importance of cracks is provided by the claim by Sobolev and Demin (private communication with R. D. Russell and M. Maxwell on a visit to Moscow, 1991) that they can reproduce the signals by fracturing a sample of sulphide ore on a knife-edge whilst passing current through the sample. My attempts at reproducing the electromagnetic signals by simply stressing the sample uni-axially were unsuccessful; the Russians also experienced this problem. This lack of success in simple laboratory tests indicates that the bulk properties of the material are not the only factor in producing RPE, the surrounding environment
and structure of the material must also be considered.

The electromagnetic responses from RPE range across the full spectrum, from VLF to HF, light and X-rays (Sobolev et al., 1982a; Sobolev et al., 1984b). In the VLF to RF portion of the spectrum the signal is in the form of a brief pulse, several microseconds in duration. This pulse may be oscillatory, with the frequency of oscillation characteristic of the type of mineral responsible for producing the electromagnetic radiation (Sobolev et al., 1982a, Sobolev et al., 1986). X-rays often result from collisions between high energy electrons and atomic nuclei (Dyson, 1973). Typical energies for X-ray production lie in the range 10 keV to 100 keV (Potts, 1993); thus, electric fields ranging from 10 to 1000 MV/m are needed (the region of acceleration must be fairly small otherwise the electrons will lose too much energy via ionizing collisions). Note that the rapid discharge of energy in the host medium by these electric fields can vaporize parts of the medium and cause acoustic disturbances. Consequently, a seismogenic process that can create enormous electric fields or potentials would explain a number of important properties of RPE.

Another significant aspect of RPE is that the electromagnetic signals arising from identical tests are not the same, and often bear no resemblance to each other (Kepic et al., 1995; Sobolev et al., 1982). Spectrograms (of the EM signals) from the Lynx Mine often displayed remarkable similarity in identical tests, but in general there is a uniqueness to each test that is inherently part of the process. It was found in the Lynx mine tests that the largest signals are the least reproducible.

In summary, I am looking for a phenomenon that can be initiated by a strong seismic disturbance, produces a large EM pulse, and can be produced in natural sulphide materials. In addition, light, X-rays, and micro-cracking may be a by-product. I expect that the process alters its environment sufficiently to inhibit the success of repeating any experiments.
Potential Mechanisms

Several possible explanations of RPE will be examined in this section. Ultimately, all but the last will be rejected because they fail to provide an adequate description of the observed characteristics of RPE.

An early suggestion put to me was the possibility of piezoelectric charging of the sulphide material followed by a very quick depolarization. This is in essence the hypothesis that Sobolev et al. (1982a) initially put forward. Two questions arise: what is the mechanism of depolarization, and are the resulting electric fields sufficient to account for associated phenomena such as X-rays. Sobolev et al. (1982a) proposed that natural thyristors, transistors and other non-linear circuit elements are produced by the impurities and crystal grain boundaries within the sulphide minerals; these will breakdown once the potential is sufficient, and allow free electrons within the semiconductor/sulphides to rapidly neutralize the polarization. Firstly, I find it difficult to believe that such an unlikely natural circuit combination should be so common throughout the orefields that I have investigated. Also, electrical circuits, natural or otherwise, tend to reproduce results when identically tested; RPE does not share this property. As to the second question, X-ray production is unlikely because the piezoelectric coefficients of natural materials are too small (of the order of pC/N) to produce sufficient polarization (under typical field conditions) for X-ray production.

A simpler mechanism would have a crack form in a charged piezoelectric material. The electrical break-down of the air gap in the crack would provide the impulsive EM signals. This mechanism can produce light, X-rays and ultrasound (Nitsan, 1977; Cress et al., 1987; Yamada et al., 1989, Enimoto and Hashimoto, 1990). It is not a particularly exotic effect; many hand-held gas lighters are based upon the electrical gas-discharge from a stressed piezoelectric. A significant drawback to a piezoelectric based hypotheses is the requirement of a piezoelectric material. The only common piezoelectric sulphide is sphalerite, but many other non-piezoelectric sulphides purportedly produce RPE.
Therefore, significant quantities of sphalerite, quartz, or the much less abundant sulphides of cadmium or arsenic would be necessary for these processes to work.

Another phenomenon that I have considered is the acoustoelectric effect found in semiconductive piezoelectric materials (Hutson and White, 1962). Hutson and White (1962) found that if a DC bias current were put across a sample of conductive piezoelectric material (CdS typically) then acoustic waves could be amplified as they propagated through the sample. Amplification only occurred once the current exceeded a critical value. The condition for amplification is that the drift velocity of the charge carriers must be greater than the acoustic velocity (White, 1962). The DC bias current sets a mean velocity to the charge carriers (the drift velocity). If an electromotive force causes the charge carriers to bunch or disperse locally then this perturbation moves at the drift velocity, \( v_d \) (Kittel, 1986),

\[
v_d = \mu E
\]  

where \( \mu \) is the mobility of the charge carriers (\( \mu = 500 \text{ cm}^2 \text{ V}^{-1} \text{ s}^{-1} \) is a typical value for many pure semiconductors) and \( E \) is the electric field. An acoustic wave in a conductive piezoelectric creates a spatial perturbation of the charge carriers, which drift via the bias current. If the drift velocity is sufficient then positive feedback amplifies the acoustic wave. Normally the space charge acts to further stiffen the solid (due to energy stored in the electric polarization), but reinforcement of the acoustic motion occurs when the space charge of one acoustic phase is transferred to an area of opposite phase. If one of these materials is stressed greatly then the resulting piezoelectric fields may bootstrap the process and the natural (phonon) modes of the crystal/sample are amplified until saturation (Wang, 1965; Pustovoit, 1969; Pozhela, 1981).

The acoustoelectric effect generates acoustic shock waves, and very high electric fields (Ridley and Wilkinson, 1969). In addition, particle emission, light and mechanical
failure have been observed (Ridley and Wilkinson, 1969). In many ways the acoustoelectric effect meets the requirements for RPE, but it fails in some critical areas. Because it relies on piezoelectric materials it does not comfortably explain why orebodies containing predominantly non-piezoelectric sulphides respond. Sphalerite, quartz or a less abundant conductive piezoelectric needs to be present. Also, it is debatable whether the seismic wave can generate enough stress many tens of meters from the shotpoint to initiate the process. Given the piezoelectric coefficient $\alpha$ and seismic stress $T$ we can calculate the resulting effective electric field, and substitute this into equation 5.1 to give the drift velocity ($\varepsilon$ is the dielectric permittivity of the material):

$$v_d = \frac{\mu \alpha T}{\varepsilon}$$

Substituting high values of stress (1 MPa), piezoelectric coefficient (2 pC/N), and mobility (400 cm$^2$ V$^{-1}$ s$^{-1}$) gives $v_d \approx 1300$ ms$^{-1}$, which is in the right ballpark (P-wave velocity), but all of these numbers are optimistically high. The most optimistic figure is the mobility, which is for perfect and pure crystals; impurities, dislocations and other defects from other minerals and metals introduce further scattering, and will reduce the mobility by at least one to two orders of magnitude (Kittel, 1986).

The above-mentioned models for the RPE process can satisfy many of the requirements for RPE, but do not fulfill all the requirements unless complex and questionable adjustments are made. My preferred model is outwardly the simplest: all that is required is that a crack open in the sulphide material. Cracks are a viable explanation because very high transfers of mechanical to electrical energy are possible via triboelectricity, and the air-gap/vacuum formed by the crack allows the entire electromagnetic spectrum of RPE observations to be explained in a fairly simple manner. Three physical effects are needed to produce RPE according to the crack model: microcracking, triboelectricity, and EM radiation from a gas discharge. The first two (mainly mechanical) effects will be considered forthwith, the latter is treated in section 5.5.
5.4 Triboelectricity and Crack Formation

Triboelectricity is the phenomenon of charge separation arising from the mechanical alteration of surfaces (Heinicke, 1984), and is the source of many naturally occurring static charges (Hays, 1991). Triboelectric effects are principally due to the creation of fresh surfaces (Figure 5.1). The fundamental process behind these effects is the transfer of charge across a boundary prior to separation; in essence a double electric layer is formed at the boundary (Heinicke, 1984). When the two new surfaces are created by splitting the material across the boundary some of the transferred charges are not returned, and the two surfaces become oppositely charged.

![Figure 5.1](image)

Figure 5.1 An illustration of the process of triboelectric charge generation. As two materials are pulled apart along a flaw or joint residual charges remain on the surfaces.

The stimulus for charge transfer can be provided in two ways: by differences in the bulk properties of the materials on either side of the boundary, or by electro-chemical means (Horn et al., 1993; Heinicke, 1984). If two materials possessing different work functions are put together then charge will be transferred across the boundary until the surface potentials are equal (Hays, 1991). The work function is defined as the potential energy to free an electron from the surface of the material, and is characteristic of the electronic properties of the material. When the two materials are separated some of the
transferred charge may not return because these charges are relatively immobile. Metals do not exhibit significant triboelectric effects because the charges are very mobile, but insulators and semi-conductors often produce notable triboelectric effects (Hays, 1991). Most triboelectric effects are due to this mechanism. Because sulphide ores tend to be rather heterogeneous in nature, with different sulphide minerals crystallizing amongst others, I expect that the likelihood for this type of triboelectric process to be fairly large. A problem with this mechanism is that the heterogeneity has to be fairly large in size (of the order of 10 cm) to produce the large EM fields observed in my field trials.

The second mechanism occurs without differences in the substrate materials. An electrical or chemical process directly provides the charge transfer mechanism across the boundary (Heinicke, 1984; Horn et al., 1993). A situation where this type of process may be important in the sulphide orebody is the passage of telluric current across the boundary of a large flaw or joint, where there is still solid contact across the joint, but some micro-cracking has already occurred. Electric potentials across the boundary causes ions/electrons to migrate to the oppositely charged surface and accumulate. If the surfaces are pulled apart then some of the displaced charges are too immobile to neutralize the opposite charges before a gap opens. This mechanism provides a good explanation for the dominance of a single polarity in the EM signals recorded in our (and Sobolev’s group) field experiments because the dipole orientation is controlled by the direction of the telluric current, which is fairly uniform over distances of several meters to tens of meters (depending on the size of the orebody and the source of the tellurics). The orientation of compositional differences are unlikely to be as uniform.

The maximum amount of charge that can be left on the crack surfaces is limited by two physical constraints: the number of atomic sites for the charges and field emission. The latter effect normally dominates, limiting the occupancy to about 10% of the available sites (Hays, 1991). Field emission of electrons occurs when the electrostatic repulsion of the excess negative charges is sufficient to overcome the binding energy to the surface
Thus, electrons will fly off the surface if there are too many of them. The electric fields near the charges must be sufficiently great that an electron will gain more energy than the binding energy. This binding energy is equivalent to the work function. Therefore, for field emission to occur the electric field must be greater than the work function divided by the distance the electron must travel before escaping the surface forces (Hays, 1991). Note that this is an oversimplification. An exact calculation involves the phenomena of electron tunneling through the Fermi/work potential, and the role of the electric field is to lower the barrier potential and increase the probability of escape (Nasser, 1971); however, both methods give approximately the same answer.

The work function \( W \) for most metals and semi-metals is about 4 eV, and the distance the electron must travel \( d \) is approximately the thickness of one atomic layer or less (approx. range of surface forces), which is approximately 0.7 nm. These numbers give field emission at about 5 GV/m (Nasser, 1971, Hays, 1991). In practice, the surface roughness limits average fields to about 1 GV/m. Once the process of field emission begins electrons will leave the surface until the mutual repulsion of the remaining electrons is insufficient to overcome the work function. If the surface is relatively flat then Gauss's Law of electrostatics can be used to find the maximum surface charge density before field emission occurs.

\[
\sigma_{\text{max}} \leq \frac{W \varepsilon}{d}
\]

Using the previous values for \( W \) and \( d \), gives a limit of about 30 mC m\(^{-2}\). This number is approximately equal to the largest triboelectric surface charges reported (Hays, 1991; Horn and Smith, 1992). Hence, field emission limits triboelectric surface charge densities to less than 30 mC m\(^{-2}\). Large values of triboelectric surface charges are in the range of 1 to 10 mC m\(^{-2}\) (Hays, 1991). I expect that if a rather unique phenomena like RPE is based upon triboelectric effects then exceptionally large amounts of charge are produced, and that 1 to 10 mC m\(^{-2}\) is a good starting point for further investigation.
The residual surface charges play an important role in the evolution of the crack. Attractive electrostatic forces from the oppositely charged surfaces resist further opening of the crack gap (Hays, 1991; Horn and Smith, 1992). This electrostatic stress is equal to

\[ T = \frac{\sigma^2}{2\varepsilon_0} \]  

Substituting surface charge densities of 1 to 10 mC m\(^{-2}\) into 5.4 gives opening pressures of 55 to 5500 kPa. If the gap is to widen then tension greater than the stress calculated by equation 5.4 must be applied. An important implication of equation 5.4 is that the surface charge density that can appear on the fresh surface is restricted by the amount of external stress available to pull the crack open, otherwise, the crack will not open.

The seismic disturbance from a small explosive charge is capable of providing forces of up to 200 kPa tens of meters from the source (Chapter 2). A stress of 200 kPa limits the surface charge densities on the crack to less than 2 mC m\(^{-2}\). A plausible scenario is that the seismic disturbance creates a crack during rarefaction (rocks are much weaker under tension than under compression: Jaeger and Cook, 1976), and then proceeds to open the gap further by transferring the seismic strain from the tension to the gap. The gap is effectively a soft spring between two stiff springs; most of the strain is taken up by the softest spring when tension or compression is applied.

The preceding scenario has a glaring weakness: the total strain obtained from the seismic wave is delivered over a period of about 1 ms. This poses a considerable problem because most of the surface charge will be lost by 1 ms, unless the medium is very resistive. A consequence of the charge leakage is that the dipole moment formed by the charges and gap will not be sufficiently large to explain the EM fields from RPE (this is discussed in detail in the next section, 5.4). If the time constant of leakage is 100 µs then the gap will only be about 2 µm wide at the time of maximum dipole moment. This gap size is far too small to reproduce the measured EM fields in my field trials. In addition, only sub-keV energies are possible for electrons traversing the gap; the gap potential
must reach at least 10 kV for electrons to produce a measurable amount of X-rays (Potts, 1993).

My preferred model of the crack process is that the seismic disturbance provides the extra tension to rock that is already under localized extensive stress to form cracks, but the pre-existing stress provides most of the work necessary to obtain large surface charge densities, larger gaps, and relatively quick opening times. Most underground rock environments are under a number of inhomogenous stresses due to overburden forces and deformative forces (Herget, 1988; Sholz, 1990). Typical horizontal stress magnitudes in underground rock range from 1 to 20 MPa for rocks 10 to 500 m deep (Roberts, 1981; Herget, 1988; Sholz, 1990; private communication with T. Urbanic of Engineering Seismic Group in Kingston, Ont.). My conjecture is that the seismic wave generally needs to add an extra 5 to 10% to a pre-existing stress to consistently generate RPE. I have many records with signals that arrive more than 30 ms after the blast, corresponding to apparent distances from the shot of more than 170 m. It is very hard to explain the EM signal amplitudes if the seismic wave has to do all of the work; even at distances of 50 m it is a difficult problem. Another advantage of relying upon pre-existing stresses is that the available surface charge density is not greatly affected by the diminishing force of the seismic wave with distance from the shot. The effect of greater distances from the shot in this model is to reduce the likelihood of opening new cracks. Maximum surface charge density is governed by the amount of stress within the rock and equation 5.4, and is independent of seismic behavior if the pre-existing stress is much greater than seismic stresses.

I am assuming that extensional forces are responsible because the tensile strength of rocks is an order of magnitude lower than the compressive strength, and is typically 3 to 10 MPa for many underground rock types (Carmichael, 1989; Jaeger and Cook, 1976). In addition, tensile cracks provide a natural way to form an electric dipole from triboelectric effects. A potential disadvantage of relying upon pre-existing stresses is that
not all of the orebody may be under tensile stresses, therefore, not all of the orebody will respond. There is no proof from my field trials that every section of the orebody does respond. In fact, the field evidence might tend to show otherwise (Chapter 4, Mobrun Mine). This property would be a liability for RPE-based exploration. Nevertheless, I cannot produce a convincing model that does not involve pre-existing stresses. It appears that pre-existing stresses must be included in a crack-based explanation of RPE. However, there is an abundance of evidence that tensile cracks exist in most geological formations (Herget, 1988; Sholz, 1990). These cracks are often associated with shear faults, folds, and joints (Sholz, 1990). My picture of an orebody is a collection of blocks separated by joints and full of cracks oriented in many directions (Herget, 1988), some of these cracks are close to failure due to the local stresses in the rock mass.

Figure 5.2  An illustration of tensile strain relief by a crack. The crack is roughly ellipsoidal in cross-section, and the air-gap is formed by the relief of strains around the material surrounding the gap. The volume of material that gives up strain to the gap is approximated by a sphere enclosing the crack.
Two principal parameters determine most of the crack properties. These parameters are the pre-existing stress, $T$, and the final diameter of the failed region, $l$. The crack is modeled as an expanding disk that grows to a diameter $l$ and with a gap of $s$. Strains from the surrounding medium, to a distance $l/2$ from the crack centre, are taken up by the gap (Figure 5.2). The maximum gap size, $s$, is

$$
s = \frac{lT}{Y}
$$

(5.5)

where $Y$ is the Young's modulus, and $T$ is the tensile stress (Jaeger and Cook, 1976). A value of 50 GPa is a representative value of Young's modulus for many igneous rocks, such as andesite and rhyolite (Carmichael, 1989; Young's modulus will change at great depths, $>1000$ m, due the closing of cracks). To progress further with this model numbers for $T$ and $l$ must be found. My chosen numbers are, $T=5$ MPa and $l=0.3$ m. The former represents a median value of tensile strength for many underground rocks (Carmichael, 1989; Jaeger and Cook, 1976), and the latter is based upon the observation of cleavage patterns in the tunnels of mines that I have conducted field trials. A crack area of about 30 cm x 30 cm is fairly large. However, it may well represent a collection of smaller fractures that connect to form a rupture. Substitution of these numbers and the various rock parameters gives

$$
\begin{align*}
T &= 5 \text{ MPa} \\
l &= 0.3 \text{ m} \\
Y &= 50 \text{ GPa}
\end{align*}
\Rightarrow
\begin{align*}
s &= 30 \mu\text{m} \\
\sigma &= 10 \text{ mC m}^{-2} \\
\Delta t &= 30 \mu\text{s}
\end{align*}
$$

(5.6)

Where $s$ is the maximum gap size, $\sigma$ is the upper bound upon surface charge density, and $\Delta t$ is the time taken to relieve the strains around the crack. The time to form the crack is sufficiently short to preserve most of the original charge. I will assume that electrostatic forces provided a substantial amount of the adhesive forces supporting the flaw before rupture, and this will be reflected in the amount of charge that is left on the surfaces. Hence, a value of 10 mC m$^{-2}$ for the surface charge density will be used for calculating the magnitude of the EM pulse.
In my model of RPE, the role of the seismic wave is to push a rock that is already under about 5 MPa of tension to the point of rupture. Subsequent cracking relieves strains from regions up to 30 cm from the centre of the crack, thus producing a large crack with a gap of 30 μm in about 30 μs. The fresh surfaces have about 10 mC m⁻² of surface charge distributed over them, which will provide the source of the EM signals observed from RPE.

5.5 EM Fields from an Expanding Crack

I propose that the EM radiation seen in my field trials and those conducted by Sobolev’s group comes from a gas discharge. As the gap widens and fills with desorbed gases then the surface charges will recombine due to electrical breakdown of the gases under the high electric fields within the gap. This electrical model of the EM process gives an intense and very short pulse of current, and can account for the brief nature of the EM pulse. However, explaining the amplitudes of the EM fields measured in my field experiments is not trivial. This is not a problem limited to my crack model. All of the rejected mechanisms share this problem because the observed EM fields from RPE are large.

To estimate the EM fields the crack is modeled as a small electrical dipole in an isotropic and homogenous medium. The crack will be treated as a parallel plate capacitor with equal and opposite charges \( Q = \sigma A \), where \( A \) is the area of the plates. At the time of the gas discharge a large amount of current flows briefly. This current (\( \dot{P} \)) provides the strong EM fields seen in the field trials.

\[
\dot{P} = \frac{\Delta Q(s)}{\Delta t}
\]  

(5.7)

The solution to the problem of the EM fields from a small current dipole is well known (Jackson, 1975; Kraichmann, 1976), and for a given dipole moment the fields are
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\[ E_y(\omega) = \frac{j\omega \mu \hat{P} \cos \theta}{\gamma} \frac{1}{4\pi r^2} \left( 1 + \frac{1}{\gamma r} \right) \exp(-\gamma r) \]

\[ E_\theta(\omega) = \frac{j\omega \mu \hat{P} \sin \theta}{4\pi r} \left( 1 + \frac{1}{\gamma r} + \frac{1}{\gamma^2 r^2} \right) \exp(-\gamma r) \quad (5.8) \]

\[ B_\phi(\omega) = \frac{\gamma \mu \hat{P} \sin \theta}{4\pi r} \left( 1 + \frac{1}{\gamma r} \right) \exp(-\gamma r) \]

where \( \gamma \) is the propagation constant (complex wave-number),

\[ \gamma = jk \quad \text{and} \quad \gamma^2 = -\omega^2 \mu \varepsilon + \frac{j\omega \mu}{\rho} \quad (5.9) \]

and the co-ordinate system is shown in Figure 5.3. These equations are valid if the distance from the source \((r)\) is large compared with size of the dipole, which is the case in the field trials. Typical resistivities \((\rho)\) of igneous underground rocks are about 1000 \(\Omega\) m (Telford, 1988).

Figure 5.3  Co-ordinate system used in describing the electric dipole fields. The dipole consists of charges +Q and -Q centered at the origin, but separated by a distance of s. The measurement of the field is at co-ordinates \((\theta, \phi, r)\).
Most of my EM measurements contained frequencies from 1 to 50 kHz. At this range of frequencies conduction currents predominate (for typical underground resistivities and dielectric permittivities), and the equations in 5.7 simplify to a form known as the near-field approximation (Kraichmann, 1976).

\[
E_r = \frac{\rho \hat{P} \cos \theta}{4\pi r^3} \exp(\gamma r)
\]

\[
E_\theta = \frac{\rho \hat{P} \sin \theta}{4\pi r^3} \exp(\gamma r)
\]

\[
B_\phi = \frac{\mu \hat{P} \sin \theta}{4\pi r^2} \exp(\gamma r)
\]

and \( \gamma^2 = \frac{j\omega \mu}{\rho} \)

Note that if the argument of the exponential is small, as is assumed here, then the form of these equations is the same in both the time and frequency domain.

To estimate the EM fields from the gap a value for the dipole moment is needed. If I assume that the gas discharge occurs soon after the crack has fully formed, and that all of the surface charges recombine then the numbers from the crack model can be used (5.6). Thus, equation 5.7 and the numbers from 5.6 can be used to calculate the current dipole. However, the gas discharge occurs in nanoseconds (Nasser, 1971), a time interval that is too short for the detectors to measure the true peak amplitude. Hence, the peak magnitude of the dipole moment must be adjusted for the limited bandwidth of the recording system, and equation 5.7 becomes

\[
\hat{P} = \frac{\Delta (Q_s)}{\Delta t} 2\pi \Delta f \Delta t \equiv 2\pi \Delta f Q_s
\]  

(5.11)

The equations in 5.10 are not particularly useful in their present form because it is the magnitudes of the EM fields that I wish to compare to the results from my field trials. Therefore, I will assume that all orientations are possible, and integrate the square of the
field magnitude to obtain the RMS. values of $E$ and $B$

$$|E| = \frac{DQs\Delta f}{3r^3}$$  \hspace{1cm} (5.12)$$

$$|B| = \frac{\mu Qs\Delta f}{6r^2}$$

The exponential, or skin depth, term is omitted because it is close to unity at 1 to 50 kHz frequencies.

With equations 5.11 and 5.12 an estimate of the EM fields from the gas discharge current can be made. Substituting the following crack model

$$A = 350 \text{ cm}^2 \quad \sigma = 10 \text{ mC m}^{-2}$$

$$Q = 350 \text{ } \mu\text{C} \quad s = 30 \text{ } \mu\text{m}$$  \hspace{1cm} (5.13)

and field experiment parameters

$$\Delta f = 50 \text{ kHz} \quad \rho = 1000 \text{ } \Omega \text{ m}$$

$$r = 80 \text{ m} \quad \mu = 1.26 \times 10^{-6} \text{ H m}^{-1}$$  \hspace{1cm} (5.14)

gives these EM field magnitudes

$$|E| = 340 \text{ nV m}^{-1} \quad \text{and} \quad |B| = 17 \text{ fT}$$

Both of these numbers are four to five orders of magnitude too small to explain typical field measurements (5 mV/m and 1 nT).

It is tempting to discard the crack model because of these low EM fields. However, none of the previously discussed models will fit the data either because it is very hard to form a sufficiently large dipole moment from the natural processes within rocks. The problem of describing the EM fields may lie with the simple current dipole model of a gas discharge. To my knowledge, there is no accepted quantitative model of a gas discharge that explains the large amounts of RF emission that accompany a spark or coronal discharge.
A possible solution to this problem is to examine the contributions from the charges on a microscopic scale. I will use the expression for the magnetic field in 5.8 as an example. In the time domain the magnetic field is (Lorrain and Corson, 1988; Jackson, 1975)

\[ B_{\phi}(t) = \frac{\mu \sin \theta}{4\pi cr} \left( \dot{P} + \frac{c}{r} \dot{P} \right) \]  

(5.15)

The terms associated with the time delay and phase distortion are not significant for frequencies below a hundred kilohertz for the field conditions stipulated in 5.14, and are not included in 5.15. In addition, I will consider \( c \), the velocity of electromagnetic waves to be constant. In fact, at low frequencies this velocity is frequency dependent in conductive media, but the velocity change with frequency does not alter the basic arguments to be presented here. There are two sources of the magnetic field in equation 5.15: radiation (\( \dot{P} \)) and induction (\( \dot{P} \)) fields. The latter was used in 5.10 because the radiation contribution is much smaller at these frequencies. However, the use of these equations assumes that a response can be produced from a linear superposition of frequencies, which is reasonable if Maxwell’s equations hold. I find it instructive to look at the microscopic scale of electrodynamics because classical field theory (Maxwell’s equations included) does not always correctly describe the behavior of charged particles.

Firstly, consider a free electron near the surface of a negatively charged plate of a parallel plate capacitor. Each plate has a surface charge density of \( \sigma \), and the plates are separated by a gap of distance \( s \). The volume between the plates contains an electric field, \( E = \sigma/e \), which will accelerate the electron toward the other plate. Acceleration stops when the other plate is reached, and soon the electron stops too because of collisions with the molecules that make up the plate (Figure 5.4). For the moment, consider the EM fields generated by the electron as it traverses the gap between the plates. The average values for \( \dot{P} \) and \( \dot{P} \) are given by (where \( e \) and \( m \) is the charge and mass of the electron)

\[ \dot{P} = \frac{\sigma e^2}{m\epsilon} \]  

(5.16)
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\[ \dot{P} = \left( \frac{\sigma e^s}{2me} \right)^{1/2} \]  

(5.17)

For small gaps the radiation will dominate. Substituting the estimated values for \( \sigma, \varepsilon, \) and \( s \) (10 mC m\(^{-2}\), \( 8.8 \times 10^{-12} \) F m\(^{-1}\), and 30 \( \mu \)m) from the crack model into 5.16 and 5.17 shows that radiation is six orders of magnitude greater than the inductive fields.

So why do induction fields dominate the near-field? In general, when the electron smacks into the other plate it decelerates, and the field seen by an external sensor is an equal and opposite field to the acceleration phase (note that the fields close to the electron may change in movement due to local polarization from the surrounding dielectric). In low energy collisions the radiation is of the form of a broadband pulse,
the bandwidth of the radiation produced is the Fourier transform of the collision event (Jackson, 1975). If we were to try and measure the EM field with a low-bandwidth sensor we would see little of the radiation as there is no net acceleration over the relatively long time periods that the sensor integrates the fields. In this case the near-field or induction approximation holds.

If the electron is accelerated so that it reaches energies of more than 10 keV then classical field theory (i.e. Maxwell's equations) does not adequately describe the EM behavior when collisions with other particles are made (Eisberg and Resnick, 1974). In a high energy collision between a fast electron (speeds with an appreciable fraction of the speed of light) and an atomic nucleus the collision is so quick that, according to classical field theory, the bandwidth covers the EM spectrum up to gamma rays (Jackson, 1975; Eisberg and Resnick, 1974). A single gamma ray would have more energy than the incident particle in many cases. In this situation a problem arises because classical field theory does not adequately describe the characteristics of high energy EM waves or photons (e.g., X-rays and gamma rays). Photons have particle-like properties, such as momentum, which are not explained by Maxwell's equations. According to classical field theory the gamma ray may have any amplitude, but quantum processes stipulate there is a single energy for a gamma photon, and that waves consist of a number of photons.

When an electron collides with an atomic nucleus photons are emitted with an energy equal to, or less than, the energy lost by the electron in the collision (Figure 5.5). This process is called bremsstrahlung, or braking radiation (Jackson, 1975; Eisberg and Resnick, 1974). In most collisions a large number of soft photons of different energies are emitted and quantum theory agrees with classical results if the classical spectrum is limited to frequencies less than the maximum allowable photon energy (Jackson, 1975). However, in a small number of collisions a single photon will take most of the energy, and an X-ray results. In this situation the momentum absorbed by the photon must be
taken into account, and the contribution to the lower frequency part of the spectrum (soft photons) is altered (Jackson, 1975). A simple picture of the process is that the electron is decelerated, but a photon rather than an opposing (broadband) field is produced. Therefore, the use of the classical field equations is not applicable in some cases even though the particle changes velocity.

Figure 5.5 An illustration of the process of bremsstrahlung, or the inverse photoelectric effect. Very fast electrons often pass through the cloud of electrons surrounding an atomic nucleus and interact with the nucleus directly. Electrostatic forces between the electron and nucleus deflect the electron, but barely alter the motion of the nucleus because of the latter's great mass. The deflection of the electron produces an EM disturbance, a photon, or a number of photons. Occasionally, the photon carries most of the momentum from the collision, producing an X-ray or gamma ray.

The implication of bremsstrahlung is that if the electrons traversing the gap reach high energies, then some will not produce (broadband) EM fields when they slow down in the solid medium. Consequently, a net amount of radiation from the acceleration of these electrons can be seen by our detectors. Because the radiation fields are so much greater than the induction fields in the small gaps of the crack model only a small percentage of electrons need to participate in bremsstrahlung for radiation to dominate. The fraction of electrons that produce X-rays is
\[ \eta = 1.1 \times 10^{-9} Z V = \beta Z \frac{\sigma s}{\varepsilon} \] (5.18)

where \( Z \) is the atomic number of the target material and \( V \) is the energy of the electron in keV. This expression was derived empirically by Compton and Allison (Massey and Burhop, 1952) to express the efficiency of X-ray production verses atomic number and accelerating voltage. The second expression in 5.18 uses the crack model parameters (5.13), and assigns a symbol for the constant (\( \beta = 1.1 \times 10^{-9} \)).

To estimate the effects of bremsstrahlung we need expressions for the radiation fields in the time domain, which neglecting conduction losses are (Lorain and Corson, 1988)

\[ E_\theta = \frac{\mu \, \ddot{P} \sin \theta}{4 \pi r} \quad \text{and} \quad B_\theta = \frac{E}{c} \] (5.19)

Note that \( c \) (which is complex in the frequency domain) is frequency dependent for frequencies where conduction currents predominate. Thus, at low frequencies (<100 kHz) the magnetic field signature will be distorted and will not resemble the original pulse (although it will be a pulse). However, the electric fields are proportional to the acceleration, barring frequency dependent attenuation due to conductive losses (which are not significant until 1 MHz for typical underground conditions and distances in my field trials) and other dispersive effects (Fuller and Waite, 1976). Integrating the magnitude squared of the electric field in 5.19 over all possible dipole orientations gives the following expression for the expected magnitude of the electric field

\[ |E| = \frac{\mu \, \ddot{P}}{6 \pi r} \] (5.20)

Substituting equations 5.16, 5.18, and adjusting for the limited bandwidth of the EM sensors (e.g. see 5.11) gives an expression that can be used to calculate the expected magnitude of the electric field from a gas discharge between the crack surfaces,
as measured by low-bandwidth detectors. Substituting in the numbers from 5.13 and 5.14 and the following values

\[
m = 1 \times 10^{-30} \text{ kg} \quad Z = 50
\]

\[
e = 1.6 \times 10^{-19} \text{ C} \quad \beta = 1.1 \times 10^{-9}
\]

into equation 5.20 gives an expected electric field magnitude of 4 mV/m. This value is equal in magnitude to many large signals from RPE seen in my field trials (about 5 mV/m typ.). To obtain a ballpark value for the magnetic field from equation 5.19 I will use the velocity of light at 10 kHz for my typical underground conditions (about 3000 km/s), which gives a predicted magnetic field amplitude of about 1.3 nT. Again, this number is equal to the observed magnitudes of large signals. Hence, this model of excess radiation from the opening of cracks provides a means to explain the large EM fields from RPE.
Many aspects of the RPE phenomenon described by Sobolev et al. (1982) have been confirmed in my field trials. These include: the pulse-like nature of the EM signature, the non-linear relationship between the seismic force and EM pulse amplitudes, poor to fair reproduction characteristics, orebody "exhaustion", and the association of RPE with sulphide minerals. Tests with explosive sources in underground field trials have shown that approximately 100-500 kPa of seismic stress needs to be applied to the sulphide orebody in order to produce RPE emissions. Furthermore, increasing the seismic force does not increase the EM signal amplitude. Instead, greater force tends to increase the number of EM pulses. Maximum electromagnetic field amplitudes of RPE pulses are typically 5 mV/m and 2 nT when measured with a 30 kHz bandwidth. With a recording bandwidth greater than 1 MHz, these pulses may reach approximately 100 mV/m and 10 nT in amplitude. The frequency content of the EM pulse is orders of magnitude greater than the seismic input, and has been observed to range from 1 kHz to 3 MHz. In summary, the characteristics of the EM pulses produced in my field experiments are exactly as described by the Russian scientists (Sobolev et al., 1982).

In addition to confirming many RPE characteristics, I have found two new properties. One of these discoveries is that the direction with which the seismic wave impinges upon the orebody can strongly influences the amount of EM activity. Typically, the orebody ceases to produce a significant EM response after several shots are fired from the same location. However, the orebody responds anew if the shotpoint is moved to a new area, where the direction to the orebody is different, until this location is also "exhausted". In addition, it was found that if no shots are fired for a period of several hours then the orebody tends to recover; this type of recovery has been mentioned to us by the Russians (personal communication to R. D. Russell and M. Maxwell in 1983 and
Chapter 6: Thesis Conclusions

1991). The other discovery is that the relatively poor replication of shot records is mainly in the ability to reproduce the largest pulses. Spectrograms from the Lynx Mine show that the periods of pulse activity can be reproduced almost exactly, but the large amplitude events were not reproduced well. Hence, the reproduction of shot data is significantly better if the largest signals are given less emphasis.

My speculation about the origin of RPE is that the seismic wave produces triboelectric charge separation due to micro-cracking, and that the EM fields are produced from a rapid recombination of these charges. Sulphides are able to provide large amounts of triboelectric charge because telluric currents and the electronic conduction mechanism within sulphide minerals causes oppositely charged electrical layers to form on flaws and joints within the rock assemblage. Tensile stresses of the order of 5 MPa are needed to open cracks with large surface charge densities \(10 \text{ mC m}^{-2}\), and this tensile stress is provided by a combination of stress from the seismic wave and pre-existing static stresses. The seismic wave from a 0.5 kg explosive cannot provide 5 MPa of stress beyond 10 m, thus, most of the stress is supplied from pre-existing static stresses for targets beyond this distance. Failure of the material under the added tensile load from the seismic wave produces a crack, which produces an electric dipole via triboelectric charge separation.

In my proposed model of RPE, rapid recombination of the triboelectric charges from a gas discharge in the crack gap causes a large EM pulse. The large magnitude of the EM field is attributed to an imbalance between the EM fields of an electron under acceleration and the deceleration process of bremsstrahlung radiation. Bremsstrahlung processes also explain the substantial amounts of light and X-rays seen by the Russians (Sobolev et al., 1982; Sobolev et al., 1984). My model of RPE explains the production of an EM pulse from the seismic wave and the non-linear, threshold relationship between the seismic force and the EM response. It also models many other aspects of RPE such as the dominance of a single polarity in the EM response, the poor-to-fair reproduction
characteristics of repeat experiments, orebody "exhaustion" and "recovery", and the coincidence of ultrasound coincident with the EM emissions (Sobolev et al., 1982a).

The Russians have provided very little information about the methods used to acquire and interpret RPE signals. Material about using RPE for exploration is particularly sparse. This thesis provides a substantial contribution in this area by fully describing a number of methods for acquiring and analyzing RPE data. Issues about the interpretation and presentation of signals that have an unknown cause and fair reproduction characteristics have been examined, and the interpretation algorithms take these issues into consideration. The methods of tomographic reconstruction, statistical analysis of arrival data, and spectrographic analysis are new methods of treating RPE data. These techniques provide a solid foundation for the treatment of RPE data for exploration purposes.

A number of significant advances were made in the area of instrumentation and field technique over the methods and instruments used in earlier tests. High velocity explosives, such as pentolite, have proven to be the best seismic source for underground work. This is due to the matching of explosive impedance with that of underground rock. A charge size of approximately 0.5 kg is able to induce RPE more than 100 m from the shotpoint. These explosives work most effectively when used in a short borehole, with a fiber optic blast sensing circuit to provide a timing signal. The fiber optic trigger has virtually eliminated the EM interference from the blast.

Digital recording of the EM signals with a 5 MHz bandwidth provided the best results in my field trials. High bandwidth recording yields better discrimination between signal and noise, and allows the use of spectrogram analysis. The use of a 12 bit digitizer or better is recommended, although, an 8 bit digitizer sufficed in my high bandwidth work. Various EM sensor configurations were tested, and both electric and magnetic field sensors were used with good results. The parallel plate dipole with a low-noise FET preamplifier delivers the best combination of bandwidth, signal-to-noise, and portability of
all the tested configurations, and this is my recommended sensor.

The field trials at the Sullivan, Mobrun and Lynx Mines provide ample demonstration of the potential application of RPE for underground exploration for massive sulphides. At the Mobrun Mine the boundaries of the 1100 lens were located, and the split structure of the lens was revealed in the grouping of arrival times of EM signals. Signal arrival times from level 14 in the Lynx Mine and from the Sullivan Mine were shown to be consistent with the location and shape of the nearby orebody mass, and the boundaries of the ore-zones could be located from this data with some geological constraints from drill data. However, the most convincing demonstration of the practical implementation of RPE was the production of a tomographic image of the orebody from data acquired on level 10 in the Lynx Mine. Based upon these field trials, I expect that with a 0.5 to 1 kg explosive charge the RPE method is capable of reliably detecting massive sulphide orebodies 100 to 150 meters from the shotpoint.

The discouraging results from the two surface trials in Queensland, Australia, indicate that the method needs improvement before it can be applied on the surface. High-bandwidth data would allow better discrimination between spherics and RPE pulses, and probably would rectify most of the acquisition problems. However, the poor seismic coupling in weathered rock may limit the application of RPE in exploration from the surface to very shallow targets, or to borehole work.

The principal conclusions that can be drawn from my thesis is that the high frequency seismoelectric phenomenon of RPE described by Sobolev, Demin, Maybuk, and Los exists, and that it can be used to explore for sulphide minerals.
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APPENDIX A

SEISMIC SOURCE CHARACTERISTICS

A.1 Explosive to Rock Coupling

To understand why pentolite works so well we need to consider how high explosives work, and how they couple to the rock. Firstly, high explosives can only be reliably detonated by means of a shock or a high pressure impulse (Davies, 1987; Tour, 1992). Elevating the temperature of high explosives does cause them to burn, but the process is the much slower process of deflagration. When a high explosive detonates the chemical reaction within the explosive proceeds along a propagating shock wave through the explosive material (Davis, 1987). With deflagration or rapid burning the diffusive process of heating the reactants to ignition temperature propagates the chemical reaction, which is a much slower process than detonation. A combination of high pressure and temperature is needed to sustain the shock wave and the rapid chemical reactions of detonation within it. The speed of the detonation shock wave is called the velocity of detonation.

Pressure from the chemical reactions within the high explosive are distributed in two ways, shock and heave. Heave is the portion of energy that is distributed throughout the spent chemical reactants, and shock is the part that accompanies the detonation front. Pentolite explosives have very high detonation pressures and a lot of shock energy. In small scale seismic applications it is the shock that contributes most to the steep wavefront of the seismic disturbance (private communication with C.I.L. seismic explosive representatives). Hence, the use of dynamite and other high velocity explosives in seismic work. This claim is supported by my underground tests because no difference in the peak seismic amplitude, as measured by a remote geophone, was found between tamped (confined) and untamped explosives. If the heave portion of the
energy were to provide a substantial boost then tamping the explosive would increase
the amplitude of the seismic wave. Heave is more important in mining and quarrying
because the relatively slow expansion of the gases completes the break-up of rock and
moves the rock mass.

In addition to having a greater portion of energy in the shock front, pentolite explosives
couple more efficiently to underground rock than do the low detonation velocity
explosives. From basic acoustic principals we know that the maximum energy transfer
across an interface occurs when the acoustic impedance's of the two materials are equal
(Zoeppitz equations, Sheriff and Geldart, 1985). Acoustic impedance \( Z \) is defined as a
product of the density and the velocity of acoustic propagation (I will only consider
acoustic or P-waves here as these are the dominant waves in underground explosions).

\[
Z = \rho v_p \\
\text{(A.1)}
\]

Coupling of the energy from explosive shock front to the rock medium follows a similar
rule. Experimental studies by Nicholls (1962) show that the condition for maximal
transfer of energy from explosive is that the product of the velocity of detonation
(V.O.D.) and density of the explosive equal the acoustic impedance of the surrounding
medium.

\[
Z_{exp} = V_D \times \rho_{exp} \\
\text{(A.2)}
\]

and for maximum energy transfer.

\[
Z_{rock} = Z_{exp} \\
\text{(A.3)}
\]

It is difficult for explosives in hard rock to satisfy A.3 because the acoustic impedance of
hard rock is high and the densities of most explosives are quite low (1100-1700 kg/m\(^3\)).
For example, consider my typical underground situation:

\[
\rho_{rock} = 2700 \text{ kg m}^{-3} \quad \text{and} \quad v_p = 5500 \text{ m s}^{-1} \quad \Rightarrow \quad Z = 14.8 \times 10^6 \text{ kg m}^{-2} \text{ s}^{-1}
\]
\[ \rho_{\text{exp}} = 1700 \text{ kg m}^{-3} \quad \text{and} \quad V_D = 7500 \text{ m s}^{-1} \quad \Rightarrow \quad Z = 12.7 \times 10^6 \text{ kg m}^{-2} \text{ s}^{-1} \]

This is a fairly good match, but the Z of pentolite (used in the example above) is the upper bound for commonly available explosives, most fall well short. For instance, the more commonly available blasting agents have properties that fall well short for our underground conditions:

\[ \rho_{\text{exp}} = 1300 \text{ kg m}^{-3} \quad \text{and} \quad V_D = 2500 \text{ m s}^{-1} \quad \Rightarrow \quad Z = 3.2 \times 10^6 \text{ kg m}^{-2} \text{ s}^{-1} \]

In summary, pentolite and other very high velocity explosives couple well because the explosive impedance matches closely that of the rock generally found underground.

A.2 Source Energy Calculations

Matched impedance does not guarantee that all the energy crossing the interface will be seen in the far field. If the impedance's of the rock and explosive are reasonably matched then the pressures in the host rock near the explosive will be similar to pressures within the detonation front, which are approximately 10 GPa for pentolite. An empirically derived expression for the peak detonation pressure for various explosives is given by Tour (1992)

\[ P_d = 450 \frac{D}{1+0.8D} (V_D)^2 \quad \text{(A.4)} \]

where \( D \) is the specific density of the explosive. Typical values of \( P_d \) (1-10 GPa) exceed the elastic limit for most rocks (300-600 MPa typical compressive strengths for granites, basalts and andesites, Jaeger and Cook, 1971; Carmichael, 1989), and as a result much of the energy dissipates in the form of cracks and heat. Note that the very competent rock typically found underground will behave elastically at much higher stresses, thus, transferring a greater amount of energy from the explosive. Nicholls et al. (1962) observed an energy transfer efficiency of approximately 5% (best) from chemical to seismic energy in fairly competent rock (Nicholls claims that higher efficiencies, up to
20%, can be obtained, but I will use the lower figures obtained by his experiment).

If it is assumed that the shock wave is responsible for most of the seismic disturbance then we can calculate the potential mechanical energy with

$$dW = V \, dP$$  \hspace{1cm} (A.5)

(Finn, 1993). Note that the peak seismic amplitude will depend upon the geometry of the explosive/host rock contact, rather than the mass of explosive (or total chemical energy). The relatively slow gaseous expansion phase (heave) contains more energy, but achieves a smaller peak amplitude because of the much greater duration of forcing.

Explosive charges used in my field trials are cylindrical in shape to match the borehole in which they are placed. Therefore, adding charges (or mass) will linearly increase the surface area presented to the rock, and seismic amplitudes will be proportional to the square root of the charge mass. However, a spherical geometry, such as that made by a small explosive for a larger charge, will result in a cube root scaling of amplitude with charge mass (Nicholls, 1962). Note that for large scale seismic work the high frequency portion of the seismic disturbance is attenuated, and the shock properties of the explosive contribute little to the response.

Equation A.5 can be used to calculate the amount of energy put into the seismic pulse. The shock pressure increase is given by equation A.4, and is about 10 GPa for pentolite. For $V$, the volume of compressed gases in the shock front, I will use the outer shell of the explosive that actually touches the borehole wall. It is this portion of the explosive that provides the impulsive force upon the host rock to produce the seismic impulse. An order of magnitude figure for the width of the detonation front ($l$) is about 1 cm (Davies, 1983). Typical dimensions of our charges are 10 -15 cm ($h$) in length and 5 cm in diameter ($d$) (225 to 450 g charge). I will assume that the energy couples completely across the boundary and then losses 95% of the energy in heat and mechanical deformation, giving a total efficiency ($\beta$) of 5%. Therefore, A.5 becomes:
Substituting my estimates of the parameters in equation A.6 gives an expected seismic energy of approximately 100 kJ from a potential energy of 2 MJ. Telford (1986) gives a 1lb (450 g) explosive 4 MJ of potential energy, which is in fair agreement with my estimate.

To estimate the stress/strain in the far field I will treat the seismic disturbance as a pulse of P-wave energy of duration $\Delta t$, and assume that the stress field is radially symmetric about the shotpoint (i.e. a homogeneous medium). Firstly, consider the energy flux, or Poynting vector ($S$), at some distance $r$ from the shotpoint (Auld, 1973)

$$S = -v \cdot T \quad (A.7)$$

where $v$ and $T$ are the particle velocity and stress respectively (for P-waves $T$ can be represented as a vector, but it is actually a tensor). The acoustic wave and the energy are propagating radially outward. For convenience, I will progress further with scalar notation with the understanding that $T$ and $v$ are co-linear and radially directed. Integrating A.7 over time and a closed surface enclosing the shotpoint gives,

$$E_s = \int dr \int_{\Delta t} dt S \quad (A.8)$$

the total energy of the source. If equation A.8 is integrated over the surface of a sphere centered upon the shotpoint, and over the pulse duration $\Delta t$, then

$$E_s = 4\pi r^2 \Delta t S \quad (A.9)$$

Substitution of a simple relationship between the average stress ($T$) and average particle velocity ($v$),

$$T = Z v \quad (A.10)$$

into A.7 to expresses the source energy (A.9) in terms of radial particle velocity at some
distance, $r$, from the shotpoint

$$E_s = 4\pi r^2 \Delta t Z v^2$$  \hspace{1cm} (A.11)

Equation A.10 is derived from the elastic stress/strain constitutive equations (Hooke's eqn, Sheriff and Geldart, 1985), and from noting that energy is exchanged from kinetic to potential energy in the seismic wave oscillations. From A.11 we can develop two useful expressions: an equation for RMS radial particle velocity at a distance $r$ as the disturbance passes by,

$$v = \frac{1}{r} \sqrt{\frac{E_s}{4\pi \Delta t Z}}$$  \hspace{1cm} (A.12)

and an expression for the RMS stress

$$T = \frac{1}{r} \sqrt{\frac{Z E_s}{4\pi \Delta t}}$$  \hspace{1cm} (A.13)

With A.13 and A.12 I now have the means to estimate seismic field amplitudes.

$$E_s = 100 \text{ kJ}$$
$$Z = 14.8 \times 10^6 \text{ kg m}^{-2} \text{ s}^{-1}$$
$$\Delta t = 3 \text{ ms}$$
$$r = 50 \text{ m}$$

$$\Rightarrow T = 125 \text{ kPa}$$
$$\Rightarrow v = 0.84 \times 10^{-2} \text{ m s}^{-1}$$

A.3 Source Parameter Measurements

An understanding of what is measured with a geophone inside the mine is needed before a comparison can be made with my theoretical estimates of particle velocity (A.14). Firstly, I use a fairly standard type of vertical (spike) geophone, similar to those used in single component exploration surveys on the surface. The geophones have a 10 Hz resonance and are critically damped. In the mine they are generally placed into the floor of the tunnel, as this is the only place that a spike can be driven in. This orientation is not ideal because the explosive is generally located at the same elevation as the geophone; therefore, the P-wave particle motion is radial, which is the direction
that the geophone is least sensitive (supposedly non-sensitive). Nonetheless, record
clear signals of appreciable amplitude are recorded. Some simple elastic theory can
explain this, and give an estimate of the radial particle velocity from the geophone
signal.

![Figure A.1](image)

Figure A.1 Vertical ground motion due to a radial P-wave. The geophone is
mounted on a stress free boundary. Radial forces on the rock beneath
the geophone causes the material to bulge upward.

Understanding the role of the tunnel is necessary to solve the problem. Consider the
forces on a cube of rock beneath the seismometer. One axis of the cube is being forced,
another axis has stress free boundaries (the geophone axis), and the last axis has
balancing forces and no particle displacement because of the adjacent rock (Figure A.2).
This model treats the seismic wave as a plane wave incident on the first axis (wavefront
curvature is not significant tens of meters away from the source). The role of the tunnel
is to provide a free surface, which the geophone rides on. Note the similarity of this
model to the definition of Poisson's ratio, \( \sigma \), the amount of strain in a direction
perpendicular to an applied stress. Using the same approach we can obtain a modified
Poisson's ratio, \(\sigma'\), (Jaeger and Cook, 1971).

\[
\sigma' = \frac{\sigma}{1 - \sigma}
\]  \hspace{1cm} (A.15)

Equation A.15 can be used to relate the vertical movement of the geophone to the radial stress/strains on the medium. For most hard-rock types (e.g., granites) \(\sigma = 0.2\), and \(\sigma' = 0.25\). Therefore, if I have a measurement of vertical particle velocity (in the tunnel), \(v'\), then the radial particle velocity \(v\), is

\[
v = \frac{v'}{\sigma'} \quad \text{or} \quad v = v' \left(\frac{1 - \sigma}{\sigma}\right)
\]  \hspace{1cm} (A.16)

With equation A.16 measured particle velocities from my underground tests can be compared to the estimates given in A.14.

The first example comes from the Lynx Mine, B.C., where I measured a peak particle velocity of \(v' = 2.7 \times 10^{-3} \text{ m s}^{-1}\) (an 80 mV signal with a 30 V s/m geophone) on a vertical geophone 75-80 meters from a 0.5 kg shot. Using equation A.15, and normalizing the amplitude for a distance of 50 meters from the shot with \(v_1 r_1 = v_2 r_2\) (so that a comparison with A.14 can be made), a particle velocity of \(v(r = 50 \text{ m}) = 1.6 \times 10^{-2} \text{ m s}^{-1}\) is obtained. This is within a factor of two of my theoretical estimate of \(v(r = 50 \text{ m}) = 0.84 \times 10^{-2} \text{ m s}^{-1}\). Note that the theoretical estimate is an average or RMS value and that the measurement is a peak value. Hence, the two figures are comparable. The peak stress (at 50 m) is approximately 250 kPa (using A.10 and \(Z = 1.5 \times 10^7\)). A similar example from the Mobrun Mine, but with a 0.22 kg charge, gives a vertical particle velocity of \(v' = 1.3 \times 10^{-3} \text{ m s}^{-1}\) at a distance of 90 meters. Once the amplitudes have been normalized for charge size (0.5 kg) and distance (50 m), peak particle velocity and stress at 50 meters are \(v = 1.3 \times 10^{-2} \text{ m s}^{-1}\) and 200 kPa respectively.

These field measurements agree with the experience of others. Data from Nicholls (1962) gives a particle velocity of 0.013 m/s at 50 m for 0.5 kg, and a calculated (from his supplied Z) peak stress of 140 kPa. Using an empirical expression developed by the U.S.
Bureau of Mines (Sheriff and Geldart, 1985), I calculate that approximately 240 kPa of stress, and a particle velocity of 0.014 m/s at a distance of 50 meters from a 0.5 kg charge can be expected.

In summary, a 0.5 kg explosive charge should be able to deliver P-wave pressures of about 200 kPa at a distance of 50 m if the explosive is well coupled to the rock.
APPENDIX B

INSTRUMENTATION DETAILS

B.1 The Fiber Optic Time-Break

The idea behind the fiber optic time-break is that the intense light emitted by high explosives can be used to obtain an accurate time break. In practice, a small length of unsheathed fiber is attached to the detonator or explosive. Some of the light from the explosion is collected by the fiber optic strand, and is transmitted through the length of the fiber optic cable to an optical receiver, which in turn translates the light into an electrical signal (Kepic et al., in press).

Optical cable made by Hewlett-Packard, the HFBR 501 series, was used as the light conduit. The fiber optic cable is an optically clear plastic fiber, 1 mm in diameter, sheathed by PVC, with the complete fiber having a total diameter of 2.2 mm. Plastic construction allows the cable to be very light (5 g per meter), flexible (it may be bent down to a 3 cm radius), and it costs only $1 per meter. Low cost is of concern because the end of the fiber link is destroyed by the explosion. I typically lose less than 1 m with 0.5 kg charges, but sometimes a greater amount is lost due to fly-rock damage. Starting lengths of the cable are between 40 and 100 meters in length. Lengths greater than 70 meters are not recommended by Hewlett-Packard because of attenuation of the light signal, but I have used 100 meter lengths with no problems (the explosion is very powerful light source). A crimp-on connector is placed on the cable at the receiving end, but the other end is used bare.

The optical receiver, a Hewlett-Packard HFBR-2522, has a 3 microsecond response time, and can detect a blast with as much as 100 m of fiber optic cable (longest length tested). The receiver produces a digital signal, which is low (0 V) if the light exceeds a small threshold and high (5 V) otherwise, and can be interfaced with TTL compatible circuits.
This signal may be used directly, but to produce a consistent signal a pulse-generator is used to produce a 300 ms pulse; otherwise, the fiber optic receiver may produce a pulse so short in duration that it is not detected by the acquisition electronics. An attenuator placed in series with the output of the pulse generator reduces signal levels on the cable connecting the optical receiver box to the recording unit, thus, reducing cable cross-talk.

The explosive end of the fiber optic cable is prepared by stripping approximately 3 to 5 cm of PVC cladding, and then taping the bare fiber to the detonator or inserting it directly into the explosive. If the end of the fiber optic cable is not stripped then there will be no signal. After use, the burnt end of the cable can be cut off and prepared for another shot. Unlike glass fiber optics, no special care is needed to put a connector on a cable; therefore, cable repair or construction is easily done in the field. The fiber optic circuit is resistant to dirt and mud, but the connection between the receiver and optical cable should be kept fairly clean. Testing the integrity of the connection and electronics between salvoes is advisable, as we have on occasion forgotten to reconnect the power or signal cable to the optical receiver electronics after moving the equipment to new area. In addition, flyrock from the explosion can ruin the cable with little or no visible signs of damage. A camera flash unit with a manual test button provides a reliable simulation of the explosion and is very portable.

A side-by-side comparison of the fiber optic and wire-break methods on a detonator showed that the fiber optic trigger occurs first, and the wire-break signal arrives 10-20 μs later. This demonstrates the great accuracy of the fiber optic time-break because the wire-break is known to be a very accurate method of determining time of detonation (Burrows, 1936).

B.2 Electromagnetic Noise from Various Triggering Methods

As mentioned previously, the large transients at time zero from blasting box currents are intolerable for seismoelectric research. An example of this type of interference is
shown in Figure B.1, which shows the response of three grounded dipoles (labeled Dipole 1, Dipole 2 and HF antenna) in an underground seismoelectric experiment to the detonation of a 50 ms delay electrical detonator in a 225 g pentolite booster. A 50 ms delay was chosen so that the initiation interference would be finished before the blast. The record shows a large transient (some of the signals are clipped because of the high gain on the recording amplifiers) when the detonator is initiated, and approximately 50 ms later the explosion occurs, producing further signals: blast EM and possibly some other seismoelectric signals. All reasonable measures were taken to reduce the transient amplitude and duration, such as keeping signal cables and detonator wires separate and the use of a blasting box that quickly disconnects once the detonator is initiated.

Figure B.1 Electric field from detonator initiation and blast. In this example a 50 ms delayed blasting cap is used to start the explosion of a 0.22 kg charge.
In Figure B.1 it is apparent that the blast is accompanied by copious amounts of electrical signals. This type of seismoelectric signal has also been seen by O'Keefe and Thiel (1992) in quarry blasts, and Endres (1982) in small explosions in an underground mine. Early blast-related EM has all but disappeared from our records since the introduction of the fiber optic time break with safety fuse detonators. Blast associated EM is thought to be due to the electrification of rocks during fracture near the shotpoint. It appears that the wires connected to the electrical detonator re-radiate much of the electrical noise from the blast area, and communicate these signals to sensors near to the wires. The reduction of the blast EM amplitude is important as it often obscures early signals from the target.

![Graph showing electrical noise versus time-break method. Results are from a small-scale test with a detonator.](image)

Figure B.2 Electrical noise verses time-break method. Results are from a small-scale test with a detonator.
At a site near Vancouver, Canada we performed a small scale test of the EM noise produced by the fiber optic method and compared it to the wire-break method. In addition, electrical and safety fuse detonators were compared. The explosive charge used in the tests consisted of a detonator plus a small pentolite booster (about 8 g). Charges were placed into water filled holes 2 cm in diameter and 30-40 cm deep; the holes were drilled into a large, partially buried boulder. Results from the tests, displayed in Figure B.2, demonstrate that a wire-break with only 0.2 mA sense current (traces A and B) creates considerable interference, and that the fiber optic with a safety fuse detonator (E) produces the least amount of noise. The fiber optic and electrical (250 ms delay) detonator combination (C) produced a significant transient, confirming our belief that long wires leading into the explosive increase the amount of blast related noise. A 2 m length of shorted twin-strand wire placed with a safety fuse and fiber optic assembly (D) produced a barely measurable transient, but demonstrates that it is best to keep conductors away from the blast area if the lowest possible amount of EM interference is desired. The excellent electrical isolation of the fiber optic method also provides another benefit, it can be safely used in areas where it would be hazardous to use electric/seismic detonators.

B.3 Optimizing Solenoidal Magnetic Sensors

There are three basic electrical arrangements for using a coil (Figure B.3). The first is the induction coil arrangement: the coil is connected to a high input impedance amplifier. A disadvantage of the induction coil is that the voltage is proportional to the derivative of the magnetic field, so higher frequencies dominate the response. This behavior can be remedied with numerical post-processing of the data, but it requires low noise (especially at the lower frequencies) and demands a large dynamic range from the electronics and digitizer to be successful. Alternatively, an analog integrator may be used before digitization. However, the greatest drawback of this scheme is the self-resonance of the coil (the parallel combination of capacitance and inductance) because it
creates a very undesirable frequency response, and dynamic range problems if underdamped.

The other two arrangements in Figure B.3 are similar; both provide a voltage proportional to the magnetic field over a limited band of frequencies. In the damped resonator design a shunt resistor is added to the coil to produce a flat response between the frequencies determined by the R/L (lower limit) and 1/RC (upper limit) poles. Not only does the resistor set the bandwidth, but it provides the role of current-to-voltage conversion. The trans-conductance (or current-to-voltage converter) amplifier design uses the feedback architecture of an operational amplifier (Op-Amp) to convert the
Appendix B: Instrumentation Details

current at the negative terminal to an equivalent voltage. In theory, the parasitic capacitance plays no part in this topology because the potential difference across the coil is maintained, by negative feedback, to be zero. In practice, the capacitance will cause stability problems with the amplifier (i.e. high frequency oscillations) and compensating capacitance is added across the feedback resistor, thus forming a pole that lowers the frequency response of the system. The trans-conductance amplifier has a relatively flat response down to the R/L pole formed by the series resistance of the coil itself. This can be an advantage since it may extend the bandwidth down to 1-10 Hz (Macintyre, 1980; Labson et al., 1985; Hauser, 1990). However, it is a severe disadvantage if there is substantial power-line noise because the amplifier will clip the signals if there is insufficient dynamic range.

I chose to implement the damped resonator design because it is fairly robust, and it has a natural high-pass response that can be designed to attenuate power-line frequencies (60 Hz and sub-kilohertz harmonics). In addition, I inherited some good designs from earlier work by B. Narod to build upon. Note that if the same value of resistor is used for the damped resonator and the trans-conductance amplifier then the two systems have the same fundamental limits on signal-to-noise (assuming a perfect noiseless amplifier). The choice between the two designs is based primarily upon the desired low frequency response.

The band-pass portion of the transfer function is obtained by substituting $\omega = 2\pi f_o$ into equation 2.4 (see Chapter 2),

$$\frac{V(\omega)}{B(\omega)} = \frac{V}{B} = \frac{NA}{R} \frac{R}{L} \quad (B.1)$$

This is equivalent to treating the resistor as the only return path of magnetic field induced current produced by the coil. In general, a well designed transducer produces the largest possible voltage for a given value of magnetic field. From equation B.24 it can be seen that by maximizing the effective area of the coil whilst minimizing the coil
inductance produces the largest signal. In addition, the shunt resistance should be maximized. However, these simple guidelines do not produce useful sensors because of trade-offs in sensor bandwidth (in adjusting L or R), and in the signal-to-noise ratio.

After completing the Sullivan and Mobrun Mine experiments in 1991 I noticed that the noise spectrum of my best magnetic sensor (UBC I) appeared to be Gaussian or white in nature. Further investigation revealed that this noise could be attributed to the damping resistor and preamplifier within the sensor. This particular sensor, designed and built by B. Narod, offers very good performance, but significant improvements could be made in this area. New sensors for broad-band measurement (1 kHz to 3 MHz) were needed (in 1991), so these sensors were optimized for low noise as well as bandwidth.

Lukoschus (1979) and Macintyre (1980) have presented analyses on optimizing induction coil magnetometers, but their treatments are too specific. Macintyre’s (1980) is tied to a particular coil geometry, and Lukoschus (1979) is concerned with optimal weight (his sensor was to be used in space). Neither give any general guidelines. In the following, I will give a treatment that leads to some fairly intuitive conclusions about solenoidal sensors. A solenoid is a long cylindrical coil, a geometry used in many geophysical instruments because it’s properties are well defined, and has a convenient shape for borehole and airborne work. The self-inductance of a solenoid is approximated by

\[ L = \mu \frac{N^2 A}{l} \]  

(B.2)

(Lorain and Corson, 1988), where l is the length, N the number of turns, A the cross-sectional area, and \( \mu \) is the magnetic permeability. Note that this approximation is accurate if the length is much greater than the diameter. Substituting equation B.2 into B.1 gives the pass-band response for a solenoid

\[ V = \frac{IRB}{\mu N} \quad \text{or} \quad V = \frac{l}{N} RH \]  

(B.3)
Appendix B: Instrumentation Details

where \( B = \mu H \). Equation B.3 is underconstrained because the low frequency corner, \( \omega_L = 2\pi f_L \), of our design has not been designated. For a given coil and low end frequency the required value of \( R \) is (eqn. 2.6, see Chapter 2)

\[
R = \omega_L L \tag{B.4}
\]

Substituting equation B.4 and B.2 into B.3 expresses the response of the solenoid given design constraints on the low frequency performance of the magnetometer.

\[
V = \omega_L \mu NAH \tag{B.5}
\]

Equation B.5 describes the magnetometer behavior well, but the really important figure of merit in optimizing the sensor is the signal-to-noise ratio (S/N), or equivalent magnetic field noise. It is more important to have low noise than high transducer gain; the latter can be compensated by good preamplifier design. The minimum noise is determined by the Johnson noise of the damping resistance (Ott, 1988).

\[
V_n^2 = 4kTR\Delta f \tag{B.6}
\]

Johnson noise has a Gaussian or white power spectrum, and it is proportional to the absolute temperature (T), resistance (R), and bandwidth (\( \Delta f \)). The constant, \( k = 1.38 \times 10^{-23} \text{ J K}^{-1} \), is Boltzmann’s constant. Substitution of equation B.4 into B.6 and equating to the square of B.5 gives the sensor noise level in terms of an equivalent magnetic field

\[
H_n^2 = \frac{4kT\Delta f}{\mu_0 \omega_L \mu_A l} \tag{B.7}
\]

Equation B.7 is the key to optimizing a solenoidal transducer. The important information that it contains is that the low end of the magnetometer pass-band should be set as high as the intended application will allow, and that it is the effective volume of the coil that is important; not, as indicated in eqn. B.5, the effective area. This is an easy concept to comprehend: a larger coil volume encompasses a greater amount of
magnetic energy (the volumetric energy density is proportional to $\mu H^2$) for a given noise power from the resistor. Noise from the preamplifier or buffer connected to the coil will add some noise, but it is generally not a problem with the wide selection of high quality amplifiers available.

### B.4 UBC Magnetic Sensor Design

The first step in designing the sensor is defining the pass-band endpoints. For my application this was set to approximately 1 kHz to 5 MHz. Next, a ferrite rod that has the greatest volume and permeability that budget and size constraints will allow is selected. Ferrite rods 510 mm long and 17 mm in diameter, type CN-20 made by Ceramic Magnetics, were used in the UBC IV and UBC V magnetic sensors. CN-20 ferrite has a bulk permeability 800 times that of free space over a frequency range of DC to approximately 4 MHz. Beyond this frequency the permeability of the ferrite falls rapidly and losses within the material increase. All of the various ferrites and magnetic alloys share approximately the same asymptotic behavior at high frequencies. Very high permeability materials reach this asymptote at lower frequencies, limiting their ability to give a flat pass-band over the desired range. Therefore, it is the upper frequency bound on the sensor that determines the material used for the core of the coil; as a result, the upper bound influences the noise performance. For example, if my requirements were only for frequencies below 100 kHz then using MN-60 ferrite would be advantageous because it has a relative permeability of 7000 (below 100 kHz), and it would reduce the noise level by a factor of 3 over CN-20.

Once the operating frequency range has been determined then it is matter of winding the wire onto the ferrite core. My requirement for a range of 1 kHz to 5 MHz leads to a resonant frequency of 71 kHz for the coil (see equations B.23). An arrangement of three pies (or bobbins) with about 500 turns each gave the required resonant frequency. The three pies were spread over 200 mm in the center of the ferrite rod and wired in series. According to equation B.30 a lower noise level can be obtained if the full length of the
rod were used, but this does not work in practice because the effective permeability is influenced by the aspect ratio of the rod (Lorain and Corson, 1988) and the amount of rod beyond the coil; the effect of the ferrite is to focus the flux through the center of the coil, and works most effectively near the center of the rod.

There are numerous subtleties in coil design for minimum stray capacitance and reducing other resonances, but I will only mention a couple of important points. One is that the innermost layer of windings on the pie can exhibit secondary resonances, which should be tuned or damped out with either a capacitor or resistor (Labson et al., 1985). A 3.3 kΩ damping resistor across each layer of windings on the pie (4 per pie of #30 AWG wire) subdued a secondary resonance at about 1.6 MHz. A shunt resistor across the coil assembly adjusts the effective resistance seen by the coil, and the pass-band characteristics of the sensor.

A split shield to reduce capacitive or electric field pick-up is placed on the coil sensors. The split shield is a conductive sheet wrapped around the sensor, but it is split so that the shield does not make a shorted winding around the rod. I have found that a shield is necessary, despite the additional parasitic capacitance, because it reduces positive feedback problems encountered when the sensor is placed near rails and other large conductors.

Calibrating broad-band magnetic sensors with test coils is problematic. Amplitude calibration is relatively easy: at the resonance frequency, or at frequency well within the pass-band, check with a known current going into Helmholtz coils or a pre-calibrated solenoid (the approach I used). Problems occur if the calibration coil has a resonance near the test frequency, or if the ferrite rod alters the test coil's characteristics. Testing the frequency response of a broad-band sensor is straightforward using a current source, and almost impossible using test coils. The magnetic field can be represented by a current source across the coil; hence, a current source across the coil may be used to simulate the effects of an external magnetic field.
A simple current source can be made from a function generator (AC voltage source) in series with a large (say 100 kΩ) resistor. Note that this arrangement requires some care as the parasitic capacitance across the resistor (< 1 pF) will limit the bandwidth of such an arrangement (<5 MHz in this case). In addition, the supplied current is should be relatively independent of load, so the resistor must be at least an order of magnitude greater than the impedance of the transducer element (the coil/resistor arrangement. Alternatively, a bridge circuit arrangement may be used (Russell and Watanabe, 1980). Note that the frequency response tests should be done with the sensor pre-amplifier and shield installed as these will influence the self-resonance of the coil.

Time domain performance was tested with a spark discharge source. The spark discharge is created with a piezoelectric driven gas-lighter. With each spark a large current flows for a very short duration, thus generating an impulsive field. Tests with a 100 MS/s digital oscilloscope and a high quality 10x probe showed that the current flows for less than 10 ns; a near perfect impulse source. Figure 2.4 (see Chapter 2) shows the time domain response and FFT of UBC IV, which is designed to have a 1.5 kHz to 3 MHz pass-band.

B.5 Long Wire Antenna: Theory of Operation

The long wire antenna can be treated as a summing charge amplifier with each segment of wire acting as a local capacitive pick-up. R.D. Russell has formalized this approach by treating the wire as a continuous conduit of current capacitively coupled from the ground (Figure B.4). The current flowing through the feedback capacitor cancels the current that ends at the summing junction due to the wire.

\[ i = -C_2 \frac{dV_2}{dt} = C_1 \frac{dV_0}{dt} \]  

(B.8)

The current flowing into \( C_1 \) is due to the capacitive coupling of the potential difference under the wire, where each segment of wire \( dx \) contributes
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\[ di = C'(x)dx \frac{d}{dt}(V(x) - V_0) \]  
(B.9)

Figure B.4 Equivalent circuit of the long wire and charge amplifier.

The capacitance per unit length is \( C' \), and to simplify the problem I will consider it constant, which is a fairly good approximation in normal use. Integrating B.9 over the length of the wire give the total current

\[ i = \int_0^l C' \frac{dV(x)}{dt} \, dx - C' l \frac{dV_0}{dt} \]  
(B.10)

Equating B.8 and B.10 and integrating with respect to time (\( x \) and \( t \) are independent of order of integration) gives

\[ (C_0 + C_w) V_0 = C_w \int_0^l V(x) \, dx \]  
(B.11)

The total capacitance of the wire is \( C_w = C' l \). Note that right-hand side of B.11 is the first moment of potential under the wire. If we denote this average potential (with respect to ground) as \( \bar{V} \) then equation B.11 becomes

\[ V_0 = \frac{C_w}{C_w + C_0} \bar{V} \]  
(B.12)
Substituting B.12 into B.8 expresses the output of the charge amplifier for a given pattern of potential underneath the sensing wire

\[ V_2 = \frac{-C_1 C_w}{C_2 (C_1 + C_w)} \bar{V} \quad \text{and} \quad \bar{V} = \frac{1}{l} \int_0^l V(x) \, dx \]  

(B.13)

Equation B.13 is normally simplified by the use of a large value (about 10 nF) for the series input capacitance \( C_1 \). In addition, if the potential gradient is relatively constant under the wire then \( \bar{V} \) can simplified and B.13 can be expressed in terms of average field strength.

### B.6 Noise Characteristics of the Parallel Plate Dipole

The high impedance and relatively small effective height of the antenna may cause signal-to-noise problems. The Johnson noise is high at the low frequency end of the spectrum, and it is often compounded by the current noise of the amplifier. Figure B.5 shows the equivalent circuit of the parallel plate dipole and the noise sources.

![Parallel dipole antenna circuit and intrinsic noise sources.](image)

The Johnson noise is modeled as a current source to aid the analysis (rather than its usual voltage source representation). Spectral density of the noise \( e_i \) in \( V / \sqrt{\text{Hz}} \) is

\[ e_i^2 = e_a^2 + (i_s^2 + i_k^2) Z^2 \]  

(B.14)

where \( Z \) is the impedance of the parallel combination of \( R \) and \( C \).
Appendix B: Instrumentation Details

\[ Z^2 = \frac{R^2}{1 + \omega^2 R^2 C^2} \quad \text{where} \quad C = C_i + C_a \quad (B.15) \]

The total voltage noise is found by integrating B.14 over the pass-band, \( f_1 \) to \( f_2 \)

\[ V_n^2 = \int_{f_1}^{f_2} e_s^2 \, df = e_s^2 (f_2 - f_1) + \int_{f_1}^{f_2} \left( \frac{R^2}{1 + \omega^2 R^2 C^2} \right) \left( \frac{4kT}{R} + i_a^2 \right) \, df \quad (B.16) \]

where \( \omega = 2\pi f \). Equation B.16 can be simplified to

\[ V_n^2 = e_s^2 (f_2 - f_1) + (4kTR + i_a^2 R^2) f_0 \left[ \tan^{-1} \frac{f_2}{f_0} - \tan^{-1} \frac{f_1}{f_0} \right] \quad (B.17) \]

where \( f_0 = (2\pi RC)^{-1} \), the high-pass corner frequency. Several observations can be made from B.17. Firstly, better noise performance is obtained if the high-pass is as set as low as possible. Note that the amplifier current noise \( i_a \) must be very low for this design to work well.

Substituting the various parameters of my system into equations 2.9 (see Chapter 2) and B.17 gives the high-pass frequency and RMS noise referred to the input.

\[
\begin{align*}
R &= 100 \text{ M\Omega} \\
C_a &= 12 \text{ pF}, \quad C_i = 8 \text{ pF} \\
e_s &= 8 \text{ nV} \sqrt{\text{Hz}}, \quad i_a = 10 \text{ fA} \sqrt{\text{Hz}} \\
\Delta f &= 1 \text{ kHz to } 5 \text{ MHz} \\
\end{align*}
\]

\[ f_0 = 80 \text{ Hz} \\
\Rightarrow \quad V_n = 18 \mu\text{V RMS} \]

The noise level of this antenna compares favorably with other sensor configurations, which have lower intrinsic source impedances. This can be attributed mostly to the performance of the high bandwidth pre-amplifier. Most high-speed devices are designed with low impedences in mind, and typically exhibit a large amount of current noise. Therefore, high speed FET devices are normally used for this type of antenna (Casey and Bansal, 1991, and Matsui, 1991) because of their low current noise.
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B.7 The Effects of Reducing Signal Bandwidth

From Parseval's theorem we have the equivalence of energy in the frequency domain and time domain

$$\int_{-\infty}^{\infty} f^2(t) dt = \int_{-\infty}^{\infty} F^2(\omega) d\omega \quad (B.18)$$

where \( f(t) \) is the time domain description of a pulse, and \( F(\omega) \) is the Fourier transform of \( f(t) \). The integrals in B.18 may be approximated by integrating over the duration of the pulse and over the bandwidth, as these are the times and frequencies with significant amounts of energy. Both \( f(t) \) and \( F(\omega) \) can be replaced with mean or RMS amplitudes \( f_0 \) and \( F_0 \), over the pulse duration and bandwidth respectively, to give an equivalent expression of Parseval's theorem for a simple pulse

$$f_0^2 \Delta t = F_0^2 \Delta \omega \quad (B.19)$$

The relationship between the duration of a pulse \( \Delta t \) and its bandwidth \( \Delta f \) is (Bracewell, 1965)

$$\Delta t \Delta f \geq \frac{1}{4\pi} \quad \text{or} \quad \Delta t \Delta \omega = 1 \quad (B.20)$$

Substitution of equation B.20 into B.19 gives

$$f_0 = F_0 \Delta \omega \quad (B.21)$$

Equation B.21 shows that the height of a simple pulse is equal to spectral density multiplied by the bandwidth. For simple, broad-band pulses \( F_0 \) is relatively constant over the frequencies spanned by \( \Delta f \). Therefore, low-pass filtering a pulse (to \( \Delta f_1 \)) below its natural bandwidth (\( \Delta f_0 \)) lowers the amplitude by

$$f_1 = \frac{\Delta f_1}{\Delta f_0} f_0 \quad \text{where} \quad \Delta f_1 < \Delta f_0 \quad (B.22)$$

White or Gaussian noise behaves differently, despite having a similar power spectrum
(Ott, 1988). If $g_0$ is the RMS value of the noise then the new RMS value of noise is

$$g_1 = \sqrt{\frac{\Delta f_1}{\Delta f_2}} g_0$$

(B.23)

Comparison of B.22 and B.23 shows that the signal-to-noise ratio $f_g$ is lowered by reducing bandwidth. In summary the bandwidth of a pulse should preserved if possible, or otherwise maximized, when the dominant noise source has a white power spectrum.

### B.8 AM Demodulation in Operational Amplifiers

Sutu and Whalen (1983, 1985), and Chen and Whalen (1981) describe this problematic interference and provide SPICE circuit models to emulate the effects, but do not explain why it happens. From my testing it appears to be mostly related to the slew-rate performance of the amplifier. The output stage of most amplifiers are class C in design, and contain a PNP and NPN transistor to put out the negative and positive portions of the outgoing waveform respectively. NPN transistors are faster than their PNP counterparts. This is usually reflected in the performance of the positive slew-rate verses the negative: the negative is nearly always slower. Therefore, a fast changing signal such as an AM radio signal tends to push the negative half cycle electronics to it's limit before the positive. This asymmetry causes rectification, hence, AM demodulation.

An early design of the T-box pre-amplifier proved to be susceptible to AM radio transmissions. Passive filtering was introduced to the front-end of the amplifier to compensate for this behavior. The passive filter is a simple RC low-pass with the resistor in series with the input. At the suggestion of R.D. Russell, an LC low-pass arrangement was avoided because inductors will often introduce more noise than they eliminate (via inductive coupling). The resistance was selected to be fairly large, about 500 kΩ, so that the frequency response of the amplifier would be relatively independent of the source impedance.
A side-effect of the filter is the extra Johnson noise introduced by this resistor. This noise contribution can be neglected in many situations, despite adding an extra 90 nV/√Hz. To put this in perspective, consider a 3 meter dipole with a T-box preamplifier. Over a 50 kHz (effective) band-width the noise amounts to a 160 μV peak-to-peak signal, and is capable of masking a 50 μV/m signal. Since the expected amplitudes of RPE signals are in the range of 1 to 10 mV/m this amount of noise is not too bothersome. However, it can obscure small signals and detail on the larger ones.

![Figure B.6](image_url)

Figure B.6  RFI demodulation effects in four common Op-Amps. The amplifiers were tested in an non-inverting configuration (gain of 6) with a test frequency of 800 kHz. Note the demodulated voltage is the refered to input voltage (i.e. adjusted for gain).

To avoid AM demodulation an amplifier should be chosen for high slew-rate performance, or a full power bandwidth of approximately 1 MHz or better (the two are essentially the same specification). Examples of good and bad amplifiers can be found in figure B.6 where the LF411 (barely) and OP-42 have suitable performance, but the OP-27,
Appendix B: Instrumentation Details

LF-441 do not. Both the OP-27 and the LF-441 were used in our initial designs of the T-box pre-amplifiers, but we had learned our lesson by the time these amplifiers were used in surface work for RPE. Now we use the OP-42 part and passive RC filters to guarantee performance.

B.9 Demodulators and Spectral Decomposition

The acquisition bandwidth of most of my field trials have been limited by the speed of the analog-to digital conversion speed, and the number of recorded channels. At it's highest rate the RCE digitizer is capable of sampling one channel at 500 kHz bandwidth (see section 2.9 for digitizer details); an order of magnitude below the upper end reached by Sobolev's group. One channel is not good redundancy. Also, I found that I needed eight channels to understand (and feel confident about) what was going on. Therefore, eight channels were used in general, giving a usable bandwidth of 62 kHz.

Sobolev et al. have long maintained that there is a lot of signal beyond 100 kHz, and that much of this energy is contained within spectral peaks (Sobolev et al., 1986) in the HF frequency range (500 kHz to 5 MHz). In fact, Demin and Maybuk were using sensors with a 100 kHz to 5 MHz pass-band (personal communication, 1992). Some RPE signals will be oscillatory because of the high Q of the spectral peaks. The purpose of the demodulator is to extract the amplitude envelope of these oscillatory signals so that the presence of these signals will be known. Broad-band pulses will be passed with much reduced amplitudes, like the direct measurements, but oscillatory signals from the demodulator will be anomalously large because of the downward frequency transformation.

The implementation of the demodulator has varied, but all my designs (see appendix) are essentially full-wave rectifiers. This type of implementation can be done with passive diodes only, but because of forward conduction problems with small signals in diodes (i.e. a pass threshold) I used an Op-Amp and a negative feedback (or a trans-
conductance) arrangement in later designs. Biased diode circuits will work well from Hz to GHz; the feedback arrangement is linear until several Megahertz, and suffices for my purposes.

Spectral decomposition is a partial solution to the problem of finding the spectral characteristics of RPE signals without having to purchase a very expensive digitizer. This solution comes at the expense of using a large number of channels. The method is based upon a number of bandpass filters, with each filter followed by a demodulator. Each bandpass filter/demodulator combination is given a channel. Eight channels allows up to eight regions within the spectrum to be monitored. I built such a system with seven pass bands 20-50 kHz, 50-100 kHz, 100-200 kHz, 200-500 kHz, 0.5-1 MHz, 1-2 MHz and 2-5 MHz. The system was used in only one experiment, near the Century deposit in Australia (see Chapter 4 for experiment detail). Unfortunately, this was a poor experiment to demonstrate the capabilities of the technique as we could not convincingly produce any RPE signals.

A demodulator allows the monitoring of activity in frequency bands normally inaccessible to the digitizing system, and yet retain a reasonable number of channels to work with. I found the demodulator to be particularly useful with the RCE digitizer, and I feel that the spectral decomposition method can be very useful if a spare low speed digitizer (such as the RCE board) is available.
APPENDIX C

ELECTRICAL CIRCUIT SCHEMATICS
Transmitter Circuit

39K
1K '
+5V
10K
1/2 LM393
2N2222
47

HFBR-1522

Power Supply may be up to 7 V. Suggested source 6V lantern battery
4.7 uF and 10 nF Decoupling capacitors used.

Receiver Circuit

+5V

47 nF Bypass capacitor used

HFBR-2522

Sig Out

More than 250 mV is needed to give a logic low, and anything less gives a logic high output. The design with 60 m of cable will work up to 400 kHz, and with 1:3 (low:high) duty cycle to 1.4 Mhz. Glitches as short as 200 nS are captured. Falling edge is slower than rising edge.

Figure C.1 Fiber optic circuit schematics.
Figure C.2  UBC V magnetic sensor circuit schematic.
Figure C.3 HBW pre-amplifier circuit schematic.

Inputs are guarded by the inverting terminal of the OPA 627 buffers.

This differential amplifier is designed to give a gain of 30 over a bandwidth of 1kHz to 5 MHz. The input consists of a HP at 1 kHz which is then buffered with low noise FET amps. A 5MHz low pass stage passes the signal to a high speed diff amp, which has a gain of 30. Lastly the signal is buffered and band-limited to 5 MHz. A optional 5 Ohm resistor can be put in series on the output to damp reflections. Note - all leads must be kept short, especially 2pF caps.

The circuit is powered through an XLR connector. Each IC has bypass capacitors and each power rail has 10uF tantalum. The OPA 627 and AD 829 have RC networks of 10 Ohms, 10nF and 4.7 Ohms, 10nF respectively on each rail. The LM 6361 has 100 nF bypass capacitors.

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Geophysical Instrumentation Group

Title: HBW amplifier
Drawn By: AK
Ver: 1.2
12 Nov 1991
Passive Demodulator

1. + 0.485 V Ideally the voltages at 1 and 2 should be 0.6 V. The original circuit came from Horowitz and Hill and used 1K resistors on the bias arrangement. So the threshold for any signal is about 110 mV before any output will occur. This necessitates a high bandwidth preamp to be put before the input from the sensor, usually a AM502.

2. - 0.485 V

This circuit consists of two biased rectifiers to form an absolute value output from the input. The outputs are to be connected to a differential amplifier or buffer. The signal must have an amplitude greater than 110 mV to give output.

Figure C.4 Demodulator circuit schematic.
Four Channel 60 Hz Notch Filter

Four notch filter sections
One for each channel

Bypass Switch

Set R1=180 K
R2=15.0 K
for 60 Hz

Pin 1
+ E

Pin 5
- E

XLR Input

0.1uF

XLR Output

0.1uF

200 K

R1

R2

20 K

15 K.

Note that the filters have a 200 K series resistance.
The filter box should not be directly connected to the RC electronics card since that has only a 20 K input impedance. So connect this to the TEK AM 502 or a high input impedance device. Also the output and input supply lines are decoupled by the above set of LC low pass filters.

Geophysical Instrumentation Group

Title : 60 Hz Notch Filter

Drawn By : AK

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Ver : 0.0

Figure C.5  Notch filter circuit schematic.
Replacement PreAmplifier for LWA and Dipole Antenna

The choice of the 499K resistor is based on requirement that the lo-pass formed with the 10 pF capacitor have a corner at approx 30 kHz and that this corner change little with various source impedences. If the noise is a problem then replace the resistor and capacitor with 47K and 100pF respectively.

Nominal Gain is 30

HP Corner for Dipole configuration is 2 Hz and for Long Wire Antenna is 160 Hz

LP Corner for both configurations is 33 kHz with 12 dB/Oct rolloff until 1 MHz

Power Supply rails are decoupled by a 100 nH inductor and a 10 uF tantalum capacitor

Figure C.6  T-Box pre-amplifier circuit schematic: