SEISMIC CHARACTERIZATION OF METHANE HYDRATE STRUCTURES

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DOCTOR OF PHILOSOPHY

By
Christine Ecker
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I certify that I have read this dissertation and that in my opinion it is fully adequate, in scope and quality, as a dissertation for the degree of Doctor of Philosophy.

(Principal Adviser)

I certify that I have read this dissertation and that in my opinion it is fully adequate, in scope and quality, as a dissertation for the degree of Doctor of Philosophy.

Amos Nur

I certify that I have read this dissertation and that in my opinion it is fully adequate, in scope and quality, as a dissertation for the degree of Doctor of Philosophy.

Biondo Biondi

Approved for the University Committee on Graduate Studies:

Thomas Uskow
Für meine Eltern,
Margret und Richard Ecker
Abstract

Bottom simulating reflectors (BSR) seem to be associated with the base of the stability zone of methane hydrates. These methane hydrates represent a potential future energy resource and might have a strong “greenhouse” effect on global climate. In this dissertation I integrate seismic and rock-physics to analyze seismic data from the Blake Outer Ridge, offshore Florida. I infer the cause of the BSR visible in the data and estimate the amount of hydrate present in the sediment structure.

Using stacking velocity analysis, 2-D impedance inversion and AVO analysis, I show that the Blake Outer Ridge BSR is caused by hydrate-bearing sediments overlying gas-saturated sediments. Furthermore, the analysis suggests that a flat reflector beneath the BSR represents the transition of gas- to brine-saturated sediments. Using the interval velocity obtained from the RMS velocities in conjunction with physical rock models, I provide a theoretical tool to quantify the amount of hydrate present. I examine three micromechanical models of hydrate formation: (A) hydrate is part of the pore fluid, (B) hydrate becomes part of the solid frame, and (C) hydrate cements grain contacts together. Model A predicts maximum hydrate saturation between 20% and 26%, model B saturations between 15% and 20%, and model C saturations of less than 1%. Subsequently, I analyze the stability of these estimates to errors in interval velocity. Such errors can cause the estimations to vary as much as ±14%, indicating that accurate velocity determination is crucial. The saturation estimation technique is furthermore validated by using known sonic velocities and porosities from two wells at the Blake Ridge. Finally, in order to differentiate between the different models,
I use forward 1-D modeling and AVO analysis. Both models A and B can qualitatively reproduce the BSR amplitude trend observed in the seismic data, however, they cannot be distinguished by means of their seismic responses.

This dissertation provides a tool for characterizing hydrate properties and estimating the amount of hydrate in the pore space. Although the study has been applied to data at the Blake Ridge, it could easily be transferred to a different geographic region.
Preface

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Several figures regarding RMS and interval velocity in chapter 2 (vrms, vint-ann, and vel1) which are marked as [ CR ] cannot be reproduced easily because of the need for additional software. The reason for the limited reproducibility is that the velocity analysis of the data has been performed in Promax. I have included a typical Promax data directory for Promax 6.1 (called Hydrate, located in the proc directory) which lists the exact flow sequence I used for the velocity analysis. If you have Promax, you can copy the directory into your /advance/data directory and thus view how I obtained the velocities.

All other figures throughout this thesis which are marked [ CR ] are limited in their reproducibility solely by the necessary computer time. I have included the original data (after they have been corrected for spherical divergence) in the data directory, so the figures can be reproduced if you are willing to wait several hours or days.

All figures in Chapter 3 which are marked as [ ER ] are reproducible since I have provided the processed data in the data directory. The processed data can be obtained by performing the calculations listed in Chapter 2. This will, however, require a computation time of several days. By providing the already processed data, all remaining calculations in Chapter 3 can be done in “real time” and in that sense fulfill the concept behind easily reproducible.
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Chapter 1

Introduction

Methane hydrate structures are increasingly recognized as being a potential future energy resource and having a possible strong "greenhouse" effect on global climate (Kvenvolden, 1993). Under standard conditions, 1 m$^3$ of methane hydrate can contain up to 164 m$^3$ of methane gas (Sloan, 1990). Assessments have shown that worldwide there might be twice as much hydrocarbon available from hydrate resources than from all recoverable and non-recoverable coal, oil and gas deposits (Kvenvolden, 1993). In order to realistically evaluate the potential impacts of these hydrate structures, a thorough understanding of their properties and characteristics is essential.

In this thesis, I analyze seismic data from the Blake Outer Ridge, offshore Florida and Georgia, to explore the reservoir potential of hydrate-bearing sediments. Recent drilling in this region has shown that alone in this area there might be as much as 4.7 x 10$^{16}$ g CH$_4$, or 35 GT, of carbon available. This quantity would meet the 1996 US natural gas consumption needs for the next 105 years (Dickens et al., 1997).

1.1 Methane Hydrates

A methane hydrate is an ice-like, crystalline lattice of water molecules in which gas molecules are trapped physically without the aid of direct chemical bonds. They are
only stable under certain pressure and temperature conditions, which are illustrated in Figure 1.1. These conditions for hydrate stability limit the occurrence of methane hydrates to two regions: polar and deep oceanic. In polar regions, hydrate structures are normally associated with permafrost both onshore in continental sediment and offshore in sediment of the continental shelves. In deep oceanic regions, hydrates are found in outer continental margins in the sediment of slopes and rises where cold bottom water is present.

Figure 1.1: Phase diagram of the stability of methane hydrate. Redrawn after Kvenvolden (1993).

Surface seismic data can image hydrate formations through the presence of bottom simulating reflectors (BSR), which are associated with the base of the hydrate stability field. BSRs parallel the seafloor at a subbottom depth of several hundred meters and are characterized by strong negative reflection coefficients and increasing subbottom depth with increasing water depth, a characteristic which is governed by the phase
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diagram of hydrate stability (Figure 1.1). They are therefore not structural reflectors, but occur at the phase transition of frozen methane hydrate to gas or water.

Two models have been proposed to account for the formation of methane hydrates and the development of BSRs: (1) the BSR is caused by hydrate overlying gas-saturated sediment and (2) the BSR is caused by hydrate overlying brine-saturated sediment. The first model assumes the local generation of methane from organic material at the depth of the hydrate. Gradual thickening and thus deepening of the hydrated zone causes it eventually to subside into a temperature region where hydrate is unstable. Consequently, free gas can be present in this region (Kvenvolden and Barnard, 1983a). The BSR is then caused by the impedance contrast at the base of the hydrated zone and the top of the gas layer. The second model, in contrast, supports the formation of methane hydrates through the removal of methane from rising pore fluids being expelled from deeper in the sediment section (Hyndman and Davis, 1992). Most of the methane is generated microbially at a depth below the hydrate stability zone, but not at depths sufficient for the formation of thermogenic methane. Therefore, free gas does not have to be present beneath the BSR. In this case, the BSR can be the consequence of the impedance contrast between overlying sediments containing substantial amounts of hydrate and underlying brine-saturated sediments.

In addition to seismic methods, several attempts have been made to recover hydrates from drilling. However, since the risk of heating and destabilizing the initial hydrate conditions during the process of drilling is considerably high, only limited information is available. Thus the core samples and well-logs do not necessarily reflect the correct in-situ hydrate characteristics and properties. Consequently, much information is inferred remotely from seismic data of the bottom simulating reflector and should be tied with the results available from drilling.

Although numerous investigations have been performed to determine the hydrate and BSR characteristics from surface seismic, they were restricted to determining only the P-wave velocity behavior around the BSR while assuming possible hydrate Poisson's ratios based on laboratory results. Little work has been done in estimating
both the P- and S-wave velocity behavior at the transition from in-situ hydrates to the underlying sediments, and subsequently in relating them to physical rock properties. The ultimate goal of interpreting seismic reflection data is to estimate the amount of hydrate present in the sediment, its distribution in the pore space, permeability and potential recoverability. This goal requires an integrated seismic analysis - rock physics approach.

1.2 Previous Work

1.2.1 Laboratory Measurements

In recent years, several laboratory measurements have been performed to determine the properties of the pure hydrate rock matrix structure. These investigations showed that the hydrate structures are very similar to pure ice structures, having a P-wave velocity of approximately 3.31 km/s and a Poisson’s ratio of 0.33 (Whalley, 1980; Pearson et al., 1986). Additionally, Stoll et al. (1979) argued that the presence of hydrate in sediments causes a significant increase in the acoustic velocities of the sediments. This phenomenon clearly establishes the potential for a sharp acoustic impedance contrast at the boundary of a sediment region containing a considerable quantity of hydrate.

1.2.2 Elastic Properties and Cause of the BSR

Numerous seismic investigations have been performed to evaluate the validity of the velocity behavior determined in laboratory measurements in in-situ hydrate structures and determine the cause of the BSRs.

Miller et al. (1991) used synthetic waveform modeling of vertical-incidence traces from offshore Peru to determine a thin free gas zone beneath the BSR. Performing waveform inversion with a constant Poisson’s ratio of 0.4 on data offshore Vancouver Island, Singh et al. (1993) concluded that the BSR in this region is overlain by
hydrate-bearing sediment having a P-wave velocity of approximately 1.7 km/s and is underlain by gas-saturated sediment having a P-wave velocity of about 1.4 km/s. Analysis on data from two drill sites offshore from Vancouver Island showed similar results (MacKay et al., 1994). Hyndman and Spence (1992), on the other hand, did not find evidence of a free gas zone in this region when they conducted a comprehensive seismic analysis of the same data used by Singh et al. (1993), including vertical-incidence waveform modeling, reflectivity modeling of AVO, and calculation of interval velocities from RMS stacking velocities. They concluded that the strong BSR amplitudes are caused by high concentrations of hydrate in a thin layer just above the BSR and normal velocity brine-saturated sediment beneath. Seismic studies in the Beaufort Sea related the BSR occurrence in that area to the presence of a free gas layer beneath the BSR (Andreassen et al., 1995). This study was conducted using forward AVO modeling with a constant Poisson’s ratio of 0.38 in the hydrate-bearing sediments.

Recent studies of the BSR occurring at the Blake Outer Ridge, offshore Florida and Georgia, strongly support the hydrate-over-gas model for this region. Rowe and Gettrust (1993) analyzed deep-towed multichannel seismic data, estimating a significantly high P-wave velocity of over 2.5 km/s in a thick hydrate layer above the BSR and a low P-wave velocity of 1.4 km/s in the gas-saturated region beneath. Similar results were obtained by Korenaga et al. (1997) using full waveform and travel time inversion of wide-angle data. Slightly lower P-wave velocities of 1.9 km/s - 2.0 km/s in the hydrate-bearing sediments were obtained by Wood et al. (1994) through travel time velocity analysis and acoustic waveform inversion; by Katzman et al. (1994) using traveltime inversion, AVO analysis and synthetic modeling of the same wide-angle data as Korenaga et al. (1997); and by Lee et al. (1994) based on 1-D seismic AVO inversion and assumed hydrate properties. Recent drilling results of ODP leg 164 (Matsumoto et al., 1996) and vertical seismic profiling (VSP) (Holbrook et al., 1996) suggest P-wave velocities of no more than 1.9 km/s in the hydrated sediment and the presence of a thick gas layer (≥ 100 m) with velocities between 1.4 km/s and 1.6 km/s. Using walk-away VSP data from one of those wells which, however, did not penetrate a visible BSR, Pecher et al. (1996) furthermore
suggested that the S-wave velocity of hydrated sediment was higher than the one in the sediments underneath.

All of the seismic investigations inferred the P-wave velocities at the BSR directly from the data, but made assumptions about the possible S-wave velocity behavior at this transition. These assumptions were mainly based on laboratory measurements which were directly translated into a possible behavior of in-situ hydrate structures. Only Pecher et al. (1996), who used converted waves visible in walk-away VSP data investigated the behavior of the shear moduli directly. Their study, however, is limited to an area of no BSR occurrence and is only a 1-D representation of an heterogeneous medium.

1.2.3 Estimation of Hydrate Saturation

In addition to investigations about the cause and properties of the BSR, several attempts have been made in estimating the amount of hydrate from seismic velocities and amplitudes. Most studies determined the desired hydrate saturations using Wyllie’s time average equation (Wyllie et al., 1958), which directly relates acoustic velocity to porosity and saturation (Sholl and Hart, 1993; Wood et al., 1994; Korenaga et al., 1997). In order to obtain a more reliable estimate in high-porosity sediments, Dillon et al. (1993) used a weighted mean of the Wyllie equation together with Wood’s equation (1941).

Wyllie’s equation has been obtained empirically for consolidated reservoir rocks and cannot be used for high-porosity unconsolidated sediments (Dvorkin and Nur, 1998). In order to apply this equation to high-porosity marine sediments, calibration is required based on extensive core measurements or well-log data. Results are modified “time-average-form” equations. It is likely that such equations can indeed link velocity to porosity and gas hydrate content if they have been derived from an extensive experimental database. However, such equations cannot be used for diagnosing sediments, i.e., inferring their internal structure from seismic.

Yuan et al. (1996) first derived a relation between velocity and porosity from core
and well-log data. They calculated a porosity profile from velocities across a BSR and subtracted from it the "normal" porosity profile of sediments without a BSR. The resulting relative porosity reduction above the BSR was attributed purely to the presence of methane hydrate in the pores, which directly translates into hydrate saturation.

All of these investigations estimated 1-D hydrate saturation profiles and assumed known a-priori porosity based on well or core information. Little work has been done in considering the actual effect of the position of hydrate in the pore space and in constructing physical rock models. Furthermore, the effect of different models on seismic waveforms and amplitudes needs to be considered in order to connect a specific hydrate saturation reliably to the seismic data.

1.3 Objective of this Thesis

Three ultimate goals arise with respect to the importance of hydrate structures as a potential future energy resource and influence on the global climate: (1) the delineation of hydrate location, (2) the mechanism for the formation of hydrate deposits, and (3) the estimation of the amount of hydrate actually present in the hydrate structures. My objective is to use an integrated study of seismic AVO analysis and modeling and rock physics analysis to estimate the hydrate and BSR properties directly from surface seismic. The reflection amplitude variation with offset can be an important indicator of free gas at an interface (Ostrander, 1984; Shuey, 1985), and should help to discriminate clearly between the two proposed BSR models and to estimate possible velocity structures around the BSR. A subsequent relation of the inferred hydrate properties with physical rock models should considerably limit the possible explanations of the physical origin of the BSR and the hydrate structures.

In the course of this study I intend to address the following key questions related to the BSR and hydrate properties:

- What causes the bottom simulating reflector?
CHAPTER 1. INTRODUCTION

- What are the elastic properties, i.e., P- and S-wave velocity, of hydrate-bearing sediments?
- Which rock physics models support these properties?
- How can we tell from these models how much hydrate is contained within the hydrate structures?

This dissertation will thus provide a tool for characterizing hydrate properties and estimating the amount of hydrate present from surface seismic. Even though the study is applied to data from the Blake Outer Ridge region, it is by no means restricted to this area.

1.4 Overview

In this thesis, I characterize the elastic properties, i.e., P- and S-wave velocities, of a hydrated structure present at the Blake Outer Ridge, offshore the southeast coast of the United States, using AVO analysis and modeling. I show how these properties can be related to the possible internal structure of the hydrated sediment by assuming physical rock models. Furthermore, I provide a theoretical tool for estimating the amount of hydrate present without prior well-log information.

Chapter 2 addresses the seismic data used in this study and the true-amplitude processing applied to them prior to AVO analysis. The actual AVO inversion and modeling is subsequently discussed in Chapter 3. The goal is to use AVO as an indicator of free gas underneath the hydrate and to delineate the behavior of P-wave and S-wave velocities across the BSR. A lot of the work described in Chapter 2 and 3 was originally performed with David Lumley and can be found in Ecker and Lumley (1993, 1994a,b). Subsequently, Chapter 4 provides a tool for quantifying the amount of gas hydrate and gas near the BSR, by linking physical rock models to the estimated interval velocities. Using three different possible schemes of hydrate deposition in the pore space, this results in different estimates depending on the used model. Since
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the use of only acoustic velocity cannot uniquely differentiate between the different models of hydrate deposition, Chapter 5 combines the rock physics models and seismic AVO analysis in order to obtain the internal hydrate structure. The work presented in Chapter 4 and 5 was performed in collaboration with Jack Dvorkin and Amos Nur from the Stanford Rock Physics Group and can be found in Ecker et al. (1997, 1998). Preliminary results are given in Ecker et al. (1996a,b).
Chapter 2

Seismic Data and Processing

2.1 Overview

In this chapter, I give an overview of the data used in this study and describe the processing steps applied to the data prior to AVO analysis. First I delineate the velocity analysis applied to the data and discuss the uncertainties both in RMS and interval velocity. I then outline the preprocessing steps of the data, the prestack time migration and succeeding amplitude calibration. The entire processing is performed in a way to preserve the true amplitude information of the data, which is essential for the subsequent AVO analysis and modeling.

2.2 Blake Outer Ridge Seismic Data

The data used in this study were recorded at the Blake Outer Ridge, offshore Florida and Georgia. A map of the region is shown in Figure 2.1, highlighting the area of known hydrate distribution as mapped from seismic bottom simulating reflectors (BSR). The part of the seismic line analyzed in this thesis is marked by the rectangular, and extends both from the hydrate region into an area without hydrate. The map also displays the position of six wells drilled at the Carolina and Blake Outer
CHAPTER 2. SEISMIC DATA AND PROCESSING

Ridge. Velocity information of wells 994, 995 and 997, which were several hundred meters deep and penetrated through zones of hydrate-saturated sediments, will be used later in this study. The sediments in the region are fairly uniform with no major lithological changes within several hundred meters of the seafloor. Most of the sediment consists of clay and nannofossils with small amounts of quartz. They are highly unconsolidated and are therefore characterized by large porosities.

Figure 2.1: Map of the region after Matsumoto et al. (1996).

The data consists of 1050 CMP gathers with a 50 m CMP spacing recorded with a non-linear streamer. This resulted in a 100 m group spacing at the near-offset traces and a 50 m group spacing at the far offset traces. The streamer length is approximately 3.6 km.

2.2.1 Stacked Data

A stacked section of the data after prestack migration and full processing can be seen in Figure 2.2. A gained and windowed stacked section is shown in Figure 2.3. The seafloor reflection occurs at about 4.3 s two-way travel time, indicating a water depth of over 3 km, and decreases in depth gradually along the line. It is followed by a strong bottom simulating reflection at a distance between 25 km and 52 km,
Figure 2.2: Stacked section after full processing and prestack migration.
Figure 2.3: Zoomed and gained stacked section after full processing and prestack migration.
suggesting the presence of methane hydrates in this region. In a small region above the BSR, there is a “quiet” zone where no diffractions or reflections are visible. Dillon et al. (1993) connected this zone, which is present above most of the BSRs in the region of the Blake Outer Ridge, to the presence of hydrate in the sediment. A fairly bright “flat spot” reflection is apparent beneath the BSR at approximately 5.4 s two-way traveltime, which does not parallel the seafloor or BSR reflection events. No BSR is visible in the first half of the seismic line, between 0 and 25 km distance, and the sediment structure appears fairly uniform. The maximum dips of the events throughout the entire length of the data are less than 2°, thus representing simple geological structures.

The zoomed and gained part of the stacked section, as shown in Figure 2.3, furthermore enhances the structure above and below the BSR. It clearly displays some events cutting across the BSR. Since the BSR is not a lithological but a phase boundary, such crossing of events can often be observed. It is a characteristic by which BSRs can be easily recognized.

2.2.2 Prestack Data

Figure 2.4 shows a constant offset section of the prestack, unmigrated data. It is windowed around the BSR. The figure shows numerous strong diffractions underneath the BSR. These diffractions can be caused by three-dimensional features such as small gas pockets. Scattering or side-slip energy from such features might be visible underneath the BSR (Lee et al., 1994).

Examining the raw prestack CMP gathers (Figure 2.5), a traveltime kink is very striking in all reflections near the central offset of the CMP gathers. This kink occurs exactly at the transition zone between the non-linear cable group spacing of 100 m at the near-offsets, and 50 m group spacing at the far-offsets.

One possible way of eliminating the traveltime kink in the data is adding the far-offset traces in pairs to simulate a constant cable spacing of 100 m at all offsets. This correction, however, causes the data to have less spatial resolution than the original
Figure 2.4: Constant offset section of the data. [proc-diffractor] [CR]

data, based on a decrease from 48 to 36 traces. Furthermore, the group summation causes a significant loss in temporal frequency content at the far-offsets because of spatial averaging of moveout delayed reflections across a twice longer effective group array. A possible way to compensate for this low-frequency group array response is to filter (deconvolve) the far-offset data in the \( k - x \) domain. However, the decreased number of summed far-offset traces results in a very short data series, which makes accurate spectral estimates and the application of spatial deconvolutional filters difficult. Another way to account for this frequency loss is to deconvolve the far-offsets as a function of Snell’s parameters \( p \) in the slant stack domain. Unfortunately, the small number of traces tends to introduce large edge effects in the slant-stack spectrum. These edge effects could be minimized by using a least-squares slant stack, which, however, would smear the notches in the \( \tau - p \) spectrum; the resulting deconvolution would overemphasize those portions of the spectrum. Therefore, it seems unreasonable to correct the raw data for the different group spacing by summing the far-offset
Figure 2.5: Upper panel shows a raw CMP gather. Lower panel shows the CMP gather after interpolating the near-offset traces, resulting in a constant group spacing of 50 m at all offsets.
traces in pairs. A simple linear interpolation of the near-offset traces after NMO, with source wavelet deconvolution and amplitude calibration to compensate for the hydrophone array attenuation, appears to be a better method for suppressing the non-linear cable effects. Figure 2.5 displays a CMP gather after application of those corrections, which will be discussed in more detail in section 2.4.

2.3 Velocity Analysis

In this section, I describe the velocity analysis performed on the data and discuss the resulting 2-D velocity fields. Furthermore, I evaluate the possible errors in both RMS and interval velocities and quantify the interval velocity uncertainty by comparing it with two vertical-seismic-profile (VSP) velocity profiles from wells 995 and 997 (see Figure 2.1), both of which penetrate a sharp BSR in the region of the Blake Outer Ridge.

2.3.1 Stacking and Interval Velocity

The velocity analysis of the data is performed in Promax, separately from the processing. The raw data is gained and decimated at the far-offsets to eliminate the traveltine kink in the reflections. I then generate supergathers at every 25th CMP location by combining 10 adjacent gathers. Since the sediment structure at the Blake Outer Ridge is fairly simple, a conventional NMO stacking velocity analysis is used to obtain good RMS velocities. The resulting 2-D velocity field can be seen in Figure 2.6.

Using Dix’s equation (Dix, 1955), the RMS velocities are converted into a physical interval velocity model, which is displayed in Figure 2.7. The velocity is characterized by an increase to approximately 1.9 km/s above the BSR. Such a velocity increase due to the presence of hydrate is a commonly observed characteristic of sediments containing methane hydrates (Rowe and Gettrust, 1993; Singh et al., 1993; Minshull et al., 1994; Andreassen et al., 1995; Yuan et al., 1996). Underneath the BSR, the
CHAPTER 2. SEISMIC DATA AND PROCESSING

Figure 2.6: RMS stacking velocities.

Figure 2.7: Interval velocities.
velocity drops to approximately 1.7 km/s, which might be indicative of the presence of free gas beneath the BSR. Consequently, the flat reflector underneath the BSR might be the transition zone from gas-saturated sediment to brine-saturated sediment. The observed drop in velocity at the BSR is supported by the negative reflection polarity at the BSR, which already suggested a velocity reversal at the transition from hydrate-bearing sediment to the sediment underneath.

Between a lateral distance of 0 and 25 km, where no BSR is visible, the velocity is characterized by a steady increase with increasing depth. No pronounced velocity anomaly is present.

2.3.2 Velocity Uncertainty

The use of the Dix equation to convert RMS stacking velocities to interval velocities is based on the assumption of horizontal layers and constant velocities between the layers. Therefore, dipping structure and vertical and lateral velocity variations can introduce significant errors into the resulting interval velocity fields (Hubral and Krey, 1980; Hajnal and Sereda, 1981; Toldi, 1985). Furthermore, the error in interval velocity is inversely proportional to the time thickness to depth ratio for the reflector interval under consideration. Because of the relatively deep water layer (more than 3 km) and the small dips of the sediments in the data used, most of the errors in the interval velocities will be caused by picking errors in the rms velocities and the small thickness between the reflectors with respect to the water depth.

RMS and Interval Velocity Errors

I evaluate the accuracy of my velocity models on one CMP gather, located above both the hydrate-saturated sediments and the gas-saturated sediments. Figure 2.8 displays its velocity scans overlain by the obtained RMS velocity trend (left and middle panel) as well as the resulting NMO corrected gather (right panel). The main reflections appear to be moved out with an accurate velocity and the velocity scans display a
Figure 2.8: The left panel shows the velocity scan overlain by the picked RMS velocity trend. The middle panel displays the velocity scan contours again overlain by the RMS velocity. The right panel shows the moveout-corrected CMP gather using the displayed RMS velocity.  \[\text{proc-rms1} \] [ER]
good agreement between the RMS picks and the maximum semblance. The interval velocity resulting from the displayed RMS velocity trend can be seen in Figure 2.9. It shows, as described in the previous section, an increase to about 1.9 km/s in the hydrated zone, and a subsequent decrease in velocity to approximately 1.7 km/s in the region where free gas might be present.

The accuracy of the interval velocity depends on the accuracy of the RMS velocity picks. RMS velocity picking errors mainly depend on the width of the maximum semblance at a reflector. Based on the width of the semblance at the BSR and the reflector underneath, I introduce a possible RMS velocity picking error of ± 10 m/s. The resulting RMS velocity trends are displayed in the left panel of Figure 2.10. The solid line represents the original RMS velocity trend, while the two dashed lines represent the velocities resulting from ± 10 m/s picking errors.

Using the Dix equation these different RMS trends are converted into interval velocities which are shown in the right panel of Figure 2.10. The solid line represents the interval velocity obtained from the original RMS velocity. The interval velocities are characterized by discrepancies which are as high as ± 200 m/s (more than 10%). The relatively small errors in RMS velocity introduced by picking can therefore yield considerable uncertainties in the interval velocities. As a consequence, the anomalous velocity zones in the hydrate and the layer underneath might be enhanced (dashed
Comparison with VSP Data

In order to properly quantify these possible errors, I compare them with two theoretical-seismic-profiling (VSP) measurements obtained at wells 995 and 997 (see Figure 2.1), both of which lie outside the outer ridge. The comparison is made on the basis of the interval velocity of the CMP gather (solid line). The measured interval velocity of 995 (dashed line) is consistent with the expected error in interval velocities, i.e., the error in the velocity trend. Figure 2.11 shows the measured interval velocity (solid line) compared with the expected error in interval velocities obtained in well 995 and well 997 (dotted line). Since the measurements are consistent, we can conclude that the expected error in interval velocities is close to the actual error. Further, it is evident that the expected error is less than the observed error. This observation is consistent with the expected error in interval velocities.

Figure 2.11: Measured interval velocity (solid line) compared with expected error in VSP data from surface to depth.

2.13 FigureCaption

Figure 2.10: The left panel shows the error in the RMS velocity trend. The solid line represents the original velocity; the two dashed lines are the velocities with picking errors of ±10 m/s. The right panel displays the interval velocities resulting from the different RMS velocity trends. It represents the expected error in interval velocities.

[proc-error] [ER]
line) or suppressed (double dashed line). In the latter case, where the velocity appears to be steadily increasing with increasing time, the large negative reflection coefficient at the BSR needs to be caused by a strong negative density contrast with simultaneous increase in velocity. Considering the lithologically fairly homogeneous sediment structure at the Blake Outer Ridge, this does not seem feasible.

**Comparison with VSP Data**

In order to further quantify these possible errors in interval velocities, I compare them with two vertical-seismic-profiling (VSP) velocity functions obtained in wells 995 and 997 (see Figure 2.1), both of which penetrated a strong BSR in the region of the Blake Outer Ridge. The comparison can be seen in Figure 2.11, where the measured interval velocity of the CMP gather in Figure 2.9 is overlain by the VSP velocity measured in well 995 (dashed line) and the one in well 997 (dotted line). Since the VSP measurements are from a region of less seafloor and, consequently, BSR depth, the high and low-velocity layers are slightly shifted in depth with regard to the one from surface seismic.

![Figure 2.11: Measured interval velocity (solid line) compared with VSP velocity from well 995 (dashed line) and the one from well 997 (dotted line).](proc-velcomp-ann)

The velocity functions displayed by the VSP data confirm the presence of a high-velocity hydrate zone and a low-velocity zone underneath. The velocity increase due to the presence of hydrate is approximately 100 m/s less in the VSP data compared to the seismic data. The VSP data, on the other hand, show a significantly stronger
decrease in velocity underneath. These discrepancies in velocities can be caused by several factors. First, the VSP and seismic measurements were not conducted in the same region; thus neither the hydrate nor the sediment structure need to be equivalent. Furthermore, both measurements are done with different frequency spectra, resulting in different resolutions. The comparison shows, however, that both the VSP and seismic data are characterized by an increase in velocity in the hydrate zone and a subsequent decrease underneath. This velocity structure at the Blake Outer Ridge was additionally confirmed by Wood et al (1994), Katzman et al. (1994) and Korenaga et al. (1997).

2.4 Processing

In this section, I describe the processing applied to the raw data. The processing steps include spherical divergence correction, source wavelet deconvolution, trace interpolation, prestack time migration and amplitude calibration.

2.4.1 Preprocessing

Divergence Correction

In the first preprocessing step, the data is corrected for constant velocity spherical divergence by multiplying each trace with a time-variant amplitude scale function of the form $A(t) = t$.

NMO and Deconvolution

Following the divergence correction, I perform a normal moveout correction (NMO) using the estimated RMS velocities. A resulting CMP gather can be seen in the upper panel of Figure 2.12. The near offsets are at the left side and increase to the right side. A significant change in shape and amplitude of the source wavelet is visible at the
Figure 2.12: The upper panel shows a CMP gather after NMO correction. The near offset is to the left, the far offset to the right. The lower panel shows the CMP gather after deconvolution, bandpass filtering, and NMO correction.
transition zone between the 100 m and 50 m group spacings. The loss in frequency and amplitude is caused by the strong hydrophone array attenuation. In the direction of the far offsets, where the hydrophone groups are shorter, the wavelet sharpens and regains its large amplitudes. These waveform shapes and amplitude changes can seriously degrade any subsequent AVO analysis. In order to regularize the wavelet as much as possible, I perform a single trace source wavelet deconvolution before NMO correction which estimates a different filter for each trace. Deep water (more than 3 km) and shallow dips in the area cause the wavefront propagation angles to change only moderately as a function of depth. For example, in Figure 2.12 at offset 2.65 km (transition between 100 m and 50 m groups spacing), the seafloor reflection propagates at 22° while the BSR reflection propagates at 20°. This small change in propagation angles makes trace-by-trace deconvolution a valid tool in this case. Although the application of single trace, as opposed to surface-consistent, deconvolution can result in undesired changes in amplitude and waveform, the Blake Outer Ridge seismic data are of high enough quality that single trace deconvolution with careful quality control gives very consistent results. After deconvolution, the data are bandpass filtered to a range of frequencies where signal is found at all offsets. This removes spurious deconvolutional high-frequency noise. Then the data are again NMO corrected.

A resulting CMP gather after deconvolution, bandpass filtering and normal move-out correction can be seen in the lower panel of Figure 2.12. The waveform has been well calibrated along all offsets. The remaining amplitude loss, especially that induced by the hydrophone attenuation, will be addressed in the amplitude calibration section.

**Trace Interpolation and INMO**

After the deconvolution and bandpass filtering, I apply a linear interpolation at the near-offset traces to even the group spacing to 50 m at all offsets and thus eliminate the traveltime kink introduced by the non-linear streamer. The data is then inverse NMO corrected and the previously applied spherical divergence correction is removed. The data is then passed to a prestack time migration.
2.4.2 Prestack Migration Processing

Numerous investigations have shown the advantages of prestack migration before AVO analysis over the conventional AVO analysis on unmigrated CMP gathers (Resnick et al., 1986; Beydoun et al., 1994; Mosher et al., 1996). Prestack migration is required in areas of strong structural dips, but also improves the data quality when the sediment structure is relatively flat: diffractions that can interfere with the primary reflections are collapsed and the lateral resolution of the data is increased as migration shrinks the size of the Fresnel zone. Furthermore, migration provides a better estimate of the subsurface reflectivity, which is proportional to the relative P-impedance contrasts in velocity and density. Therefore, I perform a Kirchhoff prestack migration/inversion on the data which is based on the method developed by Lumley (1993). This method estimates the reflection coefficients and the reflection angles directly from the prestack data and compensates internally for spherical divergence. In order to avoid spatial aliasing, I interpolate the data to a CMP spacing of 12.5 m and use a smoothed 2-D RMS velocity field as the migration velocity.

Since it is essential for any subsequent AVO analysis that the reflector moveout is very flat, I apply an additional residual moveout correction (RNMO) to the data. In order to improve the coherency of the events prior to this correction, I also apply the first part of the amplitude calibration (normalization of the amplitudes) which will be described in the next section. An example of a prestack migration/inversion reflectivity (left) after RNMO and the reflection angles (right) is shown in Figure 2.13. The angle contours start at 5 degrees at the near offset and increase in 5 degree increments. The maximum angle coverage in this gather is approximately 25 degrees.

2.4.3 Amplitude Calibration

After prestack migration and RNMO correction, the data need to be calibrated for the loss of amplitudes with offset. The use of 100 m group spacing at the near offsets and 50 m group spacing at the far offsets introduced a strong array attenuation at
Figure 2.13: Migrated gather after RNMO and reflection angles. Angle contours are displayed in 5 degree intervals. The first contour represents an angle of 5 degree.

Mapping and CM modeling

BSR
the central offsets of the data. The resulting loss in amplitude is clearly visible in Figure 2.12. Further loss in amplitude can be caused by different source strength and receiver gains, source array effects, transmission loss, attenuation, noise and so on. Ideally, these effects should be compensated for in a deterministic way, which would require the accurate calculation of the amplitude affecting components, assuming the different amplitude affecting processes are clearly separable. Very often, however, the accurate parameters are difficult to calculate or estimate.

Calibration Scheme

The high quality of the Blake Outer Ridge data and the relatively deep water (more than 3 km) enable me to use a more statistical amplitude calibration: I use the seafloor reflection as a reference reflection to calibrate the amplitudes for the amplitude distorting effects relevant to the data. This approach does not attempt to compensate for all effects that diminish amplitudes individually. Rather, it corrects empirically by predicting the seafloor AVO and calibrating the data to the theoretical curve at all offsets. Amplitude correction by means of a reference reflector has been described by Chiburis (1984, 1987) and Ross and Beale (1992).

The calibration is done by assuming a functional form of the seafloor AVO response based on reasonable contrasts in P- and S-impedance at this interface. First, I normalize the data so that the seafloor peak amplitudes are equal to the near-offset seafloor amplitudes. I pick the seafloor amplitude at each offset \([A(x)]\) and normalize the trace amplitudes \([\text{trace}(t, x)]\) at each offset by \(\frac{A(0)}{A(x)} \text{trace}(t, x)\). Then each trace is scaled to match the predicted seafloor AVO trend. This method will preserve the magnitude of the zero-offset seafloor amplitude variations, which are indicators of changing lateral lithology, but predetermine the shape of the seafloor AVO behavior.

By calibrating the amplitudes in this way, the described hydrophone attenuation will be automatically corrected. Given the thick water layer and small structural dips in the area, the propagation angles at the different reflectors are only about 2-3° apart. Therefore, such an offset-depending calibration procedure will be valid for the
reflectors beneath the BSR.

This offset dependent amplitude calibration, however, does not compensate for the effects of possible amplitude loss underneath the seafloor. Inelastic attenuation in the sediments and thin layering effects are therefore not included. In case of frequency dependent attenuation, higher frequencies will be attenuated faster than lower frequencies, thus yielding lower frequencies and amplitudes at the BSR in comparison with the seafloor. This would result in still undercorrected amplitudes at the BSR after calibrating with the seafloor AVO response. Frequency attenuation, however, will also have an effect on the hydrophone attenuation. Waves with less frequency content will be attenuated less. Thus, the BSR amplitudes would be affected less by the array attenuation than the seafloor. Consequently, correcting with the seafloor as a reference reflector would yield slightly overcorrected amplitudes at the center offsets where the array attenuation is strongest. Some of these effects might be visible in the data (see Figure 2.15), but are treated as negligible in the later AVO analysis.

**Calibration Parameters and Scaling Function**

In order to determine the theoretical seafloor AVO response, I need to assume elastic properties for the water layer and the near-seafloor sediments. The reflection amplitude is generated by an impedance contrast averaged one Fresnel zone above and one Fresnel zone below the reflector, in this case the seafloor. After migration, the Fresnel zone collapses to 1/4th of a seismic wavelength \( \lambda = v/f \), where \( v \) is the rms velocity of the reflection event and \( f \) is the dominant frequency. In the case of the Blake Outer Ridge data, the dominant frequency is approximately 30 Hz and the velocity at the seafloor is 1.5 km/s. Therefore, the seafloor reflection amplitude is generated in a region approximately 12.5 m above and 12.5 m below the seafloor.

The water can be assumed to have a P-wave velocity of 1.5 km/s, zero S-wave velocity, and a density of 1.0 g/cm\(^3\). The properties of the near-surface sediments are estimated based on results of the drilling at sites 994, 995 and 997 (Matsumoto et al., 1996) and on observations by Hamilton (Hamilton, 1976). Since the marine
sediments are highly unconsolidated, I chose a P-wave velocity of approximately 1.55 km/s, a S-wave velocity of about 0.1 km/s and a density of 1.3 g/cm$^3$ to represent the near-surface marine sediments. Based on these elastic parameters, I construct a theoretical seafloor AVO response using the Zoeppritz equations (Aki and Richards, 1980).

The Zoeppritz equations (Aki and Richards, 1980) describe the amplitudes of transmitted and reflected P- and S-waves in the case of plane waves incident on a reflector. Assuming small layer contrasts and angles coverage well within the pre-critical region, Aki and Richards (1980) linearized these equations. The resulting acoustic reflection coefficient can be described as

$$ R(\theta) = \frac{1}{2 \cos^2(\theta)} I_p - 4 \gamma^2 \sin^2(\theta) I_s + (2 \gamma^2 \sin^2(\theta) - 0.5 \tan^2(\theta)) D \quad (2.1) $$

where $I_p$, $I_s$ and $D$ are the relative contrasts in P-impedance, S-impedance and density, $\theta$ is the reflection angle and $\gamma$ is an estimate of the background shear to compressional velocity ratio $v_s/v_p$.

Using the estimated elastic parameter changes across the interface from water to near-surface sediment, I can thus determine the functional form of the seafloor AVO response using equation 3.5. The appropriate scaling function of the traces after they have been normalized using the maximum seafloor amplitudes is therefore $A(\theta) = \frac{R(\theta)}{R(0)}$ and can be calculated from equation 3.5. The result is shown in Figure 2.14. It shows that the scaling function preserves the near-offset amplitudes and increases the far-offset amplitudes by less than 2%. This will result in nearly constant seafloor amplitudes with increasing offset.

**Final Corrected Gathers**

All migrated gathers are scaled using the previously displayed scaling function of the seafloor. A resulting final gather after preprocessing, migration, residual NMO
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AVO Correction Function

Figure 2.14: AVO scaling function.

Final Migrated Gather after Amplitude Calibration

Figure 2.15: Final CMP gather after amplitude calibration.
correction, and amplitude calibration is shown in Figure 2.15. The final gather displays smoothly varying amplitudes with increasing offset, much improved compared to those shown before amplitude calibration (Figure 2.12). In this gather, the bottom simulating reflector at the base of the hydrate stability field is characterized by a strong, negative reflectivity that increases negatively with increasing offset. This BSR amplitude behavior has also been observed by Lee et al. (1994) and Katzman et al. (1994).

After the final amplitude correction of the data, they can now be used as input into an impedance inversion and AVO modeling technique in order to evaluate the elastic properties of the hydrate-bearing sediments and those underlying the BSR. This is the content of the next chapter.

2.5 Conclusions

In this chapter, I have discussed the seismic data used in this dissertation, their origin, and their appearance. I have described the used velocity analysis in Promax, which yielded an increase in velocity in the hydrate-saturated sediments to about 1.9 km/s and a decrease in the sediments underneath to approximately 1.7 km/s. This drop in velocity at the BSR might be indicative of the presence of free gas in the region. Possible RMS velocity picking errors of $\pm 10$ m/s can result in as much as $\pm 200$ m/s error in the interval velocities, either enhancing or suppressing the observed high and low velocity zones. The presence of an increase in velocity in the hydrate and a subsequent decrease in a gas-saturated zone is, however, confirmed by VSP measurements in the region. Furthermore, the observed negative reflection polarity at the BSR is in good agreement with a velocity reversal at the BSR.

Subsequently, I have outlined the true-amplitude processing scheme of the data which is essential in preserving the amplitude information required for the subsequent AVO analysis. The processing scheme consisted of source wavelet deconvolution, trace interpolation, prestack migration/inversion, RNMO correction and amplitude calibration. The resulting gathers display strong, negative BSR reflection coefficients.
Such amplitude trends appear to be observed at most places showing a strong BSR (Hyndman and Spence, 1992; Katzman et al., 1994; Andreassen et al., 1995)

After these processing steps, the final gathers are used as input into AVO inversion and modeling, which will be described in the next chapter.

### 2.6 Acknowledgements

I would like to thank Keith Kvenvolden, Myung Lee and Bill Dillon from the USGS for providing us with a copy of the seismic data used in this study. Originally, a significant part of this analysis was performed in collaboration with David Lumley. I would like to thank him for many helpful discussions and suggestions. Furthermore, I would like to thank him for the permission to use his prestack migration code as described in section 2.4.2.
Chapter 3

AVO Inversion and Modeling

3.1 Overview

In this chapter, I use the seismic amplitude information extracted by the processing described in Chapter 2 to determine the characteristics and possible causes of the BSR reflection. Using 2-D P- and S-impedance contrast inversion, the BSR appears to be characterized by negative P-impedance contrast and mostly positive S-impedance contrast. The negative P-impedance contrast is in good agreement with the velocity reversal determined in Chapter 2. Local investigation of the BSR reflection character and amplitudes shows that the BSR is underlain by many small structural layers which contribute laterally to significant tuning effects. In areas without tuning, the BSR is characterized by a strong increase in amplitude with increasing offset, while the other areas exhibit more constant amplitudes with offset. In the case of increasing amplitudes with offset, forward modeling using the Zoeppritz equations supports the BSR model of hydrate overlying free gas-saturated sediment. The more constant amplitudes with offset, on the other hand, would indicate the presence of brine-saturated sediment underneath the BSR. Although a laterally heterogeneous structure of hydrate and gas/brine distribution is physically plausible, tuning appears to be the more important factor in producing constant or distorted amplitudes in the area. This suggests that the BSR is caused byhydrate-bearing sediment overlying gas sediment.
This is in good agreement with the flat reflector underneath the BSR, which appears to mark the bottom of the gas zone and the transition to brine-saturated sediments. The presence of a thick gas layer in the region of the Blake Outer Ridge has furthermore been confirmed by recent drilling results (Matsumoto et al., 1996).

3.2 AVO Theory

The variation of reflection and transmission coefficients with incident angle and thus offset is commonly known as offset-dependent reflectivity. The Zoeppritz equations (Zoeppritz, 1919) describe the reflection and transmission coefficients as a function of incident angle and elastic media properties (density, P-wave velocity, and S-wave velocity). They apply to a reflection of plane-waves between two half-spaces, and do not include wavelet interferences due to layering. Furthermore, amplitudes are a measure of the reflection coefficient only when effects that cause amplitude distortions have been removed. Thus, preprocessing to remove transmission loss, source and receiver effects, spherical divergence, multiples, and so on, is essential to the successful recovery of the reflection coefficients.

The variation of the P-wave reflection coefficient with offset and angle, which is the basis of AVO analysis, can be used as a direct hydrocarbon indicator (Ostrander, 1984). Therefore, it should help in delineating the possible causes and characteristics of the BSR in the data from the Blake Outer Ridge.

3.2.1 Linearized Zoeppritz Equations

Since the Zoeppritz equations are highly nonlinear with respect to velocities and density, many approximations have been made in order to linearize them. Aki and Richards (1980) assumed small layer contrasts, simplifying the equations to the following form:

\[
R(\theta) = \frac{1}{2 \cos^2(\theta)} I_p - 4 \gamma^2 \sin^2(\theta) I_s + (2 \gamma^2 \sin^2(\theta) - 0.5 \tan^2(\theta)) D \tag{3.1}
\]
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where

\[ I_p = \left( \frac{\Delta v_p}{v_p} + \frac{\Delta \rho}{\rho} \right), \]  
\[ I_s = \left( \frac{\Delta v_s}{v_s} + \frac{\Delta \rho}{\rho} \right), \]  
\[ D = \frac{\Delta \rho}{\rho}, \] 

with

\[ \Delta v_p = v_{p2} - v_{p1}, \]
\[ \Delta v_s = v_{s2} - v_{s1}, \]
\[ \Delta \rho = \rho_2 - \rho_1, \]
\[ v_p = \frac{(v_{p2} + v_{p1})}{2}, \]
\[ v_s = \frac{(v_{s2} + v_{s1})}{2}, \]
\[ \rho = \frac{(\rho_2 + \rho_1)}{2} \] 

And

\[ \gamma = \left( \frac{v_s}{v_p} \right) \]

The relative contrasts in P-impedance, S-impedance and density are given by \( I_p, I_s \) and \( D \), respectively. \( \theta \) is the reflection angle and \( \gamma \) is an estimate of the background shear to compressional velocity ratio \( v_s/v_p \).

The reflectivity curves corresponding to either a unit perturbation in P-impedance contrast \((I_p = 1, I_s = 0, D = 0)\), S-impedance contrast or density contrast can be seen in Figure 3.1. For a unit perturbation in relative P-impedance contrast, the P-impedance inversion curve dominates at small angles of incidence and increases with increasing offset. For a unit perturbation in relative S-impedance contrast, the S-impedance inversion curve is zero at normal incidence and is increasingly negative with increasing offset. Over the conventional range of surface reflection data acquisition geometry illumination, which is typically 0° to 35°, the density inversion curve is not significant, as most of the density contrast contributes to the reflection AVO through the impedance contrasts alone. As the reflection amplitudes are mostly a combination of the P- and S-impedance contrast inversion curves, reflectors with P- and S-impedance contrasts of the same polarity and magnitude are expected to show approximately constant amplitude versus offset. On the other hand, reflectors with
P- and S-impedance contrasts of opposite polarities, indicating a transition zone of changing rock pore fluid properties, should show increasing amplitudes versus offset.

![Theoretical Impedance Contrasts](image)

Figure 3.1: Theoretical P, S, and Density impedance inversion curves.

3.2.2 Inversion Technique

Reflectivity data can be inverted for changes in P- and S-impedance across an interface and therefore for possible pore fluid transitions. After adequate amplitude recovery through preprocessing, the Blake Outer Ridge data should approximate the reflection coefficient and can thus be used in an elastic parameter inversion using equation 3.5. The P- and S-impedance contrasts at each subsurface position are estimated by applying a least-squares elastic parameter inversion method (Lumley, 1993b). This technique fits seismic gathers at each pseudo-depth and surface position to the theoretical P- and S-impedance curves (see Figure 3.1). The method uses bootstrapping with offset and angles based on the properties of the theoretical impedance contrast curves displayed in Figure 3.1. First, $I_p$ is estimated for angles $\theta \leq 15^\circ$. Then a least-squares estimate is found for $I_s$ for angles $10^\circ \leq \theta \leq 40^\circ$ using the estimate of
I_p as a constraint. Finally, if the data contain angles above 35°, the density contrast D can be estimated using both I_p and I_s as constraints.

3.2.3 Expected BSR AVO Responses

Changes in the AVO responses, as shown in Figure 3.1, can be directly related to changes in Poisson’s ratio, indicating pore fluid transitions. Three different possibilities can be imagined for the transition from hydrate-bearing sediment to the sediments underneath:

- **Case 1**: Poisson’s ratio in the hydrate layer (layer 1) is smaller than Poisson’s ratio in the layer underneath (layer 2)
- **Case 2**: Poisson’s ratio in the hydrate layer is approximately the same as the one in the layer underneath
- **Case 3**: Poisson’s ratio in the hydrate layer is larger than the one underneath

Figure 3.2 shows the theoretical AVO curves corresponding to these three different scenarios. Since the BSR is characterized by both a negative reflectivity and a drop in P-wave velocity (see Chapter 2), the zero-offset reflection amplitude and the resulting P-impedance contrast can be expected to be negative. In Chapter 2, I have shown that the presence of hydrate increases the P-wave velocity of the sediment. If, additionally, hydrate-bearing sediment possesses a high S-wave velocity with respect to the surrounding sediment, amplitudes should decrease as angle increases (case 1). Consequently, the BSR would be caused mainly by the strong effect of the hydrate on the elastic moduli of the saturated in-situ sediments. The BSR would thus be characterized by both a strong negative P-impedance contrast and S-impedance contrast. If, on the other hand, hydrate-bearing sediment does not have a significant impact on Poisson’s ratio and is underlain by brine sediment with approximately the same Poisson’s ratio, the amplitudes should remain constant with increasing angle
In this case, the BSR would be the consequence of hydrate overlying brine-saturated sediment, and would be characterized by a negative P-impedance contrast and a slightly negative S-impedance contrast. Domenico (1976) showed that gas saturation can significantly decrease Poisson’s ratio with respect to brine saturation. Such a decrease in Poisson’s ratio should manifest itself in an increasingly negative amplitude versus offset (Ostrander, 1984). This assertion is based on the drop in P-wave velocity and density caused by the gas saturation and the simultaneous slight increase in S-wave velocity (case 3). In this case, the BSR would be the result of hydrate-saturated sediment overlying gas-saturated sediment and be characterized by negative P-impedance and positive S-impedance.

3.3 Impedance Inversion of the Data

In this section, I describe the results of the 2-D inversion for P- and S-impedance contrasts of the data from the Blake Outer Ridge. Then I focus singularly on the
reflection corresponding to the bottom simulating reflector (BSR) and evaluate its amplitudes locally. This minimizes the errors that might have been introduced into the 2-D inversion by uncorrected residual moveout and yields good insight into the lateral and vertical elastic property variations at the transition from hydrate-bearing sediments to the sediments beneath.

3.3.1 2-D Impedance Estimation

The 2-D impedance inversion is performed on the preprocessed, migrated, and amplitude-calibrated gathers and reflection angles described in Chapter 2. Since the angles of incidence for the Blake Outer Ridge data are restricted to values of less than 35°, I do not include the density contrast $D$ (see equation 3.5) in the inversion. Therefore, the linearized Zoeppritz equations are only solved for relative changes in P- and S-impedance using the method described in the previous section (3.2.1).

The resulting P-impedance contrast section is displayed in Figure 3.3. It clearly shows that the seafloor and the BSR have P-impedance contrasts of opposite polarity, but approximately the same magnitude. The BSR contrast is strongest between 44 km and 50 km lateral distance, and strongly discontinuous between 36 km and 44 km. The BSR's negative P-impedance contrast suggests a velocity decrease at the transition from hydrate-bearing sediment to the sediment underneath. This observation matches the velocity information as described in Chapter 2, section 2.3. The flat reflector below the BSR gives a strong impedance contrast of the same polarity as the seafloor.

The S-impedance contrast section is displayed in Figure 3.4. The seafloor and the BSR appear to be characterized by S-impedance contrasts of the same, positive polarity. This indicates an S-impedance contrast across the BSR of opposite polarity than the P-impedance contrast. Underneath the BSR, the S-impedance contrast enhances the structure dipping against the BSR. The flat reflector, which was strongly visible in the P-impedance contrast map, has nearly disappeared in the S-impedance contrast map, suggesting that this transition zone has a very small S-impedance contrast.
P-Impedance Contrast

Figure 3.3: P-impedance contrast section. [avo-pmp-ann] [ER]
Figure 3.4: S-impedance contrast section.
The windowed P- and S-impedance contrast sections are shown in Figure 3.5. As described before, the figure clearly displays the differences in the polarity of the P- and S-impedance contrasts across the BSR. While the P-impedance contrast appears to be negative, the data show a positive S-impedance contrast. Furthermore, the S-impedance contrast section appears to enhance structural interference across the BSR from underlying features.

![P-Impedance Contrast](image1)

![S-Impedance Contrast](image2)

Figure 3.5: Windowed P-impedance contrast (left panel) and S-impedance contrast (right panel).

A simple multiplication of the P- and S-impedance contrast sections produces a P*S anomaly map shown in Figure 3.6. P- and S-impedance contrasts of the same sign are plotted as black, while contrasts of opposite sign are plotted white. The anomaly map shows the same contrasts at the seafloor. Since the transition from water to
sediments is characterized by an increase in P-wave velocity, S-wave velocity and density, both a positive P-wave impedance contrast and a positive S-wave impedance contrast can be expected. In a small section above the BSR, there is a "quiet" zone where no diffractions or reflections are visible that might be due either to the presence of disseminated methane hydrate in the sediments (Lee et al., 1994) or to the naturally low reflectance of a uniform sedimentary section at the Blake Outer Ridge (Holbrook et al., 1996). The BSR reflection displays mostly opposite P and S impedance polarity. The velocity analysis in Chapter 2 has shown that the BSR is characterized by a significant decrease in velocity, and thus a negative P-impedance contrast. P- and S-contrasts of the opposite polarity would indicate a positive S-impedance contrast at the base of the hydrate zone. The S-impedance contrast is the sum of the relative changes in S-wave velocity across an interface and the corresponding density changes. Assuming small changes in density across the BSR, the positive S-impedance contrast would be the result of a positive S-wave velocity contrast, which is clearly opposite to the present negative P-wave velocity contrast. A hydrate-gas transition zone would explain such P- and S-impedances, whereas a hydrate-brine transition zone could be expected to have both negative P- and S-impedances (see Figure 3.2). This seems to indicate the presence of free gas beneath the BSR. Moreover, the strong reflector beneath the BSR, which is characterized by a strong, positive P-impedance contrast, seems to have nearly no S-impedance contrast. This small change in shear impedance suggests the transition from gas- to brine-saturated sediment.

The structure beneath the BSR as seen in Figure 3.3 and 3.4 can interfere with the actual BSR amplitudes and cause strong distortions in both waveform and amplitude, and invalidate the inverted P-and S-impedance contrasts. Furthermore, possible uncorrected residual moveouts in the data can add to errors since the 2-D inversion depends upon perfectly flat reflectors. Therefore, I evaluate the BSR amplitude responses locally, both to obtain an insight into the effect of the structure under the BSR, and to minimize possible residual moveout errors.
Figure 3.6: P*S impedance contrast map.
3.3.2 Local BSR AVO Response

The local examination of the transition across the BSR and possible interference effects proceeds in two steps. First, I reexamine the migrated and calibrated gathers. Then I pick the maximum (negative) amplitudes of the BSR reflection as a function of offset and invert them for impedance contrasts. This eliminates any error that might have been introduced into the 2-D inversion due to possible uncorrected residual moveouts of the events.

Reflectivity Gathers

I pick several spots along the BSR to examine the amplitudes that correspond to the BSR reflection. The corresponding gathers can be seen in Figures 3.7 and 3.8. The first panel of Figure 3.7 represents a gather from the continuation of the BSR at 28 km distance, the second one shows a gather from 33 km distance. The third panel at 37 km distance represents the part of the BSR that appears highly discontinuous. The gathers displayed in Figure 3.8 are from the strong parts of the BSR at 42 km, 44 km and 46 km, respectively. The reflection events that should correspond to the BSR are indicated by an arrow.

In the case of clearly separable reflection events from the top and bottom of different layers, the BSR should bear a wavelet shape similar to the seafloor but with reversed polarity. The gathers along the continuation of the BSR (see the left and middle panel in Figure 3.7) do not show a strong, negative BSR reflection, but instead a wavelet shape that appears to be indicative of thin layering (tuning). Depending on the time-thickness of the interfering events, the wavelets of both reflections are being superimposed and influence both shape and amplitude in the recorded reflection (Yilmaz, 1987). Figure 3.3 shows that the BSR in this region between 25 km and 35 km lateral distance appears to be wedged against the flat reflector beneath the BSR, which is still dimly visible. This can yield a superposition of both reflection events. The P- and S-impedance contrasts across the flat layer beneath the BSR indicate that this flat reflection might be the transition from gas- to brine-saturated sediment.
Figure 3.7: Reflectivity gathers showing the assumed BSR reflections at lateral distances of 28 km, 33 km and 37 km.
Figure 3.8: Reflectivity gathers showing the BSR reflections at lateral distances of 42 km, 44 km and 46 km.
This would suggest that the BSR between 25 and 35 km lateral distance might be underlain by a thin free gas zone.

The right panel of Figure 3.7 shows a gather representing the part of the BSR which appears highly discontinuous (see Figure 3.3 between 35 km and 44 km). There is no clearly identifiable negative BSR reflection, but it appears to be an increasingly positive event. Among the factors potentially contributing to this drastic disappearance are: disturbance of the hydrate stability field, fractures and thin layering.

The gathers between 42 and 46 km, as displayed in Figure 3.8, correspond to strong BSR reflections and show clearly identifiable BSR wavelets. However, the gathers at 42 km and 44 km distance also appear to display some interfering effects with the layering underneath the BSR. Comparison with Figures 3.3 and 3.4 shows that the location of these gathers (42 km and 44 km) coincides with strong underlying structure. At a lateral distance of 46 km, the gather in the right panel of Figure 3.8 shows a clear BSR response that has increasingly negative amplitudes with increasing offset. This BSR amplitude behavior is consistent with the BSR amplitude characteristic observed in this region by Lee et al. (1994) and Katzman et al. (1994).

The structure and thin layering intersecting with the BSR can cause a strong distortion of the original BSR amplitudes. Therefore, the amplitudes and resulting impedance contrasts have to be evaluated carefully.

**Local BSR AVO Inversion**

After qualitatively examining the BSR amplitudes in several gathers, I quantify the actual AVO responses along the BSR by first picking the maximum (negative) amplitude along the BSR, and then inverting locally for the P- and S-impedance contrasts. In this way, I minimize errors that might have been introduced by the 2-D inversion due to uncorrected residual moveouts. Furthermore, this analysis results in a direct insight into the transition from hydrate-bearing sediment to the sediment underneath and the possible effect that tuning might have on the resulting impedance contrasts.
The analysis in the previous section suggested that between 25 and 38 km lateral distance the BSR might be underlain by a thin layer. Thus, the BSR amplitudes might suffer from tuning effects. Therefore, I examine only the BSR amplitudes between 40 and 52 km lateral distance.

**BSR Impedance Contrasts**

The resulting P- and S-impedance contrasts across the BSR can be seen in Figure 3.9, with P-impedance contrast plotted as (o), and S-impedance contrast plotted as (*). The local impedance contrasts display well-constrained, negative P-impedance contrasts, as already expected. However, at about 41.8 km and 44 km, the P-impedance contrasts appear “spiky”. Looking back at Figures 3.5, 3.6, and 3.8, this behavior seems to be associated with the structure underneath the BSR that is interfering with the BSR response. Therefore, possible tuning effects might account for the anomalously large P-impedance contrasts.

The S-impedance contrast across the BSR displays a laterally more scattered response. Since the limited angle coverage of the data causes the inversion for changes in S-impedance to be less well constrained, more variance can be expected. Most
of the S-impedance contrast appears to be positive, but there is significant lateral heterogeneity. In areas of structural interference, the S-impedance contrast is strongly negative, i.e. at 42 km and 44 km. As mentioned before, this amplitude anomaly is most likely based on wavelet interference effects.

The actual BSR AVO function that produced these impedance contrasts is shown in Figure 3.10. The figure displays the BSR amplitude difference between the near-offset traces and the far-offset traces. A positive difference means that the amplitudes at the BSR are increasingly negative with increasing offset, while a negative difference represents decreasing amplitudes with increasing offset. The zero crossing (constant amplitudes with increasing offset) is represented by the solid line. As expected from the P- and S-impedance contrasts, the AVO behavior shows decreasing amplitudes with offset at 42 km and 44 km, as well as at about 46.5 km and after 50 km. These regions probably represent a superposition of structure and BSR amplitudes. The BSR AVO behavior shows mostly positive amplitude differences, thus indicating increasing amplitudes with increasing offset, or regions where the difference becomes fairly small, suggesting nearly constant AVO in these regions. Excluding the very apparent tuning effects at 42 km, 44 km and 46 km, the lateral BSR amplitudes thus display two different trends:

- increasingly negative amplitudes with increasing offset
- nearly constant amplitudes with constant offset

The increasingly negative amplitudes with offset represent the positive S-impedance contrasts, while the constant amplitudes with offset yield small positive or negative contrasts, depending on the density structure. Figure 3.2 shows that these amplitude variations probably predict the existence of both brine- and gas-saturated sediment underneath the BSR. However, thin layering might be distorting the BSR amplitudes. Increasingly negative amplitudes with increasing offset have been attributed to the BSR appearance in the region of the Blake Outer Ridge (Lee et al., 1994; Katzman et al., 1994). Furthermore, Ostrander (1984) has shown that a thin gas layer can cause
amplitudes to be diminished, thus causing either decreasing or constant amplitudes with offset. Therefore, the regions displaying constant amplitudes with offset and thus appearing to indicate the presence of hydrate-bearing sediment overlying brine-saturated sediment might also be caused by tuning effects. AVO modeling of the two different AVO responses observed in the data should help us to distinguish these effects.

### 3.4 Confidence of Zero-Offset Reflectivity

In order to conduct AVO modeling of amplitudes, it is important to have a measure of the actual zero-offset reflection amplitude. The prestack migration described in
Chapter 2 should have resulted in a good estimate of the subsurface zero-offset reflectivity. This reflectivity is both input into the impedance inversion and the subsequent AVO modeling. The inverted P-impedance contrast is approximately twice the size of the actual zero-offset reflection.

The zero-offset reflectivity can also be obtained from the first multiple of the seafloor. In a layered medium, the reflection coefficient of the seafloor and of every subsequent layer can be calculated after Warner (1990):

\[
R_w = \frac{A_m}{A_w},
\]

\[
R_i = \frac{A_i}{A_w} R_w
\]

where \( R_w \) is the reflection coefficient of the seafloor, \( A_w \) is the amplitude of the seafloor reflection and \( A_m \) that of the first seafloor multiple. The reflection coefficient of the \( i \)th layer is given by \( R_i \), and \( A_i \) is the amplitude of the \( i \)th reflection. The original Blake Outer Ridge data was recorded to about 12 s two-way travel time, and thus includes the first seafloor multiple.

Comparison of the reflection amplitudes obtained from both methods will tell me how much confidence to give the reflectivity resulting from the migration and hence the zero-offset amplitudes that are input into the AVO modeling. Because in the case of using the first seafloor multiple the reflection coefficients of all subsequent reflectors are based on the seafloor reflection coefficient, I compare the two results on the seafloor.

Figure 3.11 displays the comparison between the reflection coefficients of the seafloor. The left panel shows the coefficients as obtained after migration, while the right panel displays the reflection amplitude after calibration with the first seafloor multiple. The reflection coefficients after migration appear better constrained than those obtained via multiple calibration. Since the first seafloor multiple occurs very deep in the section (more than 9s two-way travel time), more errors due to decrease in the S/N noise ratio and picking can be expected. The overall trend of the reflection
coefficients is, however, fairly similar. Therefore, I conclude that the zero-offset reflection amplitude of my data does indeed yield a good approximation of the subsurface reflectivity.

3.5 AVO Modeling

After having evaluated the different impedance structures across the BSR, I perform AVO modeling of the BSR amplitudes in order to obtain possible velocity models that can explain the cause of the BSR. These models can be used to examine the actual predictions based on the impedance inversion.

Using the interval velocities presented in Chapter 2, I explore the effects of different impedance structures on the BSR AVO response in an attempt to reproduce the observed seismic data. Several models are constructed based on the possible causes...
of the BSR shown in Figure 3.2. The velocity models are constrained to preserve the average interval velocity of the macro layers. To avoid possible tuning effects in this basic modeling approach, the layers are assumed to be thicker than a quarter of a wave length. Synthetic AVO responses are then estimated for the individual models using Zoeppritz equations (3.5) and compared with the amplitude responses observed by the seismic data.

3.5.1 BSR AVO Modeling

Increasing Amplitudes with Offset

Here I analyze one of the two amplitude effects observed in the data: increasing amplitudes with increasing offset. Figure 3.12 shows a reflectivity gather and the corresponding BSR AVO trend picked along the BSR reflection. The reflectivity gather displays a fairly well-resolved BSR wavelet. The AVO trend shows clearly the increasingly negative amplitudes with increasing angle. The offsets were converted into angles by using the reflection angles at the BSR, which were computed by the prestack migration (see Chapter 2). For incidence angles between 17° and 18°, the amplitudes are anomalously low as a result of uncorrected amplitude effects as described in Chapter 2, section 2.4.3 (i.e. hydrophone array attenuation at central offsets). Therefore, these two points are given small weights in the following AVO modeling procedure.

The initial P- and S-wave velocity used in the modeling are shown in Figure 3.13. The P-wave velocity above and below the BSR is calculated directly by averaging the interval velocity obtained from velocity analysis. The S-wave velocity is determined by assuming a Poisson's ratio of 0.45, which is consistent with the brine-saturated, highly unconsolidated sediments typical of this region.

In the first attempt to model the observed AVO amplitudes, these initial velocities are used as input velocities. The resulting AVO curve is obtained using Zoeppritz equations and is compared with the one observed in the data (Figure 3.14). The comparison of both curves shows that with nearly constant amplitudes with increasing
angles, the initial velocity model fails not only to reproduce the zero-offset reflection coefficient, but also to reproduce the general AVO trend. Assuming negligibly small density contributions, the near-offset amplitudes are mainly dependent on the P-wave velocity contrast at the reflector, while the AVO trend is characterized primarily by the S-wave velocity contrast. Thus, the AVO response resulting from the initial velocity model implies the use of both incorrect P- and S-wave velocities at the BSR.

Based on this result, the subsequent modeling attempts to increase the P-wave velocity contrast across the BSR in order to recreate the observed zero-offset reflection amplitudes. The required increase is obtained by increasing the velocity in the hydrate layer and simultaneously decreasing the velocity in the layer underneath (Figure 3.15). This yields a thinner hydrate layer over brine sediment. The S-wave velocity is again determined using a Poisson’s ratio of 0.45.

The resulting AVO trend is shown in Figure 3.16. The comparison of the modeled AVO response with the observed response indicates that this model can successfully
Figure 3.13: Average interval velocities across the BSR for the case of increasing amplitudes with angle.

Figure 3.14: AVO curve obtained from the initial P-wave velocity and assumed brine Poisson’s ratio (solid line) compared with the one observed in the data (*).
Figure 3.15: Interval velocities across the BSR for a thin hydrate-over-brine model in the case of increasing amplitudes with angle. The solid line represents the initial velocities. The arrows indicate if the modeled velocities have to be increased or decreased. [avointl-vp] [NR]

Figure 3.16: AVO curve obtained by increasing the P-wave velocity contrast across the BSR (solid line) compared with the one observed in the data (*). [avo-fit1-vp] [ER]
reproduce the zero-offset data. This suggests a P-wave velocity of 2.05 km/s in the hydrate and 1.58 km/s in the underlying sediments might resemble the actual conditions at the BSR. However, the calculated AVO trend is still contrary to the observed one, displaying nearly constant amplitudes with increasing angles. Hence, a change in Poisson’s ratio seems to be required at the transition from hydrate-bearing sediments to the sediments underneath.

The observed AVO trend of the data suggests that the hydrate-bearing sediment has a higher Poisson’s ratio than the sediment underneath (see Figure 3.1). Since it does not make sense physically to increase the hydrate Poisson’s ratio above 0.45 (fluids have Poisson’s ratios of about 0.5), I decrease the Poisson’s ratio in the layer underneath to simulate the drop in Poisson’s ratio. This yields the same P-wave velocities as described before, but an increase in S-wave velocity of about 0.25 km/s across the BSR (Figure 3.17). The resulting velocities represent a Poisson’s ratio of 0.45 in the hydrate and approximately 0.28 in the sediments underneath. Instead of keeping the hydrate Poisson’s ratio constant, I could have increased the S-wave velocity there as well. This would have yielded a lower Poisson’s ratio in the hydrate and would have required an even lower Poisson’s ratio than 0.28 in the sediments underneath. Nonetheless, both cases would require a strong increase in S-wave velocity across the BSR. Since the drilling at the Blake Outer Ridge (Matsumoto et al., 1996) has shown that the overall shear structure in the sediments is very weak, increasing the model S-wave velocities too much would generate conditions that do not resemble in-situ conditions.

The significant drop in P-wave velocity at the BSR as well as the decrease of Poisson’s ratio suggests the presence of free gas underneath the hydrate layer. Domenico showed that the presence of gas can cause a decrease in Poisson’s ratio down to 0.1 (Domenico, 1976). However, because the sediments at the Blake Outer Ridge are highly unconsolidated and have shale contents of more than 50% (Matsumoto et al., 1996), I would not expect the Poisson’s ratio to drop as low as 0.1 in the presence of free gas.

A comparison of the synthetic AVO curve obtained from the model in Figure 3.17
Figure 3.17: Interval velocities for hydrate-bearing sediments overlying gas-saturated sediments. The initial velocities are represented by the solid lines. The arrows indicate whether the modeled velocities have to be decreased or increased.

Figure 3.18: AVO curve obtained by the hydrate-over-gas velocity model (solid line). It is compared with the AVO behavior observed in the real data.
with the AVO trend observed in the data is shown in Figure 3.18. The synthetic curve agrees well with the real data for both near and far offsets. Thus, a strong increase in S-wave velocity and a simultaneous decrease in P-wave velocity at the transition from hydrate-bearing sediments to sediments containing free gas is required to explain the seismic data. The inferred velocity contrasts are, furthermore, in good agreement with the prediction based on the negative P-impedance contrast and positive S-impedance contrast obtained from the impedance inversion.

**Nearly Constant Amplitudes with Offset**

Here I analyze the second amplitude effect observed along the BSR: nearly constant amplitudes with increasing angle. A representative reflectivity gather and the picked BSR amplitude trend are shown in Figure 3.19. Since some residual amplitudes are left at the central offsets between 16° and 20° (Chapter 2, section 2.4.3), those amplitudes are edited manually to be in better agreement with the general amplitude trend.

As described before, I determine the initial velocity models by averaging the interval velocity determined in Chapter 2 and assuming a Poisson's ratio of 0.45, representative of brine-saturated sediments. This initial velocity model is shown in Figure 3.20.

The synthetic AVO curve corresponding to this velocity model is again calculated using Zoeppritz equations. It is compared with the AVO trend observed in the data in Figure 3.21. The comparison shows that this first velocity model fails to reproduce the zero-offset reflection coefficient, while the general AVO trend appears to be in fairly good agreement. This already suggests that this AVO trend might not represent a hydrate-over-gas model.

In order to fit the data, the P-wave velocity across the BSR must be increased and the S-wave velocity contrast slightly decreased (Figure 3.22). The result is similar to the P-wave velocity contrast described in the previous section: the hydrate velocity increases to approximately 2.07 km/s while the velocity in the underlying sediment
Figure 3.19: Reflectivity gather shown on the left panel and the picked BSR AVO trend on the right panel. The amplitudes are nearly constant with increasing offset.

Figure 3.20: Initial velocity models across the BSR for the case of constant amplitudes with angle.
Figure 3.21: Synthetic AVO curve obtained from the initial velocities (solid line) compared with the AVO trend observed in the data (*). [avo-fit3] [ER]

decreases to about 1.55 km/s. The slight decrease in S-wave velocity contrast results in a slightly smaller Poisson’s ratio of about 0.43 in the sediments below the hydrate, compared to a value of 0.45 in the hydrated sediment itself. This small decrease in Poisson’s ratio is consistent with a hydrate-over-brine model, and may be explained by the heterogeneity of the sediments, which can cause not only variations in P-wave velocity but also in Poisson’s ratio.

The resulting AVO curve is shown in Figure 3.23. The comparison of the modeled AVO curve with the one observed indicates that this model successfully reproduces the zero-offset reflection coefficient and the AVO trend. This good agreement suggests that there exists a small decrease in S-wave velocity across the BSR. This indicates that the model of hydrate over brine-saturated sediment would be required to reproduce the nearly constant amplitudes with increasing offset.

The analysis of the nearly constant amplitude trend with increasing angle indicates that the BSR amplitudes require a significant negative P-wave velocity contrast and a small negative S-wave velocity contrast. Based on the fairly unchanged Poisson’s ratio at the transition, this would suggest a hydrate-over-brine sediment model for these amplitude variations. However, the significant decrease in the velocities underneath
Figure 3.22: Interval velocities in the case of slightly decreasing Poisson’s ratio in the sediments beneath the hydrate. The solid line represents the initial velocities. The arrow indicates if the modeled velocities have to be increased or decreased.

Figure 3.23: Synthetic AVO curve obtained from the interval velocity (solid line) compared with the amplitudes observed in the data (*).
the hydrate, consistent with drilling results and similar seismic analyzes in the region, strongly suggests the presence of free gas (Matsumoto et al., 1996; Holbrook et al., 1996; Lee et al., 1994; Katzman et al., 1994; Wood et al., 1994). Consequently, two interpretations are possible:

- brine- and gas-saturated zones are distributed irregularly underneath the BSR, and the amplitude trend does in fact correspond to a hydrate-over-brine region.
- thin layering underneath the BSR distorts the amplitude response, causing the amplitude to resemble a hydrate-over-brine trend instead of a hydrate-over-gas trend.

A heterogeneous distribution of gas and brine saturation is physically plausible in marine sediment. The gas can be trapped in thin layers that serve as permeability barriers, yielding a patchy gas distribution (Dillon et al., 1993). However, the resolution of localized patches of brine and gas saturation from surface seismic is highly questionable. Therefore, it is more reasonable to assume that the constant amplitudes result from thin layer tuning effects, rather than patchy brine saturation beneath the BSR.

3.6 Conclusions

In this chapter, I have described the applied impedance inversion on the data from the Blake Outer Ridge and the subsequent AVO modeling analysis. The 2-D impedance inversion has shown that the BSR is characterized by a strong negative P-impedance contrast, as was expected from the observed decrease in interval velocity across this interface (Chapter 2). Furthermore, the BSR has a mostly positive S-impedance contrast. These impedance contrasts strongly suggest the presence of free gas beneath the BSR. The amplitudes along the BSR are distorted at several locations laterally along the BSR, probably due to thin layering underneath the BSR. The flat layer beneath the BSR has a positive P-impedance contrast and a very weak S-impedance contrast,
indicating a transition from gas-saturated sediment to normal brine-saturated sediment. Because this flat reflector appears to be “wedged” against the portion of the BSR between 25 and 35 km lateral distance, the presence of a thin free gas layer underneath the BSR is implied. The subsequent AVO modeling confirmed the model of hydrate overlying free gas. A strong negative P-wave velocity contrast and a simultaneous increase in S-wave velocity is required to reproduce the observed increasing amplitude with increasing angle. This implies the presence of hydrate to increase the P-wave velocity in the hydrate-bearing sediments, but not drastically change the shear structure of the saturated sediment. Furthermore, a second AVO trend connected to the BSR, showing nearly constant amplitudes with angle, suggests the existence of some brine-saturated regions below the BSR. However, looking at the wavelet shape and the significant decrease in P-wave velocity, it might be more reasonable to assume that this trend was produced by thin bed tuning. Ostrander (1984) showed that thin bed tuning can affect amplitudes more than 30%. This would explain the distorted AVO trend.

3.7 Acknowledgements

The initial part of this study was done in collaboration with David Lumley, whom I would like to thank for his many helpful discussions and suggestions. Also, I would like to thank Sergey Fomel for the use of his inversion library.
Chapter 4

Hydrate Saturation Estimation

4.1 Overview

In this chapter, I link the information available from seismic with physical rock models. This enables me to address an important question that arises after having evaluated the cause of the BSR and the properties of the sediments across the BSR: how much methane hydrate is present in the sediment? This is an essential question to answer if we are to realistically evaluate the possible impact methane hydrates might have as a future energy resource. Here, I provide a theoretical tool for quantifying the amount of gas hydrate and gas around the BSR at the Blake Outer Ridge. Since there is no direct well control of the data, all estimations are based solely on information available from surface seismic. I examine three different micromechanical models of hydrate formation: (A) hydrate is part of the pore fluid; (B) hydrate becomes part of the solid frame, thus reducing porosity and weakly affecting the stiffness of the sediment; and (C) hydrate cements grain contacts and therefore strongly reinforces the sediments. Using the interval velocities obtained from velocity analysis together with the rock-physics models, I obtain lateral maps of hydrate and gas saturation across the BSR. Model A predicts maximum hydrate saturations between 20% and 26%, model B saturations between 15% and 20% and model C saturations less than 1%. Maximum gas saturation is between 1% and 2%. Subsequently, I analyze the
stability of these estimates to errors in the interval velocities. Such errors can cause the estimations to vary as much as ± 14% (note that % refers to the saturation itself and not to the percentage of saturation). Therefore, accurate velocity determination is crucial for correct reservoir characterization. Finally, I validate the technique by using known well-log velocities and porosities from wells 994 and 995 at the Blake Outer Ridge.

4.2 Rock-Physics Models

In this section, I describe the rock-physics models which are used to determine the internal structure of the hydrated sediment and the amount of hydrate present in the sediment. These physical rock models link the elastic wave velocities in high-porosity sediments to density, porosity, effective pressure, mineralogy and water, gas and hydrate saturation.

Figure 4.1: Hydrate models used in this study. Model A represents hydrate being part of the fluid. In model B, hydrate becomes part of the solid frame. Model C assumes that hydrate cements the grains evenly.

In order to estimate the amount of hydrate present in the sediment, I examine three different models of possible hydrate deposition in the pore space (Figure 5.1). In the first model, I assume that hydrate is suspended in the water, thus contributing
only to the bulk modulus of the pore fluid (Figure 5.1A). In the second model, hydrate becomes part of the solid (Figure 5.1B). This causes a reduction of porosity and an additionally weak stiffening of the sediment structure. The third model assumes that hydrate cements grain contacts (Figure 5.1C), significantly changing the rock stiffness and again reducing the porosity. This last model is probably not likely to apply physically to the Blake Outer Ridge, since the sediments in this region are highly unconsolidated (Matsumoto et al., 1996). However, since I want to examine the effect of the different micro-models on the saturations and since this third model (C) might be applicable to regions other than the Blake Outer Ridge, I include it in this study. Strictly speaking, the cementation theory is only valid for porosities less than 40%. However, it can be used to approximately estimate the elastic properties of granular aggregate of higher porosities.

4.2.1 Sediments without Gas Hydrate

The dry sediment moduli without hydrate can be calculated using a modified Hashin-Shtrikman-Hertz-Mindlin theory (Dvorkin and Nur, 1996). The subsequent saturated sediment moduli are calculated using Gassman’s equations. The calculations can be applied both to the brine-saturated sediments and the gas-saturated sediments at the Blake Outer Ridge.

Dry Sediment Properties

In order to relate the velocities in marine sediments without gas hydrates to porosity, saturation, mineralogy, and effective pressure, I use a modified Hashin-Shtrikman-Hertz-Mindlin theory first introduced by Dvorkin and Nur (1996). This theory first calculates the effective bulk and shear moduli at critical porosity ($\phi_c \approx 40\%$) using the Hertz-Mindlin theory (Mindlin, 1949). Critical porosity separates the mechanical and acoustic behavior into two distinct regions (Nur et al., 1995): for porosities lower than $\phi_c$, the mineral grains are load bearing, while for porosities greater than $\phi_c$, the sediment becomes a suspension, with the fluid phase load-bearing. The effective
moduli at critical porosity are given by:

\[ K_{HM} = \left[ \frac{n^2 (1 - \phi_c)^2 G^2}{18 \pi^2 (1 - \nu)^2} P \right]^{\frac{1}{3}}, \]

\[ G_{HM} = \frac{5 - 4 \nu}{5 (2 - \nu)} \left[ \frac{3 n^2 (1 - \phi_c)^2 G^2}{2 \pi^2 (1 - \nu)^2} P \right]^{\frac{1}{3}}, \tag{4.1} \]

where \( K \) and \( G \) are the bulk and shear moduli of the mineral making up the rock. The Poisson’s ratio is given by \( \nu \), \( P \) is the effective pressure and \( n \) is the average number of grain contacts, taken to be 8.5 (Murphy, 1982). The effective pressure \( P \) is calculated as follows:

\[ P = (1 - \phi) (\rho_s - \rho_f) g h \tag{4.2} \]

where \( \rho_s \) and \( \rho_f \) are the solid and fluid density, respectively; the depth below the seafloor is given by \( h \), and \( g \) is the gravity acceleration.

If the sediment rock consists of a mixed mineralogy, the bulk and shear moduli \( K \) and \( G \) of the rock can be determined using a Hill’s average formula:

\[ K = \frac{1}{2} \left[ \sum_{i=1}^{m} f_i K_i + \left( \sum_{i=1}^{m} \frac{f_i}{K_i} \right)^{-1} \right], \]

\[ G = \frac{1}{2} \left[ \sum_{i=1}^{m} f_i G_i + \left( \sum_{i=1}^{m} \frac{f_i}{G_i} \right)^{-1} \right], \tag{4.3} \]

where \( m \) is number of different mineral components, \( f_i \) is the volumetric fraction of the \( i \)th component in the rock, and \( K_i \) and \( G_i \) are the bulk and shear moduli of the \( i \)th component, respectively.

Subsequently, the dry moduli of the solid phase can be calculated for porosities above and below the critical porosity \( \phi_c \) using a modified Hashin-Strikam upper and lower bound (Dvorkin and Nur, 1996; Ecker et al., 1996b).

**Porosity \( \phi \) below critical porosity \( \phi_c \):**

For sediment porosities below the critical porosity, the dry moduli are determined
CHAPTER 4. HYDRATE SATURATION ESTIMATION

by the following equations:

\[
K_{\text{dry}} = \left( \frac{\phi/\phi_c}{K_{HM} + \frac{4}{3} G_{HM}} + \frac{1 - \phi/\phi_c}{K + \frac{4}{3} G_{HM}} \right)^{-1} - \frac{4}{3} G_{HM},
\]
\[
G_{\text{dry}} = \left[ \frac{\phi/\phi_c}{G_{HM} + Z} + \frac{1 - \phi/\phi_c}{G + Z} \right]^{-1} - Z,
\]
\[
Z = \frac{G_{HM}}{6} \left( \frac{9 K_{HM} + 8 G_{HM}}{K_{HM} + 2 G_{HM}} \right)
\] (4.4)

Porosity \( \phi \) above critical porosity \( \phi_c \):

If the porosity is above the critical porosity of 40%, the dry moduli can be calculated as follows:

\[
K_{\text{dry}} = \left[ \frac{(1 - \phi)/(1 - \phi_c)}{K_{HM} + \frac{4}{3} G_{HM}} + \frac{(\phi - \phi_c)/(1 - \phi_c)}{\frac{4}{3} G_{HM}} \right]^{-1} - \frac{4}{3} G_{HM},
\]
\[
G_{\text{dry}} = \left[ \frac{(1 - \phi)/(1 - \phi_c)}{G_{HM} + Z} + \frac{(\phi - \phi_c)/(1 - \phi_c)}{Z} \right]^{-1} - Z
\] (4.5)

Saturated Sediment Properties

After having determined the dry properties of the solid phase, the saturated rock properties can be calculated at seismic frequencies using Gassman’s equations. These equations relate the effective moduli of a dry rock with those containing fluid. The saturated bulk and shear moduli \( K_{\text{sat}} \) and \( G_{\text{sat}} \) are given by

\[
K_{\text{sat}} = K \frac{\phi K_{\text{dry}} - (1 + \phi) K_f K_{\text{dry}}/K + K_f}{(1 - \phi) K_f + \phi K - K_f K_{\text{dry}}/K},
\]
\[
G_{\text{sat}} = G_{\text{dry}},
\] (4.6)

where \( K \) is the bulk modulus of the mineral making up the rock, \( K_{\text{dry}} \) and \( G_{\text{dry}} \) are the dry bulk and shear moduli of the rock, and \( K_f \) is the bulk modulus of the saturating fluid. In the case of purely brine-saturated sediments, \( K_f \) is identical to the bulk modulus of water. If the sediment is homogeneously saturated with free gas,
K_f becomes an average of the brine and gas fluid moduli:

\[ K_f = \left( \frac{S_w}{K_w} + \frac{(1-S_w)}{K_g} \right) \]  \hspace{1cm} (4.7)

where \( K_w \) and \( K_g \) are the bulk moduli of water and gas, and \( S_w \) is the water saturation.

The elastic velocities \( v_p \) and \( v_s \) and the bulk density \( \rho_B \) can then be determined with the following equations:

\[
\begin{align*}
\rho_B &= (1-\phi) \rho_s + \phi \rho_f, \\
v_p &= \sqrt{\left(\frac{K_{sat} + \frac{4}{3} G_{sat}}{\rho_B}\right)}/\rho_B, \\
v_s &= \sqrt{\frac{G_{sat}}{\rho_B}};
\end{align*}
\]  \hspace{1cm} (4.8)

where \( \rho_s \) is the bulk density of the solid phase and \( \rho_f \) the density of the pore fluid. Both the solid density and the fluid density can be obtained as an arithmetic mean of the volumetric fractions of their components.

4.2.2 Sediments with Gas Hydrates

I evaluate three different possibilities of hydrate deposition in the pore space. All three of them can be seen in Figure 5.1.

Hydrate is Part of the Fluid

In this first model (Figure 5.1A), the hydrate is treated as part of the pore fluid, thus having no effect on the sediment properties. This is equivalent to a fluid substitution in purely brine-saturated sediments: instead of being 100% brine-saturated, the pore fluid consists of a mixture of brine and hydrate. The bulk modulus of the fluid can therefore be calculated as an average of the water and hydrate moduli:
where $K_h$ is the bulk modulus of the hydrate, $K_w$ the bulk modulus of water and $S_w$ the water saturation. The required dry and saturated rock properties are then calculated using equations 4.1 through 4.8.

**Hydrate becomes Part of the Solid**

In this case, I assume that hydrate becomes part of the solid frame. This has two effects: porosity reduction and change of the solid bulk and shear moduli, thus slightly stiffening the sediment frame. The reduced porosity $\phi_r$ can be calculated from the actual sediment porosity without gas hydrates:

$$\phi_r = \phi (1 - S_h),$$

where $S_h$ is the hydrate saturation of the pore space. The bulk and shear moduli of the solid phase are now a mixture of the sediment solid and the hydrate and can be calculated from the Hill average:

$$K = \frac{1}{2} (f_h K_h + (1 - f_h) K_s + [f_h/K_h + (1 - f_h)/K_s]^{-1}),$$

$$G = \frac{1}{2} (f_h G_h + (1 - f_h) G_s + [f_h/G_h + (1 - f_h)/G_s]^{-1});$$

where $K_s$ and $G_s$ are the solid bulk and shear moduli of the sediment without hydrate (as calculated from equation 4.3) and $f_h$ is the volume fraction of hydrate in the solid phase. It can be calculated as follows:

$$f_h = \frac{\phi S_h}{1 - \phi (1 - S_h)}$$

The dry and saturated moduli can then be determined using equations 4.1, 4.4, 4.5, 4.6 and 4.8.
Hydrate cements the Grain Contacts

In this model I assume that hydrate is formed at grain contacts and thus strongly reinforces the sediment. As model 2, it yields a porosity reduction which is equivalent to the one in equation 4.10. Strictly speaking, the cementation theory is only applicable for porosities less than 40%. It can, however, be used to approximate the elastic properties of granular aggregate at higher porosities. The effective bulk and shear moduli of the dry rock frame cemented by gas hydrate can then be calculated using the cementation theory by Dvorkin and Nur (1993):

\[ K_{\text{dry}} = \frac{1}{6} n (1 - \phi) (K_n + \frac{4}{3} G_h) S_n, \]
\[ G_{\text{dry}} = \frac{3}{5} K_{\text{dry}} + \frac{3}{20} n (1 - \phi) G_h S_r; \]  

(4.13)

where \( S_n \) and \( S_r \) are proportional to the normal and shear stresses of a cemented two-grain combination and depend on the amount of hydrate at grain contacts, and on the sediment and grain moduli. The exact formulas for \( S_n \) and \( S_r \) are given in Appendix A. The saturated rock properties are again calculated using Gassman’s equations (4.6).

4.2.3 Sediment Mineralogy and Calculation Parameters

The proposed models for brine, gas and hydrate-saturated sediments require knowledge of the mineralogy of the sediments. Mineralogy information can be obtained either from geological interpretations of an area or from well-logs. The seismic data from the Blake Outer Ridge that are the basis of this analysis do not have direct well control. However, well information is available from several kilometers distance (see Figure 2.1). Since the drilling has shown that the sediments at the Blake Outer Ridge are very uniform and lithology does not change significantly (Matsumoto et al., 1996; Holbrook et al., 1996), I feel confident about relating mineralogy information obtainable from the well-logs to the seismic data at hand. Because of the absence of
quantitative core values of the exact sediment mineralogy, the actual mineralogy was inferred by fitting the well-velocities, excluding areas of hydrate and gas saturation (Dvorkin et al., 1997).

The so-inferred mineralogy model used in this study consists of 60% clay, 35% calcite and 5% quartz. These three mineralogy components and the high amount of clay is consistent with the drilling description of the sediments. The elastic properties of this mineralogy and the water, gas and pure hydrate properties needed to calculate the saturated rock properties are summarized in Table 4.1.

<table>
<thead>
<tr>
<th>Substance</th>
<th>Bulk Modulus [GPa]</th>
<th>Shear Modulus [GPa]</th>
<th>Density [g/cm$^3$]</th>
</tr>
</thead>
<tbody>
<tr>
<td>Calcite</td>
<td>76.8</td>
<td>32</td>
<td>2.71</td>
</tr>
<tr>
<td>Clay</td>
<td>20.9</td>
<td>6.85</td>
<td>2.58</td>
</tr>
<tr>
<td>Quartz</td>
<td>36</td>
<td>45</td>
<td>2.65</td>
</tr>
<tr>
<td>Water</td>
<td>2.5</td>
<td>0</td>
<td>1.032</td>
</tr>
<tr>
<td>Pure Hydrate</td>
<td>5.6</td>
<td>2.4</td>
<td>0.9</td>
</tr>
<tr>
<td>Gas</td>
<td>0.1</td>
<td>0</td>
<td>0.235</td>
</tr>
</tbody>
</table>

Table 4.1: Properties of sediment mineralogy, water, gas and pure hydrate.

### 4.3 Estimation of Hydrate and Gas Saturation

From seismic alone, I can detect the presence of methane hydrates by means of bottom simulating reflectors (BSR). A migrated section of the data from the Blake Outer Ridge (Figure 4.2) clearly shows a BSR, which is connected to the bottom of the hydrate stability field. Furthermore, I can use surface seismic to determine the elastic contrasts across the BSR and infer the cause of the BSR (see Chapter 3). However, in order to estimate the actual amount of hydrate and gas present in the sediments above and below this BSR, respectively, seismic information alone is not sufficient. I need to link the seismic with the developed rock-physics models.

The information directly available from seismic is the interval velocity. Figure 4.3 shows the interval velocity which was derived in Chapter 2. The velocity was converted into depth by using a simple vertical stretch from time to depth. This depth
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Figure 4.2: Stacked section after prestack migration.

Figure 4.3: Interval velocity section.
conversion is required since the subsequent calculation will require depth as an input. The interval velocity is overlain by a wiggle plot of the migrated section. Above the BSR, the velocity increases to about 1.9 km/s while it decreases to approximately 1.6-1.7 km/s beneath the BSR. As shown in Chapter 3, the BSR is mostly the consequence of hydrate-bearing sediments overlying gas-saturated sediments. The transition of gas-saturated sediment to brine-saturated sediment is visible as a flat reflector beneath the BSR. In the region between 0 and 25 km lateral distance, where no BSR is visible, the velocity is uniformly increasing with increasing depth.

4.3.1 Methodology Description

In order to link the seismic interval velocity with the proposed rock-physics models, we need to know either the porosity of the sediments or the actual brine, gas and hydrate saturations. Solving for both porosity and saturation simultaneously would pose a non-unique problem: a known interval velocity could be reproduced from any combination of porosity and saturation, however unphysical. Since there is no direct well-control on the seismic data from the Blake Outer Ridge, both porosity and saturation are unknown. Therefore, an initial assumption has to be made about either one of them. Since there is no real physical relationship that would govern the behavior of the porosities throughout the sediments, I base my first analytical step on fixing the saturation: I assume the entire sediment section to be 100% brine-saturated. Physically, this assumption will be true everywhere but for the regions of hydrate and gas saturation, where it will produce porosity anomalies. I then calculate what I call the baseline porosities by posing an "inverse" problem: using a saturation of 100% brine, I change the porosity until the rockphysics model can reproduce the seismic interval velocities. This results in anomalies where the sediment is not 100% brine-saturated. The porosity is underestimated in areas of hydrate saturation, while it is overestimated when free gas is present.

In order to relate these anomalies to the actual amount of hydrate and gas saturation, a measure of the anomaly size is needed. The anomalous baseline porosity
has to be related to a porosity without the effects of both hydrate and free gas. I offer two techniques of determining such a reference “normal” porosity:

- calculating the average porosity trend from the region without BSR
- calculating the porosity trend on a trace-by-trace basis by fitting polynomials to the porosities above the hydrate and below the gas (thus assuming that those parts of the sediment is fully brine-saturated)

Both of those methods have pitfalls. In the first, I will neglect any possible lateral variations of porosity based on actual lithology changes. Furthermore, any error in interval velocity which was directly mapped into the porosities will be propagated. Also, by calculating the average trend from the region without BSR, I assume that those sediments do not contain gas hydrates. This assumption, however, may or may not be true for the sediments at the Blake Outer Ridge, as the drilling has shown (Matsumoto et al., 1996). The trace-by-trace approach does include lateral lithology variations and does not propagate velocity errors. However, it suffers from the lack of reflectors below the water bottom towards the end of the seismic line (see Figure 4.2, close to 50 km lateral distance). Combining both methods, however, should enable me to minimize the errors in the hydrate and gas saturation estimates and to obtain reasonable upper and lower bounds of possible saturations.

After having obtained estimates of the size of the porosity anomalies caused by the presence of hydrate and gas, I can relate them to the actual hydrate and free gas saturation. This is done by using the original baseline porosities, obtained by assuming 100% brine saturation everywhere, as a starting porosity model. I then pose again an “inverse” problem: I decrease the assumed brine saturations and increase hydrate or gas saturation until I can minimize the porosity anomalies, calculated by either the averaging or the trace-by-trace technique, and until I can fit the seismic interval velocities. Negative anomalies are considered to be caused by the hydrate, thus the hydrate saturation is increased in these regions. In areas of positive anomalies, on the other hand, the gas saturation is increased.
4.3.2 Baseline Porosity

As described in the preceding section, I first calculate the baseline porosities by assuming 100% brine saturation of the sediments and fitting the observed interval velocities. This leads to anomalies in regions where the assumption fails.

![Baseline porosity curves](sat-base-curves-ann)

Three baseline porosity trends along the seismic line are shown in Figure 4.4. The three porosity curves represent the porosities at a lateral distance of 12.5 km, a part of the line where no BSR is visible; at a distance of 34 km, a part of the line where the BSR might be underlain by a thin gas zone; and at a distance of 45 km, where the hydrate is overlying a thick gas-saturated zone (Chapter 3). These baseline porosities indicate the slow appearance of anomalies in the porosity trends. In the region without BSR, the porosity is monotonously decreasing with depth. This represents the expected decrease in porosity with depth due to the increasing consolidation of the sediments caused by the increase in pressure. The appearance of hydrate in the sediments above the BSR is indicated by the negative porosity anomaly in the middle panel. The next panel shows the presence of hydrate as even more pronounced. Furthermore, it displays a strong positive porosity anomaly.
immediately beneath the one caused by hydrate.

In order to relate these anomalies to actual hydrate and gas saturations, I need to determine the magnitude of the anomalies. This requires knowledge of a normal "background" porosity without the effects of hydrate and gas saturation.

4.3.3 Average Normal Porosity

In this first approach, I assume that the seismic section without a BSR gives a good approximation to the normal background porosities without the effects of hydrate or gas saturation. Consequently, I use the baseline porosities, determined in this region by assuming 100% brine saturation, and average them into an averaged normal porosity trend.

The resulting normal porosity is overlain with the anomalous baseline porosities in Figure 4.5. The first panel shows that there are misfits between the normal and baseline porosities in regions assumed to be hydrate and gas free. Such misfits can result from lateral lithology variations, as well as from some uncertainties in the velocities which are directly translated into errors in porosity. The middle panel shows that the anomaly which is related to the presence of hydrate slightly deviates from the average normal trend. It might be hard, however, to separate this hydrate anomaly from those solely related to curve misfits (see left panel). The right panel displays a clear negative hydrate anomaly and a strong positive gas anomaly. Since the anomaly caused by the presence of gas is considerably larger than the one caused by the hydrate, it will result in a higher detectability of the areas of gas saturation.

The size of the anomalies, or deviations between the normal porosity and the anomalous baseline porosity, can be obtained by subtracting the two porosity trends. The resulting trace residuals are displayed in Figure 4.6. The gas-saturated section underneath the BSR yields a very strong positive residual, as expected from Figure 4.5. The residuals attributed to the presence of hydrate, however, appear to be of the same magnitude as some of the residuals in the region without BSR. These residuals are caused by local porosity variations that can represent lateral variations of lithology.
or variations introduced by velocity errors. A clear separation of some of the hydrate-related residuals from those caused by these other effects might be difficult, especially for the hydrate anomaly between 32 km and 42 km lateral distance.

In order to examine the actual size of the residuals in more detail, I show a histogram of the porosity differences (Figure 4.7). The solid line represents the residuals between a lateral distance of 0 and 27 km (region without BSR), while the dashed line represents the residuals in the region with BSR (27 km to 52 km). The comparison between both curves indicates the clear impact gas has on the residuals. It significantly increases the porosity anomalies and is therefore clearly identifiable. The negative residuals caused by the hydrate also appear to be separate from the residuals caused by local curve misfits. However, there is an overlap of the size of the small hydrate anomalies between 32 km and 42 km (as shown in the middle panel of Figure 4.5) and the anomalies caused by the curve misfits. In this investigation, I chose to take anomalies less than -0.05 to be due to hydrate and anomalies larger than 0.08 to be gas-related. If I had taken a higher uncertainty for the hydrate anomalies, i.e.,
Figure 4.6: Residuals obtained by subtracting the average normal porosity from the anomalous baseline porosities.
Figure 4.7: Histogram of the porosity residuals obtained using the average porosity trend. The solid line represents the residuals for the line between 0 and 27 km (without BSR). The dashed line represents the residuals for the line between 27 and 52 km (with hydrate and gas). [sat-histo1] [CR]
values higher than -0.05, I would not be able to resolve the hydrate-related anomalies between 32 km and 42 km lateral distance.

4.3.4 Trace-By-Trace Normal Porosity

In this second approach, I obtain the normal porosities on a trace-by-trace basis. The normal trend is determined locally at each surface position by fitting polynomials to the near-surface and deep-sediment baseline porosities which are assumed to be free of gas and hydrate. The fitting is done using an iterative least-squares approach and second order polynomials.

Figure 4.8 shows the anomalous baseline porosities overlain by the polynomial-fitted normal porosity curves. As before, the first panel represents the porosities in an area without a prominent BSR. Both porosity lines are in good agreement. Since the second order polynomial can adequately reproduce the baseline porosity profile, this polynomial is an appropriate functional form of the porosity-depth profile of the presumably brine-saturated sediments. The middle section shows the beginning of a small hydrate anomaly and a clearly identifiable hydrate anomaly is visible in the third panel, as well as a considerable gas-related anomaly of opposite polarity.

After having determined the reference normal porosity trend at each surface position, I again calculate the actual magnitude of the gas and hydrate anomalies by subtracting normal and anomalous baseline porosities. The resulting residuals are shown in Figure 4.9. Compared to the averaging technique, both the hydrate and the gas-related anomalies are clearly enhanced with respect to small local anomalies caused by curve misfits.

In order to examine the residuals caused by the hydrate and the gas in more detail, I again plot a histogram of the porosity differences (Figure 4.10). The residual histogram shows the residuals for the part of the line without BSR (sold line) compared to the one with BSR (dashed line). The hydrate and gas anomalies are clearly identifiable and separable from other misfits. The relatively high positive residuals in the region without BSR are generated close to the seafloor by fitting a stiff polynomial
4.3.5 Hydrate and Gas Saturations

After having determined the magnitude of the anomalies caused by the presence of hydrate and gas in the pore space, I can calculate the actual hydrate and gas saturation. This is done by again assuming 100% brine saturation everywhere in the sediment. The assumed brine saturation is decreased and the hydrate or gas saturation increased until I can minimize the porosity anomalies, as calculated by either the averaging or the trace-by-trace technique, within their described uncertainty bands and until I can fit the seismic interval velocities. The resulting estimates in hydrate and gas saturations using the residuals of the averaging method are shown in Figure 4.11, while those obtained using the residuals of the trace-by-trace approach are displayed in Figure 4.12.
Figure 4.9: Residual obtained by subtracting the trace-by-trace normal porosities from the anomalous baseline porosities.
Figure 4.10: Histogram of the porosity residuals obtained using the trace-by-trace porosity trend. The solid line represents the residuals between 0 and 25 km (without BSR). The dashed line represents the residuals between 25 and 52 km.
The averaging method shows maximum hydrate saturation above the BSR between 45 km and 52 km lateral distance. Hydrate model A, in which hydrate is part of the pore fluid, results in a maximum hydrate saturation of approximately 26%, while model B (hydrate becomes part of the solid) yields a more conservative estimate of about 20%. Since this model slightly stiffens the sediment frame, less hydrate is required in order to increase the velocity in the hydrate-saturated sediments. In the case of hydrate cementing the sediment grains, less than 1% hydrate saturation is required to satisfy the seismically-obtained velocities. Since cementation has a drastic effect on the stiffness of the sediment, only very little hydrate can considerably increase the velocity. Small hydrate concentrations of approximately 7% are obtained in the region between 32 km and 42 km lateral distance. Some anomalies caused by trace misfits which are of the same magnitude as the hydrate-related anomalies between 32 km and 42 km, appear as hydrate anomalies as well (between 0 km and 5 km, and between 25 km and 27 km). These “anomalous” zones, but also the hydrate zone between 32 and 45 km, would have been suppressed had I chosen to allow a higher uncertainty for the hydrate residuals (as described in section 4.3.2). The gas saturation beneath the BSR is approximately 1% to 2% throughout the low velocity zone.

The trace-by-trace method, on the other hand, results in clear hydrate saturations both between 32 km and 42 km, and 45 km and 52 km lateral distance. Model A gives a maximum hydrate saturation of approximately 20%, model B of about 15%, and model C again of less than 1%. In the region between 32 km and 42 km, the hydrate saturation is about 10%. Additionally, the gas saturation is estimated to be about 1% to 2% and is thus stable in both the averaging method and the trace-by-trace method.

A comparison of the average and the trace-by-trace method shows that the latter yields hydrate estimates that can be up to 5% more conservative. Both methods result in maximum lateral hydrate saturation in the area between 45 and 52 km. The trace-by-trace method results in clearly distinguishable hydrate anomalies between 32 km and 42 km. Combining both method appears to be a good way to estimate a
Figure 4.11: Hydrate and gas saturation obtained using the average normal porosity trend.
Figure 4.12: Hydrate and gas saturation obtained using the trace-by-trace normal porosity trend.
reasonable range of possible hydrate saturations and give upper and lower uncertainty bounds on the estimates.

4.4 Uncertainties introduced by Interval Velocities

In this section, I discuss possible uncertainties introduced into the hydrate and gas estimations by errors in the used interval velocities. Based on the estimated errors in interval velocity (see Chapter 2), I recalculate the hydrate and gas saturations. This yields an insight into the robustness of the hydrate and gas estimates.

4.4.1 Error in Interval Velocity

The evaluation of velocity uncertainties in Chapter 2 has shown that ± 10 m/s picking errors in the RMS velocity can cause as much as ± 200 m/s errors in the interval velocities. Since I have discussed the issue in more detail in Chapter 2, I simply show these resulting errors in interval velocity again in Figure 4.13. The solid line represents the velocity determined from the original RMS velocity. It shows clearly the velocity increase connected with the presence of hydrate in the sediment and the subsequent velocity decrease because of the gas saturation. Picking errors give rise to the velocities represented by the dashed lines. The errors can either enhance the anomalous velocity zones in the hydrate and gas layer (dashed line) or suppress them (double dashed line). It can be imagined that these velocity trends will produce different estimates of the hydrate and gas anomalies.

4.4.2 Saturation Uncertainties

Using the above described interval velocities obtained after introducing a reasonable picking error into the RMS velocities, I repeat the method of estimating hydrate and gas saturations. This will yield the robustness of the hydrate and gas saturations to errors in velocity.
Figure 4.13: Errors in interval velocity resulting from picking errors in the RMS stacking velocities. The solid line is the original velocity. [sat-vint-error] [ER]

The previous analysis (section 4.3) showed that in the case of hydrate cementing the grains, only very little (≤ 1%) hydrate saturation is required to fit the seismic data. I do not anticipate that errors in velocity will considerably change these estimates based on the drastic effects only small amounts of hydrate have on the sediment stiffness. Therefore, I only investigate the effect of velocity errors on the saturations for hydrate models A and B.

As described in section 4.3, I assume the sediment to be 100% brine-saturated and use the interval velocities to calculate the baseline porosities. These baseline porosities are then compared with the normal reference porosity in this region, i.e., a porosity which is not affected by the presence of hydrate and gas in the pore space. I chose to compare the baseline porosities with the calculated average porosity in the region (calculated in section 4.3). I also could have compared the baseline porosities with the normal porosity calculated by the trace-by-trace method. However, the sensitivity of both the average and the trace-by-trace method to the velocity errors is similar. Therefore, I only show the results using the averaging method.
The three baseline porosities obtained from the three different interval velocities (see Figure 4.13) are shown in Figure 4.14. They are overlain with the average porosity trend that was calculated in section 4.3. The left panel shows the baseline porosity calculated from the original interval velocity. The middle panel represents the porosity in the case of enhanced hydrate and gas velocity zones. It shows an increased hydrate and gas anomaly. The right panel represents the porosities obtained from the velocity, which increased steadily. The anomaly caused by the gas disappears completely, while there is still a hydrate-related anomaly in the very near-surface sediments. However, additional velocity errors in this region could make this anomaly disappear as well.

Figure 4.14: Baseline porosities for the three velocities shown in Figure 4.13, overlain by the average average porosity calculated in section 4.3. The left panel represents the porosity trend obtained from the original velocity. The middle panel represents the porosity determined from the velocity which enhances both the hydrate and gas velocity zones; and the right panel shows the porosity calculated from the velocity which suppresses the velocity increase due to hydrate or decrease due to gas.

From these new hydrate and gas porosity anomalies, I calculated again the hydrate and gas saturation. The results, using hydrate model A and hydrate model B, can be seen in Figure 4.15. The left panel shows the saturations obtained for hydrate
model A, in which hydrate is solely part of the fluid. The left panel represents the saturations obtained for hydrate model B, in which hydrate is assumed to become part of the solid sediment frame. The solid curve represents the saturations obtained from the original velocity; the dashed line the saturations obtained from the velocity trend that increases the hydrate and gas velocity anomalies; and the double dashed line the saturations for the velocity trend that suppresses both gas and hydrate anomalies. The saturations show that there is an approximately a 14% discrepancy between the original saturation and the one using enhanced velocities. The saturations resulting from the suppressed velocity trend (double dashed line) still display hydrate saturation in the near-seafloor sediments. Additional errors in velocity in that region could, however, cause this saturation to drop all the way to zero. Therefore, it appears to be reasonable to assume that a ± 14% uncertainty can theoretically be introduced into the hydrate saturations by errors in the velocity. The gas saturation displays an error of about ± 2% (note that % does not refer to percentage of saturation but to saturation itself.)

This uncertainty assessment has shown that the hydrate saturation is sensitive to errors in interval velocity. The saturation values can theoretically vary up to ± 14%. A Comparison with VSP velocities (see Chapter 2, section 2.3.2) and with other seismic velocity measurements in the region (Wood et al., 1994; Katzman et al., 1994; Korenaga et al., 1997) suggests, however, that my initially obtained interval velocities represent the overall velocity trends in the region of the Blake Outer Ridge fairly well. Thus, an error of ± 14% in saturation represents an upper uncertainty bound, but the actual errors might be much smaller.

4.5 Well Log Comparison

In order to validate the technique of estimating hydrate and gas saturations using only interval velocities, I apply it to the known well-log velocities from wells 994 and 995. Besides known velocities, the wells also provide known porosity functions from cores which represent the normal porosities without the effect of hydrate or gas. Therefore,
Figure 4.15: Saturation estimates for hydrate model A (left panel) and hydrate model B (right panel) resulting from errors in the interval velocity. The solid line represents the saturations based on the original velocity. The dashed line represents the saturations calculated from the velocity enhancing both hydrate and gas velocity anomalies; and the double dashed line represents the saturations for the velocity trend suppressing hydrate and gas anomalies. [sat-sat-error] [CR]
the application of the technique without known porosities can be compared to the equivalent analysis using the known core porosities.

The smoothed well velocities are shown in Figure 4.16. They are used together with the rock-physics models to invert for a baseline porosity trend by assuming 100% brine-saturation throughout the sediments. Figure 4.17 shows the baseline trend overlain by the fitted porosity trend and the true (core) porosity trend. The comparison shows that for well 994, the true normal porosity (dotted line) and the inferred normal porosity (polynomial fitting) are in reasonable agreement. At well 995, there is a discrepancy between the true normal porosities and the one obtained by fitting a polynomial to the baseline porosity.

**Smoothed Well Velocities**

![Smoothed Well Velocities](image)

Figure 4.16: Smoothed well-log velocities from well 994 (solid line) and well 995 (dashed line).

Using the porosities displayed in Figure 4.17, I calculate an estimate of the hydrate anomaly by (1) subtracting the true porosity from the baseline porosity and (2) subtracting the fitted porosity from the baseline porosity. The resulting anomalies are then used to calculate hydrate saturation.
Figure 4.17: Porosities for well 994 (left panel) and well 995 (right panel). The real core porosities are given by the dotted lines, the baseline porosities by the solid lines, and the fitted normal porosities by the dashed lines.
The resulting saturations for hydrate models A and B can be seen in Figure 4.18. The saturations obtained using the true porosity are represented by the solid lines, while the ones determined using purely the velocities are given by the dashed lines. Since the cementation model (model C) would again result in both cases in having less than 1% hydrate saturation, I did not investigate it here. The comparison of the results indicates that there is a good fit between the saturations in well 994, both for hydrate model A and hydrate model B. The maximum hydrate saturation is at about 30% for hydrate model A and 21% for hydrate model B. The range of these estimates is consistent with the results I obtained from the seismic data. As expected from the porosity misfits at well 995, the resulting saturations agree less than those in well 994. The discrepancy is less than 10% above a depth of 3.3 km. The more significant discrepancy below 3.3 km can be attributed to the lack of velocity data in the lower portion of the well which affects the accuracy of the polynomial fitting. It has to be noted, however, that the overall magnitude of the hydrate saturation is again consistent with the results obtained from seismic. Therefore, the methodology offered to estimate hydrate saturation from only velocity appears to be quantitatively accurate.

4.6 Conclusions

In this Chapter, I provided a theoretical tool for estimating the amount of hydrate and gas from seismic interval velocity. I developed rock-physics models that link the elastic wave velocity in high-porosity marine sediment to density, porosity, effective pressure, mineralogy, and water, gas and hydrate saturation. Three micromechanical models of hydrate deposition in the pore space were examined: (A) hydrate is part of the pore fluid, (B) hydrate becomes part of the solid frame, and (C) hydrate cements grain contacts. Using the interval velocities obtained from the stacking velocity analysis described in Chapter 2, I calculated lateral maps of hydrate and gas saturation. The calculations were based both on an averaging approach and a trace-by-trace approach, resulting in an upper and lower bound for the possible hydrate saturation.
Figure 4.18: Hydrate Saturations for well 994 (upper figures) and well 995 (lower figures). The solid line represents the saturations obtained using the real porosities. The dashed line represents the saturation obtained using the technique of inverting for porosity. [sat-wellsat] [ER]
I find considerably large lateral variations in hydrate saturation. Model A results in a maximum hydrate saturation between 21% and 26% and model B between 15% and 20%. In case of hydrate cementing the sediments, only 1% of hydrate is required in the pore space to increase the velocity corresponding to that observed in the seismic data. The estimated gas saturation is approximately 1%-2%. Subsequently, I evaluated the robustness of these hydrate and gas saturation estimates with respect to the velocity errors that were determined in Chapter 2. The analysis suggests that the maximum errors introduced into the hydrate saturation estimates can be up to ± 14%, while the gas saturation shows maximum uncertainties of ± 2%. The uncertainty estimates are upper bounds on possible errors. The comparison with VSP data (see Chapter 2, section 2.3.2) and other seismic velocity investigations (Wood et al., 1994; Katzman et al., 1994; Korenaga et al., 1997) suggests that the error in interval velocity is smaller than those upper bounds which would shrink the uncertainties on the hydrate saturation. Using additional velocity and porosity information from well-logs 994 and 995, I evaluated the validity of the proposed technique and the used models. This investigation suggests that the technique is quantitatively accurate.

4.7 Acknowledgements

This study was done in collaboration with Jack Dvorkin and Amos Nur of the Stanford Rock-Physics Group, and I would like to thank them for the many discussions on the topic. I would also like to thank Gary Mavko for helpful suggestions, especially on the relevance of uncertainty assessment. Furthermore, I would like to thank Sergey Fomel for the use of his inversion library.
Chapter 5

Internal Structure from AVO

5.1 Overview

In chapter 4, I linked seismic interval velocities with physical rock models in order to estimate the hydrate and gas saturations across the BSR. The use of acoustic velocity information alone enabled me to estimate hydrate saturations for different micromechanical models of hydrate deposition in the pores. However, it cannot be used by itself to differentiate between the models and, ultimately, the obtained differences in saturation. Consequently, I use additional amplitude information, in particular AVO analysis, in an attempt to differentiate between the proposed rock-physics models of internal hydrate deposition. This analysis is performed by simple 1-D modeling. Based on given saturations, which were calculated in chapter 4, and assuming a porosity of 45%, I calculate elastic velocities and densities. Using Kirchhoff modeling, I generate synthetic seismograms and investigate the BSR AVO response obtained from the different models. This investigation shows that seismic amplitudes cannot differentiate between hydrate being part of the fluid (model A) and hydrate becoming part of the frame (model B), but give a distinctly different AVO trend in case of hydrate cementing the grains (model C). The actual seismic amplitudes can be qualitatively reproduced by either model A or B. These models cause the sediment structure at the Blake Outer Ridge to be weak and might cause the hydrate to clog large pore
space conduits, since both models A and B are less permeable than model C. This could explain why free gas is trapped underneath the BSR: the hydrate presents an impermeable barrier.

5.2 Rock-Physics Models

In this section, I briefly review the examined mechanisms of hydrate deposition. Following this, I explore the elastic velocities and densities that are connected with those different models. From those properties, it is possible to predict the BSR AVO behavior.

5.2.1 Micromodels of Hydrate Deposition

I evaluated three different models of possible hydrate deposition in the pore space. In the first model (model A), I assume that hydrate is suspended in the water and thus only contributes to the bulk modulus of the pore fluid (Figure 5.1A). In this case, the formation and deposition of the hydrate in the sediment would have no effect on the stiffness of the sediment and could be treated as pure fluid substitution. In the second model, hydrate becomes part of the solid sediment frame (Figure 5.1B). This causes a reduction in porosity and a weak stiffening of the sediment structure. The third model assumes that hydrate cements grain contacts (Figure 5.1C), thus significantly changing the rock stiffness and simultaneously reducing the porosity. Even though this last model is not likely to apply to the sediments at the Blake Outer Ridge due to their high porosity, I include it into this study to show the different effects of these models not only on saturation but also on the elastic properties and seismic amplitudes. Even though, strictly speaking the cementation theory applies only for porosities of less than 40%, it can be used to approximately estimate the elastic properties of granular aggregate of higher porosities.
5.2.2 Seismic Velocities

Using the different schemes of hydrate deposition, it is possible to calculate the effect these models will have on the actual velocities and thus on velocity contrasts from hydrate-bearing sediments to the underlying sediments. The theoretical calculations are based on the rock-physics theory discussed in detail in Chapter 4. The calculation parameters, i.e., the sediment mineralogy, is, as before, 60% clay, 35% calcite and 5% quartz. The actual calculation parameters are given in Table 4.1.

I determine the effect of increasing hydrate saturation on the sediment elastic properties, i.e., P-wave velocity and S-wave velocity. The resulting velocities for hydrate deposition models A, B, and C can be seen in Figure 5.2. They are determined by assuming a porosity of 45% and a BSR depth of approximately 500 m, which is consistent with the seismic data. The left side of Figure 5.2 represents the saturated sediment velocities for model A, the middle shows the velocities for hydrate model B, and the left shows those of model C. Maximum hydrate saturation is about 50%. It is obvious, that model A and B affect the P-wave and S-wave velocities similarly at small hydrate saturations. In both models, the presence of hydrate increases the P-wave velocity of the sediment slowly with increasing hydrate saturation. Since model B affects the stiffness of the sediment frame, its effect on the P-wave velocity...
becomes stronger than that of model A. Whereas the P-wave velocity increases, the S-wave velocity slightly decreases with hydrate saturation for model A. It is practically independent of hydrate saturation. While the S-wave velocity produced by model B also does not change much for small hydrate saturation, it is clearly increasing with increasing saturation. By becoming part of the solid, the hydrate not only affects the P-wave velocity but also, slightly, the S-wave velocity. In the cementation model, the hydrate acts to considerably increase both P- and S-wave velocities. This effect is well-pronounced already at very small hydrate saturations.

![Figure 5.2: P-wave (solid line) and S-wave velocities (dashed line) for hydrate models A, B, and C versus hydrate saturation.](image)

The increase in P-wave velocity in all three models is in good agreement with the existing seismic velocity analysis performed in Chapter 2. This impact of hydrates on the acoustic velocity was used in Chapter 4 to invert for the amount of hydrate present in the pore space. While a sufficient increase in velocity can be obtained due to only very small amount in the cementation model, both model A and model B require larger amounts of hydrates to obtain a similar P-wave velocity increase.
Since the P-wave velocity cannot be used to discriminate uniquely between the different hydrate deposition models, the S-wave velocity becomes important. The different effects of the models on the S-wave velocity might be detectable in AVO analysis and thus allow a discrimination of the different models and their resulting hydrate saturation estimates.

5.3 Synthetic Modeling

In this section, I discuss the 1-D modeling performed with the different hydrate models. I first describe the creation of the model and the actual modeling technique. Subsequently, I evaluate the synthetic seismic gathers and compare them with the real reflectivity gather.

5.3.1 1-D Model

I create simple 1-D models for the three schemes of hydrate deposition by assuming that (1) the hydrated sediment is overlain by brine-saturated sediment and underlain by gas-saturated sediment or (2) the hydrated sediment is both overlain and underlain by brine-saturated sediment. I explore those two possibilities because although there is clear evidence from the seismic data at the Blake Outer Ridge that there is free gas beneath the hydrate, the possibility of lateral brine patches underneath cannot completely be excluded. Therefore, including both scenarios in the forward modeling will yield the different amplitude responses connected to either brine or gas saturation. The elastic sediment properties are calculated based on the actual hydrate and gas saturations estimated in Chapter 4. I chose one surface position at about 49 km to represent the model base, and use its hydrate and gas saturations to calculate the elastic, saturated sediment properties using the rock-physics scheme described in detail in Chapter 4. In this way, I base the modeling on the actual different saturation estimates connected with the different models. All sediments are assumed to have a porosity of about 45%, which approximates the porosity in the hydrate zone at the
surface position under consideration. The thicknesses of the layers are adjusted to match the zero-offset travel times of the real reflectivity gather at the surface position of 49 km. In this way, a direct comparison between the synthetics and the real gather is possible. The resulting model is summarized in Table 5.1.

<table>
<thead>
<tr>
<th>Layer</th>
<th>Thickness [km]</th>
<th>Saturation (Model 1)</th>
<th>(Model 2)</th>
<th>(Model 3)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Water</td>
<td>3.315</td>
<td>100% brine</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Brine sediment</td>
<td>0.15</td>
<td>100% brine</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Hydrate sediment</td>
<td>0.35</td>
<td>25%</td>
<td>20%</td>
<td>0.4%</td>
</tr>
<tr>
<td>Gas sediment</td>
<td>0.3</td>
<td>1% gas or 0% gas</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Brine sediment</td>
<td>0.5</td>
<td>100% brine</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Table 5.1: Model used to calculate synthetic seismograms.

The properties listed in Table 5.1 are used to calculate the actual saturated P-wave, S-wave and density of the different models of hydrate depositions. The results for the case of hydrate overlying gas-saturated sediments can be seen in Figure 5.3. The properties of hydrate model A are represented by the solid line, those of model B by the dashed line and those of model C by the dotted line. The calculated saturated sediment properties display the expected increase in P-wave velocity in the hydrate-bearing zone and the decrease in P-wave velocity due to the saturation of free gas underneath. The effect on density is very small for all three hydrate models. The most pronounced difference is, as hoped, visible in the S-wave velocities of the models. While hydrate as part of the pore fluid does not affect the S-wave velocity at the estimated hydrate saturation, hydrate cementing the frame does significantly increase the S-wave velocity in the hydrate sediments with respect to the sediments underneath. This already indicates that this model is not what might apply to the Blake Outer Ridge data since the AVO analysis discussed in Chapter 3 did not provide evidence of such a pronounced increase in shear velocity in the hydrate zone. Hydrate being part of the fluid and hydrate being part of the solid appear to be seismically indistinguishable from these elastic properties. When the hydrate becomes part of the frame, it slightly increases the shear wave velocity in the hydrated zone. This increase, however, is so minimal that it will not be resolvable from seismic.

The P-wave velocity, S-wave velocity and density properties for hydrate models
Figure 5.3: S-wave, P-wave, and density for models A, B, and C in the case of hydrate-bearing sediment overlying gas-saturated sediment. [model-models] [ER]

A, B and C in the case of hydrate being underlain by brine-saturated sediments can be seen in Figure 5.4. The figure shows that in this case there is a less pronounced P-wave velocity contrast at the transition from hydrate to the sediments underneath. Both S-wave velocity and density do not show a strong, visible change in regard to the previous gas-saturated sediments.

5.3.2 Real Reflectivity Gather

The real reflectivity gather at the subsurface position of 49 km is shown in Figure 5.5. The BSR reflection and the one from the bottom of the gas underneath are strongly visible. The BSR displays the characteristic behavior of increasing amplitudes with increasing offset.

The actual amplitude picks along the BSR as a function of offset are shown in Figure 5.6. Anomalous amplitudes between offsets 2.25 and 2.6 km represent some residual effects present after amplitude calibration (see Chapter 2, section 2.4.3). This
Figure 5.4: S-wave, P-wave, and density for models A, B and C in the case of hydrate-bearing sediment overlying brine-saturated sediment.

displayed AVO behavior was related to the presence of free gas-saturated sediment underneath the hydrate (Chapter 3). I furthermore concluded from the AVO analysis on the real data that the hydrate does not have a pronounced effect on the shear properties of the in-situ sediment. This already indicates that hydrate cementing the sediment cannot explain the seismic data from the Blake Outer Ridge.

By calculating synthetic seismograms based on the saturations inferred from seismic at the distance of 49 km and by subsequently comparing the synthetic gathers with the one in Figure 5.5, I should be able to get an better insight into the internal structure of the hydrated sediment.

5.3.3 Synthetic Gathers

I calculate the synthetic seismograms using the elastic properties displayed in Figure 5.3. The forward modeling is done using a generalized Kirchhoff body force scattering
Figure 5.5: Real reflectivity gather. [model-realcmp] [ER]

Figure 5.6: BSR AVO picks. [model-realavo] [ER]
theory by Lumley and Beydoun (1993). This technique combines Zoeppritz plane-wave reflection and Rayleigh-Sommerfeld elastic diffraction responses and generates correct AVO responses for locally planar reflectors. The method computes only primary reflections, which is, however, sufficient for this case. Because of the thick water layer and limited offset range, the real data do not show multiple contamination or converted shear waves.

The resulting synthetic gathers for hydrate models A, B and C in the case of hydrate being underlain by free gas are shown in Figure 5.7. All three gathers display strong, negative BSR reflections and reflections from the bottom of the gas zone underneath. It is obvious that both models A and B result in increasingly negative BSR amplitudes with increasing offset. The cementation model, on the other hand, yields decreasing BSR amplitudes with increasing offset, as expected from the way the hydrate was affecting the elastic sediment properties. All the models show clear reflections off the top of the hydrate and the bottom of the gas.

Figure 5.8 shows the synthetic seismograms for the case of hydrate-bearing sediments overlying brine-saturated sediments. As before, all three schemes of hydrate deposition in the pore space produce clear reflections from the top and bottom of the hydrate zone. However, the amplitudes appear significantly weaker as was the case with underlying gas sediments. The cementation model (model C) again shows decreasing amplitudes with offset, while both models A and B seem to produce more constant amplitudes with offset.

A comparison of these synthetic gathers with the real reflectivity gather (Figure 5.6) makes it obvious that the amplitudes produced by model C (the cementation model) do not match the general amplitude trend displayed by the data. This suggests that the hydrate is most likely to be deposited either as part of the fluid or as part of the solid in the pore space. It also suggests, that, at least at the surface location of this CMP gather, the amplitudes generated by hydrate overlying brine appear too weak to reproduce the data. An explicit look at the AVO behavior at the BSR should further support these indications. Moreover, the seismic data do not show a strong or clearly identifiable response from the top of the hydrate. Reflections at
Figure 5.7: Synthetic gathers for hydrate models A, B, and C in the case of hydrate overlying gas-saturated sediment. [model-model-ann] [ER]
Figure 5.8: Synthetic gathers for hydrate models A, B, and C in the case of hydrate overlying brine-saturated sediment.
about 4.5 s two-way traveltime could be attributed to the top of hydrate, but proof for such cannot be given from seismic. Furthermore, the top of hydrate has not been resolved in any seismic data at present. This would suggest that the top of hydrate is gradational.

5.3.4 BSR AVO Response

In order to emphasize the effect of the different models on the amplitudes at the BSR, I evaluate the AVO response directly at the BSR. This is done by picking the maximum (negative) BSR amplitudes.

The resulting AVO curves for the case of gas underlying hydrate sediments are shown in Figure 5.9. The AVO behavior on the upper panel represents those obtained from hydrate models A and B. The AVO behavior in the lower model represents the one obtained from hydrate model C. Neither model A (hydrate is part of the fluid) and model B (hydrate becomes part of the solid frame) can be distinguished based on their amplitudes. Both of them result in an increasingly negative amplitude with increasing offset. Model B does affect the shear wave velocity in the hydrated sediment differently. However, since this model requires less hydrate saturation to reproduce the acoustic interval velocities, the effect on the shear properties become less pronounced as when both model A and model B would require the same amount of hydrate saturation. The cementation model, on the other hand, displays a clearly opposite amplitude behavior. The amplitudes are decreasing with increasing offset.

The picked AVO curves for the case of hydrate underlain by brine-saturated sediment can be seen in Figure 5.10. As before, the upper panel represents the AVO behavior of hydrate models A and B, while the lower panels shows the AVO behavior of model C. Once again, models A and B cannot be differentiated from seismic. They show a slightly less increasing amplitude trend with increasing offset. The most pronounced difference from the gas-saturated case is the considerable decreased zero-offset reflection coefficient, which can be clearly attributed to the presence of brine underneath the hydrate.
Figure 5.9: Synthetic BSR AVO picks for the case of hydrate overlying gas. The upper panel shows the picks for hydrate models A and B. The lower panel shows the picks for hydrate model C.
Figure 5.10: Synthetic BSR AVO picks for the case of hydrate overlying brine sediments. The upper panel shows the picks for hydrate models A and B. The lower panel shows the picks for hydrate model C.
Comparison of these AVO trends with that of the real data (Figure 5.6) shows first of all that hydrate underlain by gas-saturated sediment could reproduce the zero-offset reflection coefficient reasonably well. Furthermore, the hydrate cementation model results in an AVO behavior which is clearly opposite to the one shown in the real data. Even though the magnitude of the amplitude increase displayed by the data cannot be quantitatively matched by either model A or B, the overall trend is in good agreement.

Based on this modeling approach, I can conclude that seismic cannot differentiate between hydrate as part of the pore fluid and hydrate becoming part of the solid. However, one of those two models or a combination of both appears to match the seismic data qualitatively. Thus, the effect of hydrate on the in-situ structure must be fairly small. The sediment frame is not significantly stiffened, as suggested by a cementation model, but appears to be only weakly, if at all, affected by the presence of hydrate.

It furthermore appears that cement positioned in the sediment may strongly affect permeability: at the same high porosity, rocks with contact cement may have higher permeability than those with pore-filling cement. This results from the numerical modeling of Cade et al. (Cade et al., 1994). The inferred sediment structure of the hydrated sediment at the Blake Outer Ridge requires hydrate to be either part of the fluid or the solid frame. Such hydrate deposition schemes can lower permeability considerably compared to the cementation model. Hydrate clogging large pore space conduits could therefore seal the BSR. This could explain why free gas is trapped beneath the BSR.

5.4 Conclusions

In this chapter, I have related the rock-physics models to the elastic sediment properties and, subsequently, AVO behavior. While using only the acoustic interval velocity enabled me to estimate hydrate and gas saturations for different hydrate deposition models, it could not uniquely differentiate between the models and, therefore, the
actual in-situ saturations. Using AVO analysis based on 1-D models calculated from the estimated hydrate and gas saturations, I generated synthetic seismograms. Those seismograms were compared with the in-situ reflectivity gather and its amplitude response along the BSR. The analysis suggests that seismic cannot differentiate between hydrate as part of the pore fluid and hydrate becoming part of the solid. Even though both models have a different effect on the elastic sediment properties, those differences are too subtle to cause differences in their amplitude behavior. AVO analysis can, however, clearly distinguish the cementation model from those two hydrate models. Based on its amplitude response, which is decreasing with increasing offset, I can conclude that this model does not represent the in-situ conditions at the Blake Outer Ridge. The amplitude behavior displayed by the real data can be matched qualitatively by either hydrate being part of the pore fluid or hydrate becoming part of the solid. The resulting sediment frame is relatively weak and hydrate might clog large pore space conduits. This inferred low permeability of the hydrated sediment explains why free gas is trapped underneath the BSR.

5.5 Acknowledgements

This study was done in collaboration with Jack Dvorkin and Amos Nur of the Stanford Rock-Physics Group. Many thanks to them for their helpful discussions and suggestions. I would also like to thank David Lumley for permission to use his Kirchhoff modeling code.
Chapter 6

Conclusions

In this dissertation, I characterized methane hydrate seismic data from the Blake Outer Ridge, offshore Florida and Georgia. I determined the cause of the BSR visible in the seismic data and provided a tool for estimating the amount of hydrate present in the pore space. This study is one of the first attempts to characterize hydrate structures by linking surface seismic and rock-physics.

The seismic data are characterized by a strong BSR with negative reflection polarity. Using stacking velocity analysis, I determined that the hydrate-bearing sediment above the BSR is characterized by an increase in P-wave velocity to approximately 1.9 km/s which decreases to about 1.6-1.7 km/s in the underlying sediment. By means of 2-D impedance inversion and AVO analysis after careful preprocessing of the data, I showed that the BSR is caused by hydrate-bearing sediment overlying gas-saturated sediment, which accounts for the observed drop in velocity. The seismic amplitudes can thus be explained by a decrease in P-wave velocity across the BSR and a simultaneous increase in S-wave velocity. Furthermore, the analysis suggests that a flat reflector underneath the BSR represents the transition of the gas-saturated sediment to brine-saturated sediment.

Using the obtained velocity information together with physical rock models, I presented a method to estimate the amount ofhydrate present above the BSR and
free gas beneath the BSR. I examined three different models of hydrate formation in the pore space: (A) hydrate is part of the pore fluid, (B) hydrate is part of the solid frame, and (C) hydrate cements grains together. Model A predicts maximum hydrate saturations between 20% and 26%, model B saturations between 15% and 20% and model C saturations of less than 1%. The saturation also vary significantly laterally along the BSR. The free gas saturation underneath the BSR is approximately 1% to 2%. Investigation of the robustness of these saturation estimates in the presence of velocity errors indicated that the saturation estimates can vary as much as 14% (note that % refers to the actual saturation itself and not to the percentage of saturation). Therefore, accurate velocity determination is crucial for the technique to give adequate results. Using additional velocity and porosity information from well-logs 994 and 995 at the Blake Outer Ridge, I evaluated the validity of the proposed technique for estimating saturations. This investigation suggests that the maximum uncertainties in the hydrate estimates are less than 10% which indicates that the technique is quantitatively accurate.

In order to differentiate between the different models of hydrate saturation in the pore space, I used forward 1-D modeling and AVO analysis. Models A and B can qualitatively reproduce the BSR AVO trend as observed in the seismic data, however, both models cannot be distinguished by means of their seismic responses. The inferred structure of the hydrated sediment suggests that the sediment is mechanically weak. The permeability is likely to be low from hydrate clogging large pore-space conduits, explaining why free gas is trapped underneath the BSR.
Appendix A

Calculation Parameters for the Cementation Theory

The Parameters $S_n$ and $S_T$ are calculated as follows:

$$S_n = A_n(A_n) \alpha^2 + B_n(A_n) \alpha + C_n(A_n), \quad A_n(A_n) = -0.024153 A_n^{-1.3646},$$
$$B_n(A_n) = 0.20405 A_n^{-0.89008}, \quad C_n(A_n) = 0.00024649 A_n^{-1.9864};$$

$$S_T = A_T(A_T, \nu) \alpha^2 + B_T(A_T, \nu) \alpha + C_T(A_T, \nu),$$

$$A_T(A_T, \nu) = -10^{-2} (2.26 \nu^2 + 2.07 \nu + 2.3) A_T^{0.079} \nu^2 + 0.1754 \nu - 1.342,$$
$$B_T(A_T, \nu) = (0.0573 \nu^2 + 0.0937 \nu + 0.202) A_T^{0.0274} \nu^2 + 0.0529 \nu - 0.8765,$$
$$C_T(A_T, \nu) = 10^{-4} (9.654 \nu^2 + 4.945 \nu + 3.1) A_T^{0.01867} \nu^2 + 0.4011 \nu - 1.8186;$$

$$\Lambda_n = \frac{2 G_h (1 - \nu) (1 - \nu_h)}{\pi G (1 - 2 \nu)}; \quad \Lambda_T = \frac{G_h}{\pi G};$$

$$\alpha = \left[ \frac{2 S_h \phi}{3 (1 - \phi)} \right]^{0.5}; \quad (A.1)$$
where $G$ and $\nu$ are the shear modulus and Poisson’s ratio, respectively, of the sediment. The hydrate properties are given by its shear moduli $G_h$ and Poisson’s ratio $\nu_h$. The parameter $\alpha$ is the ratio of the cemented contact radius to the grain radius in case of hydrate evenly enveloping the sediment grains.
Bibliography


BIBLIOGRAPHY


Hubral, P., and Krey, T., 1980, Interval velocities from seismic reflection time measurements: SEG.


Toldi, J., 1985, Velocity analysis without picking: SEP–43.


