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A Seismic Reflection Survey Over the Wayne-25 Oil Field in Cass County, Michigan

Paul D. Horton

Western Michigan University

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A SEISMIC REFLECTION SURVEY OVER THE WAYNE-25 OIL FIELD
IN CASS COUNTY, MICHIGAN

by

Paul D. Horton

A Thesis
Submitted to the
Faculty of The Graduate College
in partial fulfillment of the
requirements for the
Degree of Master of Science
Department of Geology

Western Michigan University
Kalamazoo, Michigan
December 1987
A SEISMIC REFLECTION SURVEY OVER THE WAYNE-25 OIL FIELD IN CASS COUNTY, MICHIGAN

Paul D. Horton, M.S.
Western Michigan University, 1987

The "optimum window" seismic reflection method was used in an endeavor to further define an oil producing structure in the Traverse Limestone. This method entails using a source-geophone offset which allows desired reflections to arrive in an undisturbed time zone.

Several sources in the study area produced seismic pulses with relatively high dominant frequency components ranging from 80 to 120 hertz. However, only small charges of dynamite produced sufficient energy to penetrate the glacial drift in the study area.

Preliminarily identified reflection arrivals from the targeted horizon were consistently masked in seismic profiling records by high amplitude, low velocity seismic arrivals. Varied source offsets, digital filtering and seismic trace stacking failed to unmask desired reflections.

Well log and seismic refraction data coupled with seismic modeling suggest a clay layer caused velocity inversion within the glacial drift which generated high amplitude, low velocity multiple phenomena, thereby masking reflections from the targeted horizon.
ACKNOWLEDGEMENTS

I am grateful to Dr. Gerry Clarkson for his guidance and support during the full length of this research project. I also wish to thank Mannes Oil Co. and Ward Kellner for their cooperation.

Bill Henderson deserves special thanks for his monumental help in the field work involved in this project. Also, Jeffrey S. Brown, Bill Morse, Eric Montgomery, Dennis Tripp, and others contributed greatly in obtaining seismic data.

Others to whom I am indebted include Dean Bojahanen, Angus Mann, and Doug Daniels for their availability and helpfulness in volunterring information.

Finally, I wish to thank my wife, Jill, for her enduring financial and moral support throughout the entire research project.

This study was supported in part by a research grant from The Graduate College of Western Michigan University and the Western Michigan University Geology Department.

Paul D. Horton
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INTRODUCTION

Purpose

The Wayne-25 oil field in Cass County, Michigan produces from a small dome in the Middle Devonian Traverse Limestone. Oil well logs and production data coupled with a seismic line run over the structure are interpreted to indicate the presence of a normal fault trending northeast, bounding the dome on the west. Determination of the actual location and orientation of the fault could be crucial in future development of the oil field. This information provided the justification for a seismic study of the oil field. The primary objective of this study was to use the "optimum window" shallow seismic reflection technique to generate a seismic profile of the Traverse Limestone intra-bedrock reflector. A secondary goal was to compare the relative effectiveness of a sledge hammer, shotgun, and a small charge of dynamite as high frequency seismic energy sources in a glacial drift-covered terrane.
Location and Geology

The study area is located in sections 25 and 36 of Wayne Township in Cass County, Michigan (Figure 1). The Precambrian basement consists of granite, felsic and mafic gneisses, extrusives, and metasediments, and has been assigned to the central basement province by Kellogg (1971) (Figure 2). Overlying the Precambrian basement is the Paleozoic sequence of sedimentary rocks which are overlain by Quaternary glacial drift deposits (Table 1).

The Wayne-25 oil field produces gas and oil from the middle Devonian Traverse Limestone. The Traverse Limestone in the southwestern part of Michigan is predominantly pure limestone with some beds of dolomite and argillaceous dolomitic limestones. Some lithographic limestone beds as well as abundant chert in the lower part of the pure limestone also occurs. The Traverse is thinnest in the southwestern part of Michigan, sometimes less than 100 feet thick.

The Traverse Limestone is overlain by the Traverse Formation shale which is composed of gray shale in the upper portion and gradually grades to more calcareous and argillaceous limestone near the base. The Traverse Formation is considered to be a transition zone between the overlying Antrim Shale and the Traverse Limestone.
Figure 1. Location Map of Study Area.
Figure 2. Basement Province Map From Kellogg (1971).
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The Antrim Shale is predominantly dark gray to black and brown, hard, thin-bedded, brittle, carbonaceous shale interbedded with lighter gray shale in the lower part. In southwestern Michigan the Antrim is sometimes divided into the upper dark Antrim Shale and the lower light Antrim Shale. The Antrim ranges in thickness from 120 feet where it has been eroded during the Pleistocene to 600 feet in parts of northern Michigan.

The Antrim shales are overlain by the Ellsworth Shale which is the uppermost Devonian Formation encountered in Michigan. The Ellsworth is greenish-gray with a considerable amount of interbeded siltstone. In western Michigan the Ellsworth ranges from 250 to 700 feet thick. The Ellsworth is overlain by the Mississippian Coldwater Shale.

The Coldwater Shale is predominantly gray to bluish-gray shale with some limestone, dolomite, siltstone, and sandstone. At the base of the Coldwater Shale is a thin bed (<20 feet thick) of red argillaceous limestone or dolomite known as the "Coldwater Red Rock." On the western side of the Michigan basin the Coldwater also contains an argillaceous dolomite zone known as the "Coldwater Lime". The Coldwater thins considerably from its thickness of 1000-1100 feet in central and eastern Michigan to 500-600 feet in western Michigan. The
Coldwater Shale is considerably thinner in the study area due to Pleistocene erosion.

Overlying the Coldwater Shale is 275 to 350 feet of Quaternary Glacial Drift Deposits. The lower portion of the drift is predominantly ground moraine deposits. The ground moraine is overlain by ice-contact stratified drift in kames and kame moraines of the Indiana Lagro Formation kame facies (Schneider & Keller, 1970).

Previous Work

Kellner (1986) of Mannes Oil Company is involved in the development of the Wayne-25 oil field. Based on oil well logs, production data, and a seismic line run by Hosking Geophysical (Figure 3), Kellner (1986) interprets the structure as a northeast trending normal fault (west side down) bounding a dome on the west (Figure 4,5).

Daniels (1986) of the Michigan Department of Natural Resources (DNR) is involved in a study for the DNR of the Traverse Limestone throughout Michigan. Daniels (1986) has mapped several lineaments trending both northeast and northwest throughout the Michigan basin based on oil well logs and production data.

Analysis of oil well logs, production data, and the seismic line supports the interpretation by Kellner (1986). However, preliminary interpretation does not
Figure 3. Seismic Line PP1-82-1 Over Study Area by Hosking Geophysical.
Figure 4. Structural Contours on the Traverse Limestone as Interpretated by Kellner (1986).
Figure 5. Cross Section K-K' Crossing the Structure in the Traverse Limestone as interpreted by Kellner (1986).
preclude the possibility of more than one fault involved in the structure.

Review of Selected Literature

Applications of engineering seismographs in comprehensive reflection surveys has increased with the advent of micro-computers. In the area of hydrology, Hunter et al. (1982b) have developed an "Optimum Window Technique" of shallow seismic common offset reflection profiling of specific targeted reflectors. Hunter et al. (1982b) used a sledge hammer source in mapping intra-drift and bedrock reflectors in glacial terranes, primarily in locating buried river channels. Luby (1982) used a sledge hammer source to obtain coherent reflections from intra-drift and bedrock reflectors in several glacial environments ranging in depth from 80 to 500 feet.

In mining applications, Singh (1983) developed a procedure using a sledge hammer and a signal enhancement seismograph to identify reflections from an irregular bedrock topography through alluvium to depths of 250 feet. Singh's results were used in placer tin reserve evaluation in the tin fields of the Kinta Valley, Malaysia.

More recently, Knapp and Steeples (1986c) successfully used an engineering seismograph and a high explosive source in a common depth point reflection survey.
over sinkholes along Interstate 70 in Kansas. Knapp and Steeples (1986c) delineated the subsurface vertical and horizontal extent of the sink holes because of the excellent acoustics of an anhydrite marker bed in the near surface.

Many others have done significant work in shallow (less than 500 feet) reflection seismics in hydrology, mining, and engineering applications (See for example, Hobson (1970), Nunn and Botzas (1977), Meidav (1969) and Warrick and Winslow (1969)). However, there is very little literature on procedures for deeper high resolution seismic profiling using engineering seismographs.
High Resolution In Reflection Profiling

The reflection wavelet seen on the seismograph record is nearly always a composite of various reflections caused by a set of closely spaced layers. High resolution reflection work is aimed at separating the constituents of these composite reflections into distinct separate reflection wavelets.

The ability to separate out the constituents, that is, obtain high resolution of the reflecting horizons, is largely determined by the frequency content of the source and reflection wavelets. The resolution obtainable depends on the distance between reflecting horizons compared to the wavelength of the seismic pulse. Thus, the shorter the wavelength of the seismic pulse, the higher the resolving power. Decreasing the wavelength demands a higher frequency content in the seismic pulse. Widess (1973) showed that beds as thin as 1/8 the predominant wavelength of the seismic pulse are capable of producing a reflection. Kallweit and Wood (1982) point out that the practical limit of resolution, however, occurs at about 1/4 the wavelength of the predominant frequency of the seismic pulse. Faults with a throw larger than 1/4 the wavelength can thus be seen fairly
clearly whereas the effects of smaller features would not be seen.

The problem of resolving thin layers and small faults thus becomes a field oriented problem depending on the ability to produce, propagate, reflect with sufficient amplitude, and record a seismic pulse with a significant amount of energy in the high frequency end of the spectrum. This dependence can be divided into four physical factors affecting the resolution obtainable in a common offset conventional coverage reflection survey: the source impulse, the earth's response, the geophone response, and the effects of processing.

The Source Impulse

The amplitude and frequency content of a reflected wave largely depends on three factors: the source impulse amplitude and frequency content, geometrical spreading of the wavefront, and division of energy at an interface (Mooney, 1984). Of these three factors, only the source amplitude and frequency content can be controlled in the field. Ideally, a truly impulsive source with a flat amplitude spectrum to high frequencies is desired (Knapp & Steeples, 1986a). For high resolution reflection work, this translates into the need of a source impulse containing a significant high frequency component.
Sledge hammer impacts on a lead plate firmly embedded in the soil can produce seismic pulses with frequencies up to 120 hertz (see Table 2). A highly repetitive seismic pulse can be obtained provided the lead plate is struck solidly in the center each time, the hammer is not allowed to bounce on the plate, and the lead plate does not become buried in the soil. This highly repetitive nature allows stacking of sledge hammer impacts to enhance the amplitudes of the seismic arrivals.

The high frequency content of the pulse produced by the sledge hammer is largely dependent on the near surface geology of the impact site. A decrease in the high frequency component is observed if the lead plate is pounded into the soil by the hammer blow. When this occurs, a large amount of the sledge hammer's energy is taken up in compaction of the soil immediately beneath the lead plate, resulting in a pulse of longer duration.

Bison Instruments Inc. has had success with a shotgun source called a "buffalo gun" as a relatively high frequency seismic source. The 12-gauge shotgun source consists of 3/4-inch steel pipe with special fittings on one end to hold a 12-guage, 1-ounce slug shotgun shell. A steel rod 5/8 of an inch in diameter with a firing pin centered on one end and a weight on the other served as the trigger for the buffalo gun. The buffalo gun is
driven and packed into a 1-inch diameter hole drilled to a depth of two feet (Figure 6). The slug impact produces an impulse that is not quite as rich in the high frequency component as the sledge hammer impact, but has greater energy. Successive slug impacts fired in the same hole are not greatly repetitive due to deformation and compaction of the soil by previous slugs and by jumping of the buffalo gun due to wearing of the shot hole. Stacking of the slug impacts will enhance the seismic record if new shot holes are drilled for each shot.

A blasting cap as a seismic source produces an impulse rich in high frequencies and low in energy making it useful only in very shallow, high resolution reflection work. Knapp and Steeples (1986b) conducted an experiment using a blasting cap versus one gram of high explosive known as detaprime. Their results showed that the blasting cap produced a spectrum with frequencies up to 400 hertz, whereas the detaprime explosive showed practically zero amplitude for frequencies above 140 hertz, and very little apparent energy above 110 hertz.

Traditional seismic sources such as dynamite produce a short, high-energy pulse rich in lower frequencies, with the dominant frequency between 40 and 60 hertz (Knapp and Steeples, 1986a). The only option to shorten the duration of the pulse and shift the predominant frequency towards
Figure 6. Sketch of Buffalo Gun Shotgun Source.
the high end of the spectrum is to decrease the charge size. The effect of increased resolution with decreased charge size was noted by Sharpe (1944).

Assuming that the radiation generated by an explosion is spherically symmetric, and that the fraction of total explosive energy which is converted into seismic energy is constant for a given type of explosive in a given medium, then the duration and amplitude of the pulse produced are both proportional to the cubed root of the charge size (Ziolkowski and Lerwill, 1979). This implies that the reduction in charge size will shift the spectrum of the pulse to higher frequencies and also decrease the amplitude by the same amount. Decreasing the amplitude will decrease the signal/noise ratio, and thus adversely affect the resolution. Therefore a charge size must be found that is small enough to produce the desired high frequency component, yet large enough to produce an acceptable signal/noise ratio. Ziolkowski and Lerwill (1979) successfully used a scaled down charge size in attempts to get higher resolution.
The Earth's Response

The Earth's response to a generated seismic pulse cannot be controlled. The overall effect of the earth on a seismic pulse travelling through it is that of a low-pass filter, attenuating the high frequencies and broadening the pulse.

Many factors affect a seismic pulse travelling through the earth. Spherical divergence of the seismic pulse decreases the energy density in an inverse proportion to the square of the distance over which the wave has travelled. Partitioning of energy at interfaces dependent on reflection coefficients and their variation with incident angles strongly affects the transmitted and reflected amplitudes. Peg-leg multiples in thin layers delay the seismic pulse and add it onto the original wave, lengthening and changing the shape of the wave.

Probably the most significant factor limiting high resolution is the loss of high frequency energy by absorption. As the frequency and distance travelled increase, absorption losses increase, resulting in a change of wave shape with distance travelled (Telford et al., 1976). Semi-saturated, loosely compacted and unconsolidated materials such as glacial drift have much higher frequency absorption coefficients than deeper
consolidated layers. Frequency attenuation by absorption is a major obstacle to high resolution reflection profiling using a high frequency source.

The Geophone Response

Geophones are motion sensitive transducers that convert ground motion to an electrical signal whose amplitude is proportional to the velocity of motion. For high resolution work, geophones with a high natural frequency act as a pre-emphasis, low-cut filter. The high frequency geophone suppresses low frequency noise and balance the spectrum of the incoming signal. Such pre-emphasis filtering reduces the detrimental effects of the low-pass earth filter, increases the significance of the recorded data, and makes the recording of high resolution data possible by attenuating to a manageable level the low frequency signal that would otherwise saturate a flat response recording system (Knapp and Steeples, 1986a).

Another factor to consider in high frequency seismic recording is the effect of geophone ground coupling on the incoming signal. Geophone ground coupling is the accuracy with which the geophones measure the actual ground motion. Geophones accurately follow the ground motion when the frequencies incoming are much less than the coupling resonant frequency of the planted geophone. Higher
frequency signals can be altered in both amplitude and phase. Krohn (1984) found that for a firmly planted geophone with a long spike, the coupling resonant frequency is kept as high as possible. Increasing the coupling resonant frequency outside the band-pass of the analog to digital conversion allows a flat response from the geophone's natural frequency to the ground coupling resonant frequency. Thus for high resolution, a high natural frequency geophone with a long spike planted in firm soil beneath the loose surface material will result in the best seismic record of the desired high frequency signals.

**Processing Effects**

Besides using the natural frequency of the geophone as a pre-emphasis, low-cut filter, the seismograph band-pass filter filters the analog signal from the geophone prior to digitizing and recording the signal. The band-pass filter is also a pre-emphasis filter, and for high resolution should be set as high as 75-100 Hertz on the low-cut side. The band-pass filter aids in attenuation of high amplitude low frequency arrivals that might otherwise swamp the digitizing system and saturate the memory, leaving no trace of high frequency reflections.

Another factor important to high resolution is that
the combining of geophone outputs in common depth point shooting tends to attenuate higher frequencies. Therefore, high resolution is obtained using single geophones in a conventional single-fold, continuous coverage survey than with common depth point shooting.

Profiling Method

The conventional method of recording seismic data utilized in this study is continuous coverage, single-fold shooting. In this method each geophone or geophone group samples a unique area in the subsurface; the responses of separate geophones are not combined.

An "end on-in line offset" source and geophone array allows continuous, single-fold coverage of the targeted reflector. The reflections from a single shot are recorded by equally spaced geophones laid out in a line with the shotpoint.

The portion of the reflector being sampled is half the length of the geophone spread. Continuous coverage profiling then requires a shift of the shot and geophone array by half the distance of the geophone spread for successive shots (Figure 7).

The shotpoint is offset from the geophones a specific distance to allow desired reflections to arrive during a
Figure 7. Common Offset Profiling Raypath Diagram Based on Well Log Depths in the Study Area.
time interval that does not overlap surface and refracted waves. This zone of time is called the "optimum window" and can be determined by a "walk away" noise test.

The Optimum Window

The optimum window is the region on a time versus distance diagram where reflections from a targeted reflector can arrive undisturbed by surface and refracted waves (Figure 8). The window should be chosen in an area as close as possible to the leading edge of the surface wave arrivals, spanning a zone where the reflector shows maximum curvature, yet not extending the window into the wide-angle zone where possible interference affects may occur (Hunter et al., 1982a). Identifying this optimum window allows a means of selecting source offset and geophone spacing to record the desired reflections in a clear time interval.

The most effective method for defining the optimum window is a walk away noise test. A walk away noise test is conducted by taking shots at increased intervals with the geophones in a fixed location (Figure 9). The walk away noise test provides a measurement of seismic response at a large number of source-geophone offsets, and gives a time vs. distance diagram from which the optimum window (source offset and geophone spread) for the targeted
Figure 8. Reflection Model and Time-Distance Graph Showing the Position of the Optimum Window Geophone Array (From Hunter et al., 1982b).

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Figure 9. Raypath Diagram Showing Subsurface Coverage Obtained During a Walk Away Noise Test.
reflector can be chosen. For cases where there are large lateral discontinuities in the near surface, source and geophone arrays may need to change to accommodate different conditions. In this case, several walk away noise tests should be done in different zones of the survey area to define any change in parameters needed.

Time Domain Corrections

Two time domain corrections applied to seismic profiling data are static and normal moveout (NMO) corrections. Static corrections are applied to correct the effects of irregularities in the near surface on arrival times of seismic events. These irregularities are primarily variations in elevation from shot to shot and local variations in the near surface velocities. In glacial terranes where there is a thick, low velocity zone and significant lateral discontinuities in the near surface velocities, statics can become a prominent source of mis-ties in correlating reflections from shot to shot.

NMO corrections account for the differences between reflection travel time due to the varying horizontal offsets of geophones or geophone groups from the source. NMO corrections can be calculated with synthetic models, or can be derived by velocity analyses of the targeted reflectors. Since NMO corrections are dynamic in nature,
involving a different correction velocity for successive reflectors, NMO corrections using the GeoPro seismograph can only be made for a specific reflector. The GeoPro seismograph does not allow movement of separate reflected events within a single trace.
FIELD SURVEY

Instrumentation

Instrumentation in this study was dictated primarily by availability. The seismograph is a Bison GeoPro 8012A 12-channel signal processing seismograph. The GeoPro is a microcomputer controlled seismograph for reflection and refraction seismology. Its programmable operating system provides a variety of options for the acquisition, display, processing, and storage of seismic data. The GeoPro system incorporates a non-saturating form of block floating point signal enhancement, and is powered by a 12-volt battery.

Capabilities include: removal of noisy or undesirable waveforms from enhanced data by waveform subtraction; keyboard control of input gains, sweep times, polarity of geophones, delayed triggering, programmed gain ranging, and pre-emphasis digital filtering; automatic or manual trigger arming; calibrated display control of individual waveform size and position; and two independent time marker cursors with independent display of time locations for individual waveforms.

Additional capabilities through applications programs include waveform mathematics, digital filtering during or
of the seismograph record can be obtained from the GeoPro printer or an office printer. An IBM compatible microcomputer with a null modem cable for interfacing with the GeoPro, and "Mirror" communications software allow permanent storage of seismograph records on computer disks.

Geophones are Mark Products, 60 hertz natural frequency, vertical component geophones with a 3-inch spike. The Mark Products geophone cable has 12 geophone take-outs at 30.5 foot (10 meter) intervals.

For sledge hammer and buffalo gun sources, an impact switch connected to the GeoPro by 2-conductor wire provides the trigger impulse. For a dynamite source, a detonator completes the 2-conductor circuit to trigger the seismograph at the time of detonation.

Field Parameters

Field parameters in a reflection survey are largely determined by two factors: the reflection survey method and the desired results combined with the local conditions of the survey area. The survey method controls parameters of source and geophone array and the type of subsurface coverage, as discussed in the reflection profiling section.
Source and Geophone Array

For this study an end on-in line offset source and geophone array was utilized with single geophones per trace. The medium depth of the targeted reflector dictates a larger geophone spacing than used in shallow seismic reflection profiling. Available equipment limited the maximum geophone spacing to 30.5 feet (10 meters) for 12 geophones. Thus, for ease in the field the maximum geophone spacing of 30.5 feet was used giving a total geophone spread length of 335.5 feet. An initial source offset of 610 feet was determined by refraction analysis of walk away noise tests and synthetic modeling. Since the sampled subsurface is roughly half the length of the geophone spread, continuous coverage requires a shift of the shotpoint and geophone spread by half the length of 335.5 feet (167.25 feet) for each successive shot. To allow some overlap of subsurface coverage from shotpoint to shotpoint, the shotpoint and geophone spread were shifted five geophone spacings (152.5 feet) for each successive shot.
Source Parameters

Three different sources were used in this study. The sledge hammer source consisted of a 16-pound sledge hammer impacted on a lead plate firmly seated in the soil on the edge of the road. The buffalo gun was tightly fitted into a 2-foot deep, one-inch diameter hole drilled with a one-inch hand auger. The blasting cap and dynamite charges of 80 grams 40% nitroglycerin were tamped into 2-inch diameter holes 3.5 to 4 feet deep, and detonated.

Seismograph Parameters

Seismograph parameters include record sweep time, record delay, filter settings, and gain settings. For walk away noise tests and reflection profiling, sweep times of both 960 and 480 milliseconds were used. The 960 millisecond sweep time gives a sampling rate of 1 millisecond and the 480 millisecond sweep samples at 2 millisecond intervals. A conclusion from sampling theory considerations is that no information is lost by regular sampling provided the sampling frequency is greater than the highest frequency component being sampled (Telford et al., 1976). The sweep times of 960 and 480 milliseconds have sampling frequencies of 500 and 1000 hertz,
respectively. Half of the sampling frequency is called the "Nyquist frequency." Any frequency present in the signal which is greater than the Nyquist frequency (fn) by the amount (f) will be indistinguishable from the lower frequency fn-f (Telford et al., 1976). This effect is called "aliasing" and must be considered in attempts at recording higher frequency waveforms. The Nyquist frequencies for the 960 and 480 millisecond sweep times are 250 and 500 hertz. These Nyquist frequencies are high enough to prevent significant aliasing of the seismic signal in the desired recording range of 100-200 hertz.

Dynamite source walk away surveys utilized a 960 millisecond sweep time, and thus aliasing of seismic signals could be a problem if the dynamite source produced a significant frequency component above the Nyquist frequency of 250 hertz. Frequency analysis in this study suggests a dominant frequency for the dynamite source of 80 hertz (See Table 2). A significant component above 250 hertz is unlikely.

Filter settings for high resolution of high frequency reflections were set at 35-200 hertz band-pass for dynamite shots, and 75-375 hertz band-pass for sledge hammer and shotgun shots. A band-reject (notch) filter of 60 hertz was applied for all shots to suppress power line generated noise. Gains were set by experimentation in the
field so that desired signals were not saturating the memory of the seismograph.

**Signal Processing**

Processing of seismic data is aimed at enhancing the desired reflections while decreasing the effects of noise such as undesirable seismic arrivals, low frequency ground noise (cars, tree roots, etc.), high frequency wind noise, and power line noise. Two processing steps were applied to the seismic records: trace mixing and digital filtering.

Trace mixing is simply a mathematical addition of each geophone response to the successive geophone response. For example, trace 1 = trace 1 + trace 2, trace 2 = trace 2 + trace 3, etc. Trace mixing enhances the seismic records by cancelling out-of-phase random noise, while enhancing the high velocity reflected p-wave arrivals. Trace mixing results in better overall correlation of the reflection wavelet from shot to shot. Walk away reflection analysis reveals that adverse effects on the high frequency component of the reflections by trace mixing are outweighed by the increase in correlatability of the reflected arrivals (see Reflection Analysis of Walk Away Data).
Digital filtering can also increase record quality. In high frequency work a band-pass filter can be applied to decrease the effects of low frequency undesirable arrivals outside the band-pass frequency range.
SOURCE ANALYSIS

Stacking Tests

Initial work using the sledge hammer and buffalo gun sources involved determining the number of stacks required to cancel out-of-phase random noise and enhance desired seismic events. A sledge hammer stack test (Figure 10) on McKenzie Road showed that after 10 stacked sledge hammer impacts, additional impacts do not significantly increase record quality.

Two stacks of buffalo gun shots using 1-ounce, 12-gauge slugs (Figure 11) are sufficient to clean up most random noise providing a new shot hole is used for the second shot. More than two stacks of 12-gauge, 1-ounce slugs becomes more expensive, and more work than the higher energy source of dynamite.

Frequency Analysis

The Bison GeoPro has a frequency range capability of 7-1000 hertz. For analysis of the dominant frequency of the seismic sources in the study area, low natural frequency 6 hertz phones were used. Idealized frequency response curves (Figure 12) of the 7 and 1000 hertz Butterworth filters indicate a flat response to seismic signals between 10 and 900 hertz for the total seismic
Figure 10. Seismograms of Sledge Hammer Stack Test Showing Effect of Multiple Stacks in Reducing Noise.
Figure 11. Seismograms of Shotgun Slug Stack Test Showing Effect of Multiple Stacks in Reducing Noise.

(b) 2 Slug Stacks

Figure 12. Idealized Frequency Response Curve of the 7 and 1000 Hertz Butterworth Filters on the Bison GeoPro Seismograph.
Frequency analysis is based on the assumption that a geophone offset 10-15 feet from the source will measure the seismic signal which represents the dominant frequency of the initially produced P-wave particular to the near surface conditions of the study area. At the small offset of 10-15 feet, losses in the energy and frequency content of the seismic pulse are due primarily to absorption in the loosely consolidated near surface layer. For a shotpoint at the surface, the dominant frequency of the seismic pulse after it has passed through the low velocity layer is the parameter of interest. Offsetting the geophones a small distance in frequency analysis allows recording of a pulse that is a better approximation of the true seismic pulse that has travelled through the thin, unsaturated, low-velocity layer.

Ideally, the first signal to arrive at a geophone a small distance from the source is a direct P-wave. The pulse recorded by the geophone is represented as a wavelet with a specific apparent frequency. With a flat response recording system, the apparent frequency will represent the dominant frequency component produced by the source.
Frequencies were taken from measurement of the period of the first seismic pulse recorded (Figure 13), where the frequency in cycles per second (hertz) is equal to the reciprocal of the period.

The dominant frequencies (Table 2) are applicable specifically to the study area, and may be generally applicable for these sources in glacial drift deposits. All frequencies are relatively high as required for high resolution reflection profiling. The dominant frequency of 80 hertz for the 80 gram charge of dynamite is significant in that traditional large charge sizes produce a dominant frequency between 40-60 hertz. Therefore, the reduced charge size increases the energy in the high frequency component of the impulse, allowing the possibility of higher resolution in seismic profiling using an explosive source.

Experimental evidence suggests that absorption coefficients are approximately proportional to frequency (Telford, 1976). For elastic waves in rocks, loss of energy is approximately exponential with distance travelled where \( (I_0) \) is the initial energy intensity in decibals, \( (\theta) \) is the absorption coefficient, and \( I \) is the energy intensity in decibals after the seismic pulse travels a distance \( (x) \).

\[
I = I_0 e^{-\theta x} \quad \text{Telford et al (1976)}
\]
Figure 13. Seismic Pulse Recorded by a Geophone Offset From the Source 10 Feet.

Table 2

<table>
<thead>
<tr>
<th>Source</th>
<th>Offset (ft)</th>
<th>Dominant Frequency</th>
<th>Absorption (db)</th>
</tr>
</thead>
<tbody>
<tr>
<td>16-lb. Sledge</td>
<td>10</td>
<td>120</td>
<td>.88</td>
</tr>
<tr>
<td>Buffalo Gun</td>
<td>10</td>
<td>100</td>
<td>.73</td>
</tr>
<tr>
<td>Blasting Cap</td>
<td>10</td>
<td>90</td>
<td>.66</td>
</tr>
<tr>
<td>Dynamite (80gm)</td>
<td>15</td>
<td>80</td>
<td>.88</td>
</tr>
</tbody>
</table>
An arbitrary value of the absorption coefficient thought to be typical is 0.25 decibals per wavelength (Telford et al., 1976). Using these relations, the results of frequency analysis and assuming a \( I_0 \) of 1.0, loss of energy intensity by absorption was calculated for each source seismic pulse travelling through near surface glacial drift (Table 2). Based on these calculations, the sledge hammer and dynamite source impulses retain their energy intensity to a greater degree. The sledge hammer and dynamite source impulses would only lose 12 percent of energy intensity compared to 17 and 24 percent for the shotgun and blasting cap source impulses, respectively.
ANALYSIS OF WALK AWAY DATA

Refraction Analysis

Walk away surveys provide two kinds of important information: near surface seismic response and definition of the optimum window for desired reflections. Refraction analysis of walk away data involved use of microcomputer software "HRASSD" produced by the New Jersey Geological Survey. This software includes a program for the transference of data from the Bison GeoPro to data disk and programs for the analysis of refractions and reflections, as well as for creating synthetic seismograms. Commercial software "REFRACT" produced by Geo-Logic Inc. was also used in refraction analysis of some of the data.

Locations of four walk away surveys (Figure 14) were largely dictated by topographic relief. Areas of low topographic relief were chosen to reduce complications in analysis of the data. S and D preceding location labels denotes sledge hammer and dynamite sources, respectively.

In refraction analyses, first arrival times plotted versus distances from the source give a Time-Distance (T-X) graph. Figure 15 shows a synthetic T-X graph with the data points (shown by '+'s) lying on approximately three
Figure 14. Location Map of Four Walk Away Noise Tests Conducted in the Study Area.
Figure 15. Synthetic Time-Distance Graph For a Three Layer Case.
straight lines. The reciprocal of the slope (m) of these lines represents three successive layer velocities. The time at which these lines intercept the 0 foot source offset line is called the intercept time (Ti) and is used in calculating the depths to the refracting layers. Using Snell's law and the simple geometry of a horizontally layered situation, the following equations can be derived for calculating the layer thicknesses.

Snell's Law \[
\frac{\sin \theta_1}{V_1} = \frac{\sin \theta_2}{V_2}
\]

\[
D_1 = V_1 T_i^2 \left[ \frac{1}{2} \left\{ \frac{V_2}{(V_2^2 - V_1^2)^{1/2}} \right\} \right]
\]

\[
D_2 = \frac{1}{2} \left[ \frac{Ti^3 - 2D_1 ((V_3^2 - V_1^2)^{1/2} / V_3 V_1)}{Q} \right]
\]

where \[
Q = \frac{V_3 V_2}{(V_3^2 - V_2^2)^{1/2}}
\]

(Mooney, 1984)

The depths calculated by this method represent the average depth of the subsurface that is sampled by the particular source and geophone array. The second and third layer velocities are apparent velocities, and could be different from actual velocities if the refracting horizon dips significantly.

**Refraction Analysis of Sledge Hammer Walk Away Data**

The seismogram from sledge hammer walk away S-1 (Figure 16) shows a 3-layer refraction case. T-X analysis results of walk away S-1 (Figure 17, Table 3) exhibit the general case where \( V_1 \) represents the velocity of
Figure 16. Seismograms From Walk Away Survey S-1 Presented As a Time-Distance Graph With Interpretation of Arrivals as a Three Layer Case.
Figure 17. Time Distance Graph of First Arrivals For Walk Away Survey S-1.

Table 3

T - X Analysis Results For Walk Away Survey S-1

<table>
<thead>
<tr>
<th>Velocities (ft/sec)</th>
<th>Intercept Times (Ti) (seconds)</th>
<th>Calculated Depths (feet)</th>
</tr>
</thead>
<tbody>
<tr>
<td>$v_1 = 1133$</td>
<td>$T_{i1} = 0.048$</td>
<td>$D_1 = 27.5$</td>
</tr>
<tr>
<td>$v_2 = 5985$</td>
<td>$T_{i2} = 0.097$</td>
<td>$D_2 = 186.5$</td>
</tr>
<tr>
<td>$v_3 = 9524$</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

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unsaturated, glacial till (see Table 7); $V_2$ represents the velocity of saturated, glacial till; and $V_3$ represents either a clay layer within the till, or the bedrock (Coldwater Shale). The analysis of the $V_3$ layer will be considered below. $D_1$ represents the water table depth, and $D_2$ represents the depth to the $V_3$ layer. Scatter of points on T-X graphs can be attributed to variations in the thickness of the low velocity layer and to the fact that the portion of the subsurface being sampled is different for successive shots in the walk away noise tests.

Seismograms from a forward and reverse sledge hammer walk away at location S-2 on McKenzie Road (Figure 18, Figure 19) show a 3-layer case. T-X analysis of the S-2 forward and reverse data (Figure 20) is interpreted using the "REFRACT" program. REFRACT analysis is based on first arrival times of the refracted seismic impulses. A file of time-distance data defined by the first arrivals is created and saved on disk. From the time-distance data, it is possible to solve a geologic model having several different layers. The number of layers in the model, as well as specific cross-over distances can be specified to constrain the solution of the model to include these points as known values. The REFRACT program least square
Figure 18. Seismograms From Walk Away Survey S-2 (Forward) Presented As a Time-Distance Graph With Interpretation of Arrivals as a 3-Layer Case.
Figure 19. Seismograms from Walk Away Survey S-2 (Reverse) presented as a Time-Distance Graph with interpretation of arrivals as a 3-Layer Case.
Figure 20. Time-Distance Graph of First Arrivals For Forward and Reverse Walk Away Surveys S-2.

Figure 21. Velocity Layer Profile Calculated From Time-Distance Graph of First Arrivals From Walk Away Survey S-2.
fits N line segments to the data points subject to these constraints: the N=1 line must pass through the origin, apparent velocities must increase, and the condition of reciprocity must be met. The condition of reciprocity simply stated is that the forward and reverse travel times must be equal for each line segment. Variance of the overall fits of the line segments to the data points is calculated by the REFRACT program. Reverse time-distance data can be used to solve for the average apparent dip of the layers. The results of the T-X interpretation are used by REFRACT to calculate true layer velocities, layer thicknesses, and layer dip presented as a geologic layer profile (Figure 21). Table 4 is a summation of the results of the REFRACT analysis of the S-2 walk away. Derivation of formulas used in solving multiple-dipping-layer refraction problems is contained in Mooney (1984).

**Refraction Analysis of Dynamite Data**

Seismograms from a dynamite walk away at location D-2 (Figures 22, 23, & Table 5) gave depths and velocities consistent with the reverse sledge hammer walk away, S-2, at the same location.

Seismograms from the dynamite walk away at location D-1 on Marcellus Road (Figure 24) again show a 3-layer case. T-X analysis (Figure 25, Table 6) gives depths and
Table 4

<table>
<thead>
<tr>
<th>Layer Velocities (ft/sec)</th>
<th>Forward Shot Layer Velocities (ft/sec)</th>
<th>Reverse Shot Layer Velocities (ft/sec)</th>
<th>Layer Dip (degrees)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1341</td>
<td>41</td>
<td>26</td>
<td>0.6</td>
</tr>
<tr>
<td>4923</td>
<td>164</td>
<td>132</td>
<td>1.8</td>
</tr>
<tr>
<td>6978</td>
<td>---</td>
<td>---</td>
<td>---</td>
</tr>
</tbody>
</table>

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Figure 22. Seismograms From Walk Away Survey D-2 Presented As a Time-Distance Graph With Interpretation of First Arrivals as a Three Layer Case.
Figure 23. Time-Distance Graph of First Arrivals From Walk Away Survey D-2.

Table 5

<table>
<thead>
<tr>
<th>Velocities (ft/sec)</th>
<th>Intercept Times (Ti) (seconds)</th>
<th>Calculated Depths (feet)</th>
</tr>
</thead>
<tbody>
<tr>
<td>$v_1 = 1017$</td>
<td>$T_{i1} = 0.046$</td>
<td>$D_1 = 23.8$</td>
</tr>
<tr>
<td>$v_2 = 5016$</td>
<td>$T_{i2} = 0.120$</td>
<td>$D_2 = 188.6$</td>
</tr>
<tr>
<td>$v_3 = 9211$</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
Figure 24. Seismograms from Walk Away Survey D-1 Presented As a Time-Distance Graph With Interpretation of First Arrivals as a Three Layer Case.
Figure 25. Time-Distance Graph of First Arrivals From Walk Away Survey D-1.

Table 6
T - X Analysis Results For Walk Away Survey D-1

<table>
<thead>
<tr>
<th>Velocities (ft/sec)</th>
<th>Intercept Times (Ti) (seconds)</th>
<th>Calculated Depths (feet)</th>
</tr>
</thead>
<tbody>
<tr>
<td>$v_1 = 1131$</td>
<td>$T_{i1} = 0.058$</td>
<td>$D_1 = 34.2$</td>
</tr>
<tr>
<td>$v_2 = 4236$</td>
<td>$T_{i2} = 0.114$</td>
<td>$D_2 = 188.2$</td>
</tr>
<tr>
<td>$v_3 = 8242$</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
velocities compatible with results of walk away surveys on McKenzie Road.

Summary of Results of Refraction Analyses

Water table depths calculated from refraction analyses correspond well with static water levels (SWL) recorded in water well logs from the study area (Table 7).

The third layer, V3, as interpreted from T-X graphs exhibits an average velocity of 8372 ft/sec which is somewhat higher than an average drift velocity given by Telford, et al. (1976), of 5000 ft/sec. Sonic log data from Allegan County oil wells indicate that the bedrock (Coldwater Shale) has a velocity of about 9000 ft/sec.

Initial analysis of the V3 layer suggests a refracted wave originating from the bedrock surface. T-X calculations of the average depth to layer V3 for the walk away surveys is 176 feet. The Coldwater Shale, however, lies at a depth of 265-350 feet throughout the study area. A general rule of thumb based on a horizontally layered situation states that the depth to the refractor being sampled is roughly equal to 1/3 of the source geophone array length (Mooney, 1984). Theoretically then, the array length of 1000 feet achieved in all walk away surveys should be great enough to allow recording of the
### Table 7
Comparison of Water Table Depths

<table>
<thead>
<tr>
<th>Walk Away Location</th>
<th>Nearest Water Well SWL (ft)</th>
<th>Calculated Depth (ft)</th>
</tr>
</thead>
<tbody>
<tr>
<td>S-1</td>
<td>35</td>
<td>28</td>
</tr>
<tr>
<td>S-2</td>
<td>43</td>
<td>34</td>
</tr>
<tr>
<td>D-1</td>
<td>41</td>
<td>35</td>
</tr>
</tbody>
</table>

### Table 8
Synthetic Refraction Walk Away Parameters

<table>
<thead>
<tr>
<th>Layer</th>
<th>Velocity (ft/sec)</th>
<th>Thickness (feet)</th>
<th>Depth (feet)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>1265</td>
<td>30</td>
<td>-</td>
</tr>
<tr>
<td>2</td>
<td>4791</td>
<td>155</td>
<td>30</td>
</tr>
<tr>
<td>3</td>
<td>8372</td>
<td>15</td>
<td>185</td>
</tr>
<tr>
<td>4</td>
<td>4791</td>
<td>95</td>
<td>200</td>
</tr>
<tr>
<td>5</td>
<td>9000</td>
<td>-</td>
<td>295</td>
</tr>
</tbody>
</table>
bedrock refracted wave as a first arrival.

A possible explanation of this anomalous data is the classic velocity inversion refraction problem in which a high velocity layer is underlain by a low velocity layer. Assuming that special case, a high velocity layer would lie in the drift at a depth of about 176 feet and be underlain by a lower velocity material.

Three oil well logs from the study area include descriptions of the glacial drift. They show an apparently continuous clay layer at 145, 155 and 185 feet, respectively, underlain by sand and gravel down to the bedrock (Figure 26).

Clays exhibit velocities from 3000-9000 ft/sec (Jakosky, 1950). The average velocity for layer V3 of 8372 ft/sec could be produced by a hard-packed, dense, clay layer. Assuming that V3 is the velocity of a clay layer, and that the velocity of the sand/gravel layer between the clay layer and bedrock is consistent with the drift above the clay layer (V2), a synthetic seismogram was calculated using the HRASSD program. The synthetic seismogram parameters (velocities, depths) were obtained from refraction analyses, and well log depths to the clay layer and bedrock (Table 8).

The HRASSD synthetic seismogram program calculates a travel-time synthetic model containing no real amplitude
Figure 26. Geologic Cross Section of Glacial Drift Across Study Area Showing a Continuous Clay Layer.
modeling. Reflection arrivals are arbitrarily assigned a wave amplitude twice that of refraction arrivals. The program calculates refractions and reflections for every layer specified, with the bottom layer considered as a half-space. Formulas are derived using Snell's law and the simple geometry of horizontal layers. Mooney (1984) derives the formulas for the multiple-horizontal-layer refraction case.

The synthetic seismogram showing only refractions (Figure 27) and the T-X analysis of the synthetic first arrivals (Figure 28, Table 9) correlates well with the observed field data. Note that the bedrock refraction is not a first arrival event in this model. Dynamite walk away D-1 (Figure 24) supports this model showing the bedrock refracted arrivals as they arrive later than the interpreted clay layer refracted arrivals in the far offset data.

A geologic and velocity layer profile of the drift in the study area (Figure 29) is based on the preceding analyses and models. The most significant detail of this interpretation is a velocity inversion caused by a high-velocity, thin, clay layer assumed to be continuous in the drift throughout the study area. This interpretation of near surface velocities and thicknesses will be used in
Figure 27. Synthetic Walk Away Seismograms Modeling the Seismic Refraction Response in the Glacial Drift of the Study Area.
Figure 28. Time-Distance Graph of First Arrivals From Synthetic Walk Away Seismograms Modeling the Glacial Drift in the Study Area.

Table 9

T - X Analysis Results For Synthetic Refraction Walk Away

<table>
<thead>
<tr>
<th>Velocities (ft/sec)</th>
<th>Intercept Times (Ti) (seconds)</th>
<th>Calculated Depths (feet)</th>
</tr>
</thead>
<tbody>
<tr>
<td>( v_1 = 1260 )</td>
<td>( T_{i1} = .050 )</td>
<td>( D_1 = 30.0 )</td>
</tr>
<tr>
<td>( v_2 = 4976 )</td>
<td>( T_{i2} = .088 )</td>
<td>( D_2 = 185.0 )</td>
</tr>
<tr>
<td>( v_3 = 8337 )</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

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<table>
<thead>
<tr>
<th>Depth (ft)</th>
<th>Weathered Zone ( V = 1265 \text{ Ft/sec} )</th>
<th>Sand &amp; Gravel Some Clay ( V = 4791 \text{ Ft/sec} )</th>
<th>Sand &amp; Gravel Some Clay ( V = 4791 \text{ Ft/sec} )</th>
<th>Coldwater Shale ( V = 9000 \text{ Ft/sec} )</th>
</tr>
</thead>
<tbody>
<tr>
<td>50</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>100</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>150</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>200</td>
<td>Clay ( V = 8372 \text{ Ft/sec} )</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>250</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>300</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>350</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Vertical Scale = Horizontal Scale \( 1'' = 100' \)

Figure 29. Geologic and Velocity Profile Model of Glacial Drift in the Study Area Based on Interpretation of Walk Away Surveys and Well Logs.
further synthetic modeling and in interpretation of seismic events recorded on reflection profiling seismograms.

Reflection Analysis

Reflections on the walk away seismograms are difficult to recognize and interpret due to power line generated noise and lack of apparent moveout. Trace mixing and digital filtering enhance the reflected arrivals allowing better correlation. The first step in identifying reflections was to create a reflection synthetic seismogram using the HRASSD synthetic program. Calculation of the synthetic seismogram, and interpretation of reflections in actual records are based on the assumption that travel-time (t) for a reflection and source-detector separation (x) are related by the simple equation:

\[ t^2 = t_0^2 + (x^2/V^2) \quad (3) \]

where \( t_0 \) is the two-way travel time for a coincident source and detector and \( V \) is a velocity. Dix (1955) showed that for a section consisting of only horizontal beds, equation 3 gives a satisfactory fit to the data for relatively small values of \( x \) (<5000 ft), and that \( V \) in the equation represents a RMS velocity (\( V_{\text{rms}} \)) where

\[ V_{\text{rms}}^2 = \frac{\langle v_i^2/t_i \rangle}{\langle t_i \rangle} \quad (4) \]
where $t^*_i$ is the travel time in the $i$th layer and $V_i$ is the average velocity with which a wave travels in the $i$th layer (Telford et al., 1976).

The layer thicknesses and velocities calculated by refraction analysis for the drift layer, coupled with sonic log data from Allegan County oil wells and formation depths from oil wells in the study area were used to define the parameters of the synthetic reflection walk away (Table 10). The synthetic walk away (Figure 30) shows refracted and reflected arrivals, and should be a fairly accurate representation of the true seismic reflection response in the study area with respect to travel times and RMS velocities.

The Traverse Formation and Traverse Limestone reflections form a couplet and are the latest reflections on the synthetic seismogram. The intercept time for the Traverse Limestone reflection is 275 milliseconds. The optimum window for the Traverse Limestone reflection occurs at a source offset of 366 to 1000 feet. A HRASSD time squared vs. distance squared ($T^2-X^2$) graph (Figure 31) of arrival times of the Traverse Limestone reflection within the optimum window zone gives a reflection RMS velocity of 7340 ft/sec. Interpretation of reflections is based on equation 3 where $t^2$ plotted versus $x^2$ yields a
Table 10
Synthetic Reflection Walk Away Parameters

<table>
<thead>
<tr>
<th>Layer</th>
<th>Velocity (ft/sec)</th>
<th>Thickness (feet)</th>
<th>Depth (feet)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>1265</td>
<td>30</td>
<td>-</td>
</tr>
<tr>
<td>2</td>
<td>4791</td>
<td>155</td>
<td>30</td>
</tr>
<tr>
<td>3</td>
<td>8372</td>
<td>15</td>
<td>185</td>
</tr>
<tr>
<td>4</td>
<td>4791</td>
<td>95</td>
<td>200</td>
</tr>
<tr>
<td>5</td>
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</tr>
<tr>
<td>6</td>
<td>10000</td>
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<td>427</td>
</tr>
<tr>
<td>7</td>
<td>9500</td>
<td>198</td>
<td>685</td>
</tr>
<tr>
<td>8</td>
<td>11400</td>
<td>43</td>
<td>883</td>
</tr>
<tr>
<td>9</td>
<td>14000</td>
<td>-</td>
<td>926</td>
</tr>
</tbody>
</table>
Figure 30. Synthetic Walk Away Seismograms Modeling the Seismic Reflection and Refraction Response in the Glacial Drift of the Study Area.
Figure 31. Time Squared-Distance Squared Graph of Traverse Limestone Reflection Arrival Times Computed by Synthetic Seismogram Program.

\[ V_{\text{rms}} = 7340 \text{ ft/sec} \]
\[ T_0 = 267 \text{ ms} \]
straight line whose slope \((m)\) equals \(1/V_{\text{rms}}^2\), and whose \(y\)-intercept equals the two way travel-time \(t_0\). Depths are then calculated using the simple equation

\[
h = \frac{1}{2} (V_{\text{rms}} t_0) \tag{5}
\]

where \(h\) is the depth to the reflector.

\(T^2-X^2\) analysis results can give velocities and depths accurate within a few percent where the records are good quality and the velocity distribution is not significantly laterally complex (Telford et al., 1976).

**Reflection Analysis of Sledge Hammer Data**

Based on the synthetic seismogram (Figure 30), the Traverse Limestone reflection is anticipated to occur somewhere between 250-320 milliseconds, exhibiting an RMS velocity of 7000-8000 ft/sec. Walk away survey S-1 (Figure 16) does not show a correlatable reflection event near the expected arrival time, even though field parameters and instrument settings were set for optimum enhancement of the desired reflection signal.

Walk away survey S-2 forward and reverse seismograms (Figure 18, Figure 19) appear to have a correlatable reflection in the expected time zone. Closer inspection reveals that these are the result of 50-60 hertz power line generated noise. In this case, the notch and 75 hertz low-cut filters used in the acquisition of the data
failed to diminish the power line noise sufficiently to increase the signal/noise ratio to an acceptable level.

Based on failure of the sledge hammer to produce a significant reflection event in the expected time zone, it was assumed that a large portion of the impact energy was being attenuated in the thick glacial drift.

**Reflection Analysis of Dynamite Data**

In the search for a source with adequate energy to penetrate the glacial drift, explosive energy became an obvious solution. Seismograms from dynamite walk away D-1 on Marcellus Road (Figure 32) have been trace mixed and digitally band-pass filtered 75-200 hertz to enhance any reflected arrivals. The second set of 12 traces between 366-701.5 feet offset are poor quality due to inadequate gain settings during field acquisition.

Located between 300 and 400 milliseconds in the third set of twelve traces (Figure 32) a high velocity event can be seen as three peaks that interfere with part of the lower velocity refracted arrival wave train. This event does not exhibit normal moveout, but appears to have the reverse hyperbolic shape characteristic of diffractions. Closer inspection reveals that the reverse hyperbolic shape of the arrivals can be attributed to a bend, an
Figure 32. Seismograms From Work Away D 1 Trace Mixed and Digitally Band-Pass Filtered 15-200 Hertz Present as a Time-Distance Graph With Interpretation of Reflection Arrivals.
increase in apparent velocity, on the last four traces of the third set of twelve (Figure 32).

A look at the events between 400 and 600 milliseconds on the fourth set of twelve traces shows the same increase in apparent velocity on the last four traces (Figure 32). The fact that this apparent velocity increase is evident on two separate sets of records recorded from the same set of geophones in the same locations suggests that the reverse hyperbolic shape of the events is an artifact of near surface conditions local to the last four geophones in this walk away geophone spread.

Therefore, assuming that the event between 250-350 milliseconds on the third set of twelve traces is a primary reflection (Figure 32), $T^2-X^2$ analysis was carried out to determine a rough estimate of the depth of origin. Hoffman (1985), found that if a reflection does not exhibit normal moveout, $T^2-X^2$ analysis will give a depth value prone to error. $T^2-X^2$ analysis (Figure 33) gives an RMS velocity of 6200 ft/sec and a computed depth based on normal incidence of 1034 feet. These calculations and the close match in arrival times with the synthetic Traverse Limestone reflection (Figure 30), led to the conclusion that this event is either the Traverse Formation or Traverse Limestone reflections, or a reflection from some other interface in the desired depth range of 900-1000
Figure 33. Time Squared-Distance Squared Graph of Interpreted Traverse Limestone Reflection Arrivals From Walk Away D-1.

\[ V_{\text{rms}} = 6189 \text{ ft/sec} \]
\[ T_0 = 343 \text{ ms} \]

Figure 34. Time-Distance Graph of High Amplitude Events Showing Velocities Similar To Refracted Wave Velocity.

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The high amplitude events between 400 and 600 milliseconds in the fourth set of twelve traces has an apparent velocity much slower than the reflection event in the third set of twelve traces (Figure 32). A T-X analysis (Figure 34) of the arrival times of these events compared to the arrival times of the refracted wave show apparent velocities and shape mimicking the refracted wave. T2-X2 analysis of these events (Figure 35) gives an average RMS velocity of 4129 ft/sec. This velocity is too low to belong to the Traverse Limestone reflection, or any reflections from greater depths.

Seismogram records from dynamite walk away D-2 on McKenzie Road (Figure 36) have been trace mixed and digitally band-pass filtered 75-200 hertz. Located between 300 and 350 milliseconds on the second and third set of twelve traces, a reflection is observed similar to the reflection event in the D-1 walk away (Figure 36). Again this event is interfered with by the lower velocity refracted arrival wave train in the last few traces of the third set (Figure 36). T²-X² analysis of this event (Figure 37) gives an RMS velocity of 6729 ft/sec and a depth of 1098 feet.

The optimum window for this reflected event
Figure 35. Time Squared-Distance Squared Graph of High Amplitude Events.

Figure 37. Time Squared-Distance Squared Graph of Interpreted Traverse Limestone Reflection Arrivals From Walk Away D-2.
Figure 36. Seismograms From Walk Away D-2 Trace-Mixed and Digitally Band-Pass Filtered 75-200 Hertz
Presented as a Time-Distance Graph with Interpretation of Reflection Arrivals.
correlates well with the window produced by the synthetic seismogram, and is located between source offsets of 450 and 1000 feet.

**Summary of Reflection Analyses**

Sledge hammer walk away seismograms do not show a reflection in the expected time zone. This fact is attributed to source energy attenuation by a thick overburden of glacial drift. Dynamite walk away seismograms are consistent with each other and exhibit a reflection interpreted as originating near the targeted zone of the Traverse Limestone. The optimum window is limited on the far offset side by a low RMS velocity, high-amplitude event resembling refracted arrivals.
INTERPRETATION OF PROFILING RECORDS

Velocity Analysis

Common offset, dynamite source profiling records from Marcellus and McKenzie Roads show a high-amplitude, low RMS velocity event covering up to 250 milliseconds of the seismogram following the refracted arrival (Figure 38, Figure 39). The seismogram records shown in figures 38 and 39 are unprocessed 612 ft source offset profiling records from the study area.

T-X plots (Figure 40) of the arrival times of the refracted wave and the high amplitude events that follow are representative of the profiling data in general. These later events do not exhibit normal moveout curvature on the T-X diagram (Figure 40). If these events are reflections, a plot of arrival times versus distance from the source should allow definition of normal moveout on a T-X graph. The synthetically calculated normal moveout curvature for the Traverse Limestone reflection is easily visible on a T-X diagram (Figure 41). The velocities of just above 8000 ft/sec for these events are similar to the refracted wave velocity $(V_3)$ of 8372 ft/sec as determined for the clay layer in the drift. These velocities are also compatible with the a bedrock refracted wave velocity of 9000 ft/sec determined from sonic logs of oil wells in
Figure 39. Dynamite Source Profiling
Seismograms From McKenzie Road.
Figure 40. Time-Distance Graph of Seismic Arrivals From Record 17 of Marcellus Road Profiling Seismograms.

Figure 41. Time-Distance Graph of Synthetic Traverse Limestone Reflection Arrivals Showing Normal Moveout Curvature.
Allegan County, Michigan.

T²-x² analysis (Figure 42) of these events gives RMS velocities of just above 4000 ft/sec. These are lower than expected reflection velocities as predicted by synthetic seismograms, and the walk away interpretation of the Traverse Limestone reflection RMS velocity of 6000-7000 ft/sec.

Seismic records from figures 38 and 39 time corrected to the first arrivals (Figure 43, Figure 44), demonstrate the near equivalent velocities of the questioned events. The events line up across the records giving the impression of continuous reflections. Sledge hammer and buffalo gun profile records from McKenzie Road have been time corrected to the first arrivals (Figure 45, Figure 46) and resemble the dynamite profile records in that the high amplitude events having similar velocities line up, giving the impression of horizontal, continuous reflectors. Profiling records shot on Marcellus and McKenzie Roads totaling over 2 miles of surface coverage failed to show any significant seismic arrivals other than these high-amplitude, low-velocity events.

The source offset was altered within the optimum window to avoid the masking of desired reflection signals by the high amplitude events. First-arrival corrected, dynamite source profiling records from McKenzie Road where
Figure 42. Time Squared-Distance Squared Graph of High Amplitude Events From Marcellus Road Profiling Seismograms.
Figure 43. Dynamite Source Profiling
Seismograms From Marcellus
Road Time Corrected to First
Arrivals.

Figure 44. Dynamite Source Profiling
Seismograms From McKenzie
Road Time Corrected to First
Arrivals.
Figure 45. Sledge Hammer Source Profiling Seismograms
From Mckenzie Road Time Corrected to First
Arrivals.
the source offset was reduced to 415 feet (Figure 47) still exhibit the low RMS velocity, high amplitude events.

Unprocessed seismic records from northern Michigan done by Hosking Geophysical of Mt. Pleasant, Michigan (Figure 48), exhibit high-amplitude, low-velocity events mimicking the shape of the refracted arrivals in the far offset data mostly between 400 and 1000 milliseconds. Mann (1986) stated that these events are fairly common in records shot in Michigan, and probably originate from within the drift, possibly at the bedrock-till interface. Mann (1986) does not speculate on the actual mechanics of their origin.

Apparently these events (Figure 48) are the same type as observed on records in the study area, and a tentative conclusion is that they are related to specific and local geologic conditions that are common within the glacial drift in Michigan.

**Summary of Velocity Analyses**

Root mean square velocities of the high-amplitude events are slower than expected for reflections from within the bedrock. T-X plots of the arrival times do not exhibit normal moveout curvature, and the velocities do not increase with time as expected for reflected arrivals.
Figure 47. Dynamite Source Profiling Seismograms From Mokenzie Road (source offset = 415 ft.) Time Corrected to First Arrivals.
Figure 48. Unprocessed Seismograms Obtained in Northern Michigan by Hosking Geophysical.
The high-amplitude coupled with the T-X velocities of these events closely resemble those of the refracted arrivals from the clay layer in the drift, or the bedrock. These facts lead to the conclusion that these events originate within, or because of, conditions in the glacial drift. Interference from these seismic events makes application of this reflection profiling technique inadequate for mapping reflections off the Traverse Limestone in the study area.

Future work in implementing this reflection survey method would benefit greatly from an understanding of these high-amplitude, low-velocity events. The origin will be considered in light of amplitude and travel times for various possible events given the velocity distribution in the glacial drift as determined by refraction analysis (Figure 20).

Amplitude Analysis

Amplitude modeling serves to distinguish the seismic events that could possibly be responsible for the large amplitude seismic events recorded. A simple model (Figure 49) is a plot of the relative amplitude (Ao/Ai) of seismic events versus the angle of incidence (θi), where Ao is the initial amplitude and Ai is the reflected or refracted amplitude. Amplitude curves were calculated by McCamy, et
Figure 49. Relative Amplitude ($A_o/A_i$) Versus Angle of Incidence ($\theta_i$) For the Model Case of $V_1/V_2 = 0.522$, Density Ratio of 0.757, Critical Angle of 31.5 Degrees (From McCamy et al., 1962).
al., (1962) for a model case of: $V_1/V_2 = 0.522$, a density ratio of 0.757, and a critical angle of 31.50 degrees. Amplitudes were calculated by numerical methods solving the simultaneous linear Zoeppritz equations which define the partitioning of the amplitude of an incident wave front at an interface (Appendix A). The general applicability of these amplitude curves by velocity contrast alone is demonstrable from studies of the velocity vs. density relation (Talwani et al., 1959).

From refraction analysis, velocity ratios of 0.572 for the drift-clay layer interface and 0.532 for the drift-bedrock interface were considered in choosing this set of curves to model the amplitude response of the glacial drift interfaces in the study area. Analysis of these curves show that the amplitude of the refracted P-wave is greater than any other event up to the critical angle. From the critical angle to 35 degrees incidence the P-wave reflected event exhibits an amplitude greater than the refracted wave in the zone preceding its approach to the critical angle. The reflected shear wave produced by an incident P-wave is often referred to as a "converted wave" and has been noted by Dohr and Janle (1980) as a distinct arrival in conditions of a high velocity contrast, such as a drift-bedrock interface. Dohr and Janle (1980) stated
that the amplitude of a converted wave is strong only at large source offsets, i.e., large angles of incidence. In the wide-angle zone between 35-55 degrees incidence the S-reflected (converted) wave has an amplitude greater than the amplitude of the reflected P-wave.

Amplitude modeling suggests that a P-wave incident between 30 and 35 degrees on the clay layer or bedrock interfaces could be responsible for a P-P reflection arrival comparable in amplitude to the refracted P-wave. Also, a P-wave incident between 35 and 55 degrees could be responsible for a P-S reflection arrival comparable in amplitude to the refracted P-wave.

Raypath Modeling

The velocity distribution model for the glacial drift as interpreted from refraction analysis (Figure 50) shows raypaths of reflected P-waves with angles of incidence of 30 and 35 degrees. This model suggests that P-wave reflections off the clay layer and bedrock interfaces between these angles of incidence could not be responsible for the events recorded at a source offset of 612 feet. If responsible, they would exhibit normal moveout and a T-X velocity of roughly 4000 ft/sec, which is about 1/2 the T-X velocity of the questioned events.

A raypath model of converted waves produced by P-
Figure 50. Velocity Distribution Model of Glacial Drift in the Study Area With Raypaths of P-Waves Incident at 30 and 35 Degrees.
Figure 51. Velocity Distribution Model of Glacial Drift in the Study Area With Raypaths of Converted Waves incident at 35 and 55 Degrees.
waves incident between 35-55 degrees (Figure 51) is calculated for a P-wave velocity/S-wave velocity ratio of 1.85. This value is taken from tabulated values of P and S-wave velocities in a glacial moraine by Parasnis (1972). Note that a converted wave reflection off the bedrock does not occur for incident angles greater than the critical angle associated with the clay layer. This model demonstrates that it is impossible for a converted wave, produced by a P-wave incident between 35 and 55 degrees, to be responsible for the high amplitude event on the profiling records, where the source offset is 612 feet.

Assuming the questioned event is not a reflection, but is a refracted arrival travelling at a velocity similar to the clay layer and bedrock velocities, an explanation of its origin is limited to multiple phenomena serving to delay refracted arrivals from the clay layer or bedrock interfaces. A raypath model of possible near-surface and interformational multiples (Figure 52) is calculated for angles of incidence less than the critical angle associated with the clay layer. Any, and combinations of all of these multiples can contribute to delays in refracted wave arrivals. However, the regular period of about 40 milliseconds exhibited by the events on the seismic records suggests a cyclic or regular delay in
Figure 52. Velocity Distribution Model of Glacial Drift in the Study Area With Raypaths of Possible Near-Surface and Interformational Multiples Incident at 35 Degrees.
the refracted arrivals.

Raypath modeling (Figure 52) suggests two simple mechanisms of multiple events capable of producing a cyclic delay given the assumed velocity distribution: a series of near-surface multiples delaying the seismic signal before or after refraction, or a series of interformational multiples in the low velocity zone between the clay layer and bedrock delaying the bedrock refraction. In each case the delays produced are equal to the travel time for one multiple. Successive multiples could produce refracted arrivals with an apparent period equal to the travel time of the multiple.

Near Surface Multiples

Assuming a velocity in the unsaturated near-surface drift of 1265 ft/sec and in the saturated drift of 4791 ft/sec along with average densities of 1.6 g/cm$^3$ and 1.8 g/cm$^3$ taken from Telford et al. (1976), the calculated reflection coefficient (R) of .38 for the water table is quite high and supports possibilities of near-surface multiples. The near-surface multiple mechanism may be significant in certain cases, but it does not explain the absence of these types of arrivals in glacial terrane where the water table is not at the surface. Based on the random occurrence of these types of events in widely
scattered seismic studies throughout Michigan by Mann (1986), this mechanism is probably not an acceptable explanation of the origin of the questioned seismic events.

Interformational Multiples

Assuming a drift velocity of 4791 ft/sec below the clay layer, the reflection coefficients for the drift-bedrock and clay layer-drift interfaces of .18 and .13 are significant, although not as large as the reflection coefficient at the water table. Based on this velocity, the multiple raypath (Figure 52) associated with this velocity layer can only bounce once before travelling as a critically refracted wave along the bedrock interface. That is, only one multiple event, instead of a series of events, would arrive at all twelve geophones offset from the source 612 feet.

Previous T2-X2 analysis (Figure 33) of the reflection as interpreted on dynamite walk away records gave a depth of origin of 1000 feet and an RMS velocity of 6200 ft/sec. This velocity is over 1000 ft/sec slower than the RMS velocity of 7372 ft/sec predicted for the Traverse Limestone reflection. The synthetic model from which this prediction came is partly based on the previous, and possibly erroneous, assumption that the velocity of the drift separated by the clay layer is the same on both
sides. A synthetic seismogram with a drift velocity below the clay layer of 3000 ft/sec gives an RMS velocity for the Traverse reflection of 6900 ft/sec.

Using the velocity of 3000 ft/sec in the raypath model (Figure 53) reflection coefficients of the clay and bedrock interfaces become .30 and .36, respectively. The critical angle is reduced to 20 degrees allowing several bounces at the critical angle capable of producing multiply-delayed, refracted arrivals with a regular period equal to the bounce time in the low velocity layer.

This mechanism of multiples in the low velocity zone above the bedrock provides a fairly simple explanation of the regular period, apparent velocity, lack of normal moveout, and relatively high amplitude of the questioned events. This mechanism is dependent on an interpreted, specific and local geologic condition in the glacial drift, namely the presence of a continuous, high velocity clay layer underlain by lower velocity glacial material to bedrock. Conditions similar to these are probably common, but not pervasive, throughout the glacial drift in Michigan.
Figure 53. Velocity Distribution Model of Glacial Drift in the Study Area With Raypaths of Possible Multiple Incident at 20 Degrees.
REMOVING THE MULTIPLES

Conventional means of dealing with high-amplitude, low-velocity seismic multiples are common-depth-point (CDP) processing coupled with F/K (velocity) filtering. Groundroll is suppressed by a combination of low-cut frequency filtering and wave-number filtering, which is simply the type of discrimination accomplished by CDP stacking.

Frequency/Reciprocal of Wavelength (F/K) filters are velocity filters for removing coherent noise on reflection records that has a different apparent moveout or apparent velocity than the desired reflection events. The time differential for a particular event between adjacent traces is the criterion on which the filtering operation is based. F/K filters are based on plots of wave numbers (reciprocal of wavelength-K) versus frequency (F) which is effectively a conventional time-distance plot rotated by 90 degrees. This is called an F/K plot and the slopes of the lines are proportional to the apparent velocities of the events, rather than the reciprocal of the apparent velocities, as on a T-X plot (Dobrin, 1976). Reflections normally have a small moveout across a record, and fall within a narrow wedge centered on the vertical axis of the F/K plot. High velocity noise generally falls within a
wedge of intermediate slope (Figure 54). F/K velocity filters provide a means of suppressing high apparent velocity noise without detriment to reflection signal quality. Thus, conventional seismic processing of CDP data automatically removes the effects of these types of multiple events.

Barring use of CDP profiling and velocity filtering, an alternative for the engineering geophysicist is to remove the effects of the groundroll. Removing the groundroll opens the time zone within the low-angle zone of incidence for reflections, allowing recording of reflections without interference from the multiply-delayed refracted wave originating from the bedrock interface. Although, the same multiples producing the multiply delayed refracted wave may then be recorded as multiple reflections. Removing the groundroll cannot be accomplished by low-cut filters alone, due simply to the magnitude of the groundroll. The summing of at least two geophone responses is required to cancel groundroll.

Knapp and Steeples (1986a) has had success in removing the groundroll by "geophone differencing." Geophone differencing involves taking the difference of the output response of two geophone elements that are vertically separated by a small distance. This technique attenuates horizontally propagating energy (groundroll) as
Figure 54. Reciprocal of Wavelength (K) Versus Frequency (F) Called an F/K Plot (From Dobrin, 1976).
it is received by the two geophones in-phase and is subtracted out. Vertically propagating energy, however, is slightly phase shifted due to the geophone separation, and differencing of the reflection signal rather than cancellation results (Knapp and Steeples, 1986d). Experimental results (Knapp and Steeples, 1986) suggest that a burial depth of 3 feet is deep enough to prevent differencing from interfering with the reflection signal. Geophone differencing attenuates the groundroll and has a gentle high-pass filtering effect on P-waves.

The application of this method for reflection profiling with limited seismic capability systems as used in this study would require the use of a small and very mobile drill rig for drilling the small diameter shallow holes necessary to bury a geophone at depth. A special set of geophone tubes would also be needed for the burying of one geophone at depth.
CONCLUSIONS

Sledge hammer, buffalo gun and blasting cap sources produce a seismic pulse rich in high frequencies of 100 hertz and above. However, in seismic profiling to locate an intra-bedrock reflector (Traverse Limestone) through relatively thick glacial drift (<500 ft), the overall energy produced by these sources is not great enough to negate attenuation of the signal by the drift in the study area.

A scaled down charge size of dynamite contains sufficient energy and is significant in increasing the high frequency component of the seismic pulse, allowing the possibility of better resolution.

The application of the optimum window technique of reflection profiling was not successful in locating intra-bedrock reflections in the study area. The method probably failed due to local conditions in the glacial drift creating a low RMS velocity, high-amplitude event masking reflection arrivals in the optimum window zone.

The local condition in the drift is interpreted from refraction analysis to be a velocity inversion caused by a relatively thin, continuous, dense clay layer within the drift. The low velocity zone below the clay layer and above the bedrock is the generator of interformational
multiples serving to delay successive refractions from the bedrock interface. The multiply-delayed refractions are interpreted as the high amplitude events which fill up the optimum window zone.

Similar conditions in the glacial drift are certainly common throughout Michigan, and could significantly deter attempts at reflection profiling using the optimum window method.

Solving the problem of the high-amplitude multiples would require abandoning the optimum window profiling method and working in the low incident angle groundroll zone. A method for cancelling groundroll by Knapp and Steeples (1986d) may be effective. Utilization of this method would require the addition of mobile drilling equipment and possibly a new geophone setup to the standard seismograph equipment.
Appendix

Zoepritz Amplitude Equations
The equations were taken after Richter (1958).

for incident $P$,\[ (A - C) \sin a + D \cos b - E \sin e + F \cos f = 0 \]
\[ (A + C) \cos a + D \sin b - E \cos e - F \sin f = 0 \]
\[- (A + C) \sin 2a + D \frac{V_1}{U_1} \cos 2b + \frac{E K}{U_1} \sin 2e - \frac{F K}{U_1} \frac{V_1}{V_2} \cos 2f = 0 \]
\[- (A - C) \cos 2b + D \frac{V_1}{V_1} \sin 2b + \frac{E K}{V_1} \cos 2f + \frac{F K}{V_1} \frac{V_2}{V_1} \sin 2f = 0 \]

and for incident $SV$,\[ (B + D) \sin b + C \cos a - E \cos e - F \sin f = 0 \]
\[ (B - D) \cos b + C \sin a + E \sin e - F \cos f = 0 \]
\[ (B + D) \cos 2b - C \frac{U_1}{V_1} \sin 2a + \frac{E K}{U_1} \frac{U_2}{V_2} \sin 2e - \frac{F K}{U_1} \frac{U_2}{U_1} \cos 2f = 0 \]
\[- (B - D) \sin 2b + C \frac{U_1}{V_1} \cos 2b + \frac{E K}{U_1} \frac{V_2}{U_1} \cos 2f + \frac{F K}{U_1} \frac{V_2}{U_1} \sin 2f = 0. \]

Where $A$ and $B$ are the amplitudes of $P$ and $SV$, incident at angles $a$ and $b$ respectively and $C, D, E, F, V_1, U_1, V_2, U_2$ are the amplitudes of the resultant rays, reflected $P$, reflected $SV$, refracted $P$ and refracted $SV$, at angles $a$, $b$, $e$, and $f$. $V_1, U_1, V_2, U_2$ are the $P$ and $SV$ velocities in the first and second layers respectively. $K$ is the ratio of the density of the second layer to the density of the first. The angles $a$, $b$, $e$, $f$ are related by Snell's Law:

\[
\frac{\sin a}{V_1} = \frac{\sin b}{U_1} = \frac{\sin e}{V_2} = \frac{\sin f}{U_2} .
\]
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