Seismic Velocity Structure Associated with Gas Hydrate at the Frontal Ridge of Northern Cascadia Margin

by

Caroll López
B.Sc., Universidad Industrial de Santander, Colombia, 2005

A Thesis Submitted in Partial Fulfillment
of the Requirements for the Degree of

MASTER OF SCIENCE

in the School of Earth and Ocean Sciences

© Caroll López, 2008
University of Victoria

All rights reserved. This thesis may not be reproduced in whole or in part, by photocopy or other means, without the permission of the author.
Supervisory Committee

Seismic Velocity Structure Associated with Gas Hydrate at the Frontal Ridge of Northern Cascadia Margin

by

Caroll López
B.Sc., Universidad Industrial de Santander, Colombia, 2005

Supervisory Committee

Dr. George Spence (School of Earth and Oceans Sciences)  
Supervisor

Dr. Roy D. Hyndman (School of Earth and Ocean Sciences)  
Co-Supervisor or Departmental Member

Dr. Ross Chapman (School of Earth and Ocean Sciences)  
Departmental Member

Dr. Earl Davis (Pacific Geoscience Centre, Sidney)  
Outside Member
Abstract

At the frontal ridge near the base of the slope off Vancouver Island, wide-angle ocean bottom seismometer (OBS) data were acquired in summer 2005, in support of the Integrated Ocean Drilling Program (IODP) Expedition 311. Marine gas hydrate is present beneath the ridge based on the observation of the 'Bottom Simulating Reflector' (BSR) that is interpreted to coincide with the base of the methane hydrate stability zone. Hydrate was also observed in downhole logs and drilling by IODP. The BSR has been identified on single-channel seismic data at ~250-260 m depth beneath the ridge crest and on its seaward slope. The OBS data have been analyzed with the objective of determining the velocity structure in the upper portion of the accretionary wedge especially the hydrate stability zone and underlying free gas. As identified by a clear refracted phase, the velocity structure above the BSR shows anomalous high velocities of about 1.95 (±0.5) km/s at shallow depths of 80 - 110 m. On vertical incidence data, high amplitude reflectors are observed near this depth. Below the BSR, the velocities increase to ~2.4 km/s at sub-seafloor depths of about 600 m. A strong refracted phase with a velocity of 4.0 km/s is generated at a depth of about 1700 mbsf. Velocities from traveltime inversion of OBS data are in general agreement with the Integrated Ocean Drilling Program (IODP) X311 downhole sonic velocities. In particular, on the log data, a layer with low porosity and high velocities of 2.4 – 2.8 km/s was observed at depths of 50 – 75 m. This probably corresponds with the 1.95 km/s layer at depths of 80-110 m interpreted from the OBS data. The refraction data thus suggest that this high-velocity layer varies laterally
through the frontal ridge region, out to distances of at least 4 km from the drillhole. BSR depths (250-280 m) estimated in the present work also agree with the IODP X311 depths. From the velocity structure, we can make estimates of hydrate concentration in a region close to the deformation front, where fluid flow velocities are expected to be large. The gas hydrates concentrations vary from ~35% for the shallow phase to ~22% for the layer above the BSR. The deep refracted phase with a velocity of 4.0 km/s at 1700 m depth indicates the presence of highly compacted accreted wedge sediments.

On the SW side of the frontal ridge, a collapse structure is observed in newly acquired multi-beam bathymetry data from the University of Washington and in seismic reflection data. The BSR is present in the region surrounding the slump. There are only weak indications of its presence within the slide region. Since hydrates may prevent normal sediment compaction, their dissociation in sediment pores is thought to decrease seafloor strength, potentially facilitating submarine landslides on continental slopes. The head wall of the frontal ridge slide is ~250 m high, extending close to the BSR depth, and the slump has eroded a ~2.5 km long section into the ridge, along strike. Migrated seismic reflection data image a set of normal faults in the frontal ridge striking NE-SW, perpendicular to the strike of the ridge and the direction of plate convergence. These faults outcrop at the seafloor and can be traced from the surface through the sedimentary section to depths well below the BSR in some locations. Seafloors scarps show that fault seafloor displacements of ~25 m to 75 m are generated. The two faults with the largest seafloor scarps bound the region of slope failure on the frontal ridge, suggesting that the lateral extent of slumping is fault-controlled.

The triggering mechanism for the slope failure may have been a combination of various effects. The possible mechanisms explored include gas hydrate dissociation, high pore pressure fluid expulsion along the faults, and salinity elevation in faults which would inhibit the formation of gas hydrates along the faults. However, an earthquake may induce initial slope failure, which can not only start gas hydrate dissociation but also increase fluid expulsion and pore pressure.
# Table of Contents

Supervisory Committee ........................................................................................................... ii
Table of Contents ...................................................................................................................... v
List of Tables ............................................................................................................................. vii
List of Figures ............................................................................................................................ viii
Acknowledgments ...................................................................................................................... x

1. INTRODUCTION .................................................................................................................. 1
   1.1 Gas Hydrates .................................................................................................................. 1
       1.1.1 Importance of gas hydrate ..................................................................................... 4
   1.2 Tectonic Setting of Northern Cascadia ................................................................. 6
   1.3 Local Geology ............................................................................................................... 8
   1.4 Northern Cascadia Gas Hydrates ................................................................................. 10
       1.4.1 Recent Geophysical Studies ............................................................................... 11
   1.5 IODP Expedition 311 ................................................................................................. 13
   1.6 Gas Hydrate Concentrations ....................................................................................... 15
   1.7 Thesis Overview and Objectives .................................................................................. 17

2. DATA COLLECTION AND PROCESSING ....................................................................... 18
   2.1 Data Acquisition .......................................................................................................... 18
   2.2 Vertical Incidence Data Processing ............................................................................ 23
       2.2.1 MCS data processing ......................................................................................... 23
       2.2.2 SCS data processing ......................................................................................... 24
   2.3 Ocean Bottom Seismometer Relocation ..................................................................... 29
       2.3.1 Errors on the relocation .................................................................................... 39
   2.4 OBS Data Processing .................................................................................................. 41

3. METHODOLOGY: RAY TRACING DATA ANALYSES .................................................. 45
   3.1 Ray Tracing Method .................................................................................................... 45
   3.2 Data description and seismic events ......................................................................... 47
   3.3 Data picking ................................................................................................................ 53
   3.4 Forward and inverse modeling of travel-time .......................................................... 56
       3.4.1 Modeling procedure ............................................................................................ 56
   3.5 Error Analyses ............................................................................................................. 63
3.6 Results: details of velocity models ........................................ 65

4. INTERPRETATION OF SEISMIC VELOCITY STRUCTURE ON THE
FRONTAL RIDGE ........................................................................ 69
  4.1 High velocities and hydrate concentrations at shallow depth ....... 72
  4.2 Depth of BSR ....................................................................... 74
  4.3 Ocean basin turbidites and hemi-pelagics beneath the frontal ridge... 76
  4.4 Hydrate concentrations ......................................................... 79
     4.4.1 Calculation of potential resource ..................................... 82

5. FRONTAL RIDGE SLOPE FAILURE: FAULT-CONTROL AND POSSIBLE
RELATION WITH GAS HYDRATE DISSOCIATION ......................... 84
  5.1 Data and Observations ......................................................... 85
  5.2 Results .............................................................................. 91
  5.3 Nature of faults ................................................................. 93
  5.4 Interpretation and discussion of the slump ......................... 95
  5.5 Summary ......................................................................... 104

6. CONCLUSIONS AND RECOMMENDATIONS .......................... 106
  6.1 Velocity analyses .............................................................. 106
  6.2 Margin-perpendicular normal faults ................................... 108
  6.3 Frontal ridge slope failure ............................................... 108
  6.4 Recommendations for future studies at the frontal ridge ........ 109

REFERENCES ........................................................................ 112

APPENDIX ............................................................................ 121
List of Tables

Table 2.1 Time and velocity parameters for Stolt migration of MCS Line PGC_CAS03 ................................................................. 23
Table 2.2 Set of time and velocity parameters for FMIG migration of SCS data ............... 24
Table 2.3 Deployment data of the grid of OBSs. ................................................................. 32
Table 2.4 Relocation data for the OBSs grid. ................................................................ 38
Table 3.1 Picking uncertainties for wide-angle arrivals .................................................. 55
Table 3.2 Inversion statistics ......................................................................................... 62
Table 5.1 Displacements and dips of major faults that bound the slump ....................... 92
List of Figures

Figure 1.1 Low-frequency seismic section from Multichannel Line PGC9902_CAS03... 3
Figure 1.2 Map of Northern Cascadia Margin gas hydrate zone........................................ 7
Figure 1.3 Migrated multichannel seismic line PGC_CAS03............................................. 8
Figure 1.4 Multichannel seismic line 85-02 across Northern Cascadia margin.................. 9
Figure 2.1 Location of Site CAS-3B ................................................................................ 19
Figure 2.2 Deployment geometry for seismic source and single channel streamer ............ 20
Figure 2.3 Single-channel seismic deployment from cruise PGC04-08 ............................... 21
Figure 2.4 Single-channel seismic PGC05-09 Line-2 after time migration...................... 25
Figure 2.5 Single-channel seismic PGC05-09 Line-3 after time migration...................... 25
Figure 2.6 Seismic images for a section of Line CAS3B-11 before and after deconvolution.... 27
Figure 2.7 Seismic images for a section of Line CAS3B-11 before and after time migration 28
Figure 2.8 Velocity-depth profile for the water column....................................................... 31
Figure 2.9 Re-located position of OBS-A ......................................................................... 34
Figure 2.10 Re-located position of OBS-H ....................................................................... 35
Figure 2.11 OBS-A residuals ............................................................................................. 36
Figure 2.12 OBS-H residuals ............................................................................................. 37
Figure 2.13 OBS relocation ................................................................................................. 39
Figure 2.14 Mean amplitude-frequency spectrum OBS-D over a time window from 1495 to 1742 ms.................................................................................................................. 41
Figure 2.15 Mean amplitude-frequency spectrum at OBS-L over a time window from 1812 to 2260 ms .................................................................................................................. 42
Figure 2.16 Recorded section for vertical-geophone on OBS-D ......................................... 43
Figure 2.17 Recorded section for vertical-geophone on OBS-L ......................................... 44
Figure 3.1 Reflected and refracted rays traveling along a profile through OBS A, D, H and G ................................................................................................................................. 48
Figure 3.2 Reflected and refracted rays traveling along a profile through OBS L, N and C ................................................................................................................................. 49
Figure 3.3 Refracted and reflected rays traveling along a profile through OBS D and L ................................................................................................................................. 50
Figure 3.4 Multichannel seismic Line PGC_CAS03 after migration.................................. 52
Figure 3.5 Migrated single-channel seismic Line CAS3B-25 ............................................. 52
Figure 3.6 Single-channel seismic CAS3B-X03 ................................................................. 53
Figure 3.7 Seafloor nmo-corrected wide-angle data along SCS Line 2 and MSC line PGC_CAS03 .......................................................... 54
Figure 3.8 Wide-angle data from OBS L and C along Line 3 and SCS data from nearby CAS3B-25 .......................................................... 55
Figure 3.9 Site U1326 sonic log ......................................................................................... 57
Figure 3.10 Observed travel-times compared with calculated travel-time along MCS line PGC_CAS03 and SCS Line 2 .......................................................... 59
Figure 3.11 Observed travel-times compared with calculated travel-time (solid lines) along SCS line CAS3B-25 and SCS Line 2. ................................................................. 60
Figure 3.12 Observed travel-times compared with calculated travel-time along SCS Line 8. ........................................................................................................... 61
Figure 3.13 Perturbations to the best fitting model. .................................................. 64
Figure 3.14 Final velocity model for SCS Line2 and MCS PGC_CAS03. ..................... 67
Figure 3.15 Final velocity model for Line3. ............................................................. 67
Figure 3.16 Final velocity model for Line8. ............................................................ 68
Figure 4.1 Final velocity model M1, along time migrated MCS PGC_CAS03 converted to depth... .......................................................... 70
Figure 4.2 Final velocity model M2, along time migrated SCS CAS3B_Line-25 converted to depth. ............................................................................. 70
Figure 4.3 Final velocity model M3, in two way time, along time-migrated line MCS 89-08. .......................................................................................................... 71
Figure 4.4 Velocity profiles from line intersection points. ........................................ 73
Figure 4.5 Resistivity log from IODP X311 Site U1326. ............................................ 74
Figure 4.6 Velocity profiles from OBS L (current survey) and OBS-1 from Waldron (1982). ................................................................................................. 78
Figure 4.7 Gas hydrate saturation calculated by effective porosity reduction. ........... 81
Figure 4.8 Map of distribution of BSR base on single-channel seismic lines from cruise PGC0408. ................................................................. 83
Figure 5.1 Multi-beam bathymetry and acoustic side-scan data from the frontal ridge. 87
Figure 5.2 Migrated single-channel seismic lines and seafloor scarps, consistent from line to line. ................................................................. 89
Figure 5.3 Map of distribution of BSR based on single-channel seismic lines from cruise PGC0408. ........................................................................... 90
Figure 5.4 Bathymetric map displaying the seafloor traces of frontal ridge faults. ....... 93
Figure 5.5 Migrated single-channel seismic lines CAS3B-13 and 15 and seafloor scarps aligned with depth. ............................................................. 95
Figure 5.6 Frontal ridge reconstruction. ................................................................. 97
Figure 5.7 Depth profiles along different transects through the ridge. ....................... 98
Figure 5.8 Ratio of shear strength to overburden pressure from IODP X311 Site 1326. 101
Figure 5.9 Seismic image for a section of SCS Line CAS3B_11. .............................. 103
Acknowledgments

I would first like to express my gratitude to my supervisor George Spence for enabling me to come and study under his supervision at the University of Victoria, and very especially for his continuous guidance and support throughout this project. Special thanks also to my committee members Roy Hyndman and Ross Chapman for their suggestions during my thesis. I would also like to thank the Chief Scientist for the cruises PGC05-09 and PGC04-08. Thanks to NSERC and the University of Victoria for fellowship funding.

I also like to thank Ross Haacke, Michael Riedel, Boris Marcaillou, Tao He, Ranjan Dash, German Yury Ojeda and Ivan Dario Olaya for helpful scientific discussions.

I especially would like to thank my extraordinary roommates Sabine, Zach and Marc for all the great moments…and dinners. Thanks to Ranjan for all the good moments at E-Hut and very especially to Bine for her total support and friendship during all this time, danke.

Mil gracias a mis amigos argentinos y mexicanos: Laura, Diego, Valeria, Carmen, Tono, Jany y Julio por su incondicional amistad y apoyo durante estos 2 años en Victoria. Mil gracias a mi familia por su total apoyo desde la distancia. Finalmente quiero agradecer a Mauricio por su invaluable compañía y apoyo a lo largo de todo este sueño.
1. INTRODUCTION

Marine gas hydrates have been identified in many areas of the accretionary prism of the Northern Cascadia subduction zone at the frontal ridge near the base of the slope off Vancouver Island. The present work at the frontal ridge involves determination of regional seismic velocities from ocean bottom seismometer data and interpretation of the velocity structure in terms of gas hydrate distribution and downward sediment compaction. As well, indicators of the presence of gas hydrates from both seismic reflection and refraction data are applied to the analysis of slope stability on the frontal ridge. As an introduction to the study, a brief overview of gas hydrates and a summary of geological and geophysical features and recent investigations on the Cascadia margin are presented below.

1.1 Gas Hydrates

Gas hydrates are ice-like solids composed of water molecule cages that enclose small-molecular-weight gas molecules. When the gas is mainly methane (>99%), these substances are named methane hydrates. The stability of gas hydrate is controlled by temperature and pressure. Gas hydrates are present in permafrost land regions and beneath the seafloor in outer continental slope regions as in the present study. Marine gas hydrates are typically found in the top few hundred meters of sediments for water depths greater than about 600 m, where sufficiently low temperature and high-pressure conditions exist. Below the base of the gas hydrate stability zone, where temperatures are too high for gas hydrate to exist, methane may occur as free gas (e.g., Kvenvolden et al., 1993).
As gas hydrates can substantially increase sediment electrical resistivity and seismic velocity, they may be detected and the concentration may be quantified by data from seismic and electrical surveys (Hyndman and Spence, 1992; Andreassen et al., 1995; Yuan et al., 1999). The presence of gas hydrates increases the seismic velocities of the sediments, since gas hydrate has high compressional velocity $V_p$ (~3300-3800 m/s) and high shear velocity $V_s$ (~1800 m/s), (Sloan, 1990). The concentration of hydrate is thus related to the increase in seismic velocity above the velocity of sediments containing no hydrate or gas. The resistivity of sediments containing hydrate is high because conductive salts are excluded from the hydrate structure.

The regional distribution of gas hydrate accumulations is inferred mainly from the presence of the Bottom Simulator Reflector (BSR) in seismic sections (Shipley et al., 1979). BSRs (Figure 1.1) are seismic reflectors resulting from a decrease in acoustic impedance from high-velocity sediments containing gas hydrate above the BSR to low-velocity sediments containing free gas below the BSR. Most of the impedance contrast appears to be caused by the presence of free gas within the sediments beneath the hydrate stability zone. BSRs commonly match closely the base of the hydrate stability zone determined from drilling data, although theoretical models indicate that, in regions where fluid and gas flux are low, the base of hydrate stability may be above the top of the free gas zone (Xu and Ruppel, 1999).
Figure 1.1 Low-frequency seismic section from Multichannel Line PGC9902_CAS03. Note that the bottom-simulating reflector (BSR), which typically marks the base of the gas hydrate stability zone, is parallel to the seafloor. Seafloor and BSR reflections have opposite polarity.

The BSR reflector can be identified by the following characteristics: (i) it emulates the shape of the ocean floor, (ii) it can cut transversely across stratigraphic reflectors, and (iii) it has an opposite polarity to the ocean bottom reflector (Kvenvolden, 1998) (Figure 1.1).

Alternatively, hydrate may be present even where there is no BSR, e.g. regions of subsidence. In these regions, slope basin formation, with accompanying sediment deposition, results in downward movement of the base of the stability field. As a result, the gas layer is transformed to hydrate and the BSR is much weakened (e.g., von Huene and Pecher, 1999; Haacke et al., 2007).
1.1.1 Importance of gas hydrate

Gas hydrates are important for three main reasons: (i) they represent a great potential energy resource, (ii) they pose a submarine slope stability risk since sediment strength decreases if gas is present (Kvenvolden, 1988), and (iii) they might result in "greenhouse" type climate effects if significant amounts of dissociated methane reach the atmosphere (Kvenvolden, 1988; Macdonald, 1990; Kennett et al., 2003).

Future energy resource:

The continuous increase in energy demand worldwide and declining conventional reserves provide motivation to explore for alternative energy resources. Unconventional deposits such as coal bed methane and oil sands are gaining more attention. Furthermore, increasing interest in natural gas is part of a worldwide campaign advocating the use of cleaner burning fossil fuels. Thus, methane stored in gas hydrates has attracted attention as an unconventional energy resource. Many methods are being tested for gas hydrate production. Depressurization is the most often considered method for commercial production of gas hydrates (Sloan, 1998), but combined depressurization and thermal stimulation have been used recently to produce a small amount of gas from the Mallik 5L-38 research well located on the Mackenzie delta, Northwest Territories, Canada (Dallimore and Collett, 2005).

The volume of methane in gas hydrates globally may be enormous, based not only on its very high methane content but also on its worldwide occurrence in continental margins. Assessments of gas hydrate, as an energy resource, may have been overly optimistic. Estimations vary from 10,000 Gt of methane carbon (Kvenvolden 1993,
Collett 2000), which is twice the mass of conventional hydrocarbon reserves, to 500–2500 Gt of methane carbon (Milkov, 2004). However, even with this uncertainty on the exact size of the reserve, the interest in this resource has been increasing substantially over the last decade.

**Impact on climate change:**

The stability of gas hydrate is critically sensitive to pressure–temperature conditions within the host sediments. A rise in bottom water temperature, a decrease in water depth or a regional uplift may cause dissociation of gas hydrates stored in marine sedimentary strata. The subsequent release of carbon to the atmosphere may enhance warming through the greenhouse effect (e.g. Maclennan and Jones, 2006). In the last decade, many studies have linked glacial maxima (low sea levels) with gas hydrate dissociation (e.g. Haq, 2000; Andersen et al, 2004, Augustin et al, 2004). Additionally, the “clathrate gun” hypothesis of Kennett et al. (2003) suggests that the release of methane from marine sediments was a major source of atmospheric methane in the Quaternary.

In general, the effects of sediment hydrate dissociation are believed to increase oceanic and atmospheric methane concentrations resulting in higher atmospheric temperatures (Paull et al., 1991). However, although methane is a greenhouse gas in the atmosphere, Kvenvolden (1999) recognized that much methane from dissociated gas hydrates may never reach the atmosphere, but rather the methane may be converted to carbon dioxide and sequestered by the hydrosphere/biosphere before reaching the atmosphere.
Slope Stability Geo-hazard:

Since hydrates inhibit sediment compaction, their dissociation in sediment pores is thought to decrease seafloor strength, potentially facilitating large submarine landslides on continental slopes. These events in fact can cause mass wasting, tsunamis, and the possible rapid release of methane into the atmosphere (Nisbet and Piper, 1998; Paull et al., 1991). These submarine geohazards, potentially caused by gas hydrate dissociation, are of immediate and increasing importance as humankind moves to exploit seabed resources in deeper waters of coastal oceans.

The causal relationship between slope failure and gas hydrate dissociation has not been proven, but a number of observations support a potential connection (Mienert et al., 2005). The Storegga Slide, off the coast of Norway, is one of the largest underwater slide complexes known, and it has been proposed as a significant source of past methane release into the atmosphere. This slide has also been associated with a tsunami on the Norwegian coast 7200 years ago (Bondevik et al., 1997).

Sediment blocks on the seafloor in fjords of British Columbia (Bornhold and Prior, 1989), and massive bedding-plane slides and rotational slumps on the Alaskan Beaufort Sea continental margin (Kayen and Lee, 1991), among other examples, have also been associated with this phenomenon and are summarized by Kvenvolden (1999).

1.2 Tectonic Setting of Northern Cascadia

The Northern Cascadia gas hydrate area is located in the accretionary prism of the Cascadia subduction zone (Figure 1.2), where the Juan de Fuca oceanic plate subducts beneath the North American continental plate at ~46 mm/year (Riddihough, 1984).
The Juan de Fuca spreading center is located several hundred kilometers offshore. The age of the oceanic plate is only 6-8 Ma at the margin. Incoming hemipelagic and turbidite sediments (~2-3 km thickness) are scraped off at the deformation front. This process of accretion, ongoing since early Eocene, has formed a large clastic sedimentary prism beneath the continental shelf and slope.

The sediments entering the accretionary prism undergo increased tectonic deformation. Thrust faulting and folding thicken the section as they progress landward. As the sediment deformation in the accretionary prism progresses, sedimentation also continues, and an irregular blanket of less deformed slope basin sediments accumulates on top of the accreted Cascadia Basin sediments. These recently deposited sediments fill local topographic lows.

![Map of Northern Cascadia Margin gas hydrate zone. IODP X311 sites are located on multichannel seismic line 89-08. The shaded region represents the gas hydrate zone with a width of ~30 km and a length of ~200 km (Hyndman et al., 2001). The red circle corresponds to X311 Site U1326. (After Expedition 311 Scientists, 2005.)](image)

**Figure 1.2** Map of Northern Cascadia Margin gas hydrate zone. IODP X311 sites are located on multichannel seismic line 89-08. The shaded region represents the gas hydrate zone with a width of ~30 km and a length of ~200 km (Hyndman et al., 2001). The red circle corresponds to X311 Site U1326. (After Expedition 311 Scientists, 2005.)
1.3 Local Geology

The study region (e.g., Hyndman, 1995) is located on the first uplifted ridge of accreted sediments on the Northern Cascadia margin of Vancouver Island (see Figure 1.1, 1.3). Site U1326, from the Integrated Ocean Drilling Program (IODP) expedition 311 (X311), is located at the top of the frontal ridge. This drill hole is the seaward-most site along the margin-perpendicular transect established during IODP X311 (Figure 1.4).

One of the most interesting features of this region is a collapse structure observed in the newly acquired bathymetry data from the University of Washington (D. Kelley University of Washington, pers. commun. 2005; Riedel, et al., 2006). The head wall of the slump feature is ~250 m high and the slump has eroded a ~2.5 km long section into the ridge. Slump material was also identified in the bathymetry data and previously in SeaMARCIIs acoustic imagery (Davis et al., 1987).

![Diagram of migrated multichannel seismic line PGC_CAS03](image)

**Figure 1.3** Migrated multichannel seismic line PGC_CAS03. Notice that the BSR is strong to the NW but almost absent to the SE. Steep normal faults penetrate the sediments above the BSR and form scarps at the seafloor.
Figure 1.4 Multichannel seismic line 85-02 across Northern Cascadia margin through ODP Site 889/890 and IODP-X311 well sites transect (after Riedel et al., 2006).

Along a margin-parallel multichannel seismic line at the top of the ridge, a series of seafloor scarps are clearly identified (Scherwath et al. 2006; Riedel et al. 2006b), (see Figure 1.3). These features are interpreted as normal faults that clearly outcrop at the seafloor and generate a seafloor displacement of ~25 m. The faults can be seismically traced from the surface down through the sedimentary section to depths below the BSR in some parts.

IODP X311, Site U1326, recovered a sequence that corresponds to slope not accreted sediments with three identified lithostratigraphic units described in the summary results of X311 (Riedel et al. 2006b). The units are characterized by fine-grained (clay to silty clay) detrital sediments with thin silty/sandy interlayers from turbidites.
1.4 Northern Cascadia Gas Hydrates

Gas hydrates at the northern Cascadia margin have been comprehensively studied using a multitude of geological, geophysical, and geochemical techniques (Hyndman et al., 2002; Spence et al., 2000). As mentioned before, the occurrence of gas hydrate has historically been inferred from the presence of a BSR in seismic images (Shipley et al., 1979). BSRs have been widely observed in seismic data collected across the northern Cascadia margin (Hyndman and Spence, 1992; Hyndman et al, 2001) and they were used to map the general distribution of gas hydrates along this margin (see Figure 1.2).

Offshore Vancouver Island, the gas hydrate-related BSR occurs in a 30 km wide band parallel to the coast beneath much of the continental slope, with its seaward limit about 5-10 km landward of the deformation front (Hyndman et al., 2001). Figure 1.4 shows a clear BSR over more than half of this region. BSRs are not evident at water depths of less than 600 m on the upper slope (Hyndman and Davis, 1992). The BSR is observed at a shallow depth of about 230-250 meters below the seafloor (mbsf). The shallow BSR depth is suggested to be the consequence of the relatively shallow water depth and high heat flow associated with the young age of the underthrusting oceanic lithosphere (Hyndman, 1995).

Seaward of the deformation front, in the deep sea of the Cascadia basin, no BSR is observed, and ODP drilling data at reference Site 888 indicated that hydrate is not evident (Westbrook et al., 1994). Interpretations vary as to whether or not hydrate is present in the Cascadia basin. Hydrates may not be present because the low-permeability of the well-bedded undeformed sediments inhibits vertical fluid and methane flow (e.g., Zuehlsdorff et al., 2000). Alternatively, hydrates may be present, but they occur in a
region of subsidence and thus BSR reflections are weak (von Huene and Pecher, 1999; Haacke et al., 2007).

These interpretations are based on the fluid expulsion model proposed by Hyndman and Davis (1992), which explains the occurrence of marine gas hydrates in accretionary prism sediments by chronic vertical fluid flux. This model proposes that the upward methane flux produces hydrate when the fluid-borne methane passes above the base of hydrate stability.

1.4.1 Recent Geophysical Studies

The northern Cascadia margin offshore Vancouver Island has been the focus of many marine geological and geophysical studies over the past two decades. Adjacent onshore studies of the LITHOPROBE program (e.g., Clowes et al., 1987a, 1987b; Davis and Hyndman 1989; Spence et al. 1991), as well as large-scale seismic reflection profiling along the margin allowed the construction of continuous onshore–offshore structural cross-sections (e.g., Hyndman, 1995). Reviews of the geophysical and geological data, the structure, and the margin tectonics have been provided by Hyndman et al. (1994), and Hyndman (1995). The extensive geophysical and geological data analyzed and integrated also include: numerous seismic reflection surveys with a variety of systems, seismic refraction, magnetics, gravity, swath bathymetry, seafloor acoustic imagery, heat flow, seismicity and piston coring. Details about reflection and refraction seismic data are presented below:
Seismic Reflection Data

Many studies have been based on the analyses of seismic reflection data. These include the following: BSR mapping and estimation of BSR reflection coefficients (Hyndman and Spence, 1992; Fink and Spence, 1999), BSR reflection waveform modeling and modeling of BSR frequency dependence (Yuan et al., 1996; Chapman et al., 2002; Spence et al.; 2000; Fink and Spence 1999), detailed interval velocity-depth profiles (Yuan et al., 1996; Chen 2006), modeling of amplitude-versus offset (AVO) or full waveform inversion (Singh et al., 1993; Singh and Minshull, 1994; Yuan et al; 1996; Yuan et al., 1999; Chen, 2006). The frequency dependence of the BSR was obtained from seismic reflection data over a wide range of frequencies (20-650 Hz) collected from 5 different seismic surveys. The frequency dependence has been modeled using synthetic seismograms, and results indicate that the thickness of the velocity gradient at the base of the hydrate stability zone is 6-10 m near ODP Site 889/890, with velocity decreasing by 250 m/s (Chapman et al., 2002).

Seismic Refraction Data

Seismic refraction methods provide the best means for determining seismic velocity, because (1) the refractions are recorded out to large offsets and so travel large horizontal distances within a given layer, (2) the refracted phase can yield velocity structures to greater depths and in regions where reflection coherency is generally poor, and (3) the largest recorded amplitudes are typically found near the critical reflection distance, and so signal-to-noise ratios are large over this range. Several deployments of OBSs have been carried out near ODP Site 889 and IODP 1327, in 1993, 1999, and 2005. These data have allowed traveltime inversion studies to provide 2-D or 3-D
velocity models in the region of the drill site (Spence et al., 1995; Hobro et al., 1998; Hobro, 1999; Hobro et al., 2005; Zykov, 2006). According to these studies, velocities directly above the BSR range from 1.87 to 1.96 km/s. The higher velocities relative to a no-hydrate reference appear to be correlated to an increase in the reflection strength of the BSR.

1.5 IODP Expedition 311

In summer 2005, a drilling transect across the accretionary prism included four sites completed during IODP expedition 311 (X311) (see Figure 1.4), Site U1326 was located at the frontal ridge, U1325 in the slope basin, U1327 near Site 889/890, and U1329 near the upper limit of water depths where hydrate is stable. The transect was directed at studying the evolution of gas hydrates across the margin, since expected fluid flow rates decrease landward whereas the total time that they flow through the sediments increases landward. The objectives of this expedition were: (i) to study the mechanism of formation of marine gas hydrates in the region, (ii) to measure gas hydrates distribution, (iii) to define the nature of the BSR, and (iv) to examine the effect of gas hydrates on the physical properties of the sediments.

The X311 boreholes transect across the accretionary prism shows significant lateral variations. At Site U1326, logging-while-drilling/measurements-while-drilling (LWD/MWD) resistivity logs confirmed the presence of gas hydrate in thin sand layers between 50 and 240 mbsf. The BSR is at 270 m. The data also yielded evidence of high gas hydrate occurrences between approximately 50 and 100 mbsf, with concentrations locally as high as 80% of the pore space. The interstitial water (IW) chlorinity profile
showed an almost constant, near-seawater baseline, and abundant local low-chlorinity anomalies associated with gas hydrate extending to 270 mbsf.

For Site U1325, LWD/MWD resistivity logs suggested that gas hydrate is concentrated in thin sand layers mainly within an interval between 173 and 240 mbsf, close to the base of the gas hydrate stability zone.

At Site U1327, where the base of the gas hydrate stability zone is 240 mbsf, high electrical resistivity and low density values were found in the depth range from 120 to 138 mbsf. Modelling of high resistivities suggested that ~50% of pore volume is filled with gas hydrate in sand-rich intervals up to 20 m thick.

Site U1329 represents the eastern limit of gas hydrate occurrence on the northern Cascadia margin. Only a faint BSR was identified in seismic data in the vicinity of this site, and no other indicators of hydrate were found in the logs or core from the drill site.

The results of X311 increased the understanding of how gas hydrates occur in this region. The combination of drilling data and seismic observations suggests strong lithologic control of the gas hydrate occurrence with preferred gas hydrate formation in sand-rich sedimentary sections. Particularly at sites U1326 and U1327, gas hydrates appear to be more concentrated in sand layers where porosities and presumed permeabilities are high.

Anomalous gas hydrate concentration intervals at shallow depths of 50-120 mbsf were not expected on the basis of existing simple models of gas hydrate formation in accretionary complexes. In Hyndman and Davis, (1992); gas hydrates were predicted to be more concentrated near the base of the gas hydrate stability zone (just above the BSR). However, these first-order models did not consider variable porosities in sediment layers
and local fluid and gas channelling, so hydrate was predicted to be evenly distributed throughout the pore space of the sediments.

1.6 Gas Hydrate Concentrations

Quantitative estimates of hydrate concentrations are available in only a few locations. Seismic velocities in the sediments can be used to estimate hydrate concentrations, since the hydrate high-velocity and grain cementing increases sediment velocity. As well, since P-wave velocity decreases significantly with even a small amount of gas, velocities below the BSR can provide a measure of gas concentration in this region.

One method to determine seismic velocity is from detailed multichannel semblance velocity analyses. Other methods at a drillhole are vertical seismic profile (VSP), and downhole sonic log velocities. Velocities of sediment samples measured in a laboratory after hydrate dissociation provide important information on the velocity-depth profile for sediments with no hydrate and no gas. The reference profile is a critical factor for quantitative estimates of hydrates and gas concentrations. According to Yuan et al. (1996), the Cascadia sediment section is very homogeneous on scales greater than a seismic wavelength, although there is fine-scale turbidite layering. Thus, the reference velocity-depth may be estimated by extrapolating the deeper velocity trend upwards. Based on this reference profile, Yuan et al. (1996) estimated that about 2/3 of the BSR velocity contrast seems to be consequence of high-velocity hydrate and about 1/3 from the low-velocity gas, relative to a no-hydrate no-gas reference velocity-depth (e.g., Yuan et al., 1996).
Other methods to estimate hydrate concentration are based on resistivity logs, seafloor electrical resistivity (controlled-source-electromagnetic, CSEM), and core sample chlorinity. At site 889 of ODP Leg 146, gas hydrate concentrations from seismic velocity estimates were calculated to be up to 20-30% of the pore space (10-15% of total sediment volume) in a layer above the BSR with a thickness of about 100 m, and 1% free gas was found in the underlying 10-20 m layer. Initial hydrate concentrations estimated from resistivity and chlorinity are even higher (Yuan et al., 1996, Hyndman et al, 1999). For marine gas hydrate occurrences, such high concentrations have not been observed on other margins.

In preparation for X311 in 2005, the gas hydrate concentrations along the northern Cascadia margin were recalculated using Leg 146 acoustic/electrical resistivity logs and pore water chlorinity/salinity data (Riedel et al., 2005; Chen 2006). The new estimates show that the concentrations could be lower, between 5% and 10% on average of the pore volume from ~130 mbsf to the BSR (~230 mbsf) (Riedel et al. 2006b).

New reference velocity profiles were also calculated from the Site 889 porosities using various published empirical relationships between porosity and velocity (Jarrard et al., 1995; Hyndman et al., 1993), and the Lee et al. (1993) weighted equation. All the newly proposed baselines from X311 were significantly shifted toward higher seismic velocities relative to the former baseline defined by Yuan et al. (1996) and Hyndman et al. (2001). However, significant uncertainty remains in the applicability of empirical parameters for each equation used (Expedition 311 Scientists, 2005).
1.7 Thesis Overview and Objectives

In support of IODP X311, wide-angle ocean bottom seismometer (OBS) data were acquired in 2005, and single channel seismic data were collected in 2004 and 2005 on the frontal ridge. A BSR was identified on the single-channel seismic data at ~240-270 m depth beneath the ridge crest and on its seaward slope. With the objective of determining the seismic velocity structure of the region and its association with the presence of gas hydrates, a travel-time inversion analysis was carried out on the OBS data. Also, structural features were interpreted over a grid of margin-parallel single-channel seismic lines, where seafloor scarps and a large slope failure were observed.

The general objective of this thesis is to help understand the processes of formation and dissociation of methane hydrate in the region. Specific objectives of this study are:

- The regional seismic velocities at the frontal ridge offshore Vancouver Island.
- Gas hydrate concentration and distribution in the region.
- 3D structure of frontal ridge faults that are observed to intersect the seafloor and produce large scarps
- Possible relation of gas hydrate to slope stability.
2. DATA COLLECTION AND PROCESSING

This chapter presents data acquisition and processing details for Ocean Bottom Seismometer (OBS) data collected in 2005, single channel seismic (SCS) data from 2004 and 2005, and multichannel seismic (MCS) data from 1999. Recent downhole data collected during IODP X311 were also examined.

2.1 Data Acquisition

*Single channel and OBS seismic data from 2005 survey*

In August 2005, Pacific Geoscience Centre, Geological Survey of Canada, the University of Victoria and the University of Toronto conducted the cruise PGC05-09 to provide geophysical information in support of drilling by IODP offshore Vancouver Island in September/October 2005.

In 2005, eight OBSSs from Dalhousie University were deployed in a grid-pattern at a nominal separation of 650 m on the frontal ridge, just landward of the deformation front (Figure 2.1). OBS D was positioned at IODP proposed Site CAS-3B, later Site U1326 of X311. The seismic source was a GI-gun with volume reducers removed, that is, the GI-gun was configured with a 105 cu. in. generator, and a 105 cu. in. injector. The GI-gun was initially fired at a 10 second rate and later at a rate of 11 seconds. The OBS internal timing device was a highly stable Seascan Inc. SAIL clock with a drift rate of < 2 ms per day. As the GI gun trigger, an Odetics/Zyfer GPS time and frequency system was used as the time standard to set and measure drift for the OBS Seascan clock. A shot log was created using Dalhousie University software. The shot table and the clock drift were
integrated onboard with the ship’s DGPS navigation and the OBS data to produce SEGY files.

Figure 2.1 Location of Site CAS-3B. Note the deployment location of OBSs, MCS and SCS data. Red dots represent the OBSs with a nominal separation of 650 m. Red stars correspond to the IODP X311 drill holes.

Single channel seismic lines were recorded simultaneously with OBS recording. Reflection data were recorded on a Teledyne Ltd. Model 178 single-channel array. The preamplifier filter in the Teledyne had a lo-cut value of 30 Hz. A Kronhite filter in the lab applied a band-pass filtering from 15-2000 Hz. Record length was 4 s for water depths < 1300 m, and 6 s for water depths near 2000 m. For further details about the deployment geometry for the seismic source and single-channel streamer, see Figure 2.2.
Eight seismic lines were collected at the frontal ridge. Three lines were oriented NW-SE (Line 1, 2, and 3), another 3 lines were NNE-SSW (Lines 4, 5, and 6), and the other two lines were NE-SW (Lines 8 and 10), see Figure 2.1. The line lengths were ~20 km for the first 7 lines and ~9 km for the last line. Unfortunately, additional line orientations could not be obtained as planned because of failure of the air compressor. However, with the variety of azimuths already collected, a sufficient number of crossing ray paths were available to allow determination of 3D structure through inversion of traveltimes.

The single-channel seismic data were very noisy, and produced poor reflection sections due to poor sea conditions (winds of 30-35 knots and wave heights of 2-3 m). The OBS records, not influenced by surface conditions, are of good quality.

![Diagram](image)

**Figure 2.2** Deployment geometry for seismic source and single channel streamer. Surveys were carried out on the ship C.S.S. John P. Tully during the 2004 and 2005 cruises.

*Single channel seismic data from 2004 survey*

In August 2004, the cruise PGC04-08, (SEICOMAG), was also conducted by the Geological Survey of Canada, the University of Victoria and the University of Toronto. A grid of 18 margin-parallel lines was collected along the ridge length (see Figure 2.3).
The seismic source was a Bolt 120 cu.in. airgun (peak frequency ~ 120 Hz). Ten lines were 10 km long, and 8 lines were 7 km long. The line separation was 50 m near the originally recorded COAMS line 2, but increased to 100 m farther away from this COAMS line. A set of 10 lines, each ~5 km long at a line separation of 500 m, was collected normal to the margin and ridge.

Figure 2.3 Single-channel seismic deployment from cruise PGC04-08. Yellow stars correspond to the IODP X311 Sites U1326 (Hole U1326A and U1326D) and U1325 drill holes. A total of 18 lines was collected at the frontal ridge.
Reflections were recorded on a 25 m Teledyne single-channel array, with a preamplifier lo-cut filter of 30 Hz and a lab Kronhite filter with a pass-band of 30-1000 Hz. The seismic data were recorded on a AGC GeoDiff system, at a sample rate of 1 ms. Record length was 4 s for water depths < 1300 m, 6 s for water depths near 2200 m, and 7 s for deep ocean basin water depths of 1500 m; no deep water delay was applied. For further details about the deployment geometry for seismic source and single channel streamer, see Figure 2.2.

Due to the poor image quality of the PGC05-09 SCS lines, the SCS lines for 2004 cruise (Figure 2.3) were used to examine small-scale faults that penetrate the seafloor on the frontal ridge just landward of the deformation front. Some lines from 2004 are near-coincident with seismic lines from the other surveys as follows: MCS line PGC_CAS03 (COAMS, 1999) and SCS Line 2 (from 2005 survey) are near-coincident with SCS line CAS3B-11; SCS Line 3 (from 2005 survey) is near-coincident with SCS CAS3B-25; and MCS Line 89-08 is near-coincident with SCS CAS3B-04.

**Multichannel seismic data**

Two multichannel seismic lines collected in previous years were also used in the present work. MCS line PGC_CAS03 (see Figure 2.1) was acquired during the summer of 1999 by the multichannel Canadian Oceanographic Acoustic Measurement System (COAMS) array, with an active length of 1140 m. A single airgun was used, with a chamber volume of 40 cu. in. MCS line 89-08 (see Figure 2.1) was acquired by the Digicon Geophysical Corporation for the Geological Survey of Canada in 1989, prior to ODP leg 146. The survey was acquired with a tuned airgun array with a total volume of
125 L (7820 in3), and a 3600 m streamer recorded shots on 144 channels. The shot point interval was 50 m, giving CDPs with 36-fold multiplicity, and a CDP spacing of 12.5 m.

2.2 Vertical Incidence Data Processing

All the processing and manipulation of the OBS and SCS data were completed using interactive 2-D seismic processing software of the GLOBE Claritas and Kingdom Suite seismic systems.

2.2.1 MCS data processing

MCS Line PGC_CAS03 (Figure 1.3, Chapter 1) was processed by M. Scherwath in 2004 and presented in Schewarth et al. (2006) and Riedel et al. (2006b). Scherwath applied Stolt migration on post-stack data, for which the nominal trace spacing between adjacent CDPs was 9.52 m. This migration allows no lateral variation in the rms velocity function, which is defined only as a function of time (see Table 2.1). To mute the shallow, far-offset data that are stretched too much after normal-move-out (NMO) application, a stretch factor of 1.0 was used.

Table 2.1 Time and velocity parameters for Stolt migration of MCS Line PGC_CAS03. The water depth range is 1800-2100 m

<table>
<thead>
<tr>
<th>Time (ms)</th>
<th>RMS Velocity (m/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0</td>
<td>1493</td>
</tr>
<tr>
<td>2550</td>
<td>1500</td>
</tr>
<tr>
<td>2900</td>
<td>1650</td>
</tr>
<tr>
<td>3500</td>
<td>2000</td>
</tr>
</tbody>
</table>
2.2.2 SCS data processing

Single channel seismic data from 2005 survey

The SCS data from cruise PGC04-09 were initially processed by a spike or gapped Wiener deconvolution on post-stack data with a gap of 10 ms and a filter length of 800 ms. After deconvolution, a finite difference time migration routine (FDMIG) with interval velocities selected from the sonic log of X311 Site U1326 was applied (Figure 2.5; Table 2.2). The FDMIG routine is based on an X-T domain implicit 45 degree migration. Despite the 45 degree algorithm, FDMIG gives reasonable results up to about 60 degrees dip (claritas 4.4.1 manual, 2001). The data were filtered with a zero-phase band-pass Butterworth filter. The low cut frequency of the Butterworth was set to 40 Hz, with a slope to zero amplitude at 20 Hz; the high cut frequency was set to 100 Hz, with a slope to 160 Hz (20-40-100-160 Hz band-pass filtering). The reflections from these data are very poor due to bad weather conditions (see Figure 2.4 and 2.5).

Table 2.2 Set of time and velocity parameters for FMIG migration of SCS data. The water depth range is 1800-2100 m

<table>
<thead>
<tr>
<th>Line Name</th>
<th>KEYVAL (Shotid)</th>
<th>T1 (ms)</th>
<th>V1 (m/s)</th>
<th>T2 (ms)</th>
<th>V2 (m/s)</th>
<th>T3 (ms)</th>
<th>V3 (m/s)</th>
<th>T4 (ms)</th>
<th>V4 (m/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Line 2</td>
<td>40378</td>
<td>2734</td>
<td>1480</td>
<td>3050</td>
<td>2100</td>
<td>3250</td>
<td>2300</td>
<td>4000</td>
<td>2400</td>
</tr>
<tr>
<td></td>
<td>40787</td>
<td>2394</td>
<td>1480</td>
<td>2702</td>
<td>2100</td>
<td>2902</td>
<td>2300</td>
<td>4000</td>
<td>2400</td>
</tr>
<tr>
<td></td>
<td>41006</td>
<td>2438</td>
<td>1480</td>
<td>2695</td>
<td>2100</td>
<td>2895</td>
<td>2300</td>
<td>4000</td>
<td>2400</td>
</tr>
<tr>
<td>Line 3</td>
<td>38520</td>
<td>2639</td>
<td>1480</td>
<td>2928</td>
<td>2000</td>
<td>3128</td>
<td>2200</td>
<td>4000</td>
<td>2400</td>
</tr>
<tr>
<td></td>
<td>38783</td>
<td>2868</td>
<td>1480</td>
<td>3094</td>
<td>2000</td>
<td>3294</td>
<td>2200</td>
<td>4000</td>
<td>2400</td>
</tr>
<tr>
<td></td>
<td>38991</td>
<td>2401</td>
<td>1480</td>
<td>2683</td>
<td>2000</td>
<td>2883</td>
<td>2200</td>
<td>4200</td>
<td>2400</td>
</tr>
<tr>
<td>CAS3B-11</td>
<td>17180</td>
<td>2400</td>
<td>1480</td>
<td>2667</td>
<td>2100</td>
<td>3000</td>
<td>2300</td>
<td>3900</td>
<td>2400</td>
</tr>
<tr>
<td></td>
<td>17420</td>
<td>2525</td>
<td>1480</td>
<td>2847</td>
<td>2100</td>
<td>3185</td>
<td>2300</td>
<td>4000</td>
<td>2400</td>
</tr>
<tr>
<td></td>
<td>17557</td>
<td>2825</td>
<td>1480</td>
<td>3120</td>
<td>2100</td>
<td>3400</td>
<td>2300</td>
<td>4500</td>
<td>2400</td>
</tr>
</tbody>
</table>
**Figure 2.4** Single-channel seismic PGC05-09 Line-2 after time migration. A linear velocity field from 1480 m/s at seafloor to 2100 m/s at BSR depth was used. Note the noisy quality of the data.

**Figure 2.5** Single-channel seismic PGC05-09 Line-3 after time migration. A linear velocity field from 1480 m/s at seafloor to 2100 m/s at BSR depth was used. Note the noisy quality of the data.
Single channel seismic data from 2004 survey

The SCS data from cruise PGC04-08 gave lower noise data than PGC05-09, because of better weather conditions. The data were processed by a spike or gapped Wiener deconvolution on post stack data with a gap of 10 ms and a filter length of 800 ms (Figure 2.6). The deconvolution was very successful in removing the strong airgun bubble pulse. After deconvolution, FDMIG migration with interval velocities selected from the sonic log of X311 Site U1326 (Table 2.2) was also applied. The migration assumed a constant trace spacing of 10 m. Band-pass Butterworth filtering of 10-20-95-115 Hz was subsequently applied. The migration was also highly successful in removing diffractions to clearly image the seafloor scarps, and in correctly re-positioning dipping reflectors (Figure 2.7).

The locations of each shot for the grid of SCS lines collected during cruise PGC04-08 were interpolated to Easting (X) and Northing (Y) along the ship’s track line. Thus, the ship’s location where the source fired was found by interpolating the time of each shot into the ship’s time along the track. The distance of each shot relative to a previous one was calculated and accumulated. Interpolation was made by a Matlab code developed by T. He, with further details found in He (PhD. Thesis, 2007). The correct position, in Universal Transverse Mercator (UTM) coordinate system, was written into Globe Claritas software headers to be subsequently exported to Kingdom suite software for further manipulation.
Figure 2.6 Seismic images for a section of Line CAS3B-11 before and after deconvolution. A. Original seismic image before deconvolution. B. Seismic image after applying spiking deconvolution with a 10 ms gap and an 800 ms filter length. Notice the significant reduction of the bubble amplitudes, 100 ms below both the sea-floor and the BSR.
Figure 2.7 Seismic images for a section of Line CAS3B-11 before and after time migration. A. Deconvolved seismic data image before applying time migration. B. Seismic image after time migration using linear velocity field from 1480 m/s at seafloor to 2100 m/s at BSR depth. Black ellipses highlight strong diffractions that are reduced by migration. For ellipse 1, the bow-tie effect is completely eliminated. For ellipse 2, the migration defines much more clearly the relation between the BSR and the strong dipping reflectivity immediately beneath it.
2.3 Ocean Bottom Seismometer Relocation

For OBSs, the coordinates of the deployment point on the sea surface are easily obtained. However, the actual rest point on the sea floor is not well known. The horizontal deviation may reach several hundreds of meters depending on the currents and the physical shape of the instrument. Additionally, the depth of the recording device is not known if the position is not known, particularly on the frontal ridge where water depth changes rapidly; as well, although regional water depths have been obtained in a recent swath bathymetry survey (Kelley, 2005 per. commun.), accuracy may be much lower than the desired accuracy for the survey geometry.

For the definition of the survey geometry, the receivers were relocated using the Dalhousie University Matlab software “obsreloc”. This program traces rays and uses traveltimes of the direct arrivals to relocate the OBS. The program uses the direct time, a water velocity profile, an initial assumed water depth for the OBS, and an initial location based on sea-surface deployment and recovery positions. It then calculates an “observed” horizontal distance between the true OBS position and each shot. It then constructs a grid around the deployment position of the OBS, and for each grid position the horizontal distance to all of the shots is simply calculated. The residual difference between the calculated distance and the observed distance and the observed distance is then accumulated over all shots. The grid position with the minimum r.m.s. residual is the estimated position of the OBS.

For the relocation on the frontal ridge, a grid was generated out to 250 m to the north and south of the OBS deployment position, and 250 m to the east and west, with a grid spacing of 2 m. A brief description of additional input information is given below:
Water velocity

A file containing depth (m) and velocity (m/s) for the water column was created. The water velocity profile from 0-1400 m on the mid-slope region was used; (Spence et al., 2002). Conductivity-Temperature-Depth (CTD) data were used to extend the water seismic velocity to depths of up to 2400 m. The average salinity used was 35 ppt, based on a CTD profile recorded by the University of Victoria in the ocean-basin ~ 30 km from the frontal ridge (Spence et al; 2001). Temperature gradients were 0.13°C/100 m for depths from 1400-2300 m (CTD offshore Oregon, collected by the University of Washington, Halvorsen et al., 2002), and 0.02°C/100 m for depths higher that 2300 m (CDT offshore Oregon at a mud volcano, collected by the University of Washington, 2001 (Halvorsen et al., 2002). Medwin’s equation was used to calculate seismic velocity (v) from temperature (t), depth (d) and salinity (s):

\[ v = 1449.2 + 4.6 t - 0.055 t^2 + 0.00029 t^3 + (1.34-0.01 t)(s-35) + 0.016 d \] (Equation 2.1)

The final velocity profile is shown in Figure 2.8.
Figure 2.8 Velocity-depth profile for the water column. Velocimeter data from the mid-slope area, ~30 km from the frontal ridge, was recorded in 2002 to a depth of 1400 mbsf. Deeper velocities were estimated using Medwin’s formula applied to CTD data collected in the ocean basin nearby.

**Shot number and direct time**

A shot table file with the deployment information for each OBS and their corresponding direct time picks was created. General information for the deployment of the OBSs is depicted in Table 2.3. For the purpose of the relocation, no filtering to the wide-angle data was applied for the picking of the direct arrival, since typical zero-phase bandpass filters can generate side lobes prior to the direct arrival time.
Table 2.3 Deployment data of the grid of OBSs. The table includes the coordinates, water depth, and time of deployment for each of the OBSs.

<table>
<thead>
<tr>
<th>OBS Name</th>
<th>A (°)</th>
<th>C (°)</th>
<th>D (°)</th>
<th>G (°)</th>
<th>H (°)</th>
<th>I (°)</th>
<th>L (°)</th>
<th>N (°)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Latitude</td>
<td>Degree</td>
<td>48</td>
<td>48</td>
<td>48</td>
<td>48</td>
<td>48</td>
<td>48</td>
<td>48</td>
</tr>
<tr>
<td></td>
<td>Minute</td>
<td>37.9229</td>
<td>37.2828</td>
<td>37.6324</td>
<td>37.0689</td>
<td>37.3285</td>
<td>37.1521</td>
<td>37.8707</td>
</tr>
<tr>
<td></td>
<td>Minute</td>
<td>3.3494</td>
<td>1.9372</td>
<td>3.0468</td>
<td>2.4221</td>
<td>2.7770</td>
<td>3.2442</td>
<td>2.5565</td>
</tr>
<tr>
<td>Water Depth (m)</td>
<td>1900</td>
<td>1800</td>
<td>1824</td>
<td>1804</td>
<td>1840</td>
<td>1828</td>
<td>1800</td>
<td>1797</td>
</tr>
<tr>
<td>Clock Reset Day</td>
<td>217</td>
<td>217</td>
<td>217</td>
<td>217</td>
<td>217</td>
<td>217</td>
<td>217</td>
<td>217</td>
</tr>
<tr>
<td>Hour</td>
<td>23</td>
<td>0</td>
<td>0</td>
<td>23</td>
<td>22</td>
<td>22</td>
<td>21</td>
<td>21</td>
</tr>
<tr>
<td>Minute</td>
<td>32</td>
<td>51</td>
<td>10</td>
<td>12</td>
<td>44</td>
<td>22</td>
<td>23</td>
<td>23</td>
</tr>
</tbody>
</table>

| Latitude | Degree | 48 | 48 | 48 | 48 | 48 | 48 | 48 |
|          | Minute | 37.7768 | 37.0544 | 37.5134 | 36.8754 | 37.2438 | 37.1819 | 37.7414 | 37.5277 |
|          | Minute | 3.2990 | 1.9846 | 2.9625 | 2.4370 | 2.7490 | 2.9998 | 2.4891 | 2.2991 |
| Final Calibration Day | 220 | 219 | 220 | 219 | 219 | 219 | 220 | 219 |
|                  Hour | 20 | 20 | 0 | 21 | 21 | 23 | 0 | 22 |
|                 Minute | 21 | 29 | 9 | 11 | 35 | 8 | 16 | 35 |
| Total Drift (ms) | -6.04 | 1.66 | 8.36 | -0.53 | -2.64 | 7.57 | -0.79 | 4.87 |
| Drift Rate (ms/day) | -2.1 | 0.91 | 4.18 | -0.27 | -1.35 | 3.72 | -0.37 | 2.38 |
Output data

The relocation process outputs the estimated position of the re-located OBS, the reference or deployment OBS position, and the root-mean-square of the difference between the range of assumed location of OBS from each shot and that of the estimated OBS location from each shot.

After processing, a diagram of contours of root-mean-square (r.m.s.) differences between the range of the assumed locations of all shots and that of the estimated locations of all shots was generated. Examples of the results for OBS-A and OBS-H are shown in Figure 2.9 and 2.10.

Examples of the distance residuals for OBSs A and H are shown in Figure 2.11 and 2.12 respectively. These OBSs show relatively uniform r.m.s. residuals through all the lines. The middle-distance offsets are the most reliable. In general it is observed that the residuals tend to increase at farther offsets, and also at locations just over the OBS. This is because rays from sources over an OBS arrive at near vertical incidence. Such rays place good constraints on the depth of the instrument, but they provide poor constraints on horizontal position. In contrast, distant sources arrive more parallel to the seafloor; these ray paths constrain the horizontal position of the instrument but provide little constraint on OBS depth.
Figure 2.9 Re-located position of OBS-A. The relocated position is a red circle-cross located at the minimum distance residual (11.6 m), and the old position (based on deployment and recovery location) is a blue star within an r.m.s. contour of 40 m. The OBS drift distance is the total distance the OBS was moved from the deployment location.
**Figure 2.10** Re-located position of OBS-H. The relocated position is a red circle-cross at the minimum distance residual (7.6 m), and the old position is a blue star just outside the 40 m r.m.s. contour. OBS-H relocation has the smallest r.m.s. error of all the OBS set.
Figure 2.11 OBS-A residuals. The average residual over all lines recorded by OBS-A is 11 m for a depth of 1923 m (see Figure 2.9 for more details). Line 2 crosses over this OBS. For each line, arrows show the shot closest to OBS-A.
Figure 2.12 OBS-H residuals. The average residual is 7 m for a depth of 1970 m and a drift of 56.3 m (see Figure 2.10 for more details). Line 3 crosses through this OBS. Arrows show the position of OBS-H.

The routine does not invert for water depth, which must be specified as an input parameter. The depth of an instrument is partially constrained by bathymetry data. The sources nearly above the OBS are not sensitive to horizontal position, so these sources were not included in the relocation process to avoid additional sources of error. However,
the calculation was still found to be very sensitive to changes in depth. A set of different depths in the range of the OBS deployment position (based on the bathymetry data) was tested to find the depth with minimum residual. The results show that the depth sensitivity is ± 5 m for a depth range of ± 60 m. The position with the lowest residuals resulting from this procedure was selected as the best fit OBS location in x, y and z.

The eight OBSs were relocated by means of this methodology. The final results of relocation are shown in Figure 2.13 and the numerical values in table 2.4. The relocation procedure for OBS I was not successful, perhaps due to an unidentified timing problem; however, no profile that included this OBS was modelled in this thesis.

**Table 2.4** Relocation data for the OBSs grid. The table includes the best-fit coordinates, water depth for smaller residuals, the final drift and residuals.

<table>
<thead>
<tr>
<th>OBS NAME</th>
<th>RELOCATION DATA</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Latitude</td>
</tr>
<tr>
<td></td>
<td>Degree</td>
</tr>
<tr>
<td>A</td>
<td>48</td>
</tr>
<tr>
<td>D</td>
<td>48</td>
</tr>
<tr>
<td>H</td>
<td>48</td>
</tr>
<tr>
<td>G</td>
<td>48</td>
</tr>
<tr>
<td>L</td>
<td>48</td>
</tr>
<tr>
<td>N</td>
<td>48</td>
</tr>
<tr>
<td>C</td>
<td>48</td>
</tr>
<tr>
<td>I</td>
<td>48</td>
</tr>
</tbody>
</table>
Figure 2.13 OBS relocation. Red dots represent the deployment position of the OBSs and yellow dots the relocated position. OBS-D is located at the Site U1326 drilled by IODP X311. Note the scarps on the top of the structure.

2.3.1 Errors on the relocation

According with Table 2.4, there are two main groups of OBSs with similar residuals. The first group consists of OBSs A, D, G and H. This group is located on the crest of the ridge. The residuals for these OBSs average 10 m, which correspond to time shifts of 15 ms. This time shift is close to but slightly larger than picking error of ± 10 ms. The other group of OBSs with similar residuals consists of OBSs L, N and C. These
OBSs have high rms residuals in comparison with those located to the west (group 1). The average distance residual is 30 m. This corresponds to a traveltime residual of >40 ms, which is several times larger than the error estimated for the picking of the direct arrival (~10 ms) (see Table 2.4). Three possible sources of error may have affected the relocation of these OBSs:

1. One of the disadvantages of this relocation software is the lack of inversion for depth. The topography of the study region is locally rugged with scarps crosscutting the seafloor. Lateral changes in depth are greatest for the OBSs located to the north-east (L, N, and C) for which the depth can change by ~60 m in 100 m lateral distance.

2. Another possible source of error is in the measured travel times for the instruments. The internal clocks of some OBSs had a large drift (e.g. ~8 ms for OBSs I; Table 2.3). Drift rates were assumed to be linear, but the drift rate may have changed during the survey.

3. In addition to time drift, the identification and picking of the direct arrival has the largest potential source of error in the relocation. In the frontal ridge region, refractions with apparent velocities slightly higher than water velocity may arrive just before the direct arrival, which makes it difficult to follow the direct arrival. This was specifically confirmed for OBS A during the later interpretation stage, and improved traveltime residuals were obtained after re-identification and picking of the direct arrival. Initial interpretation after high pass filtering did not give good results.
2.4 OBS Data Processing

Refraction and wide-angle reflection data recording from three-component geophone OBSs were processed using spike deconvolution and band-pass filtering. Only OBS vertical components were used in the modeling since this component measures the compressional velocity of the section in the Z direction. To suppress the bubble pulse effects, a predictive statistical spiking deconvolution was applied (PSDECON Claritas module). PSDECON acts as a constrained trace-by-trace filter that results in smaller spatial and lower signal/noise ratio without damaging weaker signals. The autocorrelation was mixed over 13 traces, and a 700 ms filter length with a gap length of 10 ms was applied for a design-gate window of 1200-4000 ms.

The amplitude spectrum of the direct arrivals extends from about 15 Hz to a high frequency limit of ~50 Hz. Butterworth band-pass filtering of 5-15-50-100 Hz was applied to all OBS data. Frequency spectra for OBS-D and OBS-L are shown in Figures 2.14 and 2.15.

![Amplitude Spectrum](image)

**Figure 2.14** Mean amplitude-frequency spectrum OBS-D over a time window from 1495 to 1742 ms and a RECODNUM range of 4896-4900.
Figure 2.15 Mean amplitude-frequency spectrum at OBS-L over a time window from 1812 to 2260 ms and a RECODNUM range of 3208-3212 (see Figure 2.17).

The deconvolution process results were not as successful as hoped on some OBSs. In particular, the refractions on OBS D (Figure 2.16) are still very ringy, with 100 ms oscillations presumably produced by the bubble pulses. However, OBS L (Figure 2.17) shows very clear refractions, with apparent velocities as high as 4 km/s and little evidence of bubble pulses.
Figure 2.16 Recorded section for vertical-geophone on OBS-D. Butterworth band-pass filtering process applied to OBS-D. (A) Raw data, (B) Data after band-pass filtering (5-15-50-100 Hz) and spiking deconvolution. Black arrows point to refraction events.
Figure 2.17 Recorded section for vertical-geophone on OBS-L. Butterworth band-pass filtering process applied to OBS-L. (A) Raw data, (B) Data after band-pass filtering (5-15-50-100 Hz) and spiking deconvolution. Notice the clear refractions before and after filtering. Black arrows point to refraction events.
3. METHODOLOGY: RAY TRACING DATA ANALYSES

The velocity structure of the frontal ridge was determined using the traveltine inversion routine of Zelt and Smith (1992). The 3D velocity structure was estimated by determining 2D velocity models along 3 profiles, each containing several OBSs. For each profile, the model was constrained by traveltimes from refractions and wide-angle reflections recorded on the OBSs, and from vertical-incidence reflections recorded on coincident single-channel seismic lines. Velocity model construction was an interactive process that combined both inversion and forward modeling.

3.1 Ray Tracing Method

The RAYINVR algorithm of Zelt and Smith (1992) traces rays in 2-D media for rapid forward modeling and inversion of refraction and reflection travel-times. The velocity model is composed of a sequence of layers that may be dipping separated by boundaries at variable depths. The velocity structure within a layer is defined by velocity values specified at arbitrary x-coordinates along the top and bottom of the layer. The depths of the upper boundary of a layer are also specified at arbitrary x.

For the purposes of ray tracing, the model is automatically broken up into an irregular network of trapezoids, each one with dipping upper and lower boundaries and vertical left and right sides. The velocities at the four corners of the trapezoid are used to interpolate a velocity field within the trapezoid, so that the velocity varies linearly along its four sides. Therefore, horizontal as well as vertical velocity gradients may exist within a trapezoid. Velocities could vary during the inversion, but the vertical velocity gradient was fixed in all layers during error analyses. A constant vertical velocity gradient
increases the resolution in places where there is insufficient ray coverage to independently determine an upper and lower layer velocity or layer thickness (Zelt and Smith, 1992).

The sources and receivers may be positioned anywhere in the model. The RAYINVR routine was originally designed for a land-based refraction survey, where one shot is recorded by many receivers. For a marine refraction survey where a single OBS records many surface shots, the modeling is accomplished simply by interchanging shots and receivers, which is permitted because of the principle of reciprocity: an OBS on the seafloor is considered as a source, while the shots at the sea surface are treated as receivers. For this model only P-waves were analyzed; no converted S-waves were included in the modeling. The seafloor and BSR phases were modeled as reflected waves, and the other phases were modeled as head waves. Further details are explained in the next section.

A plot of the model and all rays traced is produced along with a plot of reduced travel time versus distance for the observed and calculated data. Two-point ray tracing was used in this modeling. The travel times correspond to any ray path which was traced and they can be either first or later arrivals. The partial derivatives of travel-time, with respect to the model parameters selected for adjustment, are calculated analytically during ray tracing; these parameters include velocities and the vertical position of boundary nodes. The travel time residuals with respect to the observed data are also calculated. The partial derivatives and travel time residuals are output to separate files and they are used later as input to the program DMPLSTSQR which updates the model.
parameters by applying the method of damped least-squares to the linearized inverse problem (Zelt and Smith, 1992).

Realistic 2D earth models can be represented by a minimum number of model parameters, i.e. the number and position of parameters specifying each layer can be adapted to the subsurface ray coverage. Layer boundaries, including the surface, may be horizontal (one parameter) or consist of numerous straight-line segments. A layer may have a constant velocity (one parameter), or the velocity structure may be defined by many upper and lower layer velocity points as in the present work. Different velocity points are specified above and below a layer boundary due to small velocity discontinuities in some areas (Zelt and Smith, 1992).

In this work a modified version of the software RAYINVR was obtained through Dalhousie University (Louden K., pers. commun. 2006). In this version, the fixed format input/output files were changed to allow specification of the velocity model to the nearest metre for depth nodes (rather than to the nearest 10 m in the original version) and to the nearest m/s for velocities (rather than the nearest 10 m/s).

3.2 Data description and seismic events

Wide-angle seismic data

A sample OBS dataset is shown for each of the three modeled profiles: OBS D along SCS Line 2 (Figure 3.1), OBS L along SCS Line 3 (Figure 3.2) and OBS L along SCS Line 8 (Figure 3.3). The first event identified (P1) corresponds to the direct arrival at the seafloor, marking the top of Layer L1. This event is sometimes difficult to identify and pick because of interference with other events, particularly shallow refractions at larger offsets.
Figure 3.1 (a) Reflected (vertical orange raypaths) and refracted rays traveling along a profile through OBS A, D, H and G. Green rays corresponds to Layer1/Layer2 interface, orange rays to the base of the layer 3 (BSR) and blue rays to layer 4/Layer 5 interface. (b) Vertical component of OBS D data along SCS Line 2 and (c) picked arrivals (solid lines) superimposed on (b), with the same color scheme as for the rays in (a). White band in the seismic data corresponds to a gap zone in the data due to air-gun failure.
Figure 3.2 (a) Reflected (vertical raypaths) and refracted rays traveling along a profile through OBS L, N and C. Green rays correspond to Layer1/Layer2 interface, orange rays to the base of the layer 3 (BSR) and blue rays to layer 4/Layer 5 interface and magenta rays to layer 5/Layer 6 interface. (b) Vertical component of OBS L data along SCS Line3 and (c) picked arrivals (solid lines) superimposed on (b), with the same color scheme as for the rays in (a).
Figure 3.3 (a) Refracted and reflected rays traveling along a profile through OBS D and L. Green rays corresponds to Layer1/Layer2 interface, orange rays to the base of the layer 3 (BSR) and blue rays to layer 4/Layer 5 interface and magenta rays to layer 5/Layer 6 interface. (b) Vertical component of OBS L data along SCS Line8 and (c) picked arrivals (solid lines) superimposed on (b), with the same color scheme as for the rays in (a).
An early event (P2) from shallow depths (<100 m), below the seafloor and above the BSR, can be traced in all the OBS data sets. With an apparent velocity of ~2.0 km/s, this phase is identified as the refraction at the top of a high velocity layer. P2 is seen out to maximum offsets of about 1.5 km to the NW and SE of the OBSs on Line 3 (Figure 3.2), ~1.5 km to the NW to the OBSs on Line 8 (Figure 3.3) and for about 5.8 km along Line 2 (Figure 3.1). No seismic event was identified for delimiting the base of L2. No refracted arrival from the BSR was observed; this is expected since it is negative velocity change downward, not allowing a refracted phase from the BSR event.

Strong high-velocity refracted arrivals, referred to as P4 and P5, were prominent on most OBSs (Figures 3.1, 3.2, 3.3). These events provide evidence for the presence of a prominent seismic discontinuity at greater depths, with an increase in apparent velocity from ~2.4 km/s above P4 to a faster phase of ~4.0 km/s below P5. P4 is evident on all OBS data sets. P5 is observed beneath SCS Line 3 and Line 8 (Figure 3.2 and 3.3), but it is not evident beneath SCS Line-2 on OBS A, D, H and G.

Vertical incidence data

The phase P2 was not identified in the vertical-incidence data. The BSR, referred to as event P3, was identified in the SCS and MCS data (Figures 3.4, 3.5). A very strong BSR is seen below OBS A and D in the MCS data (Line PGC_CAS03 between CDP 2800-3400, see Figure 3.4). There is a strong lateral variation in the appearance of the BSR on SCS Line 3 and SCS line CAS3B-25. The clearest BSR is seen only below OBS C on line CAS3B-25 SHOTID 5800-6200, while it is almost absent below OBS N and L on this line (Figure 3.5). Along the line perpendicular to the margin, Line 8 (Figure 3.6), coherent reflectivity is not seen in the core of the ridge. However, the BSR is observed in
the middle part of the structure, but it is very difficult to observe in both the seaward and landward flanks of the structure.

**Figure 3.4** Multichannel seismic Line PGC_CAS03 after migration. The MCS line is coincident with Line2. Notice that the BSR is strong to the NW, but to the SE it is almost absent.

**Figure 3.5** Migrated single-channel seismic Line CAS3B-25, collected in 2004, is very close to 2005 Line 3. Notice that the BSR is almost absent with the exception of shots 5800-6200.
Figure 3.6 Single-channel seismic CAS3B-X03, collected in 2004, is very close to 2005 Line 8. Notice the strong BSR in the middle part of the structure (after Scherwath et al. 2006)

Reflections corresponding to the deep phases (P4, P5) are also not evident beneath the ridge on the vertical incidence data (SCS and MCS data). At vertical incidence data, two-way time expected for the reflectors are ~3.2 s for P4 and ~3.9 s for P5 (see Figures 3.5, 3.6). These phases were strongest on OBS-L and D along SCS Line 3 and 8.

3.3 Data picking

Refractions and wide-angle reflections were picked on the vertical component of OBS data. The picking accuracy was limited by effects such as the presence of noise, interference with other events, and possible phase shifts at critical incidence angles due to the complex structure of the study region.

The wide-angle OBS data were displayed both after a seafloor NMO correction (Figure 3.7) and after static corrections with a linear reduction velocity, typically 4 km/s
(Figure 3.8). To help the BSR identification, the zero-offset OBS data were compared to the vertical incidence reflection data at the location of the OBS. The portion of the MCS data lying between two OBSs was inserted between the positive offset half of one OBS dataset and the negative offset half of the other, such that the seafloor arrivals on the MCS and OBS data sets coincided (Figure 3.7 and 3.8). Hence, the subsequent reflection events near zero offset align automatically. The BSR reflection was identified in most of the wide-angle data sets at near offsets, but continuation of the event to large offsets was often difficult.

![Diagram](image)

**Figure 3.7** Seafloor NMO-corrected wide-angle data along SCS Line 2 and MCS line PGC_CAS03. Note the correlation of BSR between both data sets. The BSR is indicated by the orange line. Apparent velocities of about 2.0 and 2.4 km/s for the first arrivals are shown by the green and blue lines respectively.

Shallow reflectors above the BSR are hard to identify in the wide-angle data because of the bubble pulse, which make their correlation with the SCS data questionable. No events other than P2 were evident above the BSR. It is also important to mention that increasing the number of layers also results in larger uncertainty in velocity estimation (Korenaga et al., 1997).
Examples of picks for the refracted phases P2, P4 and P5 are shown in Figures 3.1, 3.2 and 3.3. Based on the dominant frequency of the data, uncertainties for the OBS picks were typically ~20 ms, except for Line3 (Figure 3.8) where the uncertainty for the BSR was 35 ms. An uncertainty of 20 ms was assigned to the MCS travel-time picks. Details about the time uncertainties are shown in Table 3.1. Picking was done manually following the positive seafloor reflection and the negative BSR reflection along the first waveform zero-crossing of the traces.

![Figure 3.8 Wide-angle data from OBS L and C along Line 3 and SCS data from nearby CAS3B-25. The BSR (orange line) is identified by the correlation of wide-angle and near-vertical data at zero offset.](image)

**Table 3.1** Picking uncertainties for wide-angle arrivals.

<table>
<thead>
<tr>
<th>Phase</th>
<th>Line2 (ms)</th>
<th>Line3 (ms)</th>
<th>Line8 (ms)</th>
</tr>
</thead>
<tbody>
<tr>
<td>P1 (direct arrival)</td>
<td>10</td>
<td>10</td>
<td>10</td>
</tr>
<tr>
<td>P2 (shallow event)</td>
<td>20</td>
<td>18</td>
<td>18</td>
</tr>
<tr>
<td>P3 (BSR)</td>
<td>30</td>
<td>35</td>
<td>20</td>
</tr>
<tr>
<td>P4</td>
<td>20</td>
<td>18</td>
<td>18</td>
</tr>
<tr>
<td>P5</td>
<td>20</td>
<td>18</td>
<td>18</td>
</tr>
</tbody>
</table>
3.4 Forward and inverse modeling of travel-time

Forward modeling refers to tracing the rays through a defined velocity model. Inversion refers to modifying the velocity model automatically, so that the misfit between the observed (travel-time phase picks) and theoretical travel-time curves is minimized. For the forward modeling and especially the inverse modeling, the selection of appropriate parameters is critical (Zelt and Smith, 1992). The inversion can produce model parameters such as strong positive or negative velocity gradients or steep layer boundary dips that are unrealistic from a physical point of view. The solution may contain many local minima in traveltome residuals, which can prevent the inversion from finding better solutions. This implies that a starting model must lie close to the optimum model, and this is achieved by forward modeling. Considerable forward modeling work was done before the inversion could work properly.

3.4.1 Modeling procedure

Iteration between inverse modeling of wide-angle traveltome data and forward modeling of vertical-incidence data was done. The SCS vertical-incidence data provided no direct constraints on velocities, but their spatially dense sampling supplies clear structural information (Korenaga et al., 1997). Vertical incidence data were used only for the seafloor and the BSR reflector; no more near-offset reflectors were consistently identifiable on both vertical incidence data and wide-angle data.

The modeling included seven OBSs along 3 seismic lines. The inversion was first done for two profiles parallel to the margin. The first profile (western side of the frontal ridge data set) has constraints from SCS Line2, CAS3B_Line-11, and from MCS PGC_CAS3B; this model is referred to as model 1 (M1). The second profile (eastern
side) has constraints from SCS Line-3 and CAS3B_Line-25, and the model is referred to
as model 2 (M2). Along SCS Line-8, perpendicular to the margin, seafloor depths and
structures varied much more rapidly than along the line parallel to the margin. To obtain
model 3 (M3) along Line-8, depths and velocities were fixed at the two intersections
(OBS D and L) with the lines parallel to the margin.

The initial velocity model of the upper, lower and compacted sediments was
constructed based on reference data from IODP X311 (sonic log site U1326, Figure 3.9),
on apparent velocities from the wide-angle data, and also on previous velocity models
from Yuan et al. (1996, 1999) at the ODP Sites 888-889 and from Chen (2006) along the
IODP X311 transect.

Figure 3.9 Site U1326 sonic log.

In forward and inverse modeling, the refracted arrivals were modeled as pure
head waves. In the algorithm of Zelt and Smith (1992), head waves are much easier to
use than turning waves. Additionally, head waves are reasonable to use because the
modeling is based on travel-times and not on amplitudes. For small velocity gradients,
traveltimes for head waves are nearly equal to traveltimes for turning waves. This is
particularly true for the OBS dataset, where the refractions are observed over only a small offset range and so they contain little information on velocity gradients.

Ray tracing with its corresponding error bars for the observed and calculated travel-time examples are shown in Figures 3.10, 3.11, and 3.12 along with the models M1, M2, and M3 respectively. In general the models show good agreement between the calculated and observed travel-times. At a given location, velocity is specified at the top and bottom of a layer. The velocity gradient at a given location was fixed in the inversion, but the layer thickness and average layer velocity were allowed to vary.

Phase P2, with an apparent velocity of ~2.0 km/s, was modeled as a head wave at the top of layer L2. It also controls the thickness of layer L1, the layer below the seafloor. P2 was modeled with a starting velocity of 1.52-1.53 km/s for layer L1 in all the models. This event was distinct and coherent enough to be identified on the OBS gathers, but confident correlation of this phase with the SCS and MCS data was not possible (Figures 3.1-3.6). Since there is not an event corresponding to the base of L2, no constraint exists about the thickness and bottom velocity of L2. The thickness of L2 (30 m) and velocity at the top of L3 (2.0 km/s) were selected by comparison with the IODP sonic data (Figure 3.9). The 30 m thickness of this layer is not constrained in this modeling since head waves are used at the interface of the different layers and there is no control on the base of this layer.

The phase corresponding to the BSR (P3) was modeled as reflections from the base of layer L3. Joint forward and inverse modeling of vertical-incidence and wide-angle travel-times was done for this layer. Velocities at the top and bottom of L3 were allowed to vary. A low velocity of ~1.6 km/s below the BSR (top of layer L4) was assumed in the starting modeling,
since velocities near this value are found in the U1326 sonic log (Figure 3.9) and these are interpreted as free gas below the BSR. The seismic travel-times for phases P3 and P4 do not provide direct constraints on this low velocity layer.

**Figure 3.10** Observed travel-times (vertical bars) compared with calculated travel-time (solid lines) along MCS line PGC_CAS03 and SCS Line 2. (a) Refracted and reflected rays (vertical rays) traveling along a profile through OBS A, D, H and G. Green rays correspond to Layer1/Layer2 interface, orange rays to the base of the layer 3 (BSR), blue rays to layer 4/Layer 5 interface and magenta rays to layer 5/Layer 6 interface. (b) observed and calculated travel-times for wide-angle data. (c) observed and calculated travel-times for vertical incidence data.
Deeper phases from below the BSR (P4, P5) were identified on OBSs L, N, C (M2), D and L (M3). These events are consistently high amplitude first arrivals.

Figure 3.11 Observed travel-times (vertical bars) compared with calculated travel-time (solid lines) along SCS line CAS3B-25 and SCS Line 2. (a) Refracted and reflected rays (vertical rays) traveling along a profile through OBS L, N and C. Green rays correspond to Layer1/Layer2 interface, orange rays to the base of the layer 3 (BSR), blue rays to layer 4/Layer 5 interface and magenta rays to layer 5/Layer 6 interface. (b) observed and calculated travel-times for wide-angle data, (c) observed and calculated travel-times for vertical incidence data.
Figure 3.12 Observed travel-times (bars) compared with calculated travel-time (solid lines) along SCS Line 8. (a) Refracted and reflected rays traveling along a profile through OBS D and L. Green rays correspond to Layer1/Layer2 interface, orange rays to the base of the layer 3 (BSR), blue rays to layer 4/Layer 5 interface and magenta rays to layer 5/Layer 6 interface. (b) observed and calculated travel-times for wide-angle data, (c) observed and calculated travel-times for vertical incidence data.
Information about the reliability of the final velocity model is given by the r.m.s. traveltime-misfit and the $\chi^2$ normalized misfit parameter provided by RAYINVR (Bevington 1969, Zelt and Smith, 1992, Zelt, 1999). Table 3.2 shows the results for each phase. A value of $\chi^2$ equal to 1 indicates that the observed data have been fit to within their assigned uncertainties (Ritzmann et al., 2002). The $\chi^2$ values for individual arrival types (refractions from, L2, L4, L5, and reflections from L3) and each data type (wide-angle and near-vertical) were monitored separately such that none would be significantly greater than unity. As a result the overall $\chi^2$ is close to 1 for each line, with values ranging from ~0.50 to 1.04 (Table 3.2). For the deeper layer, L5, $\chi^2$ is relatively small. This result is not unexpected because of the lack of control for this layer since the P5 event was recorded over only a short offset range and also only from a single direction.

<table>
<thead>
<tr>
<th>Model</th>
<th>Phase</th>
<th>Points used</th>
<th>RMS travel-time residuals (ms)</th>
<th>$\chi^2$ error</th>
</tr>
</thead>
<tbody>
<tr>
<td>M1</td>
<td>P2</td>
<td>248</td>
<td>25</td>
<td>1.038</td>
</tr>
<tr>
<td></td>
<td>P3</td>
<td>706</td>
<td>24</td>
<td>1.547</td>
</tr>
<tr>
<td></td>
<td>P4</td>
<td>196</td>
<td>24</td>
<td>0.624</td>
</tr>
<tr>
<td></td>
<td>BSR(vert. inc.)</td>
<td>48</td>
<td>21</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Total</td>
<td>1198</td>
<td>18</td>
<td>0.958</td>
</tr>
<tr>
<td>M2</td>
<td>P2</td>
<td>317</td>
<td>15</td>
<td>0.828</td>
</tr>
<tr>
<td></td>
<td>P3</td>
<td>354</td>
<td>19</td>
<td>0.865</td>
</tr>
<tr>
<td></td>
<td>P4</td>
<td>229</td>
<td>16</td>
<td>1.019</td>
</tr>
<tr>
<td></td>
<td>P5</td>
<td>220</td>
<td>11</td>
<td>0.446</td>
</tr>
<tr>
<td></td>
<td>BSR(vert. inc.)</td>
<td>52</td>
<td>22</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Total</td>
<td>1172</td>
<td>16</td>
<td>0.802</td>
</tr>
<tr>
<td>M3</td>
<td>P2</td>
<td>219</td>
<td>16</td>
<td>0.946</td>
</tr>
<tr>
<td></td>
<td>P3</td>
<td>251</td>
<td>18</td>
<td>0.784</td>
</tr>
<tr>
<td></td>
<td>P4</td>
<td>84</td>
<td>14</td>
<td>0.818</td>
</tr>
<tr>
<td></td>
<td>P5</td>
<td>56</td>
<td>12</td>
<td>0.558</td>
</tr>
<tr>
<td></td>
<td>BSR(vert. inc.)</td>
<td>15</td>
<td>18</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Total</td>
<td>625</td>
<td>16</td>
<td>0.829</td>
</tr>
</tbody>
</table>
3.5 Error Analyses

In order to estimate uncertainty in the solution, a sensitivity analysis was performed by perturbing the depth parameters (Katzman et al., 1994). For a given layer, the depth of the layer base was perturbed by constant shifts of ± 10, 20, and 30 m; the interface depth and other parameters were held fixed, and an inversion was performed for the layer velocity. The RMS travel time residuals increase as the magnitude of the perturbation becomes larger. As observed in Figure 3.13, the residual/depth curves show clear minima in traveltime residuals indicating the depth range for the best fit. The width of the curves provide approximate estimate of errors on parameters. However, this error analysis is a subjective estimate and not a formal statistical estimate.

An arbitrary error bound of 1 ms for the base of L1, L4 and L5 was determined as the amount that modeled travel-time pick misfits can exceed the minimum travel-time residuals (see Figure 3.13). Due to larger picking error for the BSR, the error bound for the base of L3 was selected slightly higher than the error bands for the other layers. An uncertainty of 3 ms for L3 allows consistency between the previous MCS data and sonic log estimates of BSR depth.

The depth error is ± 10 m for L1, ± 15 m for L3, ± 15 m for L4 and about ± 50-100 m for L5. The BSR depth (depth to base of layer L3) changes from M1 to M2, and the errors are estimated to be higher for M2 since the picking error (~35 ms) and the uncertainty in the correlation of the BSR from OBS data to vertical incidence data are bigger. The velocity error corresponding to the depth changes is automatically generated after inversion, since an increase in depth also requires an increase in velocity to satisfy
the observed data; i.e. for the maximum depth of a given layer, the velocity is also required to be at the maximum.

Figure 3.13 Perturbations to the best fitting model. Depth was varied in increments of 10 m, and the velocities were inverted. The purpose of these perturbations was to explore the range of permissible solutions. An arbitrary range of 1 ms rms residual for layer 1 (a1, a2) and 4 (c1, c2), and 3 ms rms residual for layer 3(b1, b2) was selected to determine the permissible depth range for the different layers (shown by gray bars).
The r.m.s. travel time residual was used as a reasonable guideline to evaluate the uncertainty and find the best fit. The smaller r.m.s. residuals and chi-squared values were selected as the best fit for the different models (Table 3.2). Improved travel-time fits come with the addition of nodes in the model for the different phases. In this work emphasis was put on minimizing the travel-time residuals provided by the program.

According to Zelt and Smith (1992), non-linear shot-receiver geometry and 3D effects increase $\chi^2$. The frontal ridge is clearly a 3D structure, and is also characterized by being highly faulted.

### 3.6 Results: details of velocity models

Five layers corresponding to four different events were modeled and the final models for lines 2, 3 and 8 are shown in figures 3.14, 3.15 and 3.16 respectively. The resultant velocity and thickness of each one is described below:

The average velocity for layer L1 is ~1.52 km/s. The thickness of this layer varies between 80 and 110 mbsf for all the models. For models M2 and M3, depths are larger in the southeastern part of the frontal ridge than in the northwest. Event P2 along the top of L2 is observed for about 5.8 km along M1 (Figure 3.14), ~2.5 km to the NW and SE of the OBSs on M2 (Figure 3.15), and ~ 1.5 km to the SW of the OBSs on M3 (Figure 3.16). No P2 event was clear enough to be identified below OBS A in M1.

Based on the P2 event, the average velocity range at the top of layer L2 is ~1.9-2.0 km/s for the three models. Refraction data provide no control on the thickness of layer L2. It was assumed to be ~30 m thick, based on a comparable high velocity layer observed in well logs from X311 Site U1326 (Figure 3.9).
The average velocity for layer L3 is $1.90 \pm 0.05$ km/s. No major lateral variation is observed. The depth of L3, just above the BSR, ranges from $\sim 240$-265 mbsf for M1 (see Figure 3.14), $\sim 270$-285 mbsf for M2 (see Figure 3.15), and about 250-280 mbsf for M3 (see Figure 3.16). The lateral coverage of the rays for P3 can be observed in Figures 3.3, 3.4, 3.5 for models M1, M2 and M3, respectively. The BSR can be observed for 3 km along M1, 2 km along M2, and for about 1 km along M3.

Deeper phases from below the BSR (P4, P5) provide a crude estimate of the velocity below the BSR, averaging over thicknesses of 400-600 m for P4, and thicknesses of 900-1200 m for P5 in M2 and M3. The average velocity for layer L4 is $\sim 2.0$ km/s and the average thickness is about $420 \pm 50$ m (see Figure 3.14, 3.15 and 3.16). Layer L5 has an average velocity of $\sim 3.2$ km/s and an average thickness of $\sim 1000$ m (see Figure 3.14, 3.15 and 3.16). P5 also controls the top of layer L6 with an average velocity of $\sim 4.0$ km/s. The lateral and vertical velocity gradients of the layers were allowed to change during the inversion modeling but they were fixed during the error analyses.
Figure 3.14 Final velocity model for SCS Line2 and MCS PGC_CAS03. Solid lines indicate the ray coverage for the different interfaces. Note the different scale with Figures 3.15 and 3.16.

Figure 3.15 Final velocity model for Line3. Solid lines indicate the ray coverage for the different interfaces.
Figure 3.16 Final velocity model for Line8. Solid lines indicate the ray trace travel time coverage for the different interfaces. Line 8 crosses the frontal ridge perpendicular to the margin.
4. INTERPRETATION OF SEISMIC VELOCITY STRUCTURE ON THE FRONTAL RIDGE

In this chapter, the seismic velocity structure from OBS traveltime inversion is compared to the structure revealed in coincident seismic reflection lines and to logging results from IODP X311 at Site U1326. To compare the consistency of the final velocity models with the vertical incidence data, the seismic reflection lines were time-migrated using the same velocity profile presented in Chapter 2, table 2.2. For migrated multichannel line PGC_CAS03, boundary depths for velocity model M1 were then superimposed on the depth-converted seismic sections (Figure 4.1). For migrated single-channel line CAS03-25, boundary depth for model M2 was superimposed on the section (Figure 4.2). Perpendicular to the margin, model M3 was converted from depth to time, and superimposed on time-migrated MCS seismic line 89-08 (Figure 4.3).
**Figure 4.1** Final velocity model M1, along time migrated MCS PGC_CAS03 converted to depth. Notice the matching of the BSR for both the velocity model and the MCS. Solid lines indicate the ray coverage for the different interfaces.

**Figure 4.2** Final velocity model M2, along time migrated SCS CAS3B_Line-25 converted to depth. Solid lines indicate the ray coverage for the different interfaces.
Figure 4.3 Final velocity model M3, in two way time, along time-migrated line MCS 89-08. Notice the correlation of both data sets in the middle part of the structure with respect to P2, P3, and P4. Solid lines indicate the ray coverage for the different interfaces. T corresponds to the sequence of turbidites and H to the sequence of hemi-pelagics.
4.1 High velocities and hydrate concentrations at shallow depth

A high velocity layer of 1.95 (±0.5) km/s was observed at 80-110 mbsf (L2). In agreement with these results, X311 logging at Site U1326A found P-wave velocities from wireline logging which showed even higher velocities than the ones estimated in this study with variable values between 1750 and > 2800 m/s (Figure 4.4). Also, a prominent especially high-resistivity section from 72 to 107 mbsf in hole U1326A (Figure 4.5) was found during X311. This interval represents intercalated layers of high (peaks > 4 Ωm) and low resistivities with likely correspondence to thin layers of high and low concentrations of gas hydrates (Riedel et al, 2006). This shallow interval is slightly different at the present model in comparison with the sonic and resistivity log data. This is not unexpected since the borehole data is a local measure while the velocity model gives an average velocity for the region. The small difference in depth for the high velocity layer may be caused by two factors: variable lithology and local faults. The high velocity layer may be controlled by high porosity sand layers intercalated with mud layers and by normal faults acting as migration paths for the free gas to flow to these shallow depths.

In the vertical incidence data, high amplitudes are observed at some locations near the depth of the L1/L2 boundary (e.g. Figure 4.1, SP 3400-3000). This level is estimated to have gas hydrate pore saturation of 60% based on gas hydrate saturation estimates from sonic log and MCS velocities (Chen 2006), and about 30% gas hydrate saturation from the porosity-reduction model at OBS D and OBS L at the intersection points of the velocity models (see discussion in section 4.4). The refraction data and subsequent travel-
time inversion modeling confirm that this high-velocity layer extends laterally throughout the frontal ridge region, out to distances of at least 4 km from the drillhole.

Figure 4.4 Velocity profiles from line intersection points. (a) Intersection of line 2 and 8, located at the drillhole Site U1326. (b) Intersection of lines 3 and 8 (closest OBS to drill site in Line3).
4.2 Depth of BSR

The BSR depth for the western model M1 (SCS Line2, SCS CAS3B-Line11, MCS PGC_CAS03 Figure 3.13 and 4.1) is about 240-260 m. The BSR becomes deeper to the east as observed in M2 (SCS Line3 and CAS3B-Line25 Figure 3.14 and 4.2), and M3 (Figure 3.15 and 4.3), where depths increase to 280 m. Since each velocity model represents a different location on the frontal ridge, the differences among BSR depths results may simply reflect regional variability on the frontal ridge slope to the east. Water depths increase from about 1800-1900 mbsf along M1 to 1850-2100 mbsf along M2. The increase in depth and thus pressure implies that gas hydrates are stable to greater depths.
For a change in depth from 1800 m to 2100 m, the pressure-temperature conditions for the GHSZ (Hyndman et al., 2001) give an increase in BSR depth by about 10 m. The increase in BSR depth may also be due partially to a difference in temperature gradient due to topographic effects because there is more lateral heat loss on the slope than on the flat top of ridges. The BSR depth estimated by travel-time inversion correlates with the depth from the sonic log at Site U1326 (Figure 4.4), but it differs from the depth calculated before the time of drilling from the multichannel data.

For layer L3, located just above the BSR (Figure 4.4), the optimal inverted seismic velocity is higher by about 100 m/s than the average sonic velocities for this layer, although the values agree within the estimated error. The sonic log gives a good estimate of local velocities, and these were used as starting parameters for the velocity modeling in the region surrounding the drillhole. The BSR was also found to be deeper to the east of the structure (M2), and the average velocity in layer L3 above the BSR was slightly lower by ~0.10 km/s. At the drill site, the velocities used to estimate the depth of the BSR during the drilling are smaller than the ones calculated here by travel time inversion (especially for the shallow level here referred as L2), and the lower velocities may be a reason to explain the relatively shallow depth obtained for the MCS BSR.

A significant increase in resistivity to ~5 Ωm just above the BSR at 250-261 mbsf is observed (see Figure 4.5). Initial interpretations suggested this resistivity peak to be caused by free gas, because of the initial depth estimate for the BSR of 234 m from seismic reflection (Riedel et al, 2006). The new velocity information not only predicts a deeper BSR but also locally high velocities of 1.9-2.0 km/s (sonic log), and ~2.0-2.1 km/s (travel-time inversion) at this depth, suggesting the presence of hydrate not free gas in
this zone. The salinity and chlorinity profiles at this site point to a deeper fluid source with a chloride concentration higher than seawater, indicative of low-temperature diagenetic reactions in the deeper parts of the site (Riedel et al, 2006b).

Alternative hypotheses have been suggested for the high velocities found by sonic log velocities in U1326 at the BSR (and possibly below depending on the BSR depth assumed). Similar to Hyndman et al. (2001), Riedel et al. (2005) suggested that it could possibly be explained by the formation of a gas hydrate annulus around the borehole in the gas zone. This might occur because cold-drilling fluids (seawater) lower the sediment temperature back into the stability field of gas hydrate. But it can also mean that the drilling allowed more gas to migrate up to the BSR increasing the concentration of gas hydrates at this level.

4.3 Ocean basin turbidites and hemi-pelagics beneath the frontal ridge

Two phases, P4 and P5, were identified below the BSR. The velocities estimated from these refraction events show that the velocities increases to ~2.4 km/s at 600 mbsf (L4/L5 interface), and to 4.0 km.s at 1700 mbsf (L5/L6 interface).

L4/L5 interface

On model M3, which extends into the deep ocean basin (Figure 4.3), the L4/L5 interface superimposed on the depth-converted seismic line 89-07 (Figure 4.3) correlates with a strong reflection event in the middle of a sequence of turbidites (Figure 4.3, CDP 4600-4750, ~4.0 s). Beneath the ridge this event is also observed at about the same depth below the seafloor as in the deep ocean basin (~600 m; Figure 4.3, CDP 4400-4300, ~3.1-3.3 s). In the multichannel seismic data, layer 4 possesses weak internal reflectors
beneath the ridge while high internal reflectivity is observed in the deep ocean basin. The L4/L5 interface is also observed in M2 for \( \sim 7 \) km along the section (Figure 4.2).

**L5/L6 interface**

Also on model M3 the interface L5/L6 is observed into the deep ocean basin (Figure 4.3, CDP 4650-4850, \( \sim 4.8-4.9 \) s). The L5/L6 interface may correspond with the interface between a rapidly deposited Pleistocene turbidite section and a pre-Pleistocene hemipelagic sediment layer (Figure 4.3), which is relatively transparent on the vertical incidence data (Davis and Hyndman, 1989).

Near the frontal ridge, Waldron (1982) calculated velocity profiles from a set of OBSs along the Cascadia margin. Their OBS 1, located 45 km seaward of the deformation front, and closest to the frontal ridge, shows similar high velocity of \( \sim 4.0 \) km.s at 1200 mbsf (see Figure 4.6). Yuan et al. (1996) suggested that below the upper turbidite sediments in Cascadia Basin, the lower hemipelagic section is distinguished by its higher velocity gradient as seen in OBS 1. This suggests that likely the similar high velocity at 1200 mbsf detected at OBS1 also corresponds to the turbidite/hemi-pelagic boundary.
Figure 4.6 Velocity profiles from OBS L (current survey) and OBS-1 from Waldron (1982). OBS L is located at the interception point between M2 and M3, and OBS -1 is located 45 km seaward from the deformation front.

The L5/L6 interface in model M3 is constrained at a location only 4-5 km from the deformation front. A change in interface depth of 500 m over 40 km corresponds to an average dip of 0.8°. This is not unreasonable, since near the deformation front the lithologic interface dip is ~1.3° (as measured from the MCS data on lines 84-04 and 89-07), and dip decreases seaward. This simple estimation supports the top of hemipelagic sediments section being correlated with the modeled phase P5 which corresponds to the L5/L6 interface.

The high velocities from layer L5 and L6 in the incoming basin section are incorporated into the ridge. Also beneath the ridge, the multichannel seismic data is highly non-reflective (Figure 4.3, CDP 4250-4400, ~4.0 s). This observation agrees with Hyndman and Davis (1989) and Spence et al (1991) in that the incoming sediment
sequence from Cascadia Basin, at the deformation front, is folded and faulted into margin-parallel anticlinal ridges down to a depth close to the oceanic crust. This event also agrees with the observations of Davis and Hyndman (1989) who found that substantial tectonic strain may be absorbed through ductile deformation as well as through folding and faulting, as evident in a lower relatively transparent section of sediments that thickens landward across the deformation front.

4.4 Hydrate concentrations

Gas hydrates replace pore fluid and so increase the sediment P-wave velocity. Hence, velocity information can be used to estimate gas hydrate concentrations. In order to relate the velocity to the amount of hydrates, good constraints on the porosity and velocity of sediments with no hydrate are needed (Katzman et al., 1994), as well as a relation between increase in velocity relative to that reference and hydrate concentration. Empirical porosity-velocity relations in conjunction with an effective porosity reduction model (Hyndman et al., 1993; Yuan et al., 1996) are used here.

The porosity reduction model assumes that gas hydrate in the sediment pore space has replaced the pore fluid and so reduces the sediment effective porosity. Yuan et al. (1996) used this model based on the fact that the P-wave velocity of pure gas hydrate is similar to that of the sediment matrix. Thus, a bulk sediment velocity measurement can be used to calculate an effective, hydrate-excluding porosity from an empirical porosity-velocity relation. The effective porosity is then subtracted from the hydrate-inclusive porosity measurement (or background porosity) to determine gas hydrate saturation, \( S_h \), by using the following equation:
\[ S_h = (\varphi - \varphi_e) / \varphi \]  

(Equation 4.1)

where \( \varphi \) is the gas-hydrate-inclusive, or background porosity, and \( \varphi_e \) is the hydrate-exclusive, or effective porosity, calculated from the porosity-velocity relation.

Hyndman et al. (1993) derived an empirical porosity-velocity relation from measurements on core porosity samples and velocity data from IODP Leg 131, Site 808, in the Nankai accretionary prism, taken from below the base of the GHSZ. The Hyndman et al. (1993) relation is also in agreement with Nankai LWD log data (ODP Leg 196) from the same site. This relation is given below:

\[ \varphi = -1.180 + (8.607/V) - (17.89/V^2) + (13.94/V^3) \]  

(Equation 4.2)

where \( V \) is the velocity in km/s.

The Nankai data are from clastic accretionary prism sediments, similar to those of Cascadia. Additionally, the data from Site 888, drilled seaward of the deformation front in Cascadia Basin during ODP Leg 146 (a site where no-hydrate/no-gas velocities and porosities have been measured) are in good agreement with the porosity-velocity relation established by Hyndman et al. (1993) (Chen, 2006). However, the data from Site 888 are more scattered than the Nankai data, probably because of the presence of sandy turbidites at Cascadia, and hence less precise in establishing a trend.

A reference velocity-depth profile calculated by Chen (2006) from no-hydrate MCS velocities at Site U1326 was used. Velocity profiles from the model at the location of OBS D (located at site U1326) and OBS L were used to form the velocity-depth profile for sediments containing hydrate (see Figure 4.4 for velocity profiles). The Hyndman et al. (1993) porosity-velocity relation (eq. 4.2) was used to calculate the
porosities for sediments with no hydrate over the depth range from 0 to 500 m. The gas hydrate saturation was calculated from equation 4.1.

The resultant gas hydrate saturation calculated by the porosity reduction model is compared with the saturations calculated from MCS velocities by Chen (2006) and the log from X311- Site U1326 (Figure 4.7).

Gas hydrate saturation estimated from the refraction data is largest in layer L2 (80-110 mbsf at OBS D), where values are ~30-38% of the pore space. The MCS gas hydrate saturations are smaller, about 10-25% of pore space. From the sonic log at IODP Site U1326, Chen (2006) used both a rock-physics model and the porosity reduction method to estimate saturations of about 60% of pore space in sand layers. For layer L3, in the interval 120-260 mbsf, saturations increased over the interval with maximum values at the base of L3 of 26% for OBS D and 32% for OBS L.

![Figure 4.7 Gas hydrate saturation calculated by effective porosity reduction. Triangles represent saturations from MCS velocities calculated by Chen (2006), dark circles are calculated along OBS D velocity profile and squares along OBS L velocity profile.](image-url)
The main sources of error in gas hydrate saturation estimates from this method are related to the uncertainty in the reference velocity profile, and the uncertainty in the site-specific porosity-velocity relation.

4.4.1. Calculation of potential resource

Calculations of potential resource of gas hydrates in the frontal ridge study area were made using the estimates of gas hydrate concentration (GH). The resources were calculated using the estimated area (A) in which hydrates are present, the thickness (T) of layers with gas hydrate, the porosity (P), and the factor of expansion of gas hydrates (E):

\[ GH = A \times T \times P \times E \]  \hspace{1cm} (Equation 4.3)

For calculating the area, the BSR was mapped along the frontal ridge in the SCS data set from 2004 (refer to Chapter 5). From the detailed seismic survey, the BSR was clearly identified over an area of \( \sim 8.4 \text{ km}^2 \). The BSR was weak and uncertain within the slide area, along the slump region of the frontal ridge, and to the NW of the structure. The uncertain-BSR extends over an area of \( \sim 4.4 \text{ km}^2 \) (Figure 4.7). L2 was assumed to be present in the area where BSR was identified.

Two layers were used for the calculation: the shallow layer (L2) with a thickness of 30 m, and the layer above the BSR (L3) with an assumed thickness of 120 m. L2 average concentrations are 32% and L3 average concentrations are 22%. The average porosities, based on the average gas hydrate saturation of fraction of pore space calculated by effective porosity reduction is 60% for L2 and 40 % for L3. Finally, the factor E of 164 represents the expansion of 1 \text{ m}^3 of hydrate to conventional gas at standard temperature and pressure conditions (Kvenvolden, 1993).
Figure 4.8 Map of distribution of BSR base on single-channel seismic lines from cruise PGC0408. Green polygons correspond to areas with BSR clearly identified; yellow polygons correspond to areas of uncertain BSR.

The results estimate 1.8 TCF of gas (50.8 billion m$^3$) for the regions with clear BSR identified, and ~0.95 TCF (26.8 billion m$^3$) for the region with uncertain BSR. In an optimistic scenario, assuming the presence of gas hydrates in the region of uncertain BSR, the frontal ridge has a total of ~2.75 TCF (77.6 billion m$^3$) possible gas resources. For a comparison with this estimate, the Canadian annual consumption of conventional gas is ~6.3 TCF (Canadian National Energy Board, www.neb.gc.ca).
5. FRONTAL RIDGE SLOPE FAILURE: FAULT-CONTROL AND POSSIBLE RELATION WITH GAS HYDRATE DISSOCIATION

Submarine landslides on continental slopes are important geologic hazards since they might cause mass wasting, tsunamis, and in regions containing gas hydrates, rapid release of methane (e.g. Paull et al., 1991; Pecher et al., 2005, Mienert et al., 2005). Furthermore, the frontal slope failures are also an important process in the construction of accretionary prism (Davis and Hyndman, 1989).

One of the largest well studied submarine slides is the Storegga Slide, which has been the site of a number of large-volume mass failures (e.g., Solheim et al, 2005). At the Storegga slide, the presence of gas hydrate has been suggested to have induced seafloor instability. Since hydrates probably inhibit sediment compaction, their dissociation in sediment pores is likely to create an under-consolidated state, to decrease seafloor strength, and to facilitate submarine landslides in areas where slopes are steep. The dissociation of gas hydrate may provide: (1) a zone of weakness at the BSR that localizes the glide plane (Dillon et al., 1998), or (2) may cause fluid and sediment liquefaction during or immediately after a slide event (Berndt et al., 2005b). Nevertheless, recent work by Brown et al (2006) does not consider gas hydrate dissociation as a primary cause of slope failures for the Storegga Slide scenario.

Mosher et al (2004) suggest the dissociation of gas hydrates as a possible triggering mechanism for sediment failure of the central Scotian slope, on the eastern Canada margin. In the USA Atlantic continental margin, where nearly 200 continental slope mass wasting features have been mapped, hydrate dissociation was also proposed as the mechanism for triggering marine mass movement (Booth et al., 1994).
On the Vancouver Island continental margin, a number of large slope failures have been identified. Most originate on the frontal anticlinal ridges, where steep slopes and probably elevated pore-fluid pressures cause unstable conditions. The failures are morphologically similar to sub-aerial landslides and slab avalanches of snow (Davis and Hyndman, 1989).

On the frontal ridge near Site 1326 of IODP X311, a large slump is observed in newly acquired multi-beam bathymetry data and in seismic reflection data, just landward of the deformation front. As described in previous sections, the region is characterized by a wide distribution of gas hydrates, based on the presence of a BSR in seismic data and on direct observations in the IODP drill holes.

In this chapter a description of slope failure based on SCS and multi-beam bathymetry data is presented. Some indications of possible triggering mechanisms for sediment failure at the frontal ridge slide are discussed and compared with observations in other settings where gas hydrates are present.

5.1 Data and Observations

This study is based on an EM 300 multi-beam bathymetry data set, processed by D. Glickson from the University of Washington (D. Kelley, pers. comm., 2005) and on single-channel seismic data from cruise PGC0408. A subset of the SCS was interpreted: eleven lines parallel to the margin (Lines CAS3B, 01, 04, 06, 11, 12, 13, 14, 15, 21, 25, 33, see Appendix A) and three lines perpendicular to the margin (Lines CAS3B_X01, X04, X07). For detailed information on the reflection data and processing see Chapter 2.
Additionally, side-scan acoustic sonar data (SeaMARC II) were acquired by the University of Hawaii Institute of Geophysics (1985) in a cooperative project with the Pacific Geoscience Centre (GSC). Only a hard-copy paper image from Davis and Hyndman (1989) is available. The Sea MARC II backscatter data were combined with multi-beam bathymetry data to correlate and identify common events (Figure 5.1). These data are particularly valuable in identifying and in delineating morphological features such as erosional events or constructional channels and slope failures (Davis and Hyndman, 1989).

The slope failure is located in the western side of the frontal ridge in the middle part of the local ridge structure (Figure 5.1). The head-wall and side-walls of the slump in the bathymetry data set can also be identified in the sidescan data. Larger blocks of material remain intact near the base of the slope. The distal portion of the debris spray extends ~ 4-5 km from its source, and is characterized by a homogeneous surface (Figure 5.1). The western slope of the frontal ridge is more affected by gravitational sliding than the eastern slope of the structure. Some alignments (NW-SE) are observed at the SE part of the frontal ridge.

The head wall of the frontal ridge slide is ~250 m high and the slump has eroded the ridge over a section ~2.5 km length. The collapsed structure covers an area of about 3.5 km² as observed in figures 5.1 and 5.2. The construction of a funnel shaped channel at the collapse structure is observed. This channel feature is starting to erode further into the frontal ridge.
Figure 5.1 Multi-beam bathymetry and acoustic side-scan data from the frontal ridge. (a) multi-beam bathymetry map. (b) side-scan image. Dark lines correspond to the head-wall and side-wall of the slump feature. Red circle corresponds to the IODP X311 site U1326. Notice the good correlation of the debris material and the northern side-wall in both data sets.
The SCS data for lines 1-6 pass over the slump. A series of seafloor scarps are observed on nearly all SCS lines (Figure 5.2). The largest seafloor displacements are observed for scarps B and C in figure 5.2. There is almost no headwall reflectivity in the upper 300-400 ms below the slump (Figures 5.2a, 5.2b), and it is difficult to pick a BSR on these lines although a faint BSR may be present (see Appendix A.1, A.2, A.3). On CAS3B_01 and CAS3B_02 some bright reflectors are seen within the slump area at depths just below the expected BSR (e.g. CAS3B_01 SP 19400-19360, at about 2.85-2.95 s, see Appendix A.1). Initially, this reflectivity was thought to be the expression of free gas below the BSR, but comparison with cross-line CAS3B_X07 and the bathymetric map shows that this bright reflector is likely the result of side-swipe from the steep NW slope of the slide feature.

From CAS3B_11 to CAS3B_33, at the eastern part of the structure, the BSR is easily identified in the region surrounding the slump (Figure 5.2c-f). The BSR is not continuous through the slump, and the picking uncertainties for this part (uncertain-BSR area in Figure 5.3) are up to ~30-40 ms. There is a wide distribution of bright reflectors below the BSR in the region on the NW side of the structure. Some examples of these events can be observed in: CAS3B_04 (SP 2038-2044, 2.73-2.85 s, Appendix A.2), CAS3B_11 (SP 17380-17280, 2.7-2.8 s, Appendix A.4), CAS3B_12 (SP 18620-18540, 2.67-2.8 s, Figure 5.3c, Appendix A.5), CAS3B_13 (SP 16680-16770, 2.66-2.8 s, Appendix A.6), CAS3B_14 (SP 17920-18020, 2.7-2.8 s, Appendix A.7), and CAS3B_15 (SP 18820-18930, 2.64-2.85 s, Appendix A.8). Another interesting feature is the presence of high reflectivity also above the BSR, evident in nearly all lines parallel to the margin.
Figure 5.2 Migrated single-channel seismic lines and seafloor scarp, consistent from line to line. (a) SCS line CAS3B_01 (b) SCS line CAS3B_06 (c) SCS line CAS3B_12 (d) SCS line CAS3B_14 (e) SCS line CAS3B_21 (f) SCS line CAS3B_33. Location of the seismic lines is shown in figure 5.4. For larger scale versions of the SCS seismic data set and the fault interpretation see Appendix A.
The BSR was mapped along the frontal ridge in the SCS data set. The BSR was clearly identified over an area of ~ 8.4 km$^2$. The uncertain-BSR (not clearly observed) occurs within the slump region, both seaward and landward of the crest of the ridge, and to the NW of the ridge, and extends over an area of ~ 4.4 km$^2$ (Figure 5.3).

**Figure 5.3** Map of distribution of BSR based on single-channel seismic lines from cruise PGC0408. Green polygons correspond to areas with BSR clearly identified; yellow polygons correspond to areas of uncertain BSR. See Chapter 4 for details about gas hydrate concentrations and probable gas reserves in the area mapped. The red circle identifies the slump on the steep slope of the frontal ridge near IODP Site 1326, just landward of the deformation front (dashed line).
5.2 Results

A set of 13 seafloor scarps, (A, B, C, D, E, F, G, H, I, J, K, L, M) have been identified on the migrated SCS sections. The scarps appear to be seafloor expressions of faults that cut through the accreted sediments of the frontal ridge (Figure 5.2, 5.4). The faults strike in the NE-SW direction, perpendicular to the margin and parallel to the direction of convergence. The faults are interpreted as normal, with extensional motion oriented NW-SE, perpendicular to the direction of compression on the margin. Disturbed sediments are confined to the vicinity of the normal faults. The faults clearly outcrop at the seafloor and can be traced from the surface through the sedimentary section to depths below the BSR in some locations (Figure 5.2, Appendix A). The seafloor traces of faults were mapped based on their seafloor expression in the SCS data set by picking the high point of seafloor scarps, where there is no ambiguity. The fault locations were then superimposed on the detailed bathymetry map (Figure 5.4). A correlation between the seismically and bathymetrically defined faults is clear where seafloor offsets are large (e.g. faults B-G)

Descriptions of fault dips and their seafloor displacement, for the four principal faults (A, B, C, and D) closest to the slide, are shown in Table 5.1. In general, the two faults with the largest seafloor scarps are B and C (Figure 5.2 and 5.4); these bound an interior region of maximum slope failure, while faults A and D bound the outer limits of failure and correlate with the side walls of the faulted region on the frontal ridge (Figure 5.4). The slope failure seems to be mostly controlled by the major faults, A and D.
Table 5.1 Displacements and dips of major faults that bound the slump.

<table>
<thead>
<tr>
<th>Fault Name</th>
<th>D</th>
<th>C</th>
<th>B</th>
<th>A</th>
<th>A</th>
</tr>
</thead>
<tbody>
<tr>
<td>Line number</td>
<td>01</td>
<td>33</td>
<td>01</td>
<td>33</td>
<td>01</td>
</tr>
<tr>
<td>Fault dips</td>
<td>29°</td>
<td>21°</td>
<td>50°</td>
<td>38°</td>
<td>18°</td>
</tr>
<tr>
<td>Seafloor Displacement (m)</td>
<td>22</td>
<td>43</td>
<td>24</td>
<td>72</td>
<td>56</td>
</tr>
</tbody>
</table>

For faults B, C and D displacements increase to east. For fault A, the biggest displacement is for line 01 in the west. The seafloor expression of fault A is barely evident in the east with a very small seafloor displacement at SCS CAS3B_LINE33 (Appendix A.11).

Secondary faults, from E to M, have the smallest displacements at the seafloor. Their seafloor expressions are more evident in the west, while identification is more ambiguous in the eastern part of the structure (see Figures 5.2, 5.4). Some of the secondary faults (from I to M), on the NW part of the slide, show a tendency to be oriented so that they follow slope contours.
Figure 5.4 Bathymetric map displaying the seafloor traces of frontal ridge faults. Red dashed lines correspond to the peaks of the seafloor scarps as identified on the SCS data. Yellow star corresponds to the U1326 from IODP X311. The blue line identifies the top of the slump headwall, which has a slope of $\sim 45^\circ$ compared to a slope of $\sim 12^\circ$ for the frontal ridge north of the slump.

5.3 **Nature of faults**

Most of the primary faults, A to H, strike at high angles to the bathymetric contours of the ridge. Thus these faults cannot be gravity sliding features. However, the secondary
faults, located in the northern part of the structure (I, J, K, L, M), show evidence consistent with gravity sliding, since their orientation tends follow the slope contours.

The primary faults are interpreted as normal faults with fault displacement in the direction of extension (NW-SE). The strongest sub-seafloor evidence of normal faulting is observed in faults C and D (Figure 5.5, Appendix A). For these faults, at ~2.6 s, stratigraphic reflectors are terminated sharply; the terminations are aligned with the seafloor scarps and some displacement is observed. Fault B was extrapolated from the seafloor to depths below the BSR (Figure 5.5). However, fault B is not clearly observed at or above the BSR but some evidence indicates its presence below the BSR. At those depths, some reflector terminations are aligned with the seafloor scarps But the association with the faults that is well defined higher in the section is not certain.

The secondary faults are more uncertain. These faults were extrapolated from the seafloor scarps to reflector terminations or small offsets. However, in a set of many short reflectors, in an anticline shape, it might difficult to identify accurately the corresponding alignment in depth for these faults (Figure 5.2).

The primary faults A and D (see Figure 5.2, 5.4) indicate that the lateral extent of slumping is fault-controlled. This observation is strongly supported by the seismic and bathymetry data, since faults A and D align closely with the sidewalls of the slump feature and thus clearly bound the region of slope failure on the frontal ridge (see also Appendix A.1, A.2, A.3).
Figure 5.5 Migrated single-channel seismic lines CASS3B-13 and 15 and seafloor scarps aligned with depth. Location of the seismic lines is shown in figure 5.4. For larger scale of the SCS seismic data set and fault interpretation see Appendix A.

This pattern of faults has not been observed in other locations along the Cascadia margin. As well, no information was found of comparable scarps and normal faults on other subducting zone margins. However, it is probable that the appropriate data sets (SCS and MCS, parallel to the margins) have not been collected along other margins.

5.4 Interpretation and discussion of the slump

In this section, several mechanisms that could have triggered the frontal ridge slope failure are discussed: (1) gas hydrate dissociation (e.g. Bugge et al., 1987; Mienert et al., 2002; Berndt et al., 2005a), (2) pore pressure fluid expulsion (e.g. Eichhubl et al; 2000, Bunz et al; 2003), (3) salinity elevation in faults, and (4) earthquakes.
Gas hydrate dissociation

At the frontal ridge, gas hydrate is widely distributed based on the presence of strong BSRs in the area surrounding the slump feature (see Figure 5.3). This is an indicator that dissociation of gas hydrates may be associated with the slope failure (e.g. Bugge et al., 1987; Mienert et al., 2002, Berndt et al, 2005a). Below the slump feature there is only a weak indication for the presence of the BSR. That is, the slump likely destroyed the BSR and there has been insufficient time since the slump occurred for the BSR to re-establish itself.

At the Storegga Slide, Berndt et al. (2005a) suggested that the coincidence of the depth of disturbed sediments and the base of the gas hydrate stability is evidence that the sediment disturbance is related to gas hydrates. However, Brown et al. (2006) carried out a careful re-calculation for depth of BSR and showed that, assuming steady-state conditions, the BSR (at ~ 300 m depth) would have been ~50 m deeper than the glide plane at the time of slope failure. They argue that the base of the gas hydrate stability zone, and any gas that may have been present, likely played only a minor role, if any, in slide initiation at that locale. However, their analysis is dependent on assumptions about ocean temperatures and seafloor depths at the time of slumping ~7500 years ago, and these are considerable uncertainties in the assumptions.

On the Vancouver Island frontal ridge, the depth of the glide plane for the slide was estimated from the detailed swath bathymetry data. First, a reconstruction of the old ridge, before the slump event, was made by following the predominant trends of the contours in order to simulate the original topography of the structure (Figure 5.6). A set of 3 depth-profiles perpendicular to the margin was extracted from the reconstruction and
from the current bathymetry (Figure 5.7): through the slump (B-B’), to the north of the slump (C-C’), and to the south of the slump (A-A’).

Figure 5.6 Frontal ridge reconstruction. (a) Present bathymetry of the frontal ridge (see Figure 5.7 for cross-sections). (b) Old frontal ridge reconstruction before the slump.
Figure 5.7 Depth profiles along different transects through the ridge. See figure 5.6 for the location of the depth profiles. Dark grey region correspond to the mass of material removed during the slump. Note the BSR, identified by a light grey bar (range of depth of 240-270 mbsf) and a dashed line (BSR average depth 255 m). BSR depth is based on the frontal ridge velocity model described in Chapter 3.

Estimates of the depth of the slump were made by overlapping the depth profile B-B' with both A-A' and C-C'. These two are the best reference for estimating the original surface of the slump, since they are close to both sides of the slump feature. The correlation with A-A' and C-C' gives the amount of material removed, a volume of ~0.62 km$^3$ with a mid-slump thickness of about 255 m. This calculation assumes that no material removed during the slide accumulated at the toe of the frontal ridge, and thus it is a conservative minimum estimate for the amount of material removed.
The basis of this hypothesis is that the sediments above the BSR are cemented by the gas hydrate and thus strengthened. In contrast, the sediments below the BSR contain free gas and are weak since no hydrate cements them. In order to test the hypothesis of the BSR acting as a glide plane for the slump failure, the BSR was traced over B-B’ (Figure 5.7). In this case an average depth of 255 m for the BSR, estimated in the velocity models from Chapter 3, was used. This simple estimate does not account for the increase in BSR depth with increasing water depth (in the case of laterally uniform heat flow), but it gives a good approximation of the depth of the BSR along the slump before and/or after the slump. The depth of the BSR is coincident with the base of the slump for ~0.5 km along the depth profile B-B’-relative to A-A’, and ~1 km along depth profile B-B’-relative to C-C’.

Pore pressure difference

In areas with high fluid flux, increased fluid seepage may result in increased pore fluid pressure. High fluid pressure, at the same time, may also increase slope instability and promote slumping and canyon incision (Orange and Breen, 1992). This behavior was observed at the Goleta head slump in Southern California, where increased seepage follows fault and anticline trends, suggesting a structural control on mass wasting imposed by upward fluid migration (Eichhubl et al, 2000).

The fluid flux at the Cascadia margin may be larger than on many other subduction margins. This is because of the great thickness of incoming sediments, and because most or all of the sediment section is scraped off and deformed at the deformation front (Davis et al, 1990). Deformation modeling indicates greatest fluid flux just landward of deformation front (Hyndman et al., 1993).
The sediment porosities at the Cascadia margin increase from ~40% in the deep ocean basin to ~45-50% at the frontal ridge as inferred from velocities by Yuan et al. (1994) at subbottom depths of ~100-300 m. Although velocities increase in the footwall sediments near the primary faults of the accretionary prism, they decrease significantly in the hanging wall; this implies an increase in porosity of the hanging wall sediments, including sediments of the frontal ridge. The high porosities were associated with undercompaction of the sediments (Yuan et al. 1994).

Sediment undercompaction is likely associated with high pore pressure at the frontal ridge. High pore pressure along the faults may locally induce anomalously higher pore pressure conditions in the immediately surrounding sediments. This can lead to slope failure since it may reduce the critical angle for slope stability to less than the angle of the steep frontal ridge slope (Davis et al. 1983; Davis and Hyndman 1990).

Fluid flow might have been more localized along faults A and D controlling the location of slump. However, the intensity of fluid flow might have varied along the collapse structure. Direct estimates of pore pressure at the frontal ridge have not been made to probe its action as the triggering mechanism for the frontal ridge slope failure. However, an indirect measure of sediment compaction, and thus pore pressure, is available from shear strength data.

Shear strength data from 0 to 120 mbsf collected at Site U1326 of IODP show an decrease with depth. Shear strength ranges from 5 kPa in sands to 300 kPa in clay, being high in the shallow subsurface. To measure the consolidation stage of the sediments, the ratio of shear strength to overburden pressure was calculated (see Figure 5.8).
Figure 5.8 Ratio of shear strength to overburden pressure from IODP X311 Site 1326. This provides information on the degree of sediment underconsolidation or overconsolidation. The ratio was calculated using automated vane shear (AVS) in comparison with handheld Torvane shear strength measurements from site U1326. The overburden pressure was calculated from MAD bulk density data (from IODP X311 Scientific results).

Shear strength to overburden pressure ratio >0.25 indicates that the sediments are over-consolidated for their depth below the seafloor. This high ratio (>0.25) was observed in the uppermost 20 m of Hole U1326C. A ratio <0.25 was observed from 20 m to 120 m (see Figure 5.6). The low ratio (underconsolidated sediments), at the lowermost 100 m of the section, indicates the weakness of the material near the crest of the frontal ridge. This may indicate that the sediments are susceptible to failure, particularly where bathymetric slopes are steep.
Shear strength to overburden pressure ratios from site U1325, U1327, U1328 and U1329 at the IODP X311 transect (landward direction) and site 888 from ODP leg 146 at the ocean basin also show underconsolidated sediments at same depths. However, the areas corresponding to these sites are not characterized for steep bathymetric slopes as at site U1326. Furthermore, significant recent sedimentation at several of these sites (e.g. U1325, U1327, 888) may contribute to high pore pressure in the sediments at these locations.

*Salinity elevation in faults*

An interesting feature of the data is the presence of high reflectivity above the BSR (see Figure 5.9). Above the GHSZ, it is expected that the methane is transformed to gas hydrate. However, the high reflection amplitudes above the BSR, on-lapping at the faults, could indicate the presence of free gas within the GHSZ. These high amplitudes may also indicate the presence of gas hydrate in sand layers. With just single-channel data, it is seismically difficult to distinguish between free gas and gas hydrate for these shallow and local reflections.

Recently, a multicomponent, multiphase, fluid and heat flow model was developed by Liu and Flemings (2007) in order to study the formation of hydrate in marine sediments. This model shows that, in zones of high gas flux, free gas supplied from depth forms hydrate, depletes water, and elevates salinity until the pore water becomes too saline to allow further hydrate formation. This event increases the instability of the material around the faults, since gas hydrates are not cementing the sediments.
Figure 5.9 Seismic image for a section of SCS Line CAS3B_11. Notice the presence of high reflection amplitudes above and below the BSR. Black ellipse highlights bright reflectors onlapping at the faults above the BSR.

For the occurrence of instability, it is necessary to have a high flux of gas. High fluid flux is expected at accretionary wedges in convergent margin environments. In this type of margin, the rate of upward fluid flow is relatively high, more than few tenths of mm per year (Haacke et al, 2007). Additionally, seawater may percolate down the faults beneath the seafloor. The seawater has very low methane concentration, and when it is in contact with hydrate, the methane in hydrate would diffuse into the seawater and thus promote the disappearance of hydrate near faults.

At the frontal ridge, a source of deep fluids associated with a high flux of gas may be the major thrust fault at the accretionary prism, forming the base (sole) of the frontal ridge (Davis and Hyndman, 1989).
Earthquakes

The Cascadia subduction zone is thought to be capable of generating major earthquakes with moment magnitudes as large as Mw = 9 at an interval of several hundred years (Hyndman, 1995). Although the subduction fault is currently locked, continuous motion of the converging plates produces tectonic loading of the locked segment, leading eventually to earthquake rupture. During the drilling by X311 at site U1326, many occurrences of turbidites were found; this material likely indicates times of active tectonism (Riedel et al., 2006).

In active tectonic margins, such as Cascadia, strong earthquake shaking may cause liquefaction, increasing the pore water pressure so that the sediment of the deformed structure loses shear resistance. This has been observed in other tectonic settings in California (Harp and Wilson, 1995), the Apulian platform of Southern Italy (Spalluto et al., 2007), and Taiwan (Weissel et al., 2001).

An earthquake is the most likely trigger mechanism for the frontal ridge slope failure on the Cascadia margin. This contrasts with the Storegga slide region, also gas hydrate related, where earthquakes are much less frequent and the maximum earthquake size probably is two orders of magnitude smaller (Mw=7; Atakan and Ojeda, 2005) than on the Cascadia margin.

5.5 Summary

The frontal ridge slope failure is clearly fault-controlled, bounded by the major faults A on the southeast and D on the northwest. The faults strike in the NE-SW direction, perpendicular to the margin and parallel to the direction of convergence. This implies that
the faults are normal, with displacements in the direction of least compressive or extensional stress. The triggering mechanism for the slope failure is possibly a combination of various effects. The mechanisms analyzed include gas hydrate dissociation, high pore pressure fluid expulsion along the faults, and salinity elevation in faults, which inhibits the formation of gas hydrates along the faults. Because of frequent earthquakes on the Cascadia margin, an earthquake is the most probable mechanism that triggered the frontal ridge slope failure. An earthquake may induce initial slope failure, which can not only start gas hydrate dissociation but also increase fluid expulsion by producing high pore pressure through transient unloading. It is not clear if margin perpendicular fault occur in other margins. However, if that is the case, they will add another component to the complexity of deformation in frontal ridge regions.
6. CONCLUSIONS AND RECOMMENDATIONS

The study of hydrate in the northern Cascadia accretionary prism frontal ridge region, that was the focus of this study, is important because it is characterized as the tectonically youngest gas hydrate occurrence on the Northern Cascadia Margin. The conclusions and recommendations of the present work are divided in two groups: velocity analyses and frontal ridge slope failure.

6.1 Velocity analyses

Travel-time inversion and modeling of wide-angle and vertical incidence data were performed to estimate the 2-D average velocity model of the frontal ridge on the Northern Cascadia Margin. An anomalous high velocity layer (L2) of $\sim 1.95 \pm 0.05$ km/s (between a layer above with velocity 1.52 km/s and a layer below with velocity from 1.70 km/s to 2.1 km/s) was identified at shallow depths of 80 to 110 mbsf. The correspondence of the high velocity layer with sonic log and resistivity data strongly suggests that high concentrations of gas hydrates are present throughout the frontal ridge region at shallow depths.

Travel-time analysis also indicates that the BSR lies 250 to 280 m below the seafloor and is overlain by a high velocity layer with an average velocity of $1.90 \pm 0.05$ km/s and with the gradient indicated above. These results agree from those from IODP Expedition 311 and from Chen (2006), who analyzed sonic logs and MCS data to obtain velocities of $\sim 1.90$ km/s just above the BSR with a depth of $\sim 260$ m on the frontal ridge at Site U1326.
Deep velocity structure in the ocean basin was also evaluated since two phases P4 and P5 were identified below the BSR. P4 shows velocities of \(~2.4\) km/s at about 600-800 mbsf (top of layer L5). P5 is present at depths up to 1.7 km with average velocities of 4.0 km/s (top of layer L6). In the Cascadia Basin, the interface between L5 and L6 lies at a depth that corresponds with the boundary between a rapidly deposited Pleistocene turbidite section and an underlying pre-Pleistocene hemipelagic sediment layer. The L5/L6 interface is also observed beneath the frontal ridge at equivalent sub-seafloor depths in the multichannel seismic data. This event implies that the incoming sediment sequence of turbidites/hemi-pelagics are incorporated conformally with the folding and faulting of the turbidite section.

Gas hydrate saturation above the BSR was estimated using a simple porosity-reduction model. The highest gas hydrate saturations of \(~30-38\)% of pore space were obtained at the top of layer L2 (80-110 mbsf). From the sonic log at IODP Site U1326, the saturation at depths of 50-75 m is about 60\% of pore space. In layer L3, in the interval 120-260 mbsf, saturations varied from 26\% for OBS D in model M1, to 32\% for OBS L in model M2.

The BSR was mapped over an area of \(~8.4\) km\(^2\). Less certain BSR along the slump region of the frontal ridge, and to the NW of the structure, was also identified. It extends over an area of \(~4.4\) km\(^2\). Gas resources were calculated based on the previous saturations and estimated areas. The results indicate 1.80 TCF of gas for the regions with a clearly-identified BSR, and \(~0.95\) TCF for the region with an uncertain BSR, for a total of \(~2.75\) TCF probable gas reserves in the frontal ridge.
6.2 Margin-perpendicular normal faults

The strike of the faults, perpendicular to the ridge (NE-SW), is parallel to the direction of tectonic compression. The eight larger faults, (A to H) are interpreted as normal faults. They strike at high angles to the bathymetric contours of the ridge and hence are inconsistent with gravity sliding behavior. Five secondary faults, located in the northern part of the structure (I, J, K, L, M), tend to follow the slope contours and thus may be associated with gravity sliding.

This pattern of faults has not been observed in other locations along the Cascadia margin. No information was found of comparable faults and scarps on other subducting zone margins. However, if that is the case, they will add another component to the complexity of deformation in frontal ridge regions.

6.3 Frontal ridge slope failure

A collapse structure observed in multi-beam bathymetry data and in seismic reflection data at the frontal ridge is fault-controlled. The region of slope failure on the frontal ridge is clearly bounded by two major faults that are associated with large 25-75 m high seafloor scarps.

At the slope failure, the distal portion of the debris spray extends ~ 4-5 km from its source. The head wall of the frontal ridge slide is ~250 m high and the slump has eroded a section of ~2.5 km long into the ridge. The collapsed structure covers an area of about 3.5 km$^2$.

The triggering mechanism for the slope failure is possibly a combination of various effects such as:
*gas hydrate dissociation:* The basis of this hypothesis is that the sediments above the BSR are cemented by the gas hydrate and thus strengthened in contrast to sediments below the BSR. Dissociation of gas hydrates may weaken the sediments above the BSR allowing them to fail. In this scenario, the base of the slump is the BSR which will act as a basal detachment zone for the slump failure.

*high pore pressure:* High pore pressure fluid expulsion along the faults may locally induce anomalously high pore pressure conditions in the immediately surrounding sediments. This can lead to slope failure since it may reduce the critical angle for slope stability to less than the slope of the steep frontal ridge face.

*salinity elevation in faults:* High gas flux, supplied from depth, forms hydrate, depletes water, and elevates salinity until the pore water becomes too saline to allow further hydrate formation. This process decreases the stability of the material around the faults, since gas hydrates are not cementing the sediments.

*earthquakes:* An earthquake may induce initial slope failure, which can not only start gas hydrate dissociation but also increase pore-fluid pressure. For these reasons an earthquake is a likely trigger factor for the frontal ridge slope failure and the base of the gas hydrate stability zone is likely to focus failure and slip.

6.4 Recommendations for future studies at the frontal ridge

Inversion of a set of lines perpendicular to the ones modeled in this work is necessary to generate a regional velocity map of the frontal ridge. The three models presented do not provide enough information to interpolate the velocity to a regional scale since there is only one intersection point between profiles.
3-D inversion of refraction and reflection data will improve the understanding of the velocity structure in the frontal ridge providing P-wave velocity constraints on the methane hydrate stability zone. For this routine it is recommended to use a combination of refraction, wide-angle reflection, and normal-incidence seismic data to provide good independent constraints on seismic velocities and interface depths.

A set of well sites in the frontal ridge to target the shallow high velocity layer will help to estimate the lateral extent, thickness and variability of this layer. Drilling to depths well below the free gas zone will help determine a better reference velocity for the GHSZ.

Shear wave analyses will provide more detail of the velocities directly related with gas hydrates. Compliance measurements made at specific frequencies are particularly sensitive to changes in sediment shear velocity, and will give good estimates of physical properties of the material in which the gas hydrates are found (Willoughby and Edwards, 1997).

It is recommended to determine the salinity along the principle faults (A and D) up to BSR depth, and compare these results with the salinities outside the fault area. The salinity may be estimated by the use of spontaneous potential (SP) logs and electromagnetic induction instruments, and may be measured in cores retrieved under in-situ pressure and temperature conditions. This information would help to understand the formation of gas hydrates in the regions surrounding the faults. Additionally, measurement of methane expulsion at the seafloor scarps in the frontal ridge will give information about fluid flow migration along the faults and possible relations with gas hydrate dissociation.
Collection of single channel and/or multichannel seismic data seaward (SW) of the slump feature will allow evaluation of the evolution of slump event. These data will also allow evaluation of the nature and evolution of the faults to the SW of the structure. Additionally, piston core data in the slump region will help to estimate the age of the slide and the evolution of gas hydrates within the slump event.

There is a series of slump features observed in the bathymetry data that extends to regions further from the frontal ridge. Analysis of these features, in the region surrounding the frontal ridge, will help understand the local tectonic environments of the region and the possible relation of slope failure to gas hydrate dissociation in the Northern Cascadia Margin. It is also recommended to compare the frontal ridge slope failure with other fault-controlled slides in different settings. Particularly promising is the method employed in Figure 5.7 to assess whether failure penetrates to the base of the gas hydrate stability zone at other locations.
REFERENCES


Program, Scientific Results, 164, 179-191.


APPENDIX

Single Channel Seismic Lines

This section presents the set of single channel seismic data from cruise PGC04-08, (SEICOMAG), collected in 2004. The SCS data set was migrated as presented in Chapter Two. The migrated seismic reflection data image a set of normal faults that clearly outcrop at the seafloor.

The faults were traced from the surface through the sedimentary section to depths below the BSR in some locations. The 12 identified faults were named alphabetically from A (SE) to M (NW). Thirteen seismic lines are presented here. The first 10 are parallel to the margin, and are shown from A.1 (CAS3B_01) to A.10 (CAS3B_33). The last 3 Figures, from A.11 (CAS3B_X01) to A.13 (CAS3B_X07) correspond to the SCS lines perpendicular to the margin. The dashed lines in the interpretation of the faults indicate a level of uncertainty in sub-seafloor fault location. For the location of the SCS data set refer to Figure 4.3, Chapter 4.
Figure A.1 Single channel line CAS3B_01 migrated. Dashed lines correspond to faults crosscutting the seafloor.
Figure A.4 Single channel line CASB 11 migrated. Dashed lines correspond to faults crossing the seafloor.
Figure A.6 Single channel line CAS3B 13 migrated. Dashed lines correspond to faults crosscutting the seafloor.
Figure A.8 Single channel line CAS3B_15 migrated. Dashed lines correspond to faults crosscutting the seafloor.
Figure A.9 Single channel line CAS3B_21 migrated. Dashed lines correspond to faults crosscutting the seafloor.
Figure A.10 Single channel line CAS3B.25 migrated. Dashed lines correspond to faults intersecting the seafloor.
Figure A.11 Single channel line CAS3B, 33 migrated. Dashed lines correspond to faults crosscutting the seafloor.
Figure A.13 Single channel line CAS3B X04 migrated.
Figure A.14 Single channel line CAS3B_X07 migrated. The apparent seafloor reflections at 2.75-2.8 s for SP 1500-1900 are reflection from the steep sidewall just to the north of this line.