A Comparative Case Study of Reflection Seismic Imaging Method

Moones Alamooti
University of Mississippi, malamoot@go.olemiss.edu

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A COMPARATIVE CASE STUDY OF REFLECTION SEISMIC IMAGING METHOD

A Thesis
presented in partial fulfillment of requirements
for the degree of Master of Science
in the Department of Geology and Engineering Geology
The University of Mississippi

by
MOONES ALAMOOTI

May 2018
ABSTRACT

Reflection seismology is the most common and effective method to gather information on the earth interior. The quality of seismic images is highly variable depending on the complexity of the underground and on how seismic data are acquired and processed. One of the crucial steps in this process, especially in layered sequences with complicated structure, is the time and/or depth migration of seismic data. The primary purpose of the migration is to increase the spatial resolution of seismic images by repositioning the recorded seismic signal back to its original point of reflection in time/space, which enhances information about the complex structure.

In this study, our objective is to process a seismic dataset (courtesy of the University of South Carolina) to generate an image on which the Magruder fault near Allendale SC can be distinguished, and its attitude can be accurately depicted. The data was gathered by common mid-point method with 60 geophones equally spaced along an about 550 m long traverse over a nearly flat ground. In the future, we will apply different migration algorithms (including finite-difference and Kirchhoff), and the results will be compared in time and depth domains to investigate the efficiency of each algorithm in reducing the processing time and improving the accuracy of seismic images in reflecting the correct position of the Magruder fault.
DEDICATION

I would have liked to dedicate this thesis to my lovely parents and kind brother for all their endless support, love, and encouragements throughout my life but since the earthquake occurred in Kermanshah, province in the western part of Iran, on 12 November 2017 at 18:18 UTC, which it caused the most significant damage and loss of life, while I was writing my thesis in Oxford, Mississippi, far from the motherland, this thesis is dedicated to the kind, patient and brave people of Kermanshah.

May the improvement in seismic knowledge and techniques provide the safer and happier life for all people around the world, especially for all kids.
# LIST OF ABBREVIATIONS AND SYMBOLS

<table>
<thead>
<tr>
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<th>Description</th>
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<tbody>
<tr>
<td>AGC</td>
<td>Automatic Gain Control</td>
</tr>
<tr>
<td>CMP</td>
<td>Common Mid Point</td>
</tr>
<tr>
<td>CVS</td>
<td>Constant Velocity Stacking</td>
</tr>
<tr>
<td>DB</td>
<td>Dunbarton Basin</td>
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<tr>
<td>DEM</td>
<td>Digital Elevation Model</td>
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<tr>
<td>DMO</td>
<td>Dip Move-out</td>
</tr>
<tr>
<td>DRG</td>
<td>Digital Raster Graphic</td>
</tr>
<tr>
<td>FFTS</td>
<td>Fast Fourier Transforms</td>
</tr>
<tr>
<td>FD</td>
<td>Finite Difference</td>
</tr>
<tr>
<td>F-K</td>
<td>Frequency-Wave number</td>
</tr>
<tr>
<td>GPR</td>
<td>Ground Penetrating Radar</td>
</tr>
<tr>
<td>MF</td>
<td>Magruder Fault</td>
</tr>
<tr>
<td>NMO</td>
<td>Normal Move-out</td>
</tr>
<tr>
<td>R</td>
<td>Receiver</td>
</tr>
<tr>
<td>RB</td>
<td>Riddleville Basin</td>
</tr>
<tr>
<td>S</td>
<td>Source</td>
</tr>
<tr>
<td>SC</td>
<td>South Carolina</td>
</tr>
<tr>
<td>S/N</td>
<td>Signal to Noise ratio</td>
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VSP  Vertical Seismic Profile
ACKNOWLEDGEMENTS

I express my appreciation to my advisor, Dr. Adnan Aydin and my committee members, Drs. Leonardo Macelloni, and Louis G. Zachos for their help and guidance. I would also like to thank the Department of Geology and Geological Engineering for giving me the graduate assistantship position during my studies.

Also, I express my gratitude to Dr. Terry L. Panhorst for his support and inspiration through my teaching assistantship. I also thank the University of South Carolina for providing access to their seismic data. I am thankful to Mr. Gary Engler for providing technical support.

Lastly, I acknowledge all the undergraduate students that I have served as their lab instructor. They enriched my life and made it more beautiful by their positive feedback and constructive suggestions.
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SEISMIC IMAGING

Seismic imaging is the seismological technique which involves inducing the seismic wave into the earth and recording the waves that are reflected from sub-surface layers. There are different types of geophysical methods as electrical resistivity tomography, ground-penetrating radar (GPR), reflection seismology, induced polarization, and magnetotellurics.

The focus of this study is the imaging of seismic reflection data from the controlled sources. If we consider the subsurface geology as a layer cake, which means that the layers are horizontally stratified, the seismic images can be produced quickly, but as these layers can vary due to erosion, changes in the lithology, and presence of any subsurface features resulted by tectonic and dynamic forces as faults and folds, then we need to modify the horizontal layered structure, considering the fractured and bent strata, so in this case the recorded response of the seismic energy consists of different elastic waves like guided waves, surface waves, refracted waves, reflections from the discontinuities, multiple reflections, interface waves which are propagated along the boundary of two different layers. Briefly, the seismic waves will be reflected, refracted or diffracted through propagation.

The seismic imaging will be a significant endeavor to characterize the underground changes in material properties, to get the accurate information about the subsurface features as the faults in geologic formations, to monitor the integrity of the crucial infrastructures as
dam and foundation, and to track the movement of CO₂ in geological carbon sequestration (Los Alamos, 2008). Moreover, to produce the seismic image of the Earth, and to get more accurate information about subsurface stratigraphic and structural features with all complexity, we need to unravel all the effects caused by diffraction, reflection, and refraction of the seismic wave propagation.

One of the essential steps in the process of seismic imaging is migration which involves relocating the seismic events (in depth or time domain) to the actual location of the event in the subsurface rather than the recorded location at the surface. This process causes the movement of the dipping reflector to their correct subsurface location (Yilmaz, 2000). The goal of migration is to reposition the seismic events to their correct spatial locations in which the true reflections appear from the structural features directly beneath the source.

There are different techniques for migrating the seismic data. The scope area of this study is Magruder fault near Allendale, SC (courtesy of the University of South Carolina). The study is conducted to produce the seismic image by applying different migration algorithms, and the results will be compared in time to investigate the efficiency of each algorithm in improving the accuracy of seismic images in reflecting the correct position of the Magruder fault. This study is limited to display the migrated section in time due to different reasons. One reason is that there is a limitation on the accuracy of the velocity modeling by picking the different velocities. Therefore the migration in depth is not entirely accurate, and it may cause some distortion in seismic images. The second reason is that the interpreters prefer to have both sections (unmigrated and migrated) in time to be able to compare them as long as lateral velocity variations are moderate (Yilmaz, 1987).
CHAPTER II
SITE GEOLOGY AND FIELD SURVEY
GEOLOGY OF THE AREA

In this study, Allendale, South Carolina is our zone of interest, and the sediments of this area are deposited within the Dunbarton and Riddleville basins (Figure 1).

The Dunbarton Triassic basin is located in the third southern part of Savannah River, SC. According to Daneils et al. (1983), western end of Dunbarton Basin (DB) is 20 km from the eastern end of Riddleville Basin (RB). Also, DB is considered as an asymmetric graben with possible normal faults on its northwest and southeast sides (Marine, 1974 and Siple, 1974).

The Riddleville Triassic basin is buried beneath coastal plain sediments of Cretaceous and Tertiary age (Herrick and Vorhis, 1963) (Figure 2). The lower basin fill is a sequence of interlayered basalt and clastic units (Petersen et al. 1984).

Magruder Fault (MF) is one of the most prominent south-east trending features of the RB. It is interpreted as a border fault and is aeromagnetically inferred as the northern boundary of RB (Zietzet al., 1980). MF dips about 50º S to a depth of at least 8 km and appears to intersect or merge into Augusta Fault (AF) at 10-12 km depth (Petersen et al., 1984). MF resembles to splay thrusts off decollements in the Valley and Ridge sedimentary province (Harris and Milici, 1977) and may have formed along a pre-Triassic thrust splaying off AF (Peterson et al., 1984) (Figure 3).
Figure 1. From Snipes et al. (1993).

Figure 2. From Petersen et al. (1984).
We regenerated the geological map of Allendale, South Carolina, based on the DEM data, aerial photography and DRG of Allendale, geological datasets of South Carolina and GPS points of source stations along the seismic line (Figures 4 and 5). We utilized the information provided by USGS, South Carolina Department of Natural Resources and Topo Quest.
Figure 4. Geological Map of Allendale.
Figure 5. Geological Map of Study Zone.
FIELD SURVEY

In this study, data acquisition is made along the State Road 304, SC. The instrument of recording seismic data is Geometrics, and the source of energy is the hammer and plate. Each trace is the result of stacking six shots. Source and receiver positions, the distance between receivers and distance between shots have been set up as the following (Figure 6). The source is moving every two receiver spacing to generate a CMP data.

![Figure 6. Survey parameters.](image)

Common mid-point (CMP) is the halfway point in the travel of a wave from a source to a flat lying reflector to a receiver (Figure 7), and fold is the measurement of the redundancy of CMP data (http://www.glossary.oilfield.slb.com).

Also, the subsurface display which is CMP spread (green lines) defined by “shotpoint location vs. receivers” is shown in Figure 8 on which we can see that the central source has the broadest range of fold number.
Figure 7. CMP.

Figure 8. Subsurface display.
CHAPTER III
BACKGROUND
SEISMIC REFLECTION DATA ACQUISITION

SEISMIC SOURCES

The acquisition of seismic reflection data depends on the various parameters, like if the data was obtained from land or marine, the accessibility of the survey area, geologic problem and the depth which needed to be investigated, etc. (Sheriff and Geldart, 1995).

We know that seismic methods are based on the transmission of the elastic waves through the subsurface. The elastic waves are produced by the seismic source and they are received by an array of the detectors. The most conventional seismic sources on land are explosive while in the marine areas, sparker or air-gun is used (Upadhyay, 2004). For most shallow seismic reflection, a hammer is an excellent source if the ground surface is relatively hard and it can create the disturbance at the point on the surface that propagates through the Earth (Burger, et al, 2006).
DETECTORS

The detectors are the sensors which are designed to record the motion produced by the seismic reflection sources, over land or offshore. The detector which is used over land is called “geophone,” while “hydrophone” is used at sea.

The geophones consist of the cylindrical magnet with a circular hole in its outer section, which holds a coil suspended by flat spring. The geophone is fixed firmly to the ground; this makes the coil remain immobile while the magnet goes along with the ground motion. The electrical voltage is produced due to the movement between coil and magnet, which is proportional to the velocity of the coil concerning the factors including the radius of the coil, the intensity of the magnetic field, etc.

The response of the geophone is calculated as the function of the frequency. The ideal feature of a geophone is a flat response for the complete frequency range of the signal.
SURVEY

The geometric arrangement of the geophones relative to the source is called the “spread” which is usually along a straight line, and may have different types: for the “in-line” spread, the source is located anywhere along the spread and preferably far from the nearest receivers due to the presence of high-level seismic energy. The “end-on” and “split” spreads are the special cases of “in-line” spread: the seismic source is located the one of the spread for the former and the center of the spread for the latter. In “cross” spread, the receivers are arranged along a line in one direction and the sources along a line perpendicular to the receivers. In “L” spread, the source is placed at the intersection of a series of geophones located in two perpendicular lines that form an L shape.

These arrangements are selected according to the anticipated geometry of the reflector. The spread length defines the effective depth of penetration, longer spread length allowing mapping of deeper reflectors.

The seismic record used in this study starts at zero time, and hence contains the arrivals from direct, air and surface waves as well as primary reflections, refractions, diffractions and their multiples.
The air, direct and surface waves have straight-line travel-time trends, which can be identified and differentiated from each other by their distinct velocities (Figure 9). Our objective is to isolate the primary reflection events (from their multiples, potential diffractions and other types of seismic arrivals) and increasing the visibility of these events by a series of image processing steps, attenuating the noise and improve the signal to noise ratio.

Figure 9. Distance versus arrival time graph for various types of waves (source; after Anstey 1970, p 80. © 1970 Gebrüder Borntraeger, Berlin Stuttgart).
DATA PROCESSING

Data processing (Figure 10) is one of the critical steps in producing the meaningful data based on minimization of the artifacts in the seismic records related to the procedure of recording the seismic data including the recording instrument and the surface on which the survey has been done to generate an accurate image of the subsurface. Through processing, the raw (unprocessed) seismic data taken in the field are turned into an interpretable image.

The main aims of the seismic data processing are:

▪ Improve the signal-to-noise ratio (by stacking)
▪ Isolate the true (reflection) signals from multiples and surface waves
▪ Obtain a higher lateral resolution (by process of migration)
▪ Obtain a realistic image (by geometrical correction)
Figure 10. Flowchart of processing the seismic data.
DEMULTIPLEXING

The field data are mostly time-sequential, which it means that the first sample is recorded for each channel before recording the second sample for the number of the channels. In the demultiplexing process, the data would be rearranged to trace-sequential which means that all data for the first channel would be recorded before the data for the other channels. SEG A and SEG B formats are multiplexed, SEG Y is a trace sequential format, and SEG D can be either way (Morton-Thompson, 1999).

On the better term, if recorded data in multiplexed format would be (scan sequential format, SEG-A, B format):

A1B1C1A2B2C2A3B3C3

Then the demultiplexed data in trace sequential format, SEG-Y, D would be:

A1A2A3 B1B2B3 C1C2C3

(Samples separated in a time sequence for each of the (3) geophones).

TRUE AMPLITUDE RECOVERY

In this step of processing, which is also known as “gain corrections” and “scaling”, the primary goal is to process the raw field data to obtain the reflection amplitudes. This process is also known as the true amplitude information which would compensate the attenuation caused by spherical divergence, transmission loss, radiation
pattern, and absorption, by adjusting the amplitude of the data. The unit of scaling corrections is defined by decibels per second. A decibel is a unit of scale defined in acoustics as $20 \cdot \log(A/B)$ where $A$ and $B$ are the amplitudes (Upadhyay, 2004).

There are different types of scaling as geometric divergence, exponential, time power, and AGC. To recover the loss caused by geometric divergence, the following correction is applied:

\[
\frac{E}{4\pi r^2} \\
A \propto E^{1/2} \\
A / 4\pi r
\]

The above equations show that energy intensity loss is proportional to the inverse square of the distance whereas the amplitude is related to the energy by its square root.

**Exponential:** This scaling function works by multiplying the traces by an exponential function of the form $= t \cdot e^{(nt)}$ which approximates the effect of absorption with some transmission loss.

**Time Power:** This type of scaling works by multiplying the traces by a power function of the form $= t^{(n)}$.

**AGC:** Automatic Gain Control AGC is the most commonly used scaling type. This step enables amplitude recovery along the trace and across the trace. AGC works by two norm equalization function of L1 or L2 by calculating the average absolute amplitude of all trace samples within a moving AGC Window. The sample at the center of the window is then multiplied by (scale/average).
The whole process moves down one sample and starts again. The muted areas at the start of a trace are avoided. The calculation starts with the first non-zero sample. At the beginning or end of the trace, the scalar used for the center sample in the AGC window for all samples before it (to the start of the trace) or all samples after it (to the end of trace). AGC works on a trace-by-trace basis.

In equation form: L1 norm equalization: \[ \frac{1}{\text{Window}} \sum_{\text{Window}} |x(t)| \]

L2 norm equalization: \[ \frac{1}{\text{Window}} \sum_{\text{Window}} x^2(t) \]

EDITING

Editing involves the removal of traces that are either dead (faulty receiver) or contains excessive noise (bad data). Such traces can set to zero, removed or replaced with interpolated traces (Bacon et al., 2003). High amplitude traces which would be possibly considered as the noise, are set to zero. After editing, the output would help seismologists to know more about the type of the attenuation needed to be applied for further processing.

CORRELATION

Cross-Correlation

The cross-correlation function is a statistical measurement to detect the similarity between two data sets. The cross-correlation function of two data sets of \(x_t, y_t\) is defined by:
\[ \phi_{xy}(\tau) = \sum_k x_{k+\tau} y_k \] where \( \tau \) is the displacement of \( x_t \) relative to \( y_t \). The lag is considered positive if the first function \( x_t \) is shifted to the left concerning other function \( y_t \). Cross-correlation in a Fourier domain is equal by multiplication of amplitude spectrum and subtraction of phase spectrum. In addition, unlike convolution, cross-correlation is not commutative, the output depends on which array is fixed and which is moved (Yilmaz, 1988).

Autocorrelation:

When the data set is being correlated with itself for a variety of time lags, is called autocorrelation by using the formula:

\[ \phi_{xx}(\tau) = \sum_k x_{k+\tau} x_k \]

The autocorrelation is a useful tool to detect the repeating periods within signals in the presence of noise. In autocorrelation function, the right shift is equal to left shift which would categorize it as a symmetrical function. The autocorrelation function is often normalized so that its maximum value at zero-time shift is 1. Both auto- and cross-correlation functions are necessary functions to suppress the effect of multiple reflections while applying the predictive deconvolution.

**STATIC OR DATUM CORRECTION**

The static correction is often required to put shots and receivers on a flat datum plane and to correct the near-surface velocity anomalies beneath the source or receiver (Kearney and Brooks, 1991). This process involves setting up a datum on which to
source and receiver to one common reference level, and then adding or subtracting the incremental time. The reference velocity would be same as the upper layer. The reflections usually come from the deep layers, and the upcoming energy path is mostly vertical. So the change in travel time is the main parameter in the static correction due to elevation (Figure 11):

\[ \Delta t = \frac{h_s}{v_1} - \frac{h_r}{v_1}, \]

which it means that the seismic record is shifted in time by a value \( \Delta t \).

The shift in time will happen if there is anomalous velocity underneath a source or detector if the weathering layer thickness changes significantly. This residual static correction is applicable for the near-surface velocity variations which cause some static distortion (Figure 12). So the equation in which we can find the shift in time is given as:

\[ \Delta t = \frac{h_a}{v_a} - \frac{h_a}{v_1} = \left( v_1 - \frac{v_a}{v_1} v_a \right) h_a \]
The first layer is frequently weathered, and it has various thickness and velocity. It is moreover ineffectively consolidated and hence is a weak transmitter of seismic energy. In exploration, it is common to drill a borehole through the weathered zone into the upper layer. Drill estimation builds up the thickness at the shot point. By having a seismometer at the surface over the shot, one can assess the velocity of the weathered zone.

DEMULTIPLE

The interpretation of recorded primary reflections point is sometimes complicated due to the strong multiples from the upper interfaces. The multiples can mask or interfere with underlying primary reflections which can have weaker amplitudes. The demultiple process attempts to remove the effect of the multiples (extra reflection events) for optimal processing and interpretation. This function is mostly used in the marine surveys because the multiples start at water bottom and ray paths of them lie entirely within the water. They have usually higher strong amplitudes which needed to be removed to isolate the primary reflection events.

MUTER

Zero out arrivals like refractions that are not primary reflections. This process is done by applying the muting which subjectively allots the values of zero to the traces through the mute interval. Sometimes “Trail Mute” is applied to zero out the airwave, and “Surgical mute” is also applied to mute the portion of the waves dominated by the ground roll.
BAND-PASS FILTER

Frequency filtering can be used in different forms like band-pass filter, band-reject, low-pass, high-pass filters. All of these filters work with the same rule. Zero-phase wavelet with the amplitude that corresponds to one of the four mentioned filters is constructed. A band-pass filter is considered as commonly used filters because, in any seismic trace, there are some low and some high-frequency noises such as ground roll and ambient noise which needed to be filtered and this filter can be applied in different stages of data processing to attenuate noise outside of the range of reflection events. The band-pass filter performs by following the equation as below:

\[ F_B(\omega) = +1, \ | \omega_1| < |\omega| < |\omega_2|, \]
\[ = 0, \ |\omega_1| > |\omega| \ or \ |\omega| > |\omega_2|. \]

The filter is discontinuous at \( \omega_1 \) and \( \omega_2 \) which can cause some ringing. Also, the band-pass filter is identical to a low-pass filter with \( |\omega_0| = |\omega_2| \), while it is equal to a high-pass filter with \( |\omega_0| = |\omega_1| \).

NOTCH FILTER

The notch filter is applied to remove excessive high-line picked and it might be used to suppress the noise at the given frequency which it means that this filter rejects the narrow band of frequencies. Notch-filter works by removing the unwanted frequencies twice, once with a filter from 0 Hertz to frequency at the start of the notch, with a half
cosine roll-off from pass-frequency, and once with a filter from the frequency at the end of the notch to a Nyquist frequency which is defined as half of the sampling frequency. The two filtered outputs are then added together.

F-K FILTER

Ground roll has small apparent velocity, low-frequency and large amplitudes that can be isolated from the reflection energy by applying the F-K filter (Yilmaz, 1988). This filter allows removing the unwanted waves like ground roll and guided waves by defining a fan-like area in which the components of the waveform are zeroed. This filter command works by performing a 2D Fourier Transform on the seismic data. Each F-K sample of seismic data is calculated and multiplied by the corresponding F-K point of the filter. Then a 2D inverse F-K Fourier transform is applied (Figure 13).

Figure 13. F-K plot. The region passed by array, frequency, and velocity filters is cross-hatched. Apparent velocity Va (Va=f/k=ω/κ). (After Sheriff, 1991).
DECONVOLUTION

Deconvolution is an inverse filtering which enhances the temporal resolution of seismic data by compressing the seismic wavelet (Yilmaz, 1988). Three significant convolution types are defined below:

Spiking deconvolution: This is a standard Wiener Levinson algorithm. The auto-correlation of the design time gate which is a segment of the trace that usually varies with offset is computed. This type of deconvolution shortens the embedded wavelet, and one of the primary purposes of this deconvolution is making the wavelet to a spike. However, it also causes the increment of noise at high frequency.

Predictive deconvolution: This one uses the later segments of the autocorrelation to eliminate some multiples, so in that case, the arrival of an event could be predicted from the earlier events. The desired output is a lagged version of the input. Hence more lags of the auto-correlation are calculated. The later delays are used as the cross-correlation of the input and desired output.

Zero-Phase deconvolution: In this deconvolution type, the time gate excludes the early part of the record which consists of the surface waves. It also eliminates that part of the record that ambient noise is dominant. In this case, the spiking operator is derived as it was explained above, then we calculate the Forward Fourier Transform and consecutively the amplitude and phase spectrum, after that we put the phase spectrum to zero and perform an Inverse Transform, so deriving the zero-phase equivalent of the minimum phase operator The zero-phase operator is then convolved with the data.
CMP SORTING

If we transform the seismic data from shot-receiver to midpoint offset coordinates, it means that we sort the common midpoints (CMP) which require the correct geometrical information. In other words, the common middle point shared between numerous source and receiver pairs is called CMP. The measurement of the redundancy of common midpoint on the seismic data is called fold which is also defined as the coverage percentage: single-fold = 100% coverage, fourty-fold = 4000% coverage.

\[
\text{Fold} = \frac{\text{Receiver Spacing} \times \text{Number of Receivers}}{2 \times \text{shot spacing}}
\]

UPHOLE STATICS

An uphole survey is utilized to find the weathering layer velocity. A borehole is bored that goes beneath the weathering layer; a few geophones are set at different known depths within the drilled pit. The geophone locations should traverse the weathering and sub-weathering layers. A shot is discharged at the surface close to the hole and the direct travel times to the geophones are recorded, which vary as a function of:

\[
t_D = 2E_D - (E_s - D_s) - (E_R - D_R) / v_b - t_{UH}
\]
where $E_D$ = datum elevation, $E_s$ = surface elevation at the shot station, $E_R$ = surface elevation at the receiver station, $D_s$ = depth of shot hole beneath the shot station, $D_R$ = depth of shot hole close to the receiver station, $v_b$ = bedrock velocity and $t_{UH}$ = uphole time.

NMO/DMO CORRECTION

In reflection seismology, normal move-out (NMO) results from the increasing source-receiver distance (offset) delaying the arrival time of the reflection from a horizontal reflector (http://www.glossary.oilfield.slb.com/Terms/n/nmo.aspx). The relationship between arrival time and offset is hyperbolic, and it is considered as a standard feature a seismologist uses to decide whether an event is a reflection or not (Sheriff and Geldart, 1995). For dipping reflectors, the moveout effect is recognized as a combination of increasing offset and dipping interface, giving dip moveout (DMO).

The normal moveout relies upon a complex combination of factors like the velocity over the reflector, offset, a dip of the reflector and the source-receiver azimuth about the dip of the reflector (Yilmaz, 2001). For a flat, horizontal reflector, the travel-time equation is:

$$t^2(x) = t^2(0) + \frac{x^2}{v^2}$$

where $x = $ offset, $v = $ velocity of the medium over the reflecting interface, $t(0) =$ travel time at zero offsets when the source and receiver are in the same place.

The NMO correction is calculated by the difference between $t(x)$ and $t(0)$

$$\Delta t_{NMO} = t(x) - t(0)$$
\[ t(0) \left\{ \left[ 1 + \left( \frac{x}{v_{\text{nmo}} t(0)} \right)^2 \right]^{1/2} - 1 \right\} \]

The NMO equation for dipping reflector would be:

\[ t^2(x) = t^2(0) + 4h^2 \cos^2 \phi \frac{\ell}{v^2} \]

where \( \phi \) is the reflector dip, \( 2h \) is offset and \( t_0 \) is the two-way zero-offset time at midpoint location \( y_n \).

The relationships between \( (y_n, t_n) \) coordinates of the NMO corrected data and \( (y_0, \tau_0) \) coordinates of the DMO corrected data are given by the following equation:

\[
\begin{align*}
y_0 &= y_n - \frac{h^2}{tn \cdot A} \left( \frac{2 \sin \phi}{v} \right)^2 \\
\tau_0 &= \frac{t_n}{A}
\end{align*}
\]

where

\[ A = \sqrt{1 + \frac{h^2}{tn} \left( \frac{2 \sin \phi}{v} \right)^2} \]
VELOCITY ANALYSIS

In velocity analysis, we estimate the stacking velocities and interval velocities. Normal move out is the basis to find the velocities from seismic data, because amount of NMO which is needed to be removed help us to maximize stacking primary reflections and to identify the lithology of the subsurface layers (Garotta and Michon, 1967; Cook and Taner, 1969; Schneider and Backus, 1968; Taner and Koehler, 1969).

Computed velocities can be utilized to modify the NMO so that reflections are lined up in the traces of a CMP gather before stacking.

Equation $t_{st}^2(x) = t_{st}^2(0) + x^2 / v_{st}^2$, where $v_{st}$ is the stacking velocity determined by finding the best fit of travel time curve $t_{st}(x)$ on a CMP gather to a hyperbola within the spreads length (Yilmaz, 1988) shows a line on $t^2(x)$ vs. $x^2$, the slope of this line is $1/v_{st}^2$ and the intercept value $x = 0$ is $t(0)$.

To detect the stacking velocity for an event, a straight line connects the points that correspond to the event, in this case, the inverse of the line’s slope is equal by the square of the stacking velocity.

The $t^2-x^2$ velocity analysis is one of the trustworthy ways to approximate the stacking velocities however the precision of that is dependent on S/N.

Claerbout (1978) suggested a technique to investigate interval velocities manually from CMP gathers. In this method, initially the slop along the line which is tangential to reflections
of the interval of the interest at top and bottom should be measured, the interval velocity between two reflectors would be equal to the square root of the product of two slope values.

\[ V_{\text{int}} = \left[ \frac{(t_2 V_{\text{rms}2}^2 - t_1 V_{\text{rms}1}^2)}{(t_2 - t_1)} \right]^{\frac{1}{2}} \]

where

- \( V_{\text{int}} \) = interval velocity
- \( t_1 \) = travel-time to the first reflector
- \( t_2 \) = travel-time to the second reflector
- \( V_{\text{rms}1} \) = root-mean-square velocity to the first reflector
- \( V_{\text{rms}2} \) = root-mean-square velocity to the second reflector.


The other possible way for velocity analysis is studying the constant velocity on the CMP gather after frequently applying NMO correction, using a range of constant velocity. Sometimes at the smaller velocities, the NMO corrected gathers are overcorrected, and it would be under-corrected at high velocities. When the event is flat on the NMO gather, the corresponding velocity would be considered as the stacking velocity for that event, and if we do the same process for different events, then we can build up a velocity function which corresponds to NMO correction of this gather.

We need to get the proper velocity function to get the best quality seismic image, so stacking velocities are usually evaluated based on the stacked event amplitude and continuity from stacked data with a range of constant velocities (Yilmaz, 1988), which it means that stacking velocities are picked from constant velocity stack (CVS) panel by
selecting the velocities that yield the best stack response at the chosen event time. In other words, this method is based on lateral continuity of the stacked traces. However, resolution of velocity estimates decreases by depth due to the reduction of moveout in depth.

We have to consider some issues while we choose the CVS, as the range of the velocities needed to stack and the spacing between the stacking velocities.

When we are selecting the range of velocities, we should know that dipping events may have high stacking velocities, also for selecting the spacing, we should examine the accretion of equal $\Delta t_{\text{nmo}}$, this would avoid oversampling of high-velocity events. The reliable way to choose $\Delta(\Delta t_{\text{nmo}})$ is to pick it, so that move out the difference between contiguous trial velocities at the maximum offset to be stacked is approximately $1/3$ of the dominant period of the data. Muting causes the short maximum offsets for shallow, while in-depth data have a sizeable dominant period (Doherty, 1986, Personal Communication).

The recent method is useful for the target zone containing the complicated structures; this would enable the seismologist to choose the stack with best the possible event continuity.

The velocity spectrum is another method of velocity analysis based on the cross-correlation of the traces in CMP gather. This process is convenient for the data with multiple reflection problems.
THE VELOCITY SPECTRUM

The velocity spectrum display is calculated by finding how well a given hyperbolic event matches real events on the central CMP gather (Taner and Koehler, 1969). First, the coherence for a range of trial hyperbolas for velocities between two values is computed at the constant intervals. The maximum amplitude of the coherent spectrum is observed where the hyperbola fits most at high amplitude seismic event. This measurement of the coherence is called semblance which is defined by the ratio of the energy of the stacked traces and available energy for stacking (Upadhyay, 2004). There are different methods for showing the semblance, but most of the times, the color contour display is used which connect the velocity and time values that highlight stacked events. The blue contours represent low semblance and red ones represent high semblance areas. In this display, velocity is on the horizontal axes and zero-offset time is on vertical axes. The semblance often calculated for the central gather of a group of more than ten to reduce noise. If we increase the gather numbers, the computation time would be increased, and it would also filter out geological changes. Choosing the broader peaks, in the deeper part of the section causes the reduction in the resolution.

There are different factors which are affecting the velocity estimates including:

1) Spread length: The large spread that cover both near and far offsets provide higher resolution in the velocity spectrum.

2) Stacking fold: High-fold number increases the resolution of seismic images by picking the accurate peaks while expanding the computation, so in some cases, it is better to choose lower fold number to save calculation.
3) S/N ratio: The precision of the velocity spectrum is limited when the signal to noise ratio is weak.

4) Muting: Muting reduces the number of the fold in the stacking process and has the adverse effect on velocity spectrum that causes debilitating the high amplitudes that are within the mute zone, this problem would be solved by multiplication of stacked amplitudes and scale factor.

   \[
   \text{Scale Factor} = \frac{\text{Actual abundance}}{\text{Number of live traces}} \text{ in the mute zone.}
   \]

5) Time gate length: The time gate is chosen between one-half and one time the dominant signal period, because if a small time gate is determined, then it drastically increases the cost of computation and if the significant time is selected, then the velocity spectrum has the problem in the accuracy of the measurement with respect to time.

STACKING

   If we sum the traces to enhance the signal-to-noise ratio, stable noise and improve seismic data quality, we are stacking the data. Traces from different shot records that have the common reflection point (CMP) data are stacked to form a single trace during seismic processing.

   Stacking reduces the amount of data by the fold number.

   We can set the stack with or without normalization. We can normalize the stacked data by $1/N, 1/\sqrt{N}$ where $N$ is the number of traces stacked. We can also normalize the stacked data by the median of the stacked traces. In some cases, that we have the noisy data set, we can also normalize the stacked data by using “Alpha Trim Mean” techniques which work by checking the
samples of several adjacent traces for each time step. The samples are examined, and the mean and the median values are calculated for that time step. Then the result is calculated as:

\[
\text{Result} = \text{Median} + (\text{Mean} - \text{Median}) \times \text{Alpha Trim Factor}
\]

**MIGRATION**

Migration is defined by correcting the reflection arrivals from non-horizontal reflectors, such as dipping layers, synclines, faults. Migration process follows seismic velocity modeling. There are two different migrations in the aspect of time and depth.

Time Migration: Process of collapsing diffractions and migrating the reflections from dipping boundaries to their actual positions in time.

Depth Migration: If the velocity structure of the earth is well known, then we can compute the shape of diffraction for the selected velocity model and finally migrate the data according to the diffraction shape.

There are three significant algorithms for migrating the seismic data in time.
F-K (STOLT) TIME MIGRATION

This algorithm works by depth conversion and depth to frequency FFTS; then the sample is reordered to trace no. vs. frequency. After that FFT to wave number vs. frequency are calculated, and seismic data is rearranged to frequency vs. wave number then migration is applied in which each constant wave number "trace" to lower frequency is mapped, and then the data is reordered to wave number vs. frequency, after this step the inverse FFT to trace no. vs. frequency is found and finally, the data is rearranged to frequency vs. trace no., inverse FFT, depth to time conversion and output.

The using velocities are RMS values at various CMP's along the line. Time to depth conversion can be done using the RMS velocities or the so-called Stolt pseudo–velocities. These are derived by calculating the time-averaged, squared RMS velocities. The square root of this quantity is then the Stolt velocity. In general, they are lower than the RMS velocity – about 95% in most cases. The W factor can range from 0 to 2 – the usual value will be close to 1.0. A value of W < 1 implies under–migration of steep dips and W > 1 implies over–migration of steep dips. Or to put it another way (Yilmaz, 1980) W < 1 compresses the impulse response on steep dips and W > 1 opens it up.
KIRCHHOFF TIME MIGRATION

The constant velocity semicircle construction is utilized to migrate a hyperbolic diffraction curve to its migrated position where the semicircles impede. An optional technique would be to sum the amplitudes along the hyperbola and place the summed amplitude at the pinnacle. This is called diffraction summation, diffraction stack or more generally Kirchhoff migration. The zero-offset segment is determined as the superposition of diffractors at each time sample (Huygen's Principal). The diffractors interfere to produce the coherent events, and individual diffractions might be noticeable at discontinuities. The amplitudes of seismic data are summed along the series of hyperbolas controlled by the velocity field. Most high amplitudes will appear at the migrated event. The amplitude will be insignificant.

FINITE DIFFERENCE MIGRATION

In Finite Difference migration, the wave equation is integrated by using the finite differences technique to move seismic waves reversely into the subsurface. This algorithm works by using steep-dip algorithms according to the continued fractions expansion to the scalar wave equation (Yilmaz, 2001). This approximation provides a theoretical dip accuracy up to 45 degrees, and it can handle the steeper dips angle without severe amplitude distortions or phase errors if we increase the degree solutions.
The solutions in finite difference migration can be used to implement downward continuation and handle the laterally varying velocity fields based on the different types of solution of 15 degree (fast), 45 degrees (more accurate) or 65 degrees (most accurate and slowest) that are considered as the implicit solutions developed after the method of solving the wave-equation by Claerbout. The further frequency domain solution which provides a 65 degrees dip limitation sometimes called FX or omega-X migration. The runtime increases with dip accuracy and the actual dip limits migrated is a dependent parameter to velocity, depth step, trace spacing and sampling interval. One of the advantages of this finite difference migration is the applicability of it to both steep dip depth migration solution and dip-limited time migration solutions.
The interpretation of seismic data means to transform the seismic data into geological information. Seismic analysis is an art. Accurate and precise knowledge of highly developed technology and the appropriate understanding of what happened beneath the earth are the requirements for the last step of the seismic method.

The two necessary parameters of seismic interpretation are 1) the properties of reflection and 2) the shape of the reflection. Seismic interpretation consists of three fundamental categories: structural, stratigraphic and lithological.

Continuity is one of the criteria on the seismic section, which is the seismic arrival of a reflection and can be distinguished on successive traces, by little changes in arrival time from one trace to another one. These reiterated pulses cause an alignment which has the continuity that can be followed. Continuity of the reflection is the coherence of two geological units one following promptly on top of the other, and the reflections created at the contact of these units where there is an interface.

Correlation is other essential criteria on the seismic section which is the identification of the pattern that might be detected by its shape, length, amplitude and also the properties of each reflection events and the spacing between them.

Correlation is defined by the shape of each signal and the sequence of reflection events and spaces with them. Among the two mentioned parameters, the sequence of the reflection event is a reliable base for correlation; contrarily the spacing between reflection
events is less valid because unconformities and any changes in seismic velocities could affect the spacing of reflections.

We can separate the seismic reflection events when the sequence is thickening or changing, at the unconformity or when the faults cut across the layers and cause the overlapping of reflections.

STRUCTURAL SEISMIC INTERPRETATION

We usually process the seismic dataset and use it to identify the structural features of the subsurface geology.

Discontinuities which strata on both sides are mostly considered as fault truncations or unconformities. When there is the discontinuity at a high angle on one side and a featureless area on the other side, which is considered as a fault, intrusive, or diapiric contacts (http://www.geo.cornell.edu/geology/faculty/RWA/structure-lab-manual/chapter-11.pdf)

STRATIGRAPHIC SEISMIC INTERPRETATION

Stratigraphic seismic interpretations enable the seismologists to create chronostratigraphic system according to the patterns of reflection events in the seismic dataset. The reflection terminal patterns and the continuity of the reflections are the essential properties to guide the seismologists to distinguish the boundaries between subsurface regions with more certainty.
LITHOLOGICAL SEISMIC INTERPRETATION

Lithological seismic interpretation concentrates on alterations in porosity, fracture rate and lithology from the obtained seismic data. Lithological seismic interpretation can give us an important clue in finding the hydrocarbons by direct hydrocarbon indicators.

Building a correct and exact model of subsurface geology leads us to eventually find the favorable hydrocarbon reservoir characteristics. However, an accurate model cannot be constructed only based on the seismic data. A realistic model that diminishes the cost of exploration and development risk in an oil or gas field must be made based on the comprehensive analysis and interpretation of all accessible information and data provided to the seismologist.

Petrophysical, historical and observer logging data should likewise be considered when constructing an accurate model of an expected target. We should always keep in mind that each bit of information can offer significant bits of knowledge and enable us to construct the reservoir model; there is a need to restrain using the unnecessary data in our model to avoid distorting our dataset with excessive noise.
CHAPTER IV
METHODOLOGY
SOFTWARE

On this study, we used the “Vista Desktop Seismic Data Processing” which is considered as one of the quickest selling seismic data processing software among the other industry software. It conveys proven algorithms for ideal quality control of seismic information in the field or the workplace. VISTA provides the useful algorithms to land, marine and VSP processing and offers the instinctive flows for handling the seismic data processing. By exploiting VISTA's adaptability, unwavering quality, and convenience, the processor increments their productivity while augmenting the constancy of their outcomes. VISTA is accessible in four modules, 2D/3D Field QC, 2D/3D Field Processing, and 2D/3D Full Processing, and VSP, which it means, this software can be used from the very beginning of the acquisition quality control to the final step which would be completing the processing the 2D and 3D datasets.

VISTA 1.0 was created in 1984 on a 10 MHz PC 8086 with 10 Mb of memory running under DOS 3.1, and it was developed in 1985 in Canada. There have been many changes from that point forward, 12 generations of PC's (8086 to Xeon and Pentium 4, Dual-Core Xeon) and eight generations of Operating systems (OS).

(https://www.software.slb.com/products/vista)
FLOWCHART

This Flowchart is tailored for this project by picking the most relevant process from a general purpose flowchart designed for complicated data sets (Figure 14).

Field Tapes(records)
Preprocessing
F-K Filter
Deconvolution
Velocity Analysis
NMO Correction
Stack
Migration

Figure 14. Processing flowchart designed for this project.

For preprocessing of the seismic data, these were selected among the other functions showed in Figure 10 for this project.

- Field Geometry
- Trace Editing
- True Amplitude Recovery
- Scaling/Gain Correction
- Band-Pass Filter
- AGC (Auto Gain Control)
- Muter
The main Seismic Window Display shows the multiple displays of seismic data and header values in different ways. In this project the “Seismic Window Display” is shown as below, displaying two shots from a raw data set (Figure 15).

![Seismic Window Display](image)

**Figure 15. “Seismic window display” for raw data.**

**GEOMETRY**

According to the seismic layout parameters, shown in Figure 16 for the unprocessed seismic data, the geometry information has not been set up yet, so we need to edit the cells regarding the geometry information (Figure 17) provided by University of South Carolina by assigning the correct values to the both geophones/receivers coordinate systems, elevation, shot point number, field record number and field station number, in order to calculate the values for offset, CMP and fold number.
Figure 16. Seismic layout parameters for raw data.

Figure 17. The source-receiver geometry data.
TRACE EDITING

In trace editing, we remove the excessive noise by applying the “Kill” function where it set all unwanted samples to zero. The display of some sequential trace numbers which needed to be removed is shown in Figure 18.

![Killed Traces - Notepad](image)

**Figure 18.** Display of the trace numbers needed to be removed.
TRUE AMPLITUDE RECOVERY

For true amplitude recovery, the exponential time power is applied by multiplying the traces by a power function of the form $t^n$ (Figure 19).

![Exponential Time Power](image)

**Figure 19. Exponential time power scale.**

SCALING

We apply Mean scaling in which each trace is scaled so that the mean amplitude is the same for all traces and also RMS scaling in which each trace is scaled so that the RMS energy is the same for all traces. We scale the seismic data in this step to check which scaling type provides a clearer display of the seismic data.

F-K FILTER

In this step, we design the F-K fan area as the following figure (Figure 20) according to F-K spectrum analysis. We select the output of mean scaling as the input for this stage and assign the values of (1), to power amplitude which attenuates the trace amplitudes, (7) to smoother traces which is a smoothing parameter for the wave number. Using a high number of smoother traces makes the output more continuous from a trace to trace.
We also assign the value \( t \) (5) to a smoother frequency which is the smoothing parameter for the frequency dimension, and it makes the output traces look cleaner or smoother when choosing a high number.

![Figure 20. F-K plot, defining the pie slice filter.](image)

**DECONVOLUTION**

We apply the three different types of the deconvolution including zero-phase, spiking and predictive deconvolution to the output obtained from the F-K filter. We assign 40 ms to the operator length which is the length of deconvolution operator (ms) and is recommend to be chosen as one and a half to two times the length of the average wavelength of the data. We also give the value 1 to the pre-whitening percentage which is equal to the amount of pre-whitening to add to the zeroth lag of auto-correlation. We apply the deconvolution to the time gate window (Figure 21).
Beside the deconvolution parameters, we explicitly define the prediction lag for predictive deconvolution which is the lag between multiples or as the period of ringing, and it is being used as de-reverberation deconvolution by 2nd zero crossing of autocorrelation.

**Figure 21. Display of time gate window.**

**BAND-PASS FILTER**

In this study, we apply the Ormsby band-pass filter which is a filter of trapezoidal shape that rejects below f1 and above f4 according to the frequency vs. amplitude analysis (Figure 22).
AGC

After band-pass filter, in this step, we apply AGC which is the most commonly used scaling type and enables amplitude recovery along the trace and across the trace. We select the command parameters as following:

Norm Equalization – L1 as the type of AGC calculation.
Length of AGC Window – 500 ms, which is the length of the moving AGC window.
Output Scale Factor – 1 that is equal to the scale factor that we apply to the entire output amplitude level.

MUTE

In this section, a mute function is designed to mute the airwave and part of the ground rolls, and the parameters are defined as below (Figure 23):
VELOCITY ANALYSIS

Before picking velocities, we generate a semblance, common offset stack, and constant velocity stack (CVS) data sets from our CMP gather by using a flow such as a Figure 24. In this flow, the raw data files are used as input with the CMP bin selection spreadsheet set up for 3 CMP’s: 2(CMP from), 180(CMP to), 20 (CMP increment) as center 2D bins.
In this flow, the zero-phase deconvolution type is chosen for the further processing, and Ormsby filter and AGC parameters are defined as Figure 25.
CVS works by taking the entire input data set, in CMP sort order, and applying the steps of NMO, CMP Stack. The NMO is performed with the first velocity from the velocity function list and will be constant for the entire trace length. The output traces for one velocity corresponds to the CMP Stack of the input traces. The whole process is then repeated for the next velocity in the velocity function list. The total number of output traces is equal to the (number of velocities) times the (number of stack traces) in the input data set. For setting the parameters of CVS, we insert 1 for dead trace number to separate one CVS "panel" from the next. In the velocity function list, we set the start velocity to 1500 m/s, end
velocity to 2500 m/s and velocity increments to 120 m/s for the CVS display. We also check the mute data check box, to supply a stretch mute 20%.

In offset sort/stack the parameters are defined as the following figure in which all the traces from a "zone" of several CMP's stacked by common offset (Figure 26).

![Offset Sort/Stack - Records](image)

**Figure 26. Offset stack parameters.**

The Semblance command works by calculating the "Semblance Function" for the whole set of input traces at each sample position across the hyperbola characterized by the current velocity. The whole procedure is repeated for the following velocity, so the quantity of output "traces" is equivalent to the number of different velocities defined. Each output "trace" has samples, whose values are the values of the semblance function at that specific time. The semblance function is defined by:
\[ S = \frac{\sum_{\text{time}} t \left( \sum_{\text{Traces}} x_{-1} \, dx t \right)^2}{N \left( \sum_{\text{time}} t \left( \sum_{\text{Traces}} x_{-1} \, dx t^2 \right) \right)} \]

Where \( N \) is the number of non-zero samples after muting. Smoothing equal to 20 Ms is applied separately to the numerator and denominator before computing this semblance quotient. Then, the semblance is set to the power of the parameter Output Power (http://seisex.tripod.com/sulist/suvelan.htm).

The input to Semblance will typically be a set of Common-Offset Stacked traces. Semblance input velocities are defined same as CVS. Mute Ramp is set to 25 samples which are the number of samples to determine the mute ramp at the start and end of the mute and 30% is the percentage that we apply to stretch muting the input data.

**NMO**

NMO function works by reading the velocities. When the input data is 2-D, the velocities are determined by CMP number and (time, velocity) pairs. That is one velocity function at indicated CMP. When the trace is read in from the previous procedure, the software determines its CMP number, and the velocity function is created at this CMP number based on linear interpolation of RMS velocities at two neighboring CMP's. Once the velocity is computed, an RMS velocity is derived for every sample of the trace. The software enables using the standard NMO equation to calculate where each output must come from.
We also check mark the mute velocity inversion which means that the output data is muted, when the velocity inversions occur.

**STACKING**

In this stage, we apply the stacking function to increases signal-to-noise ratio: as offset traces are stacked, the signal is enhanced while the random noise tends to be stable. Stacking also helps the attenuation of multiples from a given set of interfaces. The flow for this stage is shown in the following figure.

![Figure 27. Stacking flow.](image)
In this flow, the zero-phase deconvolution is selected as the input data set and CMP stacking works by stacking traces received in CMP sorted order. A new stack trace is started when a new CMP ensemble is encountered in the input. We set the stack normalization as 1/N which means normalize the stacked data by 1/N where N is the number of traces stacked.

**MIGRATION**

**FK (Stolt) 2-D Post Stack Time Migration**

The command in this stage works by calculating the 2-D Fourier Transform of the input stack traces after undergoing the Stolt’s pseudo-depth conversion. FK points are then mapped to lower frequencies along the lines of constant K according to the published equations (Yilmaz, 1987). The output is then transformed back to the time, and the depth conversion is reversed to give the final 2-D time migrated section. Stolt's original classic paper appeared in 1978 and Chun and Jacewitz wrote their most cited paper on the physical interpretation of the F–K migration method in 1981. The parameters for this stage is determined as W-Factor of 1 which is described by Stolt for variable velocity, the base velocity of 1600, 1610, 1690 m/s respectively for each velocity model, and the trace distance of 4.5 which is the CMP distance.

**Kirchhoff 2-D Post Stack Time Migration**

This algorithm works by the conventional diffraction sum method which is summing energy along diffractions corresponding to the local velocity. The input velocity file contains RMS values at various CMP's along the line.
This algorithm linearly interpolates velocities for each time sample position and then uses those to linearly interpolate a velocity "trace" corresponding to every input trace (CMP). We use 100% RMS velocity. The trace distance is 4.5 m which are the distance between CMP, and the Max Dip Angle is 35 which is the maximum dip angle to migrate, and we define the maximum lateral extent of the migration operator regarding traces by the number of 15.

Finite difference 2-D Post Stack Time Migration

This algorithm works by using the different degree solution which is an approximation to the wave equation. We set the trace distance to 4.5 m for this migration algorithm too, and taper pad is 20 traces which helps to avoid unwanted edge effects, and we apply different types of the solution by choosing 15 degrees (fast), 45-65 degree (more accurate). This will help us to compare the efficiency of each algorithm in enhancing the precision of seismic imaging in displaying the correct position of Magruder fault.
CHAPTER V

RESULTS AND DISCUSSION
In this section, the results of each stage of Magruder fault’s seismic data processing are displayed according to flowchart designed for this project (Figure 14).

**GEOMETRY OF SEISMIC DATA**

The seismic layout parameters after setting the geometry information of data are as the following figure.

![Figure 28. Seismic layout parameters after geometry correction.](image-url)
In order to calculate the correct CMP and fold number, for the bin grid we use the crooked line binning using the Inline spacing as $\sim \frac{1}{2}$ the distance between the receivers (~9.2 m), and we move the spacing by 2.5 m to better center the shots/receivers along the bin grid, we also increase the bin spacing for x-line as 15 to visually be able to see the bin color values on the grid, so the 2-D geometry display would be as the following figure.

![2-D Geometry Display](image)

**Figure 29.2-D geometry display.**
Also, as there is not the considerable difference between the shots and receivers’ elevation (Figure 30), static correction is not required or essential.

**EDITING**

After geometry correction, we can remove the noisy traces by setting them to zero (Figure 31).
EXPONENTIAL TIME POWER SCALING

We use the output of the previous step, as the input for applying the exponential time power in which we can observe increment in the amplitude of the traces (Figure 32).

Figure 32. Before and after exponential time power scaling.
MEAN / RMS SCALING

We can also apply mean and RMS scaling to the result obtained from exponential time power to compare if the application of these types of scaling would provide a uniform image or not (Figures 33 and 34). We see the mean scaling could more efficiently impact the deeper part of the seismic section and also the far-offsets.

Figure 33. Before and after mean scaling.

Figure 34. Before and after RMS scaling.
**F-K FILTER**

In this step, we would apply the F-K filter that we have defined (Figure 20) to output from the mean scaling. So according to Figure 35, the differences between the input and out are significant in the first traces of each shot points.

![Figure 35. Input data vs. difference input, F-K applied.](image)

**ZERO-PHASE DECONVOLUTION**

The result after performing the zero-phase deconvolution shows the clarity.

![Figure 36. Before and after zero-phase deconvolution.](image)
SPIKING DECONVOLUTION

The spiking deconvolution doesn’t show that much difference in seismic image comparing to zero-phase deconvolution (Figure 37).

![Before and after spiking deconvolution](image)

**Figure 37. Before and after spiking deconvolution.**

PREDICTIVE DECONVOLUTION

The result of predictive deconvolution causes losing the information about the probable reflection event (Figure 38).

![Before and after predictive deconvolution](image)

**Figure 38. Before and after predictive deconvolution.**
Since the result obtained from zero-phase deconvolution gives us the more precise image, due to the noise attenuation and amplitude equalization, according to frequency analysis shown in Figure 39, we use the output of its function for the next stages of seismic data processing.

Figure 39-a. Frequency analysis before zero-phase deconvolution for far offset (trace no 96).
Figure 39-b. Frequency analysis after zero-phase deconvolution for far offset (trace no 96).
Figure 39-c. Frequency analysis before zero-phase deconvolution for middle offset (trace no 60).
Figure 39-d. Frequency analysis after zero-phase deconvolution for middle offset (trace no 60).
Figure 39-e. Frequency analysis before zero-phase deconvolution for near offset (trace no 9).

Figure 39-f. Frequency analysis after zero-phase deconvolution for near offset (trace no 9).
ORMSBY BAND-PASS FILTER

In this step, we would see how the Ormsby filter that we have defined in Figure 22 improves the clarity by removing the high frequency and low amplitude events (Figure 40). In addition, the frequency analysis before and after applying the Ormsby filter is shown in Figures 41-a and -b.

Figure 40. Before and after applying Ormsby filter.

Figure 41-a. Frequency analysis before applying Ormsby filter.
AGC

We use the result of previous step (Ormsby-filter) and scale it by applying the AGC (Figure 42) in which we can observe the amplitude recovery that happens along and across the traces. We can also observe the effect of AGC in the equalization of the amplitude by frequency analysis (Figure 43-a, b).

Figure 41-b. Frequency analysis after Ormsby filter.

Figure 42. Before and after applying AGC.
Figure 43-a. Frequency analysis before AGC.

Figure 43-b. Frequency analysis after AGC.
MUTING

As we would like to isolate the reflected wave, we apply the muting to the output of AGC to remove the waves which are not the primary one like airwave (Figures 44 and 45).

Figure 44. Display of different seismic events based on the arrival pattern.

Figure 45. Before and after muting.
CMP SORTING

Since the seismic data are shot-ordered from the beginning of the processing to this stage, we sort them based on CMP before stacking (Figures 46-a and -b). The CMP –sorted data is shown in individual CMP number (a) and continuous CMP sort order (b) before stacking in which TRC in horizontal axis represents the trace numbers.

Figure 46-a. CMP display based on individual CMP number.

Figure 46-b. CMP display based on continuous CMP sort order.
VELOCITY ANALYSIS

By picking the different velocities, we construct more than twenty different velocity models. We select three models (Figure 48-a,b,c) among them which show more uniformity with the geology of the area. For velocity analysis, the semblance technique is used by finding how well a given hyperbolic event matches real events on the central CMP gather. The semblance analysis is shown in Figure 47 for CMP = 142. The contour display is used which connect the velocity and time values that highlight stacked events. The blue contours represent low semblance, and green ones represent high semblance areas. In this display, velocity is on the horizontal axes and zero-offset time is on vertical axes.

As an example, the velocity analysis of each model is shown in the following figure (Figure 49-a,b,c) for CDP = 2, on which the blue is \( V_{\text{interval}} \), red is \( V_{\text{rms}} \), green is previous and yellow is next picking velocity. The figure shows that in the first model, we attempt to pick the \( V_{\text{rms}} \) in short time step while we increase the time step for selecting the \( V_{\text{rms}} \) in the 2\(^{nd}\) and 3\(^{rd}\) models. The significant difference between 2\(^{nd}\) and 3\(^{rd}\) model is that in the 2\(^{nd}\) model we consider only the high velocities and for the 3\(^{rd}\) model, we also include the low-velocity events.
Figure 47. Velocity analysis for 1st model.
Figure 48-a. 1\textsuperscript{st} velocity model.

Figure 48-b. 2\textsuperscript{nd} velocity model.
Figure 48-c. 3rd velocity model.

Figure 49-a. Analysis of 1st velocity model.
Figure 49-b. Analysis of 2nd velocity model.

Figure 49-c. Analysis of 3rd velocity model.
After constructing the velocity model, we can also convert the velocity model to the depth scale in each model by selecting the input data pairs of time/RMS Velocity (Figure 50- a,b,c). In the following figures the black color shows high amplitude and red shows low amplitude:

Figure 50-a. Depth section of 1st velocity model.
Figure 50-b. Depth section of 2\textsuperscript{nd} velocity model.

Figure 50-c. Depth section of 3\textsuperscript{rd} velocity model.
NMO CORRECTION

After velocity analysis, we apply NMO to each velocity model by using the velocity functions that we have selected in our velocity analysis.

**Figure 51.** NMO correction for 1st velocity model.

**Figure 52.** NMO correction for 2nd velocity model.
Figure 53. NMO correction for 3rd velocity model.

CMP STACKED TRACES

In this step, the CMP of seismic traces is stacked for different velocity modeling according to the flow defined in figure 27 on which we can start observing the geological feature (fault) on the seismic image at 160 ms with the dip angle almost 35º south (Figure 54,55,56).

Figure 54. Stacked traces of 1st model, whole record time (right) first 800 ms (left).

Red circle is representative of the geological feature.
Figure 55. Stacked traces of 2nd model, whole record time (right) first 800 ms (left).

Red circle is representative of the geological feature.

Figure 56. Stacked traces of 3rd model, whole record time (right) first 800 ms (left).

Red circle is representative of the geological feature.
TIME MIGRATION

As we have applied three different migration algorithms to three velocity modeling, we can compare the results to see which time migration algorithm on which velocity model gives the accurate and more explicit information about the Magruder fault while we know any minor change in velocity modeling could mainly affect the outcome.

In the 1st velocity model, the image derived from FK-Stolt time migration (Figure 57-a) shows the consistency with the geology of the area. The Kirchhoff time migration produces an image with smoother boundaries but appears to lose the contrast with depth. The FD time migration algorithms seem to create better images of the subsurface, and they show the faults grow most efficiently by the coalescence of different smaller faults (e.g., Peacock and Sanderson, 1991; Willemse, 1997; Cartwright et al., 1995). So as we see the results obtained from the FD migration with different solutions, the fault appears on the seismic image at 180 ms and its shape shows undulation with a dip angle of 30º south and depth of almost 140 m.
Figure 57-a. F-K time migration of 1st model. Blue circle represents the Magruder fault area.
Figure 57-b. Kirchhoff time migration of 1st model. Blue circle represents the Magruder fault area.
Figure 57-c. FD-15° time migration of 1st model. Blue circle represents the Magruder fault area.
Figure 57-d. FD-45° time migration of 1st model. Blue circle represents the Magruder fault area.
Figure 57-e. FD-65° time migration of 1st model. Blue circle represents the Magruder fault area.
Figure 57-f. Depth section of the FD-15° time migrated of 1st model. Blue circle represents the Magruder fault area.
In the 2\textsuperscript{nd} velocity model, while we picked the high-velocity events for constructing the velocity model, the image that we obtained from FK-Stolt time migration didn’t show us the precise shape of the fault. The FD time migration algorithms appear to produce the images with more details about the subsurface. Also, the 45-degree solution provides the more accurate shape of the fault while it doesn’t distort the information of the deeper layers. In addition, in the depth section, the fault’s location is at almost 120 m (Figure 58).

![Figure 58-a. F-K time migration of 2\textsuperscript{nd} model. Blue circle represents the Magruder fault area.](image.png)
Figure 58-b. Kirchhoff time migration of 2nd model.
Blue circle represents the Magruder fault area.
Figure 58-c. FD-15° time migration of 2nd model. Blue circle represents the Magruder fault area.
Figure 58-d. FD-45° time migration of 2nd model. Blue circle represents the Magruder fault area.
Figure 58-e. FD-65° time migration of 2nd model. Blue circle represents the Magruder fault area.
Figure 58-f. Depth section of the Kirchhoff time migrated of 2nd model. Blue circle represents the Magruder fault area.
In the 3\textsuperscript{rd} velocity model, the image that we obtained from FK-Stolt time migration shows us the better shape of the fault, and it also gives us the information about the other probable geological features. The Kirchhoff migration provides the smoother boundaries, but it doesn’t show the dip angle of the fault. The FD time migration algorithms appear to produce the images that provide more information about subsurface, while they tend to split the fault line to small segments, and it seems that the obtained results are over migrated which decreases the depth location of the fault to almost 130 m with the dip angle of 25\textdegree south.

Figure 59-a. F-K time migration of 3\textsuperscript{rd} model. Blue circle represents the Magruder fault area.
Figure 59-b. Kirchhoff time migration of 3rd model.

Blue circle represents the Magruder fault area.
Figure 59-c. FD-15° time migration of 3rd model. Blue circle represents the Magruder fault area.
Figure 59-d. FD-45° time migration of 3rd model. Blue circle represents the Magruder fault area.
Figure 59-e. FD-65° time migration of 3rd model. Blue circle represents the Magruder fault area.
Figure 59-f. Depth section of the F-K time migrated of 3rd model. Blue circle represents the Magruder fault area.

Our trials showed that the depth migration results in distortion of the time-migrated images which is consistent with the geology of the area.

In addition, according to the results of the time migration algorithms on three velocity modeling, we can conclude that the first velocity model gives us the better result as we picked.
the $V_{\text{rms}}$ in short time step. The presence of high-velocity layer in the first 100 ms is due to the presence of coastal plain sediments with the velocity of 2050-2256 m/s. Then we would see sudden decreasing in the velocity due to the presence of Pliocene unit containing the sand and shale in the 1st layer.

The velocity range for sand varies between 1900 m/s-1500 m/s. At 140 ms, we would observe the probable Magruder fault that causes an increase in velocity with depth. The layer of shale would be observable at 410 ms with the velocity range of 2110 m/s-1940 m/s.

At 580 ms, the 2nd layer is observed which belongs to Eocene which means the contact would be noticeable at this time. Sand and shale are also the major materials of this unit and we would see that the velocity range for shale varies between 1963 m/s-2246 m/s and for sand it varies between 1926 m/s -1635 m/s.

As the applied migration algorithms do not create significant improvement in the image quality, we would like to use different and novel optimization algorithms on this velocity model in future in order to improve the existing migration algorithms in the sense of increasing the accuracy of the result while providing more information about the subsurface stratigraphic and structural features.
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VITA

MOONES ALAMOOTI

No1, Brevard Hall, University of Mississippi 38677, University, Mississippi (662) 202-7737 • malamoot@go.olemiss.edu

EDUCATION

M.Sc., Engineering Geology, University of Mississippi, May 2018
   Thesis: A Comparative Case Study of Reflection Seismic Imaging Method
B.Sc., Geology, University of Tehran, July 2010

TEACHING EXPERIENCE

Teaching Assistant in “Rock Mechanics”. Fall Semester of 2013
Teaching Assistant in “Mineralogy-Petrology”. Fall Semester of 2013
Teaching Assistant in “Geophysics” Spring Semester of 2014
Teaching Assistant in “Historical Geology”. Spring Semester of 2014
Teaching Assistant in “Mineralogy-Petrology”. Fall Semester of 2014
Teaching Assistant in “Geophysics”. Spring Semester of 2015
Teaching Assistant in “Historical Geology” Spring Semester of 2015
Head Teaching Assistant in “Mineralogy-Petrology” Fall Semester of 2015
Teaching Assistant in “Rock Mechanics”. Spring Semester of 2016

Teaching Assistant in “Physical Geology” Summer Semester of 2016

Teaching Assistant in “Rock Mechanics”. Fall Semester of 2016

Teaching Assistant in “Physical Geology”. Fall Semester of 2016

Teaching Assistant in “Physical Geology”. Winter-Intersession 2017

Teaching Assistant in “Rock Mechanics”. Spring Semester of 2017

Teaching Assistant in “Physical Geology” Fall Semester of 2017

AWARDS

Ranked top 0.5 % of Iran’s nationwide university entrance exam for Bachelor of Engineering and Science among more than 400,000 participants.

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CONFERENCE ABSTRACTS AND POSTERS

Symposium on the Application of Geophysics to Engineering and Environmental Problems (SAGEEP), March 2017

Geological Society of America (GSA) Section Meetings, March 2017


81st Annual Mississippi Academy of Sciences Meeting, February 2018.

CONFERENCE PAPERS
