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Seismic Characterization of a Gas Hydrate System in the Gulf of Mexico - A Novel Approach for Evaluating High-Resolution Wide-Aperture Data

by

Priyank Jaiswal

A Thesis Submitted In Partial Fulfillment of the Requirements for the Degree Masters of Arts

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ABSTRACT

Seismic Characterization of a Gas Hydrate System in the Gulf of Mexico - A Novel Approach for Evaluating High-Resolution Wide-Aperture Data

by

Priyank Jaiswal

Gas hydrates were discovered in a mud diapir in the leased block Mississippi Canyon 798 - Gulf of Mexico, through piston coring. Subsequently, a seismic experiment was set up to investigate the dynamics behind the hydrate formation. Wide aperture seismic travelt ime data obtained from the experiment have been inverted to estimate 2D P-wave velocity models of the five shot lines. The results from modeling indicate the presence of free gas in regions that show up as zones of high reflectivity on the reflection profiles. The topography of the study area suggests presence of active salt bodies, which in turn, makes it plausible for the gas in the Mississippi Canyon 798 to have deeper sources.
Acknowledgements

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Finally, I gratefully acknowledge and dedicate this thesis to those known and unknown researchers who have been working tirelessly over the ages to unveil the mystery of Nature. It is this pursuit of knowledge that has led and will lead humankind down the path of progress and prosperity.
Table of Contents

Abstract ii
Acknowledgements iii
Table of Contents iv
List of Illustrations vi
List of Tables xii
List of Acronyms xiv
1. INTRODUCTION 1
2. BACKGROUND 9
   2.1 Physical and Chemical Properties of gas Hydrates 9
   2.2 Occurrences 9
   2.3 Importance of studying hydrates 10
   2.4 Seismic Methods for Characterizing Hydrates 12
   2.5 Gulf of Mexico 15
      2.5.1 Salt Tectonics in Gulf of Mexico 17
      2.5.2 Geological and Geophysical Studies in the Gulf of Mexico 18
      2.5.3 Gas Hydrates in Gulf of Mexico 19
3. DATA 28
   3.1 Multi Channel Seismic Near-Vertical Data 28
   3.2 Ocean Bottom Seismometer Wide-Angle Data 30
   3.3 Data Adjustments 31
## List of Illustrations

1. **INTRODUCTION**
   - Figure 1.1. Location of the study area over the Mississippi  
     - Page: 6
   - Figure 1.2. Location of MC 798 over the shelf area.  
     - Page: 7
   - Figure 1.3. Experiment design in MC798.  
     - Page: 8

2. **BACKGROUND**
   - Figure 2.1. Inferred thickness of the GHSZ.  
     - Page: 22
   - Figure 2.2. Estimation of depth of BSR.  
     - Page: 23
   - Figure 2.3. Single-channel seismic data on Blake-Ridge.  
     - Page: 24
   - Figure 2.4. Outline of the Gulf of Mexico basin.  
     - Page: 25
   - Figure 2.5. Canopy complexes.  
     - Page: 26
   - Figure 2.6. Cenozoic structural provinces of the Gulf of Mexico.  
     - Page: 27

3. **DATA**
   - Figure 3.1a. Unmigrated SCS data along Line 1.  
     - Page: 34
   - Figure 3.1b. Unmigrated SCS data along Line 2.  
     - Page: 35
Figure 3.1c. Unmigrated SCS data along Line 3. 36

Figure 3.1d. Unmigrated SCS data along Line 4. 37

Figure 3.1e. Unmigrated SCS data along Line 5. 38

Figure 3.2. Comparison of stack with the SCS for Line 1. 39

Figure 3.3a. OBS A hydrophone channel data from Line 3. 40

Figure 3.3b. OBS B hydrophone channel data from Line 1. 41

Figure 3.3c. OBS C hydrophone channel data from Line 4. 42

Figure 3.3d. OBS D hydrophone channel data from Line 2. 43

Figure 3.3e. OBS F hydrophone channel data from Line 5. 44

Figure 3.3f. OBS E hydrophone channel data from Line 2. 45

Figure 3.4a OBS A z component data from Line 3. 46

Figure 3.4b. OBS B z component data from Line 1. 47

Figure 3.4c. OBS C z component data from Line 4. 48

Figure 3.4d. OBS D z component data from Line 2. 49

Figure 3.4e. OBS F z component data from Line 5. 50
Figure 3.4f. OBS E hydrophone channel data from Line 2.  

4. METHODOLOGY  

Figure 4.1. Example of model parameterization.  

5. DATA MODELING  

Figure 5.1a. Events identified in the data along Line 3.  

Figure 5.1a. (continued)  

Figure 5.1b. Events identified in the data along Line 2.  

Figure 5.1b. (continued)  

Figure 5.1c. Events identified in the data along Line 4.  

Figure 5.1d. Events identified in the data along Line 1.  

Figure 5.1e. Events identified in the data along Line 5.  

Figure 5.2. Starting models for each line.  

Figure 5.3a. Comparison of the observed wide-angle.  

Figure 5.3a. (continued)  

Figure 5.3a (continued)
Figure 5.3a. (continued)

Figure 5.3b. Comparison of the observed near-vertical traveltimes.

Figure 5.4a Ray diagram of the reflection from the base of L1.

Figure 5.4b Ray diagram of the reflection from the base of L2.

Figure 5.4c Ray diagram of the reflection from the base of L3.

Figure 5.4d Ray diagram of refracted arrivals from the base of L1.

Figure 5.5a. 1D velocity profiles below the intersection points.

Figure 5.5b. 1D velocity profile below OBS A and OBS E.

6. RESULTS AND INTERPRETATION

Figure 6.1a. Final model for Line 1.

Figure 6.1b. Final model for Line 2.

Figure 6.1c. Final model for Line 3.

Figure 6.1d. Final model for Line 4.

Figure 6.1e. Final model for Line 5.
Figure 6.2a. Extra picks that estimate the LVZ in Line 4. 116

Figure 6.2b. Alternate model of Line 4. 117

Figure 6.3a. Salt map of northern Gulf of Mexico. 118

Figure 6.3b. Regional dip section across the Gulf of Mexico. 119

Figure 6.4. Distribution of gas hydrates at the MC853 site. 120.

Figure 6.5a. Contour plot of velocity in L2. 121

Figure 6.5b. Contour plot of depth of the base of L2. 122

Figure 6.6a. Contour plot of velocity in L3. 123

Figure 6.6b. Contour plot of depth of the base of L3. 124

Figure 6.7a. SCS data from Line 1. 125

Figure 6.7b. SCS data from Line 2. 126

Figure 6.7c. SCS data from Line 3. 127

Figure 6.7d. SCS data from Line 4. 128

Figure 6.7e. SCS data from Line 5. 129

Figure 6.8a. Comparison of data and velocity model of Line 1. 130
Figure 6.8b. Comparison of data and velocity model of Line 2. 131

Figure 6.8c. Comparison of data and velocity model of Line 3. 132

Figure 6.8d. Comparison of data and velocity model of Line 4. 133

Figure 6.8e. Comparison of data and velocity model of Line 5. 134

Figure 6.9. Reverse polarity of the reflection from top of the wedge. 135

Figure 6.10. Solubility curves and phase relationships. 136

Figure 6.11. Interpretation by Coleman et al (1983). 137

Figure 6.12. Interpretation by Goodwin and Prior (1989). 138
List of Tables

Table 2.1. Physical Properties of water ice and methane hydrates 9

Table 5.1. Velocity function of the starting models for all five lines. 67

Table 5.2.1 OBS locations for Line 1 69

Table 5.2.2. Iterative Inversion for Line 1 69

Table 5.2.3. OBS locations for Line 2 70

Table 5.2.4. Iterative inversion for Line 2 70

Table 5.2.5. OBS locations for Line 3 71

Table 5.2.6. Iterative inversion for Line 3 71

Table 5.2.7. OBS locations for Line 4 72

Table 5.2.8. Iterative inversion for line 4 73

Table 5.2.9. OBS locations for Line 5 73
Table 5.2.10.  Iterative inversion for Line 5

Table 5.3.  Wide-angle and near-vertical $\chi^2$ errors for five lines

Table 5.4.  $\chi^2$ error statistics for individual phases

Table 5.5.  Summary of inversion performed fixed velocity nodes for Line 2.
### List of Acronyms

<table>
<thead>
<tr>
<th>Acronym</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>BSR</td>
<td>Bottom Simulating Reflector.</td>
</tr>
<tr>
<td>CMP</td>
<td>Common Mid-Point.</td>
</tr>
<tr>
<td>GHSZ</td>
<td>Gas Hydrate Stability Zone.</td>
</tr>
<tr>
<td>GOM</td>
<td>Gulf of Mexico.</td>
</tr>
<tr>
<td>HRZ</td>
<td>High Reflectivity Zone.</td>
</tr>
<tr>
<td>L1</td>
<td>Layer 1.</td>
</tr>
<tr>
<td>L2</td>
<td>Layer 2.</td>
</tr>
<tr>
<td>L3</td>
<td>Layer 3.</td>
</tr>
<tr>
<td>L4</td>
<td>Layer 4.</td>
</tr>
<tr>
<td>LVZ</td>
<td>Low Velocity Zone.</td>
</tr>
<tr>
<td>MC798</td>
<td>Leased Block Mississippi Canyon 798.</td>
</tr>
<tr>
<td>MCS</td>
<td>Multi-Channel Seismic.</td>
</tr>
<tr>
<td>OBS</td>
<td>Ocean Bottom Seismometer.</td>
</tr>
<tr>
<td>SCS</td>
<td>Single-Channel Seismic.</td>
</tr>
</tbody>
</table>
1. Introduction

Gas hydrates are naturally occurring solids comprised of water molecules that form a rigid lattice of cages around gas molecules of low molecular weight. Although other gases (e.g. CO₂, H₂S, etc.) can and do form hydrate when they are available, methane is the dominant gas, making up probably >99% of naturally occurring hydrates [Kenvolden, 2000]. The ability of natural gases to form gas hydrate deposits was first found out in the Soviet Union [Vasil’ev et al., 1970]. Since then, hydrates have been discovered and quantified in many places around the world.

Reports on gas hydrates in the Gulf of Mexico (GOM) can be traced to as early as 1984 [Brooks et al., 1984]. These discoveries were commonly made by the analysis of high-resolution shallow seismic datasets (also known as geohazard datasets) collected by the industry to determine the seabed and near subsurface conditions before deploying offshore facilities [Antonie, 1975; Sieck and Self, 1977; Prior and Coleman, 1981]. Neurater and Bryant (1989) analyzed a geohazard dataset from the leased block Mississippi Canyon 798 (Figures 1.1 and 1.2), (hereafter referred to as MC798) and found evidence for the presence of hydrates in a small topographic mound in the SE region of the block. The mound seemed to be acoustically amorphous in an otherwise generally well-stratified region. It has 10 m of relief, stands beneath a water depth of 810 meters and has a radius of 275 m. A 10 m piston core taken at the crest of the mound revealed that it was a mud diapir. The recovered core had approximately 5 m of greenish-
gray to black silty clay with white ice-like chunks of gas hydrates disseminated in the matrix [Neurater and Bryant, 1989].

In 1998, the U.S. Geological Survey, the University of Mississippi, and the Department of Energy jointly conducted a high-resolution seismic survey on the R/V Tommy Munro. The aim of the experiment was to collect and analyze data to understand the properties of hydrate-bearing sediments in the leased block MC798. Primary focus of this experiment was acquisition of high-resolution multi-channel seismic (MCS) data through a twenty-four-channel streamer. Simultaneously, an ocean bottom seismometer (OBS) experiment was designed to collect wide-angle data. The experiment layout consisted of six OBSs equipped with 4.5 Hz geophones deployed in two north/south lines across the mud diapir (Figure 1.3). The spacing between neighboring OBSs along each line was 1.5 km. No OBS was deployed on top of the diapir because brine pools (observed commonly near similar diapirs in the gulf) may have put the OBS at risk. Five shooting profiles (two north/south, two east/west and one northwest/southeast) approximately 10 km long (Figure 1.3), each were acquired with a 575/575 cm³ generator/injector (GI) airgun. The deployment of seismometers was done in such a way that the area bounded by the OBSs has dense enough ray coverage to allow two-dimensional (2D) traveltme inversion along each line.

The main goal of this thesis is to derive 2D P-wave velocity models below the five shot lines by inverting the reflected and refracted arrivals from the OBS and MCS data simultaneously. After the 2D models of all lines were estimated and adjusted for
consistency at the intersection points, they were compared with 2D models of nearby regions from all other studies and interpreted geologically.

To estimate a 2D P-wave velocity model of each line, the OBS and MCS data were inverted using an algorithm for joint traveltime inversion that is applicable to any type of body-wave seismic data [Zelt and Smith, 1992]. The algorithm will hereafter be referred to as ZS92. The following features of the ZS92 algorithm make it more applicable to the present dataset over other popular methods such as: (1) the spacing and number of model nodes specifying velocity and interface depth can be irregular, (2) any or all of the model parameters can be selected for inversion, (3) the velocities are ‘tied’ to the layer interfaces, and (4) simultaneous inversion of all arrival types is possible. Examples of applying the ZS92 inversion method to refraction and wide-angle reflection data can be found in Holbrook et al. (1994,1999), and Hughes et al. (1998).

Chapter 2 describes the geological setting of MC 798 and provides the background information that is required in this thesis to model the data. This chapter summarizes the physical and chemical properties of gas hydrates, their detection procedure, their geological association, their seismic characteristics and their worldwide discoveries. The chapter provides a brief description of occurrences of the gas hydrate stability zone (GHSZ). GHSZs are zones below the ocean bottom within which the P-T conditions allow hydrates to exist. A small discussion follows on bottom simulating reflectors (BSRs), their origin and their seismic characteristics. Next, there is a discussion about the significance of understanding gas hydrate systems. The chapter concludes with
a brief discussion about the GOM and cases where hydrates have been discovered in the gulf.

Chapter 3 describes various features of the OBS and MCS data. The chapter points out peculiar patterns in the reflection profiles (such as chaotic and/or high amplitude reflections) that are interpreted in subsequent chapters. The chapter discusses the limitations and benefits of modeling wide-angle and near-vertical datasets. It also discusses various acquisition problems associated with each of the datasets and the ways the data were corrected to overcome them.

Chapter 4 presents the method used for traveltime inversion. It describes the joint inversion scheme, which is the main focus of this thesis. It provides the mathematical background of ray theory and damped least-squares inversion. It also describes the practical methodology applied to the data and the limitations of the methodology as well. The computer programs used for traveltime inversion are presented.

Chapter 5 presents the application of the ZS92 algorithm to the modeling of the wide-angle and near-vertical datasets simultaneously. It describes the technique used for identifying common events across the MCS and OBS data. Subsequently, it discusses the bootstrap method of constructing a starting model and picking traveltimes.

Chapter 6 discusses the criteria for a geologically reasonable and acceptable model. It discusses various features of the final models estimated from traveltime inversion. It also discusses the possible presence of low velocity zones (LVZs) in the
region. Finally, the chapter interprets the models of all five lines in the context of the regional geology.

Chapter 7 provides concluding remarks on the thesis and gives suggestions for the direction of future work.
Figure 1.1. Location of the study area over the Mississippi Trough (Ferebee and Bryant, 1979). Contours are bathymetry in meters.
Figure 1.2. Location of MC 798 over the shelf area. Bathymetry of shelf is contoured and marked every 100 m near the study area. MC 798 stands in a water depth of 700-900 m. Each grid represents 4.82 km X 4.82 km block. Latitude and longitude are marked every one degree.
Figure 1.3. Experiment design in MC798. Layout of the seismic lines is shown as solid lines. The positions of OBSs A, B, C, D, E and F are shown as black circles. The mud diapir is shown as a white circle. Line numbers are circled and shown at the beginning of the lines. The seafloor bathymetry is contoured every 50 m and labeled every 100 m. Latitude and longitude are marked every 0.05 degrees.
2. Background

2.1 Physical and Chemical Properties of Gas Hydrates

Gas hydrates can occur in three crystalline structures: I (body-centered packing), II (diamond packing in cubic system) and H (hexagonal crystallographic system). Of these, structure I occurs most commonly in nature [Sloan, 1998]. A comparison of physical properties of structure I with water is presented in Table 2.1.

Table 2.1. Physical Properties of water ice and methane hydrates [after Kvenvolden, 2000].

<table>
<thead>
<tr>
<th>Property</th>
<th>Water</th>
<th>Hydrate</th>
</tr>
</thead>
<tbody>
<tr>
<td>P wave velocity $V_p$ (km/s)</td>
<td>3.8</td>
<td>80</td>
</tr>
<tr>
<td>Velocity Ratio $V_p/V_s$ at 272 K</td>
<td>1.88</td>
<td>1.95</td>
</tr>
<tr>
<td>Poisson’s Ratio</td>
<td>0.33</td>
<td>~0.33</td>
</tr>
<tr>
<td>Bulk Modulus at 272 K (Gpa)</td>
<td>8.8</td>
<td>5.6</td>
</tr>
<tr>
<td>Bulk Density (g/cm$^3$)</td>
<td>0.916</td>
<td>0.912</td>
</tr>
</tbody>
</table>

2.2 Occurrences

Gas Hydrates have been found in two general sedimentary environments: (1) clay rich, high porosity ocean and sea bottom sediments; and (2) arctic onshore sands [Dvorkin et al., 2000]. Among the different types of ocean bottom sediments, methane accumulates in continental margin sediments most rapidly. This is because at continental margins (1) the flux of organic carbon to the seafloor is greatest, and (2) the sedimentation rate is highest. High fluxes of organic carbon into sediments ensure
generation of methane from microbial activity and high sedimentation rates bury organic materials before they are aerobically oxidized. Prerequisites of high hydrostatic pressure (>5 bars) and low bottom water temperature (<7°C) for the stability of gas hydrates mean that hydrates occur mostly below 530 m water depth in the low latitudes, and generally below 250 m depth in high latitudes [Haq, 2000]. Typically they are stable in continental margin sediment zones that can be viewed as a seaward-thickening prism (Figure 2.1). The shape and thickness of the prism depends on several factors like salinity, topography, geotherm, gas composition, etc. The present study area is in a shelf setting located at a depth of 700-900 m below sea level. An appropriate diagram for the stability of methane hydrates in the study area is shown in Figure 2.2.

2.3 Importance of studying hydrates

Gas hydrates are now being studied exhaustively due to various economic, and environmental reasons.

Potential energy resources

It is being increasingly hypothesized that gas hydrates can contribute immensely to the world’s energy requirement if free gas can be extracted from them with low economic and environmental costs [Collett, 2000]. One volume of methane hydrate can yield up to 164 volumes of methane gas under standard temperature and pressure [Davidson et al 1978]. The estimated amount of gas in hydrates greatly exceeds the volume of total conventional gas reserves [Collett, 2000]. The estimated range is over three orders-of-magnitudes.
The potential volume of gas contained in the hydrates can be calculated using five main parameters: (1) areal extent of the gas hydrate occurrence, (2) thickness of the hydrate layer, (3) sediment porosity, (4) degree of gas hydrate saturation, and (5) the gas hydrate yield volumetric parameters (which define how much free gas at STP is stored in the hydrates) [Collett, 1993]. Countries like Japan, India, South Korea, etc., who are dependent on foreign countries for their energy resources, are initiating ambitious programs to access the potential of gas hydrates as a future source of energy [Collett, 2000].

Influence on climatic change

As mentioned earlier, hydrates are only found within the GHSZ, whose thickness depends on ambient temperature, pressure, salinity and the gas type. An increase in the mean temperature and/or drop in atmospheric pressure can therefore cause the hydrates to dissociate and emit methane into the atmosphere. Methane is ten times more potent than carbon dioxide as a greenhouse gas [Haq, 2000]. Although its residence time in the atmosphere is short, methane released from hydrates can play a significant role in climate and ocean change [Dickens et al., 1997]. Paleoclimatic records the from recent geological past gleaned from ice cores from Antarctica and Greenland show that at the onset of a glaciation cycle there is a decrease in the CO₂ and CH₄ content of the atmosphere [e.g., Jouzel et al., 1993; Petit et al., 1999]. As suggested by Haq (2000), the released methane during sea level drop can provide a negative response to glaciation and thus form a climatic feedback loop. This feedback loop would close when the sea level is once again
high enough to stabilize the residual clathrates and encourage the genesis of new deposits of hydrates.

*Potential of generating submarine slope failures*

The growth of gas hydrates strengthens the sediments. Disturbances within GHSZs lead to the dissociation of hydrates into water and free gas, which decreases the shear strength of the sediments and makes them prone to failure. In places like the GOM, where the hydrate stability field is close to the seafloor, the up dip end is subjected to the maximum pore pressure alterations when gas hydrates decompose. This decomposition leads to slope failures. Slope failures are commonly recognized by headwall scars [*Nisbet and Piper*, 1998], which have in turn been identified in numerous seismic profiles around the world [e.g., *Summerheys et al.*, 1979; *Embley*, 1980; *Carpenter*, 1981; *Rothwell et al.*, 1998]. On a smaller scale, it is necessary to understand the mechanical behavior of the sediments containing hydrates as they impact seafloor-drilling, deployment of underwater instruments, etc.

2.4 **Seismic Methods for Characterizing Hydrates**

Seismic methods help to characterize the structure and stratigraphy of gas-hydrate deposits. In general, seismic images provide maps of:

1. Layers that are known to contain gas.
2. The "source" of gas, and clues as to whether the gas originates in a deep thermogenic reservoir or in a more distributed biogenic source zone.
3. Conduits that supply gas and fluid to the GHSZ.
4. Conduits that allow gas to escape from the GHSZ.

5. Possible traps of gas.

In this thesis, the seismic data have been used to estimate the acoustic properties of zones that are potential gas reservoirs.

Acoustic properties of Hydrate bearing sediments

Hydrates and free gas affect the elastic properties of the host sediment in ways that are seismically detectable. Hydrate formation can also cause blanking of the sediment acoustic stratigraphy through cementation of the sediment structure [Miles, 2000] (Figure 2.3). Partial replacement of the pore fluids with hydrates can increase the P-wave velocity of the sediments from ~1.6 km/s to ~2.5 km/s or more [Miles, 2000]. The presence of free gas, on the other hand, decreases the P-wave velocity in the sediments depending on its concentration [Wood and Ruppel, 2000]. Because of the low P-wave velocity of the sediments in the presence of gas, sediment $V_p$ becomes sensitive to even small amounts of gas in the pore spaces [e.g. Domenico, 1976]. Therefore, a small change in gas concentration can bring about a large change in sediment $V_p$.

Predicting the P-wave velocities of the gas hydrate and gas bearing sediments is not straightforward because (1) the sediments consist of multiphase aggregations of minerals (e.g. clay minerals, quartz and calcite), pore fluid (typically saline water), and hydrate or gas, (2) the rock physics of hydrate bearing sediment is still not properly understood, and (3) the estimation of many parameters (mineral properties, porosity, effective pressure and pore fluid compressibility, etc.) for advanced geophysical modeling requires in situ measurements through drilling and sampling, which are rarely available. The absence of a
hydrate-free "reference" velocity of the sediments causes a problem in quantifying the hydrate and gas contents.

**Bottom Simulating Reflectors**

Gas hydrates have historically been inferred on the basis of the presence of bottom simulating reflectors (BSRs). BSRs are high amplitude reversed polarity events on seismic reflection records that mark the base of the GHSZ [e.g. Shipley et al., 1979; Holbrook et al., 1996]. The depth of the base of the GHSZ depends mainly on the geothermal gradient, salinity, and P-T conditions of the subsurface. Since the P-T contours in the sediments commonly mimic the seafloor, the base of the GHSZ also mimics the seafloor. At the base of the GHSZ, where the curve of stability conditions crosses the phase boundary, free gas can exist. If the free gas is in contact with the hydrates at the base of the GHSZ, it gives rise to a strong event in the seismic reflection records that often simulates the ocean bottom and hence the name "Bottom Simulating Reflector" (Figure 2.3).

Typically, it has been inferred in a hydrate-gas system that hydrates overlie free gas [Xia et al., 2000; Lodolo et al., 2002]. The underlying free gas causes a drastic decrease in \( V_p \), which in turn causes a sharp contrast in acoustic properties. The underlying free gas has much more contribution to the strength of reflections from the phase boundary than the overlying hydrate wedge [e.g. Korenaga et al., 1997]. Occasionally though, in the absence of free gas, weaker BSRs have been found to occur due to hydrates only [e.g., Minshull et al., 1994; Pecher et al., 1996b]. However, the
presence or absence of BSRs does not allow any immediate conclusions about the presence or absence of hydrates. Although the presence of hydrates has been confirmed in the study area through shallow coring, no reflection event on the seismic records from near the diapir could be identified that had the properties of a “conventional” BSR. In the absence of a conventional BSR, one of the goals of this thesis is to explain the reasons behind the occurrence of hydrates in the diapir.

2.5 Gulf of Mexico

The GOM basin is a roughly circular structural basin approximately 1,500 km in diameter. It is filled in its deeper part with 10 – 15 km of sedimentary rocks that range in age from Late Triassic to Holocene. The Florida Carbonate Platform and the Yucatan carbonate platform flank the basin to the east and south, respectively (Figure 2.4). The western limit of the basin has been placed roughly at the foot of the Chiapas massif and the Sierra Madre Oriental of Mexico, and along the eastern edge of the Coahuila platform. To the north, the structural limits of the basin correspond, from west to east, with the basinward flanks of the Marathon uplift, the Ouachita orogenic belt, the Ouachita Mountains, the Central Mississippi deformed belt, and the southern reaches of the Appalachian Mountains [Salvador, 1991] (Figure 2.4).

Late Quaternary sedimentation

Before the late Jurassic, deposition in the GOM consisted of Louann Salt (Middle Jurassic), and mostly nonmarine red beds. Since the late Jurassic, the drainage basins of the Mississippi River system have been delivering sediments to the GOM [Worzel and
Bruke, 1978]. It was during the latest Jurassic that the terrigenous clastic influx from
the tectonically elevated northern and western continental interior began to overwhelm
the predominant carbonate environments that previously encircled the gulf [Garrison and
Martin, 1973]. The Quaternary deposits of the northern GOM are unusually thick. As
much as 3600 m have accumulated beneath the present shelf in offshore Louisiana and
Texas, and up to 3,000 m of sediments have accumulated in the deep GOM basin in the
vicinity of the present Mississippi Fan. The sediments in the Mississippi Canyon have
been deposited within the last 27,000 years [Coleman et al., 1983].

Mississippi Canyon

The Mississippi Canyon or Trough was originally interpreted by Shepard (1937),
and later by Gealy (1955), as resulting from massive submarine slumping in combination
with diapir placement. Coleman et al. (1983) interpreted it to have formed by the
Canyon’s proximity to salt diapirs, subaerial erosion, and canyon cutting by sand-rich
density flows. Coleman et al (1983) used high-resolution MCS profiles and soil borings
to study the Mississippi Canyon area and to determine the ages of seismic horizons by
radiocarbon dating and estimated the canyon to have been formed post 25 -27,000 yr B.P.
by massive shelf-edge failures followed by retrogressive slumping of sediments around
the edges of the canyon. Infilling of the canyon begun around 20,000 yr B.P. and was
complete by 10,000 yr B.P. Coleman et al. (1983) noted that the removal of 1,500-2,000
km³ of sediment occurred in a short period of time (5000 yr or less) and that infilling of
the canyon was almost contemporaneous with its formation.
The present-day expression of the Mississippi Canyon extends from a water depth of about 50 m on the shelf, through the shelf margin, and across the upper continental slope [Goodwin and Prior, 1989]. The morphology of the canyon on the outer shelf is quite different from that of the slope segment. The shelf-indenting portion has a maximum width of about 25 km and a maximum relief of about 350-400 m, declining towards the shelf. Canyon wall gradients average 7 to 8 degrees, but locally exceed 12 degrees. The upper canyon edge is dominated by alternating bayments or re-entrants, inverting ridges, and isolated knolls. The knolls and ridges have been interpreted as the result of diapirism [Febree, 1978]. The study area for this thesis lies on the western flank of the canyon.

2.5.1 Salt tectonics in the Gulf of Mexico

It has long been recognized that the northern GOM contains a vast amount of allochthonous salt [e.g., Martin, 1980]. This generally takes the form of complexes of diapirs and small sheets as well as several vast, laterally extensive, sub-horizontal allochthonous salt bodies [Simmons, 1992]. Allochthonous salt bodies with multiple stems were termed canopies by Peel et al. (1995). The structure of the sedimentary units in the present study area is believed to be formed due to adjacent salt emplacement. Canopies and canopy complexes shown in Figure 2.5 represent the original maximum extent of the salt.

Peel at al. (1995) recognized large provinces of similar structural characters in the GOM that also share a common tectonostratigraphic history and general history of
canopy emplacement. The province in which the present study area lies is called "Eastern province" (Figure 2.6). The Eastern province is characterized by a combination of the following features: (1) a major linked system (principally of middle-late Miocene age) connecting extension that is probably located under the present day shelf, to contraction in the Mississippi Fan forebelt; (2) a large salt canopy (Canopy I) on the present-day middle-slope that formed in middle-late Miocene; and (3) a largely evacuated salt canopy (Canopy VI), under the present day shelf that was emplaced in the Paleogene [Peel et al., 1995]. Canopy I lies on the border of the Mississippi Canyon and Atwater protraction areas. The canopy is largely intact, having relatively few salt withdrawal features over it. The base of the salt is commonly well imaged in this area [Wu et al., 1990b]. Canopy VI is interpreted to be a paleo-canopy, from which the majority of the salt has been removed by salt withdrawal. Today it is an extensional salt wedge from which diapirs rise.

2.5.2 Geological and Geophysical Studies in the Gulf of Mexico

During the first half of the 19th century, individual scientists engaged in broad reconnaissance journeys and conducted the initial geological work in the GOM. The second half of the 19th century witnessed the growth of geological survey societies in the states bordering the GOM. The effort was mainly limited to interpreting the rock outcrops along the periphery of the basin. The successful discovery of oil along the coastal plains surrounding the GOM during the early years of the 20th century marked the birth of a strong and aggressive petroleum industry in the region. Following this, geophysical
methods were introduced. First came the torsion balance (gravity) and refraction seismograph to delineate the salt domes. Then came the reflection seismographs in the late 1920’s, followed by various advanced techniques over time. Blending of geophysical data with subsurface geological information from thousands of wells over the last 85 years led to a progressively better interpretation of the local and regional aspects of the stratigraphy, structure, and geological history. The geophysical surveys of the central deeper part of the GOM basin mark the most recent stage in the study. The refraction and reflection surveys, conducted by oil companies, the USGS and academic institutions have allowed a better understanding of the sedimentary nature of the basin and the crust below it.

2.5.3 Gas Hydrates in the Gulf of Mexico

The GOM is increasingly being treated as a separate end-member for gas hydrate systems: a focused-methane-flux environment. Focused methane-flux-environment is an environment where the supply of gas to specific zones within the GHSZ is through faults as opposed to an environment where gas is present throughout the GHSZ in quantities enough to sustain a hydrate-gas system. The northern deep-water GOM slope consists largely of mini basins from salt withdrawal that are separated by ridges above the crest of sub-surface salt structures [e.g., Nelson, 1991]. Above those ridges and along the flanks of salt domes, extensive faulting occurs, providing numerous deeply rooted fluid migration paths. Vent sites, such as mud diapirs and volcanoes, are ubiquitous where these faults intercept the seafloor [e.g., Neurauter and Roberts, 1994]. Gas hydrates have been observed frequently as outcrops or in shallow sediment cores in the vicinity of vent
sites [Neurauter and Bryant, 1989; MacDonald et al., 1994]. Locally high rates of methane flux are required to maintain shallow gas hydrates at or close to the seafloor. This is because the methane undersaturation of the ocean promotes diffusion of gases from beneath the seafloor and subsequent gas hydrate dissociation [Xu and Ruppel, 1999]. In the vicinity of fault systems, ample methane and other hydrocarbon gases are supplied for gas hydrate formation from deep thermogenic sources in addition to biogenic methane from shallower sediments [Brooks et al., 1994]. This is different from the mostly biogenic, relatively shallow sources (although often beneath the base of the gas hydrate stability) that are assumed to supply most of the methane for hydrate formation in both low- and high-methane-flux settings [e.g., Paull et al., 1996; Westbrook et al., 1994]. Gas hydrate systems in the focused-methane-flux environment (like the GOM) generally seem to lack the conventional ways in which BSRs are supposed to manifest themselves in seismic data [Bangs et al., 1993; Holbrook et al., 1996; MacKay et al., 1994]. One of the plausible reasons may be that in petroleum generating areas like the gulf, where other gas hydrate forming gases leak up from great depths through fault systems, the base of the GHSZ is quite perturbed and therefore the BSRs do not behave in a conventional way [Paull et al., 2000]. Shallow gas hydrates near fault systems in the focused-methane-flux environment of the GOM have only been studied in the last ten years [e.g., MacDonald et al., 1994]. We know very little, however, about the extent of the gas hydrate outcrops below the seafloor, about possible free gas beneath the base of the GHSZ, and about gas hydrates, if present at all, in the mini basins away from the faults.
As a first step towards interpreting the hydrate-gas system in the study area, evidence for the presence of free gas is sought through traveltime modeling. Due to the absence of any well log or core data, it is not possible to establish the presence of free gas directly and therefore, indirect suggestions are considered, such as: (1) features in the data (like HRZs, chaotic reflections, etc.) that can be associated with presence of free gas, or (2) anomalies (like lowering of velocities) in the estimated P-wave velocity models, or (3) a combination of both. If the presence of free gas in the study area is established, further arguments can be made about certain events in the data being a BSR based on their characteristics (as explained under the section on Bottom Simulating Reflectors). Based on the presence or absence and the extent (if present) of the BSR during the interpretation, the dynamics of the hydrate-gas system in the study area can be established.
Figure 2.1. Inferred thickness of the GHSZ in sediments of a continental margin assuming a typical geothermal gradient. After Kvenvolden and Barnard (1982).
Fig 2.2. Estimation of depth of BSR. Stability of methane hydrate in seawater as defined by temperature (T) and Pressure (P, indicated as water depth). The seawater equilibrium curve is plotted with data obtained from laboratory experiments. Variations of stability with variations in salinity are not taken into account in this diagram. The data on the equilibrium curve is shown as cross marks. Horizontal line at 800 m depth is the seafloor. Dashed lines are geotherms calculated with a bottom water temperature of 8°C. The intersection points of the geotherms with the equilibrium curve is the expected depth of BSRs at 800 m water depth. According to this diagram the thickness of the GHSZ can vary anywhere from 50 to 150 m below sea floor depending on the geotherms taken into consideration. The actual thickness in the study area can be more or less depending on the lateral variations of the bathymetry, geotherm, salinity and bottom water temperature [Sassen et al., 2001].
Figure 2.3. Single-channel seismic data on Blake Ridge, depthconverted using velocity-depth functions determined from traveltime inversion of vertical seismic profile data. Solid lines show positions of drill holes 994D, 995B, and 997B. Reflection event interpreted as BSR is shown. Depth is in meters below sea level (mbsl). Above the BSR the amplitude of the reflections is reduced (from 3100 to 3200 mbsl). This phenomenon is known as "blanking". After Holbrook et al. (1996).
Figure 2.4. Outline of the Gulf of Mexico basin. After Salvador [1999]. The location of the study area is marked with a solid rectangle. Second-order structural features within the basin: 1, Macuspana Basin; 2, Villahermosa uplift; 3, Comalcalco Basin; 4, Isthmus Saline basin; 5, Veracruz Basin; 6, Cordoba platform; 7, Santa Ana massif; 8, Tuxpan platform; 9, Tampico-Misantla basin; 10, Valles-San Luis Potosi platform; 11, Magiscatzin basin; 12, Tamaulipas arch; 13, Burgos basin; 14, Sabinas basin; 15, Coahuila platform; 16, El Burro uplift; 17, Peyotes-Picachos arches; 18, Rio Grande embayment; 19, San Marcos arch; 20, East Texas basin; 21, Sabine uplift; 22, North Louisiana salt basin; 23, Monroe uplift; 24, Desha basin; 25, La Salle arch; 26, Mississippi salt basin; 27, Jackson dome; 28, Central Mississippi deformed belt; 29, Black Warrior Basin; 30, Wiggins uplift; 31, Apalachicola embayment; 32, Ocala uplift; 33, Southeast Georgia embayment; 34, Middle Ground arch; 35, Southern platform; 36, Tampa embayment; 37, Sarasota arch; 38, South Florida basin.
Figure 2.5. Major allochthonous salt canopies and genetically related canopy complexes. Numbers I to VI correspond to canopies I to VI. Canopies I and VI have been defined in the text. The areas outlined are large single canopies and sets of genetically related sheets. Additional large canopies and canopy complexes may exist farther north. The age of formation and geological age of the canopies are relatively well defined for I and II but more poorly constrained farther west and north. Heavy lines south of canopies I and IIb and both east and west of canopy III indicate fold axial traces in the Mississippi Fan, Port Isabel, and Perdido foldbelts, respectively. Solid lines are seismic lines through this area not discussed in this thesis. After Peel et al. (1995).
Figure 2.6. Division of the Northern Gulf of Mexico margin into Cenozoic structural provinces. Some of the major characteristics of each province are labeled. Hatched lines denote Lower Cretaceous shelf edge and Florida Escarpment offshore. After Peel et al., (1995). The study area is indicated as a solid rectangle.
3. Data

The inversion of traveltime data provides estimations of the 2D P-wave velocity and interface structure of the subsurface. This property, often in conjunction with other geological and geophysical data and assumptions, can be interpreted in terms of the composition or the physical state of a geological system [e.g., Zelt and White, 1995; Katzman and Holbrook, 1994]. In this thesis, MCS and OBS traveltime data have been inverted jointly using the ZS92 algorithm to estimate the geological model of a speculated hydrate-gas system.

3.1 Multi-Channel-Seismic Near-Vertical Data

The MCS data have a near trace offset of 40 meters. The near trace from all the MCS shots is collectively termed the single channel seismic (SCS) data and has been used mostly for display purposes in this thesis because: (1) the SCS data undergoes the least amount of processing and therefore it is least susceptible to processing artifacts, and (2) it shows all the structural features relevant to this thesis.

Numerous reflectors could be identified in each of the SCS reflection profiles (Figure 3.1). Among them, only three events E1, E2 and E3 were distinct, coherent, and identifiable on all lines. The first of these events (E1) marks the base of an acoustically transparent layer (L1), possibly a pelagic drape of silty clays [Neurauter and Bryant, 1989]. In this thesis, this layer was referred to as L1; "L" in the nomenclature stood for "Layer". The second and third reflection events (E2 and E3) form the base of sedimentary packages L2 and L3 that could not be identified geologically, i.e., their
lithology was unknown in absence of any well logs or core samples. No reflection below L3 was coherent or continuous enough such that it could be identified in all the five lines. Besides the three main events, numerous other acoustically strong events were identified, representing different time horizons (feature labeled R in Figure 3.1). But, the continuity of these reflectors at various depths diminished to the point of having a hazy or obscure character. The diapir manifests itself as chaotic reflections (labeled C in Figures 3.1a-3.1d) on the lines passing near it. Thick chaotic seismic sequences show up as high reflectivity zones beneath E2 at different locations (feature Z in Figures 3.2). Immediately below the mound, the chaotic reflection character seems to extend up to the seafloor (best illustrated in Figure 3.1a). Reflections in some sections are upturned along the flanks of the amorphous zones (E2 beside HRZs in Figure 3.1). The lowermost event (E3) seems to have domed up beneath the terrace area over which the diapir is situated (Figures 3.1).

The MCS data from all the shot lines were converted into a six-fold CMP stack using conventional CMP processing techniques [P. Hart, personal communication 2001]. The reflection events in the stacks were better resolved. A visual comparison of the quality of the reflection events in the stack and the SCS data from Line 1 is made in Figure 3.2. The trace spacing in the stack is 14.5 m.

3.2 Ocean Bottom Seismometer Wide-Angle Data

It was intended during the design of the OBS experiment to record signals with frequencies greater than 100 Hz. The source used to shoot the seismic lines therefore had
an unusually high dominant frequency (~120 Hz). Each of the six OBSs were equipped with 4.5 Hz geophones and collected four component data (three displacement components and one pressure component). The OBS data have a sampling interval of 1.434 ms. Only inline data (2D) is used in the present study. The wide-angle data has up to 7 km offsets on some records although the usable part is mostly within 4.5 km. This thesis uses the pressure component and one displacement component (the vertical component) of the OBS data for modeling. In this thesis, the pressure component of the OBS data is recorded by the hydrophone channel and is termed the hydrophone channel data (Figure 3.3) and the vertical component is termed the z-component data (Figure 3.4). Due to clipped amplitudes of the direct arrivals and problems of low-frequency high-amplitude noise, the peak frequency of the hydrophone channel data is low (~20 Hz). The z-component data have a dominant frequency of 40-60 Hz. Due to a lack of strong coupling to the subsurface, the z-component data show a lot of ringing (Figure 3.4). Reflections from the base of L1, L2, and L3 that were identified in the MCS data could be identified in the hydrophone channel and the z-component data (Figures 3.3 and 3.4).

The hydrophone channel and the z-component data were used simultaneously while picking wide-aperture traveltime events. The hydrophone channel data, due to their high amplitudes and less reverberation compared to the z-component data, were used to pick events near zero offsets. As the dominant frequency of the z-component data was higher than the peak frequency of the hydrophone channel data, the traveltime events at far offsets could be better identified compared to the hydrophone channel data at the
same offsets and therefore the z-component data was used for picking events at far
offsets. Moreover, at far offsets the reverberations in the z-component data also abated,
further enhancing the clarity of the events.

Even for data acquired using closely-spaced source and receiver arrays, the
constraints on velocity and interface structure provided by the traveltimes are inherently
limited. This is because (1) the subsurface is inadequately sampled by the surface-to-
seafloor wave propagation between the sources and receivers used in the experiment, (2)
the waveform healing phenomenon, and (3) the fact that only a small portion of the
seismogram (traveltimes) is used [Zelt, 1999]. In our case, the resolution in parts of the
model is limited where there is (1) an inadequate angular distribution of the ray paths, or
(2) a lack of reciprocity of the ray paths.

3.3 Data Adjustment

Originally, it was planned to record shots with an independent OBS shot logger in
addition to the MCS shot logger. However, the OBS shot logging failed. It is suspected
that their failure was due to high humidity/temperature and clock drift [Data Report,
1999]. The MCS shot logger recorded only with an accuracy of 0.01s. To solve the
problem of shot logging, three parameters are required: the position (latitude and
longitude values) of the OBS, the depth of the OBS, and the water velocity.

From work done by other scientists in this region, a velocity value of 1.49 km/s
was used for water [I. Pecher, personal communication 2001; NSF proposal, 1999]. Each
shot line had two or three inline OBSs. For each OBS in the hydrophone channel record, the trace that appeared to have the minimum arrival time was assigned “zero-offset”. The navigation time of this trace (from its header) was compared to the times in the MCS navigation file and the nearest trace in the MCS data (corresponding to the shot fired above that particular inline OBS) was identified. A position (in latitude/longitude) was also assigned to this trace from the navigation files. The depth of this OBS was estimated from the two-way traveltime of the seafloor reflection in the trace from MCS data. Thus, for each OBS record, using the calculated value of the depth and the designated value of the water velocity, offsets (from the zero offset trace) of all other traces were computed. Theoretically, when an OBS record is plotted with a reduced velocity of 1.49 km/s (i.e. the water velocity), the direct arrivals should align horizontally. Repeating the process iteratively, the value of the depth and position of the OBS that gave the most horizontal look to the direct arrivals was chosen. For the intersecting lines, the latitude was determined with the help of one inline OBS and the longitude with the help of the other inline OBS. OBS A and E therefore have only their longitudes well defined.

Direct arrivals were picked on the hydrophone channel records. These were the observed one-way traveltimes. The seafloor reflections were picked on the stack and using the water velocity, the bathymetry was computed. The OBSs were positioned appropriately on the seafloor and rays were shot to the sea surface (airgun positions). Thus, the one-way traveltimes of direct arrivals were predicted and their differences from the observed traveltimes were computed. The observed traveltimes were obtained by
picking direct arrivals on the hydrophone channel data. The traces were shifted such that the observed direct arrivals coincided with the predicted ones. After adjusting the data using the procedure described above, the uncertainty associated with the traveltime picks could be lowered to as much as 6 ms. Unfortunately though, the correction for the shot-logging problem did not prove effective at offsets beyond the crossover distance from the direct arrivals to the sub-seafloor events. Beyond the crossover distance, the waveforms begin to overlap and the direct arrivals could not be picked accurately.

The process of computing time corrections for the OBS data assumed constant water velocity. This is an acceptable simplification that is usually made while modeling a marine traveltime dataset. The effect of this simplification was taken into account when the static corrections were computed by the correction procedure described above. Partly due to the system delay while recording the MCS data and partly due to the static corrections in the OBS data, the seafloor reflections in the MCS data were no longer twice the direct arrival times in the OBS data. The MCS data were therefore shifted upwards by 40 – 45 ms to make them consistent with the OBS data.
Figure 3.1a. Unmigrated SCS data along Line 1. The location of the diapir is indicated. Position of inline OBSs B and D are shown as black circles. (L1) is the Holocene drape. (L2) and (L3) are the sedimentary packages for the purpose of modeling. Labels on the arrivals are as follows: (E1) reflection from base of L1, (E2) reflection from base of L2, (E3) reflection from base of L3, (R) reflectors that are only locally continuous and coherent, (C) chaotic reflections below the diapir, and (Z) HRZ below L2. (E2) and (E3) are upturned near (Z). The Holocene drape thickens in the center of the line. E3 domes up below OBS D at the approximate location of the diapir.
Figure 3.1b. Unmigrated SCS data along Line 2. Position of inline OBSs D, F and E are shown as black circles. Symbols have same meanings as in Figure 3.1a. E3 domes up below OBS D at the approximate location of the diapir. (E2) and (E3) are upturned near (Z). The Holocene drape thickens in the center of the line.
Figure 3.1c. Unmigrated SCS data along Line 3. Position of inline OBSs A, B and C are shown as black circles. Symbols have same meanings as in Figure 3.1a. (E2) is upturned near (Z) and (E3) domes up below OBS C which is near the diapir. The Holocene drape thickens in the center of the line.
Figure 3.1d. Unmigrated SCS data along Line 4. Position of inline OBSs C and D are shown as black circles. Symbols have same meanings as in Figure 3.1a. (E2) is upturned near (Z). (E3) domes up from both ends towards the collapse feature in the center. The Holocene drape thickens in the center of the line.
Figure 3.1e. Unmigrated SCS data along Line 5. Position of inline OBSs F and B are shown as black circles. Symbols have same meanings as in Figure 3.1a. (E2) is upturned near (Z). (E3) domes up from both ends towards the collapse feature in the center below OBS F. The Holocene drape thickens in the center of the line.
The labels E1, E2, and E3 are placed below the respective reflection events. The arrivals are clearer in the shallower region.

Figure 3.2: Comparison of stack (above) with the SCS (below) for Line 1. The locations of the OBS are shown with solid circles.
Figure 3.3a. OBS A hydrophone channel data from Line 3. Plot made with a reducing velocity of 2 km/s. Labels on the arrivals are as follows: (D) direct arrivals, (E1) reflection from base of L1, (E2) reflection from base of L2, (E3) reflection from base of L3, (rf) refraction from base of L1, and (m) multiples. The amplitudes of the direct arrivals are clipped to about 3 km on either side.
Figure 3.3b. OBS B hydrophone channel data from Line 1. Plot made with a reducing velocity of 2 km/s. Symbols have same meaning as in Figure 3.3a. The amplitudes of the direct arrivals are clipped to about 3 km on either side.
Figure 3.3c. OBS C hydrophone channel data from Line 4. Plot made with a reducing velocity of 2 km/s. Symbols have same meanings as in Figure 3.3 a. The amplitudes of the direct arrivals are clipped to about 3km on either side.
Figure 3.3d. OBS D hydrophone channel data from Line 2. Plot made with a reducing velocity of 2 km/s. Symbols have same meanings as in Figure 3.3a. The amplitudes of the direct arrivals are clipped to about 3km on either side.
Figure 3.3e. OBS F hydrophone channel data from Line 5. Plot made with a reducing velocity of 2 km/s. Symbols have similar meanings as in Figure 3.3a. The amplitudes of the direct arrivals are clipped till about 3km on either side.
Figure 3.3f. OBS E hydrophone channel data from Line 2. Symbols have same meanings as in Figure 3.3a. The amplitudes of the direct arrivals are clipped to about 3km on either side.
Figure 3.4a OBS A z component data from Line 3. Plot made with a reducing velocity of 2 km/s. Labels on the arrivals are as follows: (D) direct arrivals, (E1) reflection from base of L1, (E2) reflection from base of L2, (E3) reflection from base of L3, (rf) refraction from base of L1, and (m) multiples. Ringing is severe near zero offset.
Figure 3.4b. OBS B z component data from Line 1. Plot made with a reducing velocity of 2 km/s. Symbols have same meanings as in Figure 3.4a. Ringing problem is severe near zero offset.
Figure 3.4d: OBS D z component data from Line 2. Plot made with a reducing velocity of 2 km/s. Symbols have same meanings as in Figure 3.4a. Ringing problem is severe near zero offset.
Figure 3.4c. OBS F z component data from Line 5. Plot made with a reducing velocity of 2 km/s. Symbols have same meanings as in Figure 3.4a. Ringing is severe near zero offsets.
Figure 3.4f. OBS E z component data from Line 2. Plot made with a reducing velocity of 2 km/s. Symbols have same meanings as in Figure 3.4a. Ringing is severe near zero offset.
4. Methodology

Traveltime inversion reconstructs the velocity structure of a body given the measurement of traveltimes of waves that have propagated through it (Snider 1989). Conventionally, in a traveltime inversion problem, a simple starting model of the earth is assumed and rays are traced through it to predict traveltimes. Traveltime residuals are computed by subtracting the observed times ($t_o$) from the predicted times ($t_p$), $\delta t = t_o - t_p$. Typically, negative residuals mean that the estimated model has features that are slower than those in the real earth, and positive means faster. Traveltimes predicted even by the best earth model typically exhibit some scatter compared to the observed data. This discrepancy is usually due to unaccounted for heterogeneities in the earth’s structure, as well as the errors in data picking [Shearer, 1999]. The residuals are used to update the starting model. This updated model is used to predict a new set of traveltimes and the process is repeated iteratively such that the predicted traveltimes eventually match the observed ones to within a tolerance, which is determined by the picking uncertainties.

The advantage of inversion, as opposed to trial-and-error forward modeling, is that it can provide better estimates of model parameter resolution, uncertainty and non-uniqueness. The data can also be fit according to a specified norm. Examples of inversion applications for laterally varying structure using wide-angle traveltimes from controlled source data include White and Clowes (1990), Lutter et al. (1990), Hole (1992), Zelt and Smith (1992), McCaughey and Singh (1997), Zelt and Barton (1998), and Lutter et al. (1999).
4.1 2D Ray Theory

Seismic ray theory is analogous to optical ray theory that has been applied for over 100 years [Shearer, 1999]. It is intuitively easy to understand, simple to program, and can be generalized to two- and three-dimensional velocity models. The theoretical basis of ray theory is derived from the eikonal equation [Aki and Richards, 1980]. Forward modeling algorithms that are based on asymptotic ray theory [Cerveny et al., 1977] have been successfully developed and tested in the past [McMechan and Monney, 1980; Cassel, 1982; Spence et al., 1984; Zeli and Ellis, 1988].

Ray theory is an approximation of finite-frequency wave propagation by infinite frequency ray propagation [Zelt, 1999]. Ray theory therefore fails at long periods or within steep velocity gradients, and it does not predict non-geometrical effects such as diffracted waves [Shearer, 1999]. Due to this limitation, when rays are traced in a model with blocky parameterizations (models with sharp structures in a boundary or discontinuous changes in the velocity or velocity gradients), traveltimes may not be predicted over certain offset ranges. This is because model edges tend to scatter or focus ray paths. The ZS92 algorithm solves this problem by simulating smooth boundaries. Smooth boundaries help trace rays to the maximum number of observation points and thus include as much data in the inversion as possible for calculating the update to the starting model.
4.2 Damped Least-Squares Inversion

The traveltime $t$ between a source and receiver along a ray path $L$ is given in an integral form for a continuous velocity field $v(x,z)$ as

$$ t = \int_L \frac{1}{v(x,z)} \, dl. $$

The discrete form

$$ t = \sum_{i=1}^{n} \frac{l_i}{v_i} $$

is used in practical applications where $l_i$ and $v_i$ are the path length and velocity of the $i^{th}$ ray segment, respectively. Though the traveltime shows a linear relationship with slowness (reciprocal of velocity), traveltime inversion is a nonlinear problem as the ray path is dependent on velocity. It is possible to solve the nonlinear problem mathematically due to Fermat’s Theorem which states that the travel time along a ray path does not change to first order when the ray is perturbed [Nolet, 1987; Ben-Menahem and Singh, 1981]. The nonlinear system of equations is solved by expanding it about a starting model, using a Taylor’s series expansion, and neglecting higher order terms. The resulting system of equations is linear and can be written as:

$$ \delta t = G \delta m. $$

(4.2.1)

Here $G$ is the partial derivative matrix ($g_{ij} = \frac{\partial t_i}{\partial m_j}$, $t_i$ is the $i^{th}$ observed traveltime and $m_j$ is the $j^{th}$ model parameter). The model adjustment vector $\delta m$ is defined as the perturbation of the updated model $m$ from the starting model $m_0$ ($\delta m = m - m_0$). Similarly, the
traveltime residual vector $\delta t$ is defined as the perturbation of the observed traveltime vector $t_o$ from the predicted traveltime vector $t_p$ ($\delta t = t_o - t_p$). As a consequence of this expansion, and the fact that the ray paths and velocity model are unknown at the outset, a starting model and iterative approach are applied.

In the present study some of the model parameters are not resolved as good as others. This is due to inadequate ray coverage of some parts of the model compared to others and the fact that in some parts of the model the rays are traveling in essentially one direction only. Ray diagrams can illustrate this problem. As a result, the system of equations under consideration can be broken into an overdetermined and an underdetermined part. Such systems of equations are termed "mixed-determined". Equation (4.2.1) is therefore mixed-determined in our case and is solved using a damped least-squares technique [Aki and Richards, 1980]. In this approach, constraints are applied to the size of the data adjustment vector and the model adjustment vector. For this, the function $\Phi$ (called the objective function) that is to be minimized, is a combination of the prediction error and solution length,

$$\Phi(m) = E + \lambda^2 L = \delta t^T \delta t + \lambda^2 \delta m^T \delta m.$$  \hspace{1cm} (4.2.2)

Here $\lambda$ is a weighting factor also called the damping parameter. $\lambda$ determines the relative importance given to minimizing the prediction error $E$ and the solution length $L$. As such, $\lambda$ controls the step length in model space, or the "amount of linearization" applied to the non-linear problem in one iteration. The predicted model from any iteration becomes the

* When a system of equations has more unknowns than data, it is called "underdetermined", and if it has too much data to obtain an exact solution, it is called "overdetermined".
starting model for the next iteration. High $\lambda$ values mean smaller steps, which in turn keeps the updated model close to the starting model. This approach of obtaining a desired traveltime fit is known as a creeping method. In this thesis, the final model is obtained within a single iteration; probable reasons are discussed in section 5.3.

The damped least-squares solution to equation 4.2.1 is given by:

$$\delta m = (G^T C_i^{-1} G + \lambda C_m^{-1})^{-1} G^T C_i^{-1} \delta t,$$

which can also be expressed as

$$\delta m = G^{-x} \delta t,$$

where

$$G^{-x} = (G^T C_i^{-1} G + \lambda C_m^{-1})^{-1} G^T C_i^{-1}$$

is the generalized inverse. In equation (4.2.4) $C_i$ and $C_m$ are the a priori data and model covariance matrices given by

$$C_i = \text{diag} \{ \sigma_i^2 \}, \quad C_m = \text{diag} \{ \sigma_j^2 \}.$$

The standard deviation $\sigma_i$ is the estimated a priori uncertainty of the $i^{th}$ traveltime measurement and $\sigma_j$ is the a priori estimate of the uncertainty of the $j^{th}$ model parameter.

During the application of the ZS92 algorithm, the value $\sigma_j$ acts to determine the relative updates of the different parameter types, velocity and boundary nodes used in this thesis. Using equation (4.2.4) for solving the mixed-determined system of equations minimizes the overall error by damping the underdeterminancy of the system. Since the number of model parameters is generally much less than the number of observations, the matrix to
be inverted in equation (4.2.4) is relatively small and not particularly sparse. Therefore the matrix inversion is performed using a simple LU decomposition [Press et al., 1986].

As described above, the solution of a non-linear problem is sought by linearizing it. The pitfall of such an approach arises from the fact that the linearized methods do not “see” the entire error surface (equivalent to $\Phi(m)$ in equation (4.2.2)). If $m_n$ is the starting model for $n^{th}$ iteration, only a part of the error surface that is in the vicinity of $m_n$ is seen by the method. The inversion process approximates the rest of the surface as a paraboloid tangent to the actual surface at that point. The new estimated model, $m_{n+1}$ (updated from $m_n$ after the $n^{th}$ iteration) is the minimum of that paraboloid. Since any minimum with continuous derivatives is locally paraboloid in shape, the method will converge to a particular minimum of the error function if and only if the initial guess is linearly close enough [Menke, 1984]. This usually raises concerns therefore as to whether a local or a global minima has been found, whether or not there exists another local minima that can fit the traveltime data equally well, and how many such local minima exist [Zelt, 1999].

Non-uniqueness, as described above, means that there could be significantly different models that satisfy the data and the specified model constraints. A sense of non-uniqueness of the predicted model can be obtained from the uncertainties associated with the data. The greater the uncertainty, the greater is the possibility of obtaining models with different features that predict the data within the same error. In our case though, the data uncertainties are used more for balancing the wide-angle dataset versus the near-vertical dataset than giving an absolute error estimate to a traveltime pick. Instead, a
better sense of non-uniqueness and error is obtained by the extent to which the values
of depth and velocity nodes below OBSs at the intersecting points of the profiles are in
agreement in the final 2D models (discussed further in section 6.3). In general, using a
priori information can help to minimize model non-uniqueness in a non-linear problem.
In this study, the problem of non-uniqueness is solved to a large extent by using the MCS
data in combination with the OBS data. The MCS data provides a tight constraint on the
range of values that depth nodes can take.

4.3 Stopping Criteria

The traveltime fits are assessed using the normalized form of the misfit parameter,
\( \chi^2 \) [Bevington, 1969]. It can be expressed as,
\[
\chi^2 = \frac{1}{N} \sum_{i=1}^{N} \left( \frac{d_{i}^{obs} - d_{i}^{pre}}{\sigma_i} \right)^2.
\]
Here \( N \) is the total number of data points and \( \sigma_i \), the uncertainty associated with \( d_i \) (\( i^{th} \)
data point). The value of \( \chi^2 \) equal to 1 indicates that the data has been fit to within their
assigned uncertainties. Theoretically, the interpreter can stop iterating after a model gives
\( \chi^2 \) equal to 1. In practice though, apart from the \( \chi^2 \) error value there are several other
factors that can prove decisive. Modeling with the ZS92 algorithm involves a strong
element of subjective input. During the whole process of traveltime inversion, beginning
from the data preparation to choosing one model over another, a judgment call is required
at every step. The interpreter closely monitors the whole process of modeling and the
visual assessment of the model and the traveltime picks through each step is critical. The
interpreter can manually alter the value of a depth or velocity node; fix the value of one
or more model parameters; continue for the rest of the modeling with only a portion of
dataset; etc. depending on what feature(s) of the model is/are to be delineated. Depending
on whether or not the desired feature in the model has been resolved, or the model allows
rays to be traced to a sufficient number of observation points, the interpreter may choose
to stop at any point. In this thesis, the author defines certain criteria on which the models
are screened. Those models that satisfy all the criteria are considered “geologically
reasonable” and the criteria are further discussed in section 6.1.

4.4 Methodology and Limitations

The ZS92 algorithm is an iterative scheme that is based on a model
parameterization and method of ray tracing suited to both the forward and inverse steps
of the inversion problem. The number and positions of the velocity and boundary nodes
can be adapted to the numbers and positions of the sources and receivers, the subsurface
ray coverage, and the desired complexity of the final model. The method uses an efficient
numerical solution of the ray tracing equations, automatic computation of the ray take-off
angles and a simulation of smooth layer boundaries. The values of the partial derivatives
of the traveltimes with respect to the model parameters (velocity and depth nodes) used in
equation 4.2.1 are calculated analytically during ray tracing. The adjustment values for
the model parameters are calculated through a damped least-squares inversion scheme.

Inverting the near-vertical data requires that each pick be treated as a coincident
source-receiver pair. This effectively increases the number of equations in the system
4.2.1 without increasing the number of unknowns and adds more constraints (mainly on
the boundaries as opposed to velocity). The ray tracer performs the zero-offset modeling as a part of its usual operation. This way both the wide-angle and near-vertical data is lumped into one system of equations solved in one simultaneous inversion and the estimated model represents contributions from both datasets.

The data have a problem of balance. The refraction picks are few in number, roughly in the ratio of 1:15 compared to the reflection picks. Also, nearly twice as many picks come from the wide-angle data as compared to the near-vertical data. If an equal contribution in the model update is desired from both the wide-angle and near-vertical datasets, they have to be weighted accordingly. In this thesis, this was done by assigning uncertainties appropriate for the necessary weightings.

4.5 Computer Programs

There are three main program modules, VMODEL, RAYINVR and DMPLSTSQ. The main modules each have an input file (vm.in, r.in and d.in respectively) to provide them with input parameters for plotting and turning various features off or on. The file v.in contains information about the values of velocity and depth nodes and is used by all the programs. Apart from them a program called ZPLOT is used to display the data and pick them interactively.

VMODEL

This program is mainly used to check the velocity model file, v.in. VMODEL performs three main tasks associated with the file v.in. First, it checks the format of the
file \textit{v.in} to ensure it is as required by RAYINVR and indicates the location and type of any detected format error. It also checks for the presence of unusually small or large velocity values, negative vertical velocity gradients, large vertical or lateral velocity gradients, low-velocity zones, crossing layer boundaries, and boundaries with large slope changes. Second, VMODEL can edit the file \textit{v.in} by re-sampling the boundary and/or velocity nodes to increase or decrease the number of model parameters. Third, VMODEL can plot:

(1) The 2D model versus depth or two-way travel time,

(2) 1D velocity profiles versus depth or time,

(3) The RMS velocity variation across the whole model or for specific layers.

\textit{RAYINVR}

This module numerically solves the 2D ray-tracing equations using a Runge Kutta method and computes the partial derivatives of traveltime with respect to the model parameters (both velocity and the vertical position of the boundary nodes). A 2D (x,z) isotropic earth model is assumed. The velocity model is composed of a sequence of layers separated by boundaries consisting of linked linear segments of arbitrary dip. Layer boundaries must cross the model from left to right, but it is possible to reduce layer thickness to zero to model pinchouts or isolated bodies. The velocity within a layer is defined by velocity values specified at arbitrary x-coordinates along the top and bottom of the layer. In general, the x-coordinates at which the layer boundaries and the upper and lower velocities are specified can be independent within and between layers. In this thesis, due to the absence of any vertical-velocity-gradient within layers, the velocity
needs to be specified only at the top of the layer; the bottom of the layer assumes the same values as above. Velocity discontinuities across layer boundaries are allowed but not required. For the purpose of ray tracing, the model is automatically broken up into an irregular network of trapezoids, each with horizontal or inclined upper and lower boundaries and vertical left and right sides (Figure 4.1). The velocities at the four corners of the trapezoid are used to interpolate a velocity field within the trapezoid so that the velocity varies linearly along its four sides. In this thesis, only horizontal velocity gradients exist within a trapezoid. A simulation of smooth layer boundaries is possible in which the incident and emergent ray angles are calculated using the slope of the smoothed boundary. The sources may be positioned anywhere in the model and rays may be directed at any angle. Ray take-off angles are determined automatically by the program for ray groups specified by the user using an iterative shooting/bisection search method. Traveltimes are calculated by numerical integration along ray paths using the trapezoidal rule [Zelt and Smith, 1992].

A plot of the model and all rays traced may be produced along with a plot of reduced traveltime versus model x-position for the observed and calculated data. The traveltimes may correspond to any ray paths which can be traced through the model, being either first or later arrivals. The traveltimes residuals with respect to the observed data are also calculated. The traveltimes and partial derivatives are linearly interpolated to the observed seismogram locations. The partial derivatives and travel time residuals are output and used later as input to the program DMPLSTSQR.
**DMPLSTSQR**

This program applies the method of damped least-squares to the linearized inverse problem (Equation 4.2.4) to update the velocity model in the file `v.in` using the partial derivatives and the travelttime residuals calculated by the program RAYINVR.

**ZPLOT**

ZPLOT is an interactive plotting and picking program for seismic data. The main features of this program are:

1. Plotting, picking, filtering, calculating power spectra, traveltime overlaying, and editing the data interactively.
2. Data that is stored in z-format for this program is obtained from SEGY-format data.
3. Plotting of data in reduced time versus offset, model position, azimuth or trace sequential number with any reducing velocity is possible.
4. The data plots can have trace-normalized (common-maximum-amplitude), true-relative-amplitude or time-varying-gain amplitude scaling.
5. The computer code allows black & white or color postscript plots of the displays on the screen to be produced.
Figure 4.1. Example of model parameterization used by the inverse method of Zelt and Smith (1992). The three-layer model is constrained by 21 independent model parameters: 11 boundary nodes (red squares) and 10 velocity nodes (blue circles). Whenever a velocity or depth node is introduced in a layer, the program automatically subdivides the layer at that point. In this figure, the model is divided into 9 trapezoidal blocks for ray tracing purposes. If the velocity is not specified at the bottom of a layer, the velocity at the top of the layer is assumed. Only one boundary/velocity node needs to be specified, at the right corner, if no lateral variation is required within a layer.
5. Data Modeling

5.1 Data Picking

The preferred final model is only as good as the traveltime picks. The heart of the problem lies in identifying common events that can be consistently picked across the OBS and MCS data. The MCS and the OBS data were merged for identifying common events across them. A simple but innovative method of merging the OBS and the MCS data is described below.

As the MCS data and the pressure component of the OBS data were both recorded using hydrophones, the seafloor reflection on both records had a kick towards the right-handside*. Due to the similar nature of the seismograms and the fact that the arrivals at the near-zero offsets were clearer on the hydrophone channel data than in the z-component data, the portion of the SCS record (which is a subset of the MCS data) record lying between two inline OBSs was inserted between the positive offset half of one and the negative offset half of the other hydrophone channel record. When the profiles were adjusted such that direct arrival on all of them coincided, subsequent reflection events aligned themselves automatically. Thus, the reflection events were correlated across the SCS into the OBS hydrophone channel data near zero offset (Figure 5.1).

After common reflection events had been identified in both datasets, SCS data from Line 3 (as a test case) was used to construct a 1D starting P-wave velocity model. For the starting models, "1D" implies that the model has no lateral velocity variation

* Events from the same horizon have opposite polarities in the hydrophone channel and z-component data.
within each layer; layer boundaries in the starting model have lateral structures as their shape (in depth) is dictated by the shape of the events (in time) picked across the stack. Picks made on Line 3 were depth converted using velocity values that have been found to be reasonable for the top 200 m sediment section for similar geological settings (P. Hart and I. Pecher, personal communications 2001). Traveltimes were predicted with different starting 1D velocity models and compared with the arrivals to about 3 km offset in the hydrophone channel of the OBS data. The set of velocity values (Table 5.1) that predicted the traveltimes closest to the observed traveltimes were chosen to depth convert picks made on the stacks along all the other lines (Figure 5.2). The fact that the same velocity function could be used to predict events below all the OBSs to about 3 km offset suggests the absence of strong lateral velocity variation throughout the region. The starting model was “under-parameterized” to facilitate incorporation of more velocity and depth nodes if required by the data later during inversion. One velocity node was placed below each OBS in all the layers and more nodes were added only if required to fit the data.

The seafloor reflections have strong amplitudes and are picked densely (~20 picks/km) in order to predict the bathymetry with high lateral resolution. Picks made on later events are sparse (~5 picks/km) to help achieve smooth boundaries in keeping with the expected lateral resolution of the model.
Table 5.1. Velocity function of the starting models for all five lines.

<table>
<thead>
<tr>
<th>Layer</th>
<th>Velocity (km/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Water</td>
<td>1.49</td>
</tr>
<tr>
<td>L1</td>
<td>1.50</td>
</tr>
<tr>
<td>L2</td>
<td>1.60</td>
</tr>
<tr>
<td>L3</td>
<td>1.70</td>
</tr>
<tr>
<td>L4</td>
<td>1.80</td>
</tr>
</tbody>
</table>

Picks on the wide-angle data were made and the model was estimated in a bootstrap fashion. In this thesis, a “cycle” refers to the author going back and forth one time between data picking and model estimation. Using the starting model, rays are traced and reflection and refraction times from the base of L1, L2 and L3 are predicted. As discussed in section 3.2, the predicted traveltimes were laid over the z-component record of the OBS data using interactive software (described later). Overlaying the predicted traveltimes gave a general sense of direction that different phases should follow and served as a guide to picking events at far offsets. As a result of overlaying, some of the earlier picks were repositioned and new picks were made at farther offsets. This new set of picks was the observed data for the next cycle. In this way, the cycles were repeated until the picks had been made at a maximum number of receiver locations and a geologically reasonable model had been estimated. Picks were made out to the farthest possible offsets because arrivals at wider angles add relatively more constraints to the velocity model. Picking arrivals at far offsets was not straightforward as the phases were
difficult to identify due to the problems of weak signal, noise, and waveform change and overlap.

5.2 Modeling

As opposed to the conventional "layer stripping" approach for modeling using the ZS92 algorithm [Zelt and Smith, 1992; Zelt and Forsyth, 1994; Zelt and White, 1995], all phases in the present dataset and all model parameters were inverted simultaneously. Layer stripping is used when the starting model is not good and the model has to be estimated in parts (starting from shallower layers to deeper one). Whole model inversion is always preferred over layer stripping as the trade-offs between all model parameters can be considered (in a single set of formulated equations). In our case, due to a robust starting model and stable inversion (found out after modeling), the whole-model approach was followed. Using the principle of travel time reciprocity, the OBSs were treated as shot points and the original shotpoints at the sea-surface as receivers while modeling the data. The shot points and the receivers have coincident location for the MCS data.

Line 1

The Line 1 model is 14 km long and 1.5 km deep. It has two inline OBSs (Table 5.2.1) and 178 shotpoints at the sea-surface. The OBSs are 1.46 km apart.
Table 5.2.1 OBS locations for Line 1

<table>
<thead>
<tr>
<th>OBS</th>
<th>Depth below sea level (m)</th>
<th>X position in the model from left (km)</th>
<th>Latitude/Longitude (degrees)</th>
<th>Corresponding CDP in stack</th>
</tr>
</thead>
<tbody>
<tr>
<td>B</td>
<td>820</td>
<td>6.620</td>
<td>28.14/-89.65</td>
<td>1405</td>
</tr>
<tr>
<td>D</td>
<td>800</td>
<td>8.608</td>
<td>28.12/-89.63</td>
<td>1806</td>
</tr>
</tbody>
</table>

Two extra velocity nodes were required at model x positions 4.5 and 10.0 km both in L2 and L3 for the solution to converge. The solution converged within a single iteration (Table 5.2.2) without user intervention.

Table 5.2.2. Iterative Inversion for Line 1

<table>
<thead>
<tr>
<th>Iteration Number</th>
<th>Number of predicted traveltimes*/observed traveltimes</th>
<th>RMS travelt ime residual (ms)</th>
<th>$\chi^2$ error</th>
</tr>
</thead>
<tbody>
<tr>
<td>0</td>
<td>2613 / 2613</td>
<td>0.006</td>
<td>1.319</td>
</tr>
<tr>
<td>1</td>
<td>2611 / 2613</td>
<td>0.004</td>
<td>0.690</td>
</tr>
</tbody>
</table>

Line 2

The Line 2 model is 13.25 km long and 1.4 km deep. It has three OBSs (Table 5.2.3) and 173 shotpoints at the sea-surface.

---

* Total number of traveltimes predicted by the model in a particular iteration, that is, the number of picks for which rays could be traced to their receiver locations.
Table 5.2.3. OBS locations for Line 2

<table>
<thead>
<tr>
<th>OBS</th>
<th>Depth below sea level (m)</th>
<th>X position in the model from left (km)</th>
<th>Latitude/Longitude (degrees)</th>
<th>Corresponding CDP in stack</th>
</tr>
</thead>
<tbody>
<tr>
<td>D</td>
<td>800</td>
<td>5.053</td>
<td>28.12/-89.63</td>
<td>1105</td>
</tr>
<tr>
<td>F</td>
<td>830</td>
<td>6.476</td>
<td>28.14/-89.64</td>
<td>1394</td>
</tr>
<tr>
<td>E</td>
<td>865</td>
<td>7.872</td>
<td>28.15/-89.64</td>
<td>1674</td>
</tr>
</tbody>
</table>

The solution converged within a single iteration (Table 5.2.4). No user intervention was required. The wide-angle reflection phase from the base of the Holocene drape could not be identified below OBS E in the hydrophone channel or the z-component data.

Table 5.2.4 Iterative inversion for Line 2

<table>
<thead>
<tr>
<th>Iteration Number</th>
<th>Number of predicted traveltimes/observed traveltimes</th>
<th>RMS traveltime residual (ms)</th>
<th>$\chi^2$ error</th>
</tr>
</thead>
<tbody>
<tr>
<td>0</td>
<td>3240 / 3240</td>
<td>0.008</td>
<td>2.218</td>
</tr>
<tr>
<td>1</td>
<td>3233 / 3240</td>
<td>0.004</td>
<td>0.806</td>
</tr>
</tbody>
</table>

**Line 3**

The Line 3 model is 12.5 km long and 1.5 km deep. It has three OBSs (Table 5.2.5) and 157 shotpoints at the sea-surface.
Table 5.2.5 OBS locations for Line 3

<table>
<thead>
<tr>
<th>OBS</th>
<th>Depth below sea level (m)</th>
<th>X position in the model from left (km)</th>
<th>Latitude/Longitude (degrees)</th>
<th>Corresponding CDP in stack</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>860</td>
<td>3.951</td>
<td>28.15/-89.65</td>
<td>866</td>
</tr>
<tr>
<td>B</td>
<td>820</td>
<td>5.628</td>
<td>28.13/-89.65</td>
<td>1209</td>
</tr>
<tr>
<td>C</td>
<td>750</td>
<td>6.710</td>
<td>28.12/-89.65</td>
<td>1497</td>
</tr>
</tbody>
</table>

Two extra velocity nodes were required at model x positions 2.0 and 8.0 km in both L2 and L3 for the solution to converge. The solution converged within a single iteration (Table 5.2.6) without user intervention.

Table 5.2.6 Iterative inversion for Line 3

<table>
<thead>
<tr>
<th>Iteration Number</th>
<th>Number of predicted traveltimes/observed traveltimes</th>
<th>RMS traveltimes residual (ms)</th>
<th>$\chi^2$ error</th>
</tr>
</thead>
<tbody>
<tr>
<td>0</td>
<td>3375 / 3375</td>
<td>0.008</td>
<td>1.971</td>
</tr>
<tr>
<td>1</td>
<td>3374 / 3375</td>
<td>0.004</td>
<td>0.605</td>
</tr>
</tbody>
</table>

**Line 4**

The model along Line 4 is 12.0 km long and 1.5 km deep. It has two OBSs (Table 5.2.7) placed 1.47 km apart and 145 shotpoints at the sea-surface.
Table 5.2.7 OBS locations for Line 4

<table>
<thead>
<tr>
<th>OBS</th>
<th>Depth below sea level (m)</th>
<th>X position in the model from left (km)</th>
<th>Latitude/Longitude (degrees)</th>
<th>Corresponding CDP in stack</th>
</tr>
</thead>
<tbody>
<tr>
<td>C</td>
<td>750</td>
<td>3.929</td>
<td>28.12/-89.65</td>
<td>767</td>
</tr>
<tr>
<td>D</td>
<td>800</td>
<td>5.406</td>
<td>28.12/-89.63</td>
<td>1060</td>
</tr>
</tbody>
</table>

Two extra velocity nodes were required at model x positions 3.0 and 6.0 km both in L2 and L3 for the solution to converge. The $\chi^2$ error goes down to 0.605 within a single iteration. The predicted reflection traveltimes from the base of L3 in the inverted model show a systematic mis-match with the observed data towards the east. The value of the velocity node at the model x position of 6.00 in L3 was changed manually from 1.64 to 1.57 to reduce the mis-match (Table 5.2.8). This helped to improve the fit to the wide-angle data but the overall $\chi^2$ error went up to 0.991 as the fit to the near-vertical data near that region deteriorated.

Line 4 was re-modeled by adding 78 more picks along the reflected phase from the base of L3 towards the east as the author was doubtful about the exact offset to which the reflection phase could be distinctly identified and classified due to the presence of noise. The value of the velocity node at model x position 6.0 km in L3 goes down to 1.48 km/s. This new model is estimated from 3096 data points, has 0.005 ms RMS traveltime residual, and a $\chi^2$ error of 0.842. A discussion of this alternate model is provided in section 6.4.
Table 5.2.8 Iterative inversion for line 4

<table>
<thead>
<tr>
<th>Iteration Number</th>
<th>Number of predicted traveltimes/observed traveltimes</th>
<th>RMS traveltime residual (ms)</th>
<th>$\chi^2$ error</th>
</tr>
</thead>
<tbody>
<tr>
<td>0</td>
<td>3072 / 3072</td>
<td>0.009</td>
<td>3.353</td>
</tr>
<tr>
<td>1</td>
<td>3064/ 3072</td>
<td>0.005</td>
<td>0.991</td>
</tr>
</tbody>
</table>

**Line 5**

The line 5 model is 11.0 km long and 1.5 km deep. It has two OBSs (Table 5.2.9) placed 1.46 km apart and 137 shotpoints at the sea-surface.

Table 5.2.9 OBS locations for Line 5

<table>
<thead>
<tr>
<th>OBS</th>
<th>Depth below sea level (m)</th>
<th>X position in the model from left (km)</th>
<th>Latitude/Longitude (degrees)</th>
<th>Corresponding CDP in stack</th>
</tr>
</thead>
<tbody>
<tr>
<td>C</td>
<td>750</td>
<td>3.929</td>
<td>28.14/-89.63</td>
<td>1077</td>
</tr>
<tr>
<td>D</td>
<td>800</td>
<td>5.406</td>
<td>28.14/-89.65</td>
<td>1370</td>
</tr>
</tbody>
</table>

Two extra velocity nodes were required at model x positions 4.0 and 7.5 km both in L2 and L3 for the solution to converge. The solution converged in a single iteration (Table 5.2.10) and no user intervention was required.
Table 5.2.10 Iterative inversion for Line 5

<table>
<thead>
<tr>
<th>Iteration Number</th>
<th>Number of predicted traveltimes/observed traveltimes</th>
<th>RMS traveltime residual (ms)</th>
<th>$\chi^2$ error</th>
</tr>
</thead>
<tbody>
<tr>
<td>0</td>
<td>2293 / 2300</td>
<td>0.007</td>
<td>1.402</td>
</tr>
<tr>
<td>1</td>
<td>2300 / 2300</td>
<td>0.004</td>
<td>0.773</td>
</tr>
</tbody>
</table>

5.2.1 Pick Uncertainties

The determination of pick uncertainties associated with the traveltimes used during the traveltome inversion was based on a combination of (1) initial time corrections applied to the data, (2) the dominant frequency of the data, and (3) the ease of convergence of the solution. After the time corrections were applied to the data using the method described in section 3.3, the traveltome picks along shot line 3 were inverted with different uncertainty values. Starting from 10 ms, the uncertainties were lowered in steps of 2 ms during each test run. Based on the dominant frequency in the near-vertical data, its associated uncertainty cannot be reduced below 4 ms. An uncertainty of 6 ms for the wide-angle data and 4 ms for the near-vertical data was found to be optimal. This was the limit that the uncertainties could be lowered such that the solution would still converge to yield a geologically reasonable model. The traveltome picks along all the shot lines were assigned the uncertainty values obtained from this test.

5.2.2 The $\lambda$ Values

The value of $\lambda$ for the application of equation (4.2.3) is determined by a trial and error method. Test runs of the inversion were made with different $\lambda$ values. During the
test runs it was found that for small $\lambda$ values the solution converged rapidly and the final model had more structure and more lateral velocity gradients compared to models obtained with larger $\lambda$ values. For relatively large $\lambda$ values, the data typically did not converge at all. A value of 3000 for $\lambda$ turned out to be optimal for all lines: the solution converged within a single iteration and the estimated model was geologically reasonable.

The fact that the solution converges within a single iteration is not typical. In this thesis, it is because (1) of the absence of strong lateral velocity gradients in the region, (2) the models consist of relatively few model parameters, (3) a good starting model was being used, and (4) in the joint inversion constraints are applied to boundary (through near-vertical data) and velocity (through wide-angle data) nodes simultaneously.

5.2.3 $\chi^2$ Error

An overall $\chi^2 < 1$ is obtained from the preferred final models for all five lines. When the wide-angle and near-vertical traveltimes are predicted from the models of all five lines, $\chi^2$ less than and greater than 1 are obtained, respectively (Table 5.3). $\chi^2 > 1$ for the near vertical data indicates that it has sampled more heterogeneities than can be resolved by the combined traveltime datasets. $\chi^2 < 1$ indicates the possibility that the wide-angle picks have less associated uncertainty than assigned currently. Figures 5.3 shows the traveltime misfits of the wide-angle and near-vertical datasets and Figure 5.4 shows the corresponding ray paths. $\chi^2$ errors associated with individual phases (Table 5.4) from the preferred final models of all five lines have $\chi^2$ error values both greater and
less that 1. Reflections from deeper parts of the model have $\chi^2 > 1$. For the refractions, $\chi^2 > 1$ indicates that later velocity variations in L1 could not be resolved appropriately. In this thesis, $\chi^2$ for individual phases (reflections from L1, L2 and L3, and refractions) and individual datasets (wide-angle and near-vertical) was monitored so that each would be as close to unity as possible, and as a result the overall $\chi^2$ is less than unity.

Table 5.3. Wide-angle and near-vertical $\chi^2$ errors for five lines

<table>
<thead>
<tr>
<th>Line number</th>
<th>Wide-angle $\chi^2$ error</th>
<th>Near-vertical $\chi^2$ error</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.472</td>
<td>1.863</td>
</tr>
<tr>
<td>2</td>
<td>0.611</td>
<td>1.440</td>
</tr>
<tr>
<td>3</td>
<td>0.529</td>
<td>1.027</td>
</tr>
<tr>
<td>4</td>
<td>0.577</td>
<td>2.01</td>
</tr>
<tr>
<td>5</td>
<td>0.786</td>
<td>1.659</td>
</tr>
</tbody>
</table>

Table 5.4. $\chi^2$ error statistics for individual phases for OBS and MCS from all five lines.

<table>
<thead>
<tr>
<th></th>
<th>Line 1</th>
<th>Line 2</th>
<th>Line 3</th>
<th>Line 4</th>
<th>Line 5</th>
</tr>
</thead>
<tbody>
<tr>
<td>Reflection from Base of L1</td>
<td>0.610</td>
<td>0.515</td>
<td>0.502</td>
<td>0.618</td>
<td>0.510</td>
</tr>
<tr>
<td>Reflection from Base of L2</td>
<td>0.547</td>
<td>0.653</td>
<td>0.693</td>
<td>0.514</td>
<td>0.886</td>
</tr>
<tr>
<td>Reflection from Base of L3</td>
<td>0.842</td>
<td>1.049</td>
<td>0.601</td>
<td>1.407</td>
<td>0.684</td>
</tr>
<tr>
<td>Head waved from base of L1</td>
<td>1.650</td>
<td>1.499</td>
<td>0.706</td>
<td>1.626</td>
<td>2.081</td>
</tr>
</tbody>
</table>
Except for the model of Line 5, no other model could trace rays to all the observation points. A model is not necessarily rejected rays cannot be traced to all the observation points because (1) the corresponding arrivals may have very small amplitudes, due to the arrivals being diffractions, and hence a ray tracing algorithm such as ZS92 will have a hard time finding these rays and/or (2) the range of take off angles needed to hit a particular receiver location may be small enough such that the ray shooting/bisection algorithm used in ZS92 was not able to trace rays to within a tolerance range around the receiver position. A model is only rejected if it is found that it fails to trace rays to roughly more than 2% of the observation points. This number is fairly typical while modeling with the ZS92 algorithm.

5.3 Model Assessment

In order to check the consistency of the results obtained by the inversion along all five lines, 1D velocity-depth profiles below each OBS at the profile intersection points were overlaid (Figure 5.5a). The profiles generally overlie exactly with occasional mismatches of 10 m/s in velocity and up to 20 m in the depth of boundaries. This suggests that the velocity nodes have been resolved to a minimum accuracy of 10m/s and the depth of boundary nodes to about 20 m. 1D profiles below OBS A and E are also overlaid (Figure 6.2b). The close match between these velocity-depth profiles suggests that the sediment properties are fairly constant in the northern part of the study area (Figure 1.3).
As shown by the final models, since the lateral velocity variations in the region are subtle, the modeling of Line 2 was re-done with only one velocity node in each layer. The depth nodes, as previous, were free to change their values (in z-direction only) if need be. The fact that the data did not converge (Table 5.5) suggests that the lateral velocity variations in the preferred final models are not merely artifacts of the inversion but are required by the data.

Table 5.5. Summary of inversion performed with the velocity nodes fixed for Line 2.

<table>
<thead>
<tr>
<th>Iteration Number</th>
<th>Number of data points used</th>
<th>RMS traveltime residual (s)</th>
<th>Normalized $\chi^2$ error</th>
</tr>
</thead>
<tbody>
<tr>
<td>0</td>
<td>3240</td>
<td>0.008</td>
<td>2.218</td>
</tr>
<tr>
<td>1</td>
<td>3239</td>
<td>0.006</td>
<td>1.420</td>
</tr>
<tr>
<td>2</td>
<td>3240</td>
<td>0.006</td>
<td>1.423</td>
</tr>
<tr>
<td>3</td>
<td>3240</td>
<td>0.006</td>
<td>1.423</td>
</tr>
</tbody>
</table>

The starting model had velocity values of 1.6 km/s and 1.7 km/s in L2 and L3. The final model had velocity values of 1.61 km/s and 1.67 km/s in L2 and L3.
Figure 5.1a. Events identified in the data along Line 3. SCS record is inserted between inline OBS hydrophone channel records.

In figures 5.1a through 5.1c, green arrows indicate reflection event from the base of L1. This event on the SCS record is marked as Event 1. Blue arrows indicate the reflection event from the base of L2. This event on the SCS record is marked as Event 2. Red arrows indicate reflection event from the base of L3. This event on SCS record is marked as Event 3.
SCS Line 3 combined with OBS A and B hydrophone channel

SCS Line 3 combined with OBS B and C hydrophone channel

Figure 5.1a (continued)
Figure 5.1 b. Events identified in the data along Line 2. SCS record is inserted between inline OBS hydrophone channel records.
Figure 5.1b. (continued)
Figure 5.1 c. Events identified in the data along Line 4. SCS record is inserted between inline OBS hydrophone channel records.
Figure 5.1 d. Events identified in the data along Line 1. SCS record is inserted between inline OBS hydrophone channel records.
Figure 5.1 e. Events identified in the data along Line 5. SCS record is inserted between inline OBS hydrophone channel records.
Figure 5.2. Starting models for each line. Layer boundaries are shown as solid lines. P-wave velocity values are labeled. OBSSs are labeled and their positions are shown as circles.
Figure 5.3a. Comparison of the observed wide-angle traveltimes with the traveltimes predicted from the final model of each line. Plots are made with a reducing velocity of 2 km/s. Colored picks are the observed traveltimes with their lengths proportional to their associated uncertainties. The predicted traveltimes are plotted as black lines. Also shown are the names and locations (as circles) of the OBSs. Color code is same as in Figure 5.1.
Figure 5.3b. Comparison of the observed near-vertical traveltimes in the stack with the traveltimes predicted by the final models along all lines. Colored picks are the observed traveltimes with their lengths proportional to their associated uncertainties. The predicted traveltimes are plotted as black points. The color code is the same as in Figure 5.1. Light blue color arrivals stands for the reflection from the seafloor that are not modeled as part of the traveltime inversion.
Figure 5.4a Ray diagram of the reflection from the base of L1. Layer boundaries are indicated as dashed lines.

For clarity every fifth wide-angle ray is shown.
Figure 5.4(b) Ray diagram of the reflection from the base of Layer 2. Layer boundaries are indicated as dashed lines. For clarity, every fifth wide-angle ray is shown.
Figure 5.4c Ray diagram of the reflection from the base of L3. Layer boundaries are indicated as dashed lines. For clarity every fifth wide-angle ray is shown.
Figure 5.4d Ray diagram of refracted arrivals from the base of L1. Layer boundaries are indicated as dashed lines. For clarity every fifth wide-angle ray is shown.
OBS B from Lines 1,3 and 5

OBS C from Lines 3 and 4.

OBS D from Lines 1,2 and 4

OBS F from Lines 2 and 5

Figure 5.5a. 1D velocity profiles below the OBSs overlaid at the line intersection points. In all plots, the profiles from Line 1 are black, Line 2 are red, Line 3 are green, Line 4 are magenta and Line 5 are blue. Vertical axis in the plots are two-way reflection times.
Figure 5.5b. 1D velocity profile below OBS A (in black) and OBS E (in red). Vertical axis in the plots are two-way reflection times.
6. Results and Interpretation

6.1 Selection Criteria for Preferred Models

Using the ZS92 traveltime inversion algorithm, P-wave velocity models of the five shot lines were estimated. The preferred final model of each line satisfied the following criteria:

1. Each model was constructed with a minimum possible number of velocity and depth nodes as required by the data to predict the traveltimes within their associated uncertainties.

2. Each model predicts all possible phases identified in the data along that line and traces rays to the maximum number of observation points.

3. Each model has only those features that are required by the data, as opposed to being consistent with the data. Due to the absence of any core, well log or geophysical data other than seismic, the extent of the interpretation that can be made from the P-wave velocity models is limited. The models have therefore been kept structurally simple to avoid over-interpretation.

6.2 Meaning of Velocities from 2D Traveltime Inversion

It is well known that over a thickness of rocks, velocity generally varies as a function of depth. During analysis and interpretation though, it is much easier to deal with constant velocity units rather than vertical velocity functions. Therefore, conventional processing techniques use the concept of interval velocity (which is essentially the average velocity within a layer) wherever such simplifications are convenient. In this thesis, as there are no vertical velocity gradients within the layers, the
velocities that traveltime inversion estimates can be related to interval velocities from conventional processing schemes. A conceptual comparison of inverted velocity, as obtained in this thesis, is made with the interval velocity.

Interval velocities can be computed from RMS velocities (using Dix’s equation), which in turn can be calculated directly from shot gathers*. Theoretically, it is the velocity for which a particular event on the stack collapses to yield the sharpest arrival on the “intuitively” best trace that can be obtained from a stack. The interval velocity for a layer is computed from the RMS velocity used to collapse reflection events from the top and bottom of that layer. Computation of the RMS velocity from shot gathers assumes a horizontally layered or a gently dipping layered earth and no lateral velocity over the width of the gather. This means that to bring out lateral heterogeneity within a layer in 2D using a conventional processing approach, (1) the earth has to be broken down into as many “1D” parts as necessary, (2) interval velocities must be determined for each of the 1D parts, and (3) the 1D models are combined by interpolation to yield a 2D model. On the other hand, the traveltime modeling approach in this thesis honors the 2D structural complexity and lateral velocity variation in the subsurface while estimating the in situ velocities. Instead of calculating velocities that make the data look like the best structural model, traveltime inversion estimates the 2D velocity-depth model of the earth itself. The inverted velocity in our case therefore represents the true velocity within a layer (to within the limitations of the data resolution and the modeling approach).

* The most common type of velocity used in conventional seismic processing is the stacking velocity obtained from common depth point stacks. The stacking velocity is often synonymous with RMS velocity [Al-Chalabi, 1974].
6.3 Model Descriptions

Line 1

Line 1 has a NW-SE orientation. The SCS data shows a HRZ below OBS D within L3. OBS D has a model x position of 8.60 km. Overall, there is a lateral lowering of velocity from 1.70 km/s at both ends of the model to 1.66 km/s below OBS D in L3 (Figure 6.1a). The model also shows a lowering in velocity from 1.61 km/s from the ends to 1.59 km/s below OBS D and B in L2.

Line 2

Line 2 has a S-N orientation. The SCS data shows a HRZ between OBSs D and F in L3. The model shows a lowering in velocity from 1.70 km/s below OBS E at the north end of the model to 1.67 km/s below OBS F to 1.66 km/s below OBS D in L3 (Figure 6.1b). The model shows a similar behavior in L2 where the velocity drops from 1.64 km/s to 1.60 km/s to 1.59 km/s below OBSs E, F and D, respectively.

Line 3

Line 3 has a N-S orientation. The SCS data shows a HRZ below OBS B in L3. The model shows a gentle lowering of velocity from 1.70 km/s below OBS A at the model x position of 3.95 km to 1.69 km/s below OBS B at a model x position of 5.62 km in L3 (Figure 6.1c). The model also shows a similar behavior in L2 where the velocity decreases from 1.64 km/s below OBS A to 1.60 km/s below OBSs B and C.
Line 4

Line 4 has a W-E orientation. The SCS data shows the presence of a HRZ in L3 from the model x position 5 km to 6.5 km. The model x position of OBS D is 5.4 km. The model shows a gradual decrease in velocity from 1.71 km/s from the western end of the line to 1.66 km/s below OBS D in L3 (Figure 6.1d). This corresponds to a velocity gradient of -0.03 s⁻¹. There is a steep increase in the velocity gradient to 0.15 s⁻¹ eastwards thereafter and the velocity decreases to 1.57 km/s at the model x position of 6 km. This sharp change in velocity indicates a rapid change in sediment properties or composition. L2 on the other hand shows a slight increase in velocity from 1.58 km/s below OBS D to 1.59 km/s at 6 km. L2 shows a gradual decrease in the velocity from 1.65 km/s at the west end of the model to 1.58 km/s below OBS D.

Line 5

Line 5 has a E-W orientation. The SCS data shows a HRZ in the model from 5 km to 7 km in L3. OBS F is located above HRZ. The x position of OBS F in the model is 6.69 km. The model shows a gradual lowering of velocity from both ends towards OBS F in L2 and L3 (Figure 6.1e). In L3, the velocity decreases from 1.71 km/s at 7.5 km to 1.69 km/s at 4 km to 1.67 km/s below OBS F. In L2 the velocity decreases from 1.61 km/s at both ends to 1.59 km/s below OBS F.

Seismic horizons have been picked on the MCS reflection stack in such a way that the HRZs always lie within L3. Along any line, one or two velocity nodes were placed
within L3 in the starting models corresponding to the location and extent of the HRZs in the data. This approach helped to better resolve the velocities within the HRZs. It can be concluded from the final models of all lines that the location of the HRZs in the near-vertical data coincides with the location of the lowest velocity in L3.

6.4 Low Velocity Zone

Low velocity zones (LVZs) represent a significant deviation from the usual increase of velocity with depth in the earth. LVZs cause geometrical shadow zones at the surface for the refracted arrivals and/or form a significant lateral heterogeneity. Therefore, unless there is strong evidence for their existence from interpreted data or their presence is absolutely necessary to fit the data, they are considered geologically unreasonable [Zelt, 1999] and are avoided. The presence of free gas can give rise to LVZs [Domenico, 1977].

As the present problem deals with a speculated hydrate-gas system where the presence of hydrates has been confirmed by shallow coring but the presence of gas can only be concluded indirectly through the interpretation of the velocity models, it would not be unexpected if LVZs were part of the final models. In fact, even an alternate model for Line 4 suggests the presence of a LVZ in the region (see section 5.2). This model was estimated by inverting the same dataset used to estimate the Line 4 model described in section 5.2 but with 78 extra wide-angle traveltime picks made on the reflection phase from the base of L3 east of OBS D (Figure 6.2a). All other features remaining the same, the velocity within the LVZ drops from 1.57 km/s to 1.48 km/s in the alternate model.
(Figure 6.2b). Though the extent of the LVZ in the alternate model coincides with the extent of HRZ in the SCS data and its presence seems to be a requirement of the data, this specific value of velocity within the LVZ is doubtful due to reasons like (1) the LVZ is controlled mainly by a single OBS, and (2) the extra picks that were responsible for the LVZ may have been made on diffracted arrivals instead of a reflected arrival. Moreover, as Lines 2 and 3 that pass nearby this region do not support the presence of such a low velocity value, it may be an artifact of the modeling.

6.5 Interpretation

The surface topography (Figure 1.3) is caused by the deformation due to the active salt domes in the vicinity of the study area (Figure 6.3). It is well known that salt tectonics in the GOM causes intense fracturing and faulting. Though such faults and fractures have not been modeled in the present study, it is possible, as in the case of MC853 [Sassen et al., 2001], that the gas present in the study area has a deeper source and comes up along salt and fault conduits (Figure 6.4). Though E3 (Figure 6.7) domes below the HRZs, it is not interpreted as the top of the salt as the velocity estimated by the traveltime inversion in L3 is too low for a salt body. The study of the relationship between salt movement and the structures found in the study area is beyond the scope of this thesis. Also, the presence of pathways that may from conduits for the migration of the gas up from the deeper sources is only a speculation based on other studies in nearby areas [e.g., Sassen et al., 2001].
The contour map of velocity within L2 shows a systematic decrease in velocity towards the diapir (Figure 6.5a). The lowest velocity of 1.57 km/s occurs beside OBS D and the diapir near Line 4. This zone is referred to as LVZ2 hereafter. The thickness of L2 at LVZ2 is ~120 m. LVZ2 possibly has the maximum amount of free gas in L2. The contour map of the base of L2 sows a systematic increase in elevation towards the diapir (Figure 6.5b). The combination of both maps (Figures 6.5a and 6.5b) suggests that LVZ2 sits on a structural high.

The contour map of the interval velocity within L3 shows a velocity distribution of 1.66 – 1.72 km/s throughout the region except in the SE part where the presence of a LVZ is indicated (Figure 6.6a). This zone has a velocity of ~1.56 km/s and occurs east of OBS D along Line 4. This zone is referred to as LVZ3 from now on. The thickness of L3 at LVZ3 is ~200 m. The contour map of the base of L3 (Figure 6.6b) shows a structural high below LVZ3. As in the case of LVZ2, it is plausible that LVZ3 has the maximum amount of free gas in L3. Unfortunately since the ray coverage around LVZ3 is not as good as around LVZ2 (LVZ2 lies partially within the 1.5 km X 1.5 km box bounded by OBSs B, C, D and F that has the maximum ray coverage), the absolute value of velocity nodes in LVZ3 may not be as reliable. Nevertheless, a lowering of velocity around LVZ3 is indicated by the models of all lines. The combination of maps (Figures 6.6a and 6.6b) suggests that LVZ3 also sits on a structural high.

It is well known that free gas tends to accumulate on the top of structures and therefore the fact that both LVZ2 and LVZ3 sit on structural highs further supports the
argument that the lowering of velocity in LVZ2 and LVZ3 is due to the presence of free gas. Whether or not the accumulation of free gas in LVZ2 and LVZ3 is also due to those regions being good quality reservoirs depends on the geological interpretation of the seismic data. For the purpose of interpreting the SCS data, they were phase-shift migrated [Stolt, 1978] using the velocities obtained from traveltime inversion (Figure 6.7). A comparison of the migrated SCS data is made with the estimated models in Figure 6.8 to illustrate the behavior of the horizons in time and depth. Understanding the nature of the reflection (labeled F in Figures 6.7a, 6.7d and 6.7e) from the top of the wedge-shaped feature (labeled W) in Figures 6.7a, 6.7d and 6.7e is crucial for the interpretation. (F) has a largely negative polarity (Figure 6.9) and it disappears before it intersects with E2 (best illustrated in Figure 6.7a). Two possible interpretations of the characteristics of the reflections from the top of the wedge-shaped feature combined with the locations of the HRZs have been suggested.

According to the first interpretation, the wedge-shaped feature is caused by overbank deposits of a channel**. This is a typical case of a channel deposit where the deposits along the channel axis are sandier and give rise to HRZs, and the overbank deposits are shalier and have the shape of a wedge. Due to deposits in the overbank being shaly, reflections from its top will have a largely negative polarity. In this case though, the sandy deposits along the channel axis are expected to have higher P-wave velocity than its surroundings. In this thesis though, the fact that HRZs are associated with a lateral lowering of velocity suggests that free gas is present along the channel axis. The

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* There is no velocity discontinuity across F as the wedge (W) is included in L2 while modeling.
** This channel should not be confused with the Mississippi River.
reflectivity along these zones are further enhanced due to the presence of free gas, as in the case of the Blake Ridge [Holbrook et al., 1996]. The sandy deposits along the channel can act as potential reservoirs for free gas that come up from a deeper source, if sealed effectively on the top and sides. The fact that the channel axis appears to be sitting on top of a structure (feature Z in Figure 6.7) and has higher elevation than it banks puts a question mark before this interpretation though. It can only be explained by assuming that the channel was situated in a structural low in the past but has been uplifted due to salt movement in due course of time. Since the ages of the horizons are unknown, the channel interpretation is difficult to defend.

According to the second interpretation, the reflection from top of the wedge-shaped feature is a BSR, i.e., it marks the gas hydrate – free gas contact. The stability diagram (Figure 2.2), constructed with a reasonable bottom water temperature (8°C) and range of geotherms (25 – 40°C/km) for this area [Sassen et al., 2001], estimates the depth of the base of the GHZS close to the top of the wedge and supports this interpretation. The shape and the nature of reflectivity of (F) can be explained by assuming that the concentration of the gas decreases laterally away from the HRZs (Figure 6.9). Due to a decrease in the gas concentration, the strength of the BSR decreases and the BSR eventually disappears when the gas concentrations are so low that the hydrates and free gas are no more in contact (Figure 6.10). Again, assuming that HRZs (containing free gas) are sitting close to the top of salt, the BSR deepens away from the HRZ (Figure 6.7a) because the salinity of the water decreases away from the salt dome. This explains the shape of (F) in Figures 6.7a, 6.7d and 6.7e. Unfortunately, in
the absence of knowledge of the exact bottom water temperature and the geotherm in MC798, nothing concrete can be concluded.

The final models derived using the minimum-parameter traveltime inversion of Zelt and Smith (1992) are compared with the models of the Mississippi Canyon obtained by Coleman et al. (1983) and Goodwin and Prior (1989) to speculate on the age of the horizons. It must be stated upfront though, that the ages estimated through this exercise are speculative and presents only one out of many possibilities. The seismic lines used in the present study are situated about 60 km SE of the lines used by Coleman et al. (1983) and Goodwin and Prior (1989). Lines 2 and 3 among the lines in the present study have a similar orientation with respect to the canyon as the models used for comparison. Line 3 has been specifically used for the comparison because of the clarity of events.

Coleman et al. (1983) developed a model of the Mississippi Canyon from a reflection profile about 60 km NW of the present study area in a water depth of about 400 m (Figure 6.11). In this model, the depth to the base of the canyon was around 1080 m below sealevel, with a sedimentary fill of about 700 m. The base of the canyon was dated to be 25,000 yrs old. Another prominent horizon, which marked the beginning of the infilling of the canyon, is dated to be 20,000 years old. The latest fill, or the pelagic sediments, were speculated to be deposited over the last 10,000 – 12,000 yrs. It was noted that even though the greatest thickness of the Holocene deposition is present along the axis of the canyon, the material drapes over the adjacent walls.
Goodwin and Prior (1989) interpreted another seismic line in the same region as Coleman et al. (1983) (*Figure 6.12*). In Borehole No. 1 along the seismic line several thin gravel zones (3 – 6 m thick) were encountered near base of the canyon. The high amplitude reflections in the seismic data suggest that such granular zones were common near the base. Borehole No. 1 also recorded the presence of an unconformity at a depth of 533.5 m. Above the unconformity all the samples had Carbon-14 ages of 19,000 yrs BP or younger. This unconformity may be the same horizon that marks the beginning of the infilling of the canyon in Coleman et al.’s model above.

As mentioned earlier, three strong events can be identified along all the SCS reflection profiles. Event 3 consists of closely spaced strong but discontinuous reflection events. These reflections may arise from the closely-spaced granular zones that were recorded by Goodwin and Prior (1989) near the base of the canyon. Also, the shape of Event 3 in time and depth in the final models in this thesis closely resembles the shape and position of the base of the canyon in the models developed by Coleman et al. (1989) and Goodwin and Prior (1989). Hence, Event 3 may be the canyon base unconformity and can be dated at 27,000 yrs BP. The channel sits on top of another strong and coherent event. This event may represent a canyon fill unconformity. In this thesis during modeling, Event 2 that was picked includes the top of the channel at its axis and bottom of the channel at the ends of the shot lines. Since this unconformity could not be related to any other events in the literature, a date could not be assigned to it.
The sustenance of a hydrate system requires a dynamic equilibrium between gas escaping into the GHSZ and gas diffusing out of it. If the first hypothesis is correct, the presence of hydrates within the diapir indicates that gas is being fed continuously into the diapir from the free gas accumulation in the axis of the channel. The axis of the channel, being more porous, acts as a reservoir but the material around the reservoir, being more shaly, is much less porous and permeable. This reduces the chances of gas being present in the study area apart from the channel axis, and therefore the hydrates may not be present anywhere else apart from the region close to the diapir. If the second hypothesis is correct though, hydrates may be prevalent in the basin. The presence of hydrates is definite above the BSR but nothing can be said about other parts of the study area. Using traveltimes only, nothing can be said about the absolute concentration of free gas in the HRZS or its relative concentration compared to any other place in the study area. This is due to the lack of any reference velocity for the sediments. Therefore, in the absence of any core samples or well data, only a qualitative interpretation was possible.
Figure 6.1a. Final model for Line 1. P-wave velocities are shown in the model at the locations of the velocity nodes. Also shown are the locations of OBSs B and D as solid circles.
Figure 6.1b Final model for Line 2. P-wave velocities are shown in the model at the locations of the velocity nodes. Also shown are the locations of OBSs D, F and E as solid circles.
Figure 6.1c. Final model for Line 3. P-wave velocities are shown in the model at the locations of the velocity nodes. Also shown are the locations of OBSs A, B and C as solid circles.
Figure 6.1d. Final model for Line 4. P-wave velocities are shown in the model at the location of the velocity nodes. Also shown are the locations of OBSs C and D as solid circles.
Figure 6.1e. Final model for Line 5. P-wave velocities are shown in the model at the location of the velocity nodes. Also shown are the locations of OBSs F and B as solid circles.
Figure 6.2a. OBS 1D hydrophone channel data showing the picks that estimate the LVZ in Line 4. Extra picks that were responsible for lowering in velocity from 1.57 km/s to 1.49 km/s in the LVZ are bounded by a solid box. Plot is made with a reducing velocity of 1.75 km/s.
Figure 6.2b. Alternate model of Line 4. P-wave velocities are shown in the model at the location of the velocity nodes. Also shown are the locations of inline OBSs C and D in solid circles.
Figure 6.3a. Salt map of northern Gulf of Mexico. The map shows the location of a regional line offshore southeastern Louisiana passing near the study area. The interpretation of this line was done by Diegel et al. (1993) and is shown in (b). The study area is shown in red.
Figure 6.3b. Regional dip section across the Gulf of Mexico, extending through southeastern Louisiana onshore to the base of the continental slope (see (a)). Section shows the regional context of the Bay Marchand-Terrebonne and Bourbon stepped counter-regional systems. The red rectangle is Area 1 from (a). Salt is shown in black. Faults below the autochthonous salt horizon (Jurassic) are speculative. Original 1:1 section has been scaled to 4:1 vertical exaggeration, substantially steepening true dips. The salt structure that effects the study area of this thesis is speculated to have similar characteristics and emplacement style as salt bodies in this line.
Figure 6.4. Sketch showing modeled subsurface distribution of gas hydrate (hatched areas) at the MC853 site (after Sassen et al., 2001). Hydrocarbons migrate vertically from deep within salt withdrawal basins along salt and fault conduits (arrows).
Figure 6.5a. Contour plot of velocity in L2. Shot lines are shown as solid lines and numbered at their beginnings. OBSs A thru F are shown as black circles. The diapir is shown as a white circle. Also shown is the location of LVZ2.
Figure 6.5b. Contour plot of depth of the base of L2. The shot lines are shown as solid lines and numbered at their beginnings. OBSs are shown as black circles. The diapir is shown as a white circle. Also shown is the location of LVZ2.
Figure 6.6a. Contour plot of velocity in L3. Shot lines are shown as solid lines and numbered at their beginnings. OBSs are shown as black circles. The diapir is shown as a white circle. Also shown is the location of LVZ3.
Figure 6.6b. Contour plot of depth of the base of L3. The shot lines are shown as solid lines and numbered at their beginnings. OBSs are shown as black circles. The diapir is shown as a white circle. Also shown is the location of LVZ3.
Figure 6.7a. SCS data from Line 1. The data is post-stack time migrated with velocities obtained from traveltime inversion. Positions of OBSs B and D on the sea floor are shown as solid circles. (W) is the wedge shaped feature, (F) is the reflection from the top of the wedge shaped feature and has largely negative polarity, and (Z) is the HRZ. E3 domes up below OBS D and is marked by closely spaced reflections. (F) dissapears before it intersects with (E2) and deepens away from HRZs. Also shown is the location of the diapir. The section has approximately 7:1 vertical exageration.
Figure 6.7b. SCS data from Line 2. The data is post-stack time migrated with velocities obtained from traveltime inversion. Positions of OBSs D, F and E on the sea floor are shown as solid circles. Symbols have same meanings as in Figure 6.7a. E3 domes up below OBS D and is marked by closely spaced reflections. The section has approximately 7.5:1 vertical exageration.
Figure 6.7c. SCS data from Line 3. The data is post-stack time migrated with velocities obtained from traveltime inversion. Positions of OBSs A, B and C on the sea floor are shown as solid circles. Symbols have same meanings as in Figure 6.7a. E3 domes up below OBS D and is marked by closely spaced reflections. (A) is the normal fault. The section has approximately 6.5:1 vertical exaggeration.
Figure 6.7d. SCS data from Line 4. The data is post-stack time migrated with velocities obtained from traveltime inversion. Positions of OBSs C and D on the sea floor are shown as solid circles. Symbols have same meaning as in Figure 6.7a. E3 domes up below OBS D and is marked by closely spaced reflections. The structure of E3 is probably formed by active salt movement in the region. The section has approximately 7:1 vertical exaggeration.
Figure 6.7e. SCS data from Line 5. The data is post-stack time migrated with velocities obtained from traveltime inversion. Positions of OBSs F and B on the sea floor are shown as solid circles. Symbols have same meanings as in Figure 6.7a. The dome shape of E3 is believed to be formed by active salt movement in this region. The section has approximately 4:1 vertical exaggeration.
Figure 6.8a. Comparison of Line 1 SCS data and estimated velocity model. The scale bar for the model (below) is same as in Figure 6.1a. Symbols have the same meaning as in Figure 6.7a. The horizontal scale above and below is the same. The vertical scale in the data (above) is two way time in seconds while in the model is depth in kilometers.
Figure 6.8b. Comparison of Line 2 SCS data and estimated velocity model. The scale bar for the model (below) is same as in Figure 6.1a. Symbols have the same meaning as in Figure 6.7a. The horizontal scale above and below is the same. The vertical scale in the data (above) is two way time in seconds while in the model is depth in kilometers.
Figure 6.8c. Comparison of Line 3 SCS data and estimated velocity model. The scale bar for the model (below) is same as in Figure 6.1a. Symbols have the same meaning as in Figure 6.7a. The horizontal scale above and below is the same. The vertical scale in the data (above) is two way time in seconds while in the model is depth in kilometers.
Figure 6.8d. Comparison of Line 4 SCS data and estimated velocity model. The scale bar for the model (below) is same as in Figure 6.1a. Symbols have the same meaning as in Figure 6.7a. The horizontal scale above and below is the same. The vertical scale in the data (above) is two way time in seconds while in the model is depth in kilometers.
Figure 6.8e. Comparison of Line 5 SCS data and estimated velocity model. The scale bar for the model (below) is same as in Figure 6.1a. Symbols have the same meaning as in Figure 6.7a. The horizontal scale above and below is the same. The vertical scale in the data (above) is two way time in seconds while in the model is depth in kilometers.
Figure 6.9. Reversed polarity of reflection from top of the wedge. (D) is the seafloor reflection, (E1) is the reflection from the base of L1, (F) is the reflection from top of the wedge (W), and (E2) is the reflection from the base of L2. The polarity of (F) is opposite compared to the polarity of (D) and therefore termed as "reversed". The reflection (F) comes from (at least) two closely spaced interfaces; the first one with a positive reflection coefficient followed by the second one with a largely negative reflection coefficient.
Figure 6.10. Schematic diagram of solubility curves and phase relationships. This diagram is relevant to understanding the presence of free gas and gas hydrates in the study area. O is the triple point. Dashed curves are the methane concentration. (1) Variation in methane concentration such that BSR is absent even though free gas and hydrate is present, and (2) variation of methane concentration such that the hydrates are in contact with free gas and BSR is formed at the contact. The strength of the BSR reduces as (2) moves towards left with decreasing methane concentration and dissapeares as soon as the concentration curve crosses the triple junction. After that point, the hydrates and free gas may be present but the BSR is absent. (1) explains the absence of BSR in the presence of hydrates and free gas, and (2) explains the presence of BSR.
Figure 6.11. Interpretation by Coleman et al (1983). Position of the seismic line is about 60 km NW of the study area. In the interpreted data, the horizons are labeled with their ages. All ages are in thousands of years before present (BP).
Figure 6.12. Interpretation by Goodwin and Prior (1989). The interpreted seismic line is located about 60 km NW of the study area. Also shown are the location of boreholes.
7. Conclusions

The objective of this thesis was to invert wide-angle and near-vertical traveltimes in 2D from five seismic lines over a speculated hydrate-gas system in MC798 of the GOM and make a geological interpretation of the final P-wave velocity models for purposes of accessing presence of free gas and gas hydrates.

2D compressional-wave velocity models of the shot lines were estimated using a minimum-structure, minimum-parameter approach. Results suggest that the minimum-parameter approach was effective in modeling the wide-angle data jointly with the near-vertical data as it allowed inclusion of as much prior knowledge as needed. The minimum structure approach helped in tracing rays to a maximum number of observations points, and prevented the model from being over-interpreted by keeping it geologically simple.

Jointly inverting the wide-angle and near-vertical data made modeling effective and simple. The near-vertical data were used to construct the starting model. The starting model had no lateral velocity variations but the layer boundaries closely resembled the shape of the corresponding horizons picked in the reflection stacks. This way, the starting model was “close” to the final model. During the subsequent travelt ime inversion, it was mainly the velocities that needed to be estimated, and they were constrained mainly by the wide-angle data. Since the near-vertical data was also being inverted simultaneously, the boundary nodes (in the z direction) could only vary with a limited range. As a result, the fit to the data along each line was obtained within one iteration.
Though the final overall $\chi^2$ was less than unity for each of the five lines, it should not be concluded that the data have been overfit. This is because the pick uncertainties, in this thesis, act more like weights balancing the contributions from the near-vertical and wide-angle data in the estimation of the final model. Also, the individual $\chi^2$ values of the wide-angle and the near-vertical data were monitored in an attempt to get each close to unity, and as a result, the overall $\chi^2$ is less than unity. Therefore, while choosing one final model over another, more emphasis was given to the visual assessment of the traveltime fits instead of $\chi^2$. Therefore, when it comes to modeling datasets like the one used in this thesis, $\chi^2$ is not the single decisive factor.

If the model of a line could predict (1) all the possible phases in the observed data, and (2) the traveltimes to a maximum number of observations points, and was constructed with the minimum possible model parameters required to fit the data, it was preferred over all other estimated models of that line. Comparing 1D velocity-depth profiles from the OBSs at the intersection points of the lines showed that the final models of all lines are consistent to within $\pm 10$ m/s error in velocity and $\pm 20$ m error in depth.

The HRZs that show up in the SCS reflection profiles were found to be associated with a lowering of velocity in the models of all lines. Two possible interpretations are made. According to the first interpretation, the HRZs on the SCS profiles were identified as the axis of a channel. The wedge-shaped feature (labeled W in Figures 6.1a, 6.1d and 6.1e) is caused by the overbank deposits. Typically, the axis of a channel has deposits that have higher velocities relative to the adjacent sediments and show up as HRZs in
reflection profiles. The association of HRZs with a lowering of velocity in our study area indicates the presence of free gas in the channel axis. The fact that velocities are higher in the sediments adjacent to the channel axis suggests two things: (1) the channel axis is sealed on its top and sides, and (2) the channel axis has a maximum concentration of free gas compared to any other part of the study area (if the free gas is actually present in any other part).

According to the second interpretation, the reflection event labeled (F) in Figures 6.7a, 6.7d and 6.7e is a BSR. The methane stability diagram (Figure 2.2) predicts the depth of the GHSZ close to where event (F) occurs in the data. The reflection strength and shape of the event (F) can be explained by arguing that the concentration of the free gas and salinity decreases laterally away from the HRZs because the HRZs are located on top of a structural high caused to salt bodies below. The decrease in the gas concentration and salinity decreases the strength of the reflection (Figure 6.10) and increases the depth of (F).

The structural features like the doming of the base of L3 below the diapir (Figure 6.7a) and the surface topography suggest ongoing deformation of the region from proximal salt bodies. Salt bodies, in general, lead to faulting and fracturing of the subsurface as they rise up. The faults and fractures act like conduits allowing gases from deeper sources to migrate up to the shallow subsurface. The gas gets trapped in zones that are locates on structural highs. It is therefore possible that the free gas that is present in the HRZs has a deeper source.
If the first interpretation is correct, the chance of gas being present anywhere else in the minibasin, apart from regions near the HRZs is low. As the sustenance of a hydrate system requires a constant and steady influx of free gas, it is possible that the hydrates may not exist anywhere else in the study area apart from the regions (like the mud diapir) that are close to the HRZs. If the second interpretation is correct, it is possible that gas hydrates may be present throughout the study area and give rise to BSRs only at places where the concentration of free gas is adequate (Figure 6.10).

*Recommendation for future studies*

The OBS data collected during the experiment was four-component. Only two components (the hydrophone channel data and the z component data) have been used in this study. Therefore, future studies could use the other two (horizontal) components for modeling. Shear wave modeling using the horizontal component data can shed more light on the properties of the sediments in the study area. The dataset used in the present study has offline OBS data that can be used in future studies to model out-of-the-plane arrivals and estimate the geometry of the channel axis in 3D. Since the hydrophone channel data has clipped direct arrivals and the z component has reverberations, the OBS and MCS data will have to be used judiciously if waveform methods are tried. The modeling and subsequent interpretation, as done in the present thesis, could have been carried further if any well log, core samples, or information about lithology were available, for example, from the oil industry.
References


