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Seismic Traveltime Inversion of Wide-Angle Data for
Strongly-Varying Structure: Central Chilean Margin and
the Subducting Juan Fernandez Ridge

by

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A Thesis Submitted
In Partial Fulfillment of the
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Master of Arts

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MAY, 2001
Abstract

Seismic Traveltime Inversion of Wide-Angle Data for Strongly-Varying Structure:
Central Chilean Margin and Subducting Juan Fernandez Ridge

By

Julia V. Naumenko

This thesis presents the results of traveltime inversion of seismic wide-angle reflection/refraction data for strongly varying media. The two-dimensional velocity structure along two lines across the central Chilean margin (near Valparaiso) has been obtained and assessed in terms of resolution, uncertainties, and non-uniqueness. The traveltime inversion method was used to model the data and assess the model reliability. A tomographic approach was used to assess the objectivity of the structures in the final models. The final models include slope sediments, the Valparaiso forearc basin, subducting sediments, an accretionary wedge, upper and lower continental crust, a two-layer oceanic crust, and uppermost mantle. The thesis results were compared with the results of forward modeling of the same dataset and with results of analogous seismic surveys across convergent margins worldwide.
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Introduction

1.1 Seismic Wide-Angle Technique for Crustal Studies

The refraction/wide-angle reflection seismic technique is widely used to study large-scale velocity variations in the Earth's crust and upper mantle. Wide-angle seismic data cover a larger angular ray-path than other types of seismic data, such as seismic reflection data, and thereby may better constrain subsurface velocity structure.

Wide-angle seismic data may be modeled using forward and inverse techniques. However, seismic traveltime inversion provides more reliable and time-efficient results compared to trial and error forward modeling. About ten years ago conventional forward-modeling methods were mostly used to interpret crustal seismic wide-angle data. However, in the past decade advances in instrument technology and numerical methods allow much larger data sets to be collected, particularly in the marine environment where densely spaced airgun shots are commonly used. For very large datasets, the increase in data quantity has surpassed the capabilities of the forward-modeling technique, whereas seismic traveltime inversion methods benefit from the increased information content. The most important advantages of an inverse method include the ability to build simpler (minimum-structure) velocity models that provide an appropriate fit to the data, and the ability to assess the final model in terms of resolution, uncertainty and non-uniqueness (Zelt, 1999).

A number of inversion methods are currently used to invert wide-angle seismic data for velocity and interface structure. Different allowances for model complexity,
model parameterization, source and receiver geometry, and the solution of the forward and inverse problems characterize these techniques.

One of the most common techniques is a simultaneous inversion for velocities and interfaces from refracted and wide-angle reflected traveltime data that was presented by Zelt and Smith (1992). The main advantage of this method is that a priori information can be incorporated by the choice of model parameterization and maintained if desired. Thus, a final model may incorporate geologically feasible lateral and vertical velocity variations, and will integrate all available information. Also, as with the other inversion methods, this technique allows for the estimation of model parameter resolution, uncertainty and non-uniqueness, and for the fitting of according to a specified norm.

1.2 Thesis Objectives

The goal of this study is to create geologically reasonable 2D crustal velocity models for two lines of wide-angle seismic data collected with ocean bottom hydrophones (OBH) across the central Chilean margin. The 2D traveltime inversion method of Zelt and Smith (1992) is employed to generate models that fit the data appropriately and are geologically feasible. For this study, the dataset is viewed as a case study to test the capability of the wide-angle inverse method in the case of strongly varying structure.

Construction of the velocity models begins with an analysis of the OBH data. Several groups of seismic waves that relate to the different structural elements of the
subsurface are selected and identified. These groups include refracted and reflected waves within the slope sediments, margin wedge, the Valparaiso Basin, the oceanic crust, the continental upper crust and the oceanic upper mantle. There are also reflected waves from a mid-continental crustal boundary and the oceanic Moho. All identified seismic phases are inverse modeled using the method of Zelt and Smith (1992) to yield a geologically reliable velocity model for both profiles. Reliability of the final models is examined using the assessment techniques of Zelt (1999).

Thesis research is focused on: 1) applying techniques for evaluating the degree of constraint, parameter resolution, uncertainties and non-uniqueness provided by the OBH data; 2) proper interpretation of seismic phases; 3) comparison of the final models with results of crustal studies of similar margins.

1.3 Outline of Thesis

The thesis consists of seven chapters describing the application of a seismic inversion technique using wide-angle seismic data for a crustal investigation in a complex geological region and an assessment and interpretation of the results.

Chapter 1 introduces the seismic wide-angle inversion technique and its application to large-scale crustal studies. This chapter discusses the motivation for the work presented in this thesis.

The second chapter gives an overview of the area of study. The overview includes the geologic and tectonic setting, and results from previous work in the area.
Theoretical aspects of traveltime inverse modeling are presented in chapter 3. As stated in the thesis objectives, the traveltime inversion method of Zelt and Smith (1992) is applied to derive the velocity models for two wide-angle seismic lines. Another inverse approach – the tomographic method of Zelt and Barton (1998) – is used for assessment of the final models. The two methods, referred to as a minimum-parameter inversion and minimum-structure inversion, respectively, are described in this chapter. The description of the programs that were developed by Zelt and used in this thesis work is presented in the third chapter as well.

The data acquisition and pre-modeling data analysis are given in chapter 4.

Chapter 5 focuses on the inverse modeling of the OBH wide-angle data and evaluating the results. The final velocity models and their interpretation are presented in this chapter. Due to non-uniqueness of the inverse solution, the assessment of the final result is an essential part of inverse modeling. The fifth chapter details the assessment of the final velocity models for both profiles in terms of resolution and uncertainty, using the assessment techniques of both the traveltime inversion method and seismic tomography method.

Chapter 6 discusses the results of the wide-angle seismic inversion for both 2-D OBH lines. Comparison of the final inverse models with previous existing models obtained by forward modeling (Flueh et. al., 1998) is given as well. Also, this chapter presents a comparison of the Chilean margin with other convergent margins worldwide.

The last chapter summarizes the main results of the thesis and suggests directions for future research.
2 Study Area

2.1 Previous Work in Study Area

The region along the continental Chilean margin is located in one of the more active seismic zones of the circum-Pacific belt (Fig.1). This seismicity is due to the convergence of the South American plate with the Nazca oceanic plate in the study area, and with the Antarctic oceanic plate to the south. The great Chilean earthquake of 1960 at latitude \( \sim 38^\circ \) S had a moment magnitude (Mw) of 9.5 and stimulated interest in the structure of this margin as a whole. In the early 1960's bathymetric profiles along the northern Chile margin showed a steep continental slope dropping away from a narrow continental shelf to a depth of 7 km in the trench axis (Hayes, 1966). Fisher and Raitt (1962) analyzed the first refraction data collected along lines seaward of the Chile trench at 23\(^\circ\)S. Hayes (1966) computed several gravity models across the Peru-Chile trench between 2\(^\circ\)N and 53\(^\circ\)S, which provided indirect evidence for high-density mantle near the trench axis. Grow and Bowin (1975) presented a two-dimensional theoretical density model over the Chile trench at 23\(^\circ\)S, which was obtained from thermal and petrologic data and then was compared with the observed gravity data of Hayes (1966) and the refraction data of Fisher and Raitt (1962). This model predicted density anomalies at different depth intervals in the descending lithosphere. The first reflection seismic survey across the Peru-Chile trench was conducted in 1967 (Scholl et al., 1970). Bathymetric work during the Nazca Plate Project (Schweller et al., 1981) provided the first comprehensive morphologic description of the Chile trench. Structural elements such as the Chile trench,
Figure 1. a) Free-air gravity map shows the Chilean margin and the Juan Fernandez Ridge intersecting the Chile Trench in the study area. b) Seismicity map for the period 1980 – 1994 shows enhanced activity where the Juan Fernandez ridge projects beneath the margin. 92% of the source depths (NEIC) within the study area are < 65 km. Plate motion vector from NUVEL-1 model (De Mets et al., 1990)
O'Higgins Seamount (a part of the Juan Fernandez Ridge), and a forearc basin near Valparaiso (named the Valparaiso Basin) were imaged.

In the 1980s and 1990s several studies were carried out in northern Chile, such as earthquake investigations and seismic refraction profiling, to better constrain the shape and structure of the Wadati-Benioff zone. Between 21°S and 24°S, onshore refraction experiments (Wigger et al. 1993) and microearthquake investigations (Comte et al. 1994) constrained seismic velocities on the South American continent.

The study area for this thesis is between 32°S and 34°S. No refraction data was collected in this area until 1995 when scientists from GEOMAR (affiliated with the University of Kiel) led a project called CONDOR (Chilean Offshore Natural Disasters and Ocean environmental Research). During the CONDOR project, seismic wide-angle measurements were made along two lines offshore of central Chile at approximately 32.75°S and 33.5°S (Fig. 2). Multi-channel seismic reflection, high-resolution magnetic and bathymetry surveys were also carried out. The objective of this project was to determine the crustal structure in this section of the Chilean margin, and in particular, to assess the influence of seamount subduction on the margin structure (Flueh 1995).
Figure 2. Location map of major tectonic structures and the two wide-angle OBH profiles in the study area. Bathymetry from von Huene et al. (1997). A star indicates the location of the 1985 Central Chile earthquake (Ms=7.8).
2.2 Geologic and Tectonic Setting

2.2.1 Structure of Convergent Margins Worldwide

According to the theory of plate tectonics, a network of rigid lithospheric plates underlies the earth's surface. These plates are created along the rift zones and consumed and recycled into the asthenosphere at the subduction zones along the trenches. In a subduction zone, a plate plunges beneath an adjacent plate to great depth until pressure and temperature cause it to melt and become part of the asthenosphere. The movement and interactions of lithospheric plates triggers a chain of tectonic and magmatic events. Non-uniform earthquake distribution is important evidence supporting the plate tectonic theory and delineates the presently active plate margins. In the vicinity of plate convergence, earthquake nucleation defines an inclined Wadati-Benioff seismic zone that may extend for several hundred kilometers into the mantle along the subducting oceanic crust.

The surface morphology depends on the type of overriding plate. When convergence occurs between two oceanic plates, the subduction zone is marked by a volcanic island arc, parallel to it, a deep trench (e.g. North-East Pacific region), and commonly a well-defined accretionary wedge. When the overriding plate is a continental one, compression between the plates generates an orogenic belt near the edge of the continental plate. Examples of such belts are the Andes on the South American plate, and Cascadia on the North American plate. The main tectonic units across the collision zone between the oceanic and continental plates are the abyssal plain, the oceanic slope, the trench, the continental slope and continental shelf. The
subsurface structures around subduction zones are characterized by complicated tectonics. As a result of collision an accretionary wedge is formed above the subducting oceanic plate. The accretionary wedge is composed of a mass of continental or oceanic material added to the margin of a continental crust by collision and welding. Worldwide studies of convergent margins show that tectonic complexity is often associated with the subduction of a seamount ridge (Nakanishi et al., 1998; Flueh et al., 1998; McIntosh et al., 1998). The subducting oceanic ridge causes changes in Moho relief, alteration of the oceanic crustal thickness, and deformation of the continental slope such as formation of synclinal structures along the slope.

### 2.2.2 Structure of the Central Chilean Margin

The thesis study is focused on the collision of the central region of Chile (near Valparaiso) with the Nazca plate (32°-34° S). In this region, the Nazca plate subducts beneath the South American plate at an azimuth of ~78° and a relative convergence rate of about 8.4 cm/yr (DeMets et al., 1990). The seismic activity and the wide range of earthquake sizes that have occurred in this section of the Chilean margin coincide with the subduction of the Juan Fernandez Ridge (Fig. 1). The area of the Valparaiso Basin along the central Chilean trench has been recognized as a mature seismic gap with a strong potential for large underthrust earthquakes based on historic seismicity (Kelleher, 1972; McCann et al., 1979; Nishenko, 1985). The epicenter of the March 3, 1985 central Chile earthquake (Mₛ = 7.8, 33.13°S, 71.87°W) is located near the
Valparaiso Basin (Fig. 2) as was the epicenter of the great 1906 earthquake ($M_s = 8.4$, $33.00^\circ$S, $72.00^\circ$W) (Christensen, Ruff, 1986).

In the vicinity of the subducting Juan Fernandez aseismic ridge, there are fundamental changes in the configuration of the Benioff Zone, volcanic arc activity, and the structure of the continental margin. These changes include subduction erosion, enhanced forearc deformation, changes in slab dip angle, seismicity patterns, and seismic coupling characteristics (Vogt et al., 1976). The association of these first order features with the subduction of the Juan Fernandez Ridge is recognized but not understood.

On the basis of previous geological and crustal studies the Andean subduction zone and orogenic structures have segmentary composition (Flueh et al., 1998). Such segments are defined by the dip of the Benioff zone, and the presence or absence of Quaternary volcanism and a central valley. The region of Valparaiso ($32^\circ$-34$^\circ$S) coincides with a transition between the two different segments of the subduction zone (von Huene et al., 1997). Offshore Valparaiso, Chile, the Juan Fernandez Ridge enters the Chile Trench at a typical segment boundary. There is no Quaternary volcanism nor a central valley to the north between $28^\circ$ and $33^\circ$S, little sediment input pond at the trench axis, and the dip of the Benioff zone flattens (Stauder, 1973). In contrast, south of $33^\circ$S volcanism is active, a thicker sediment wedge exists at the trench, and a central valley and a steeply dipping Benioff zone characterize a normal convergent margin. The Juan Fernandez Ridge and its seamounts interrupt an otherwise monotonous Pacific Ocean basin. This 900-km-long low rise supports a chain of eleven seamount groups that extend from the present hot spot west of Alexander
Selkirk Island to the O'Higgins Seamounts (Fig. 1). According to gravity data analysis, the normal oceanic crust entering the Chilean subduction zone includes a well of thicker crust beneath an east-west hot spot chain of seamounts (von Huene et al., 1997). Juan Fernandez Ridge is entirely on the Nazca plate so its subducted end has no mirror image west of the east Pacific Rise as occurs for the Nazca Ridge (Pilger, 1981). Before being subducted, the oceanic plate is flexed and therefore many normal faults appear on the oceanic slope, parallel to the trench axis.

The trench axis changes its trend at latitude 33°S from almost north-south to N25°E. This latitude coincides also with an abrupt increase of sediment thickness in the trench axis from north to south. The first reflection seismic data in this area (Scholl, 1970) show a major barrier to axial sediment transport from the Juan Fernandez Ridge crest. According to Schweller et al. (1981), north of 33°S the sediment fill is only a few hundred meters thick, whereas in the south, it may be more than 1 km thick.

The fracture trend and an elongated ridge on the slope indicate an offset and a change in strike direction of the Juan Fernandez Ridge east of O'Higgins Seamount (Figs. 1 and 2). There, morphology, gravity, and magnetic data indicate a change in strike to N65°E (von Huene et al., 1997). This change in strike is significant to the tectonic history of the central Chilean margin. With an essentially east-west strike and normal plate convergence, the point of ridge subduction remains fixed through time. However, the oblique trend of the Juan Fernandez Ridge relative to the direction of plate convergence results in southward ridge migration. Nazca Ridge migration is inferred to have caused the low dip of the Wadati-Benioff zone in this area (Nur and
Ben Avraham, 1981; Pilgar, 1981). Magnetic anomalies show the nearly orthogonal orientation of the Juan Fernandez chain with respect to the trend of the Cenozoic seafloor spreading anomalies. This relation indicates a bend between the western younger and eastern older partly subducted portions of this ridge (Fig. 3). From the modeling of the seafloor spreading anomalies it follows that the Juan Fernandez Ridge jumped during anomaly 16 at 37.4 Ma. This implies that a 60-km wide lithospheric plate was transferred from the Pacific to the Nazca plate (von Huene et al. 1997). The age of the oceanic plate at the survey area varies in the range of 36-38 Ma. The central Chile volcanic gap extends about 570 km north of the current position of the Juan Fernandez Ridge crest beneath the arc.

The continental crust extends to the middle-lower slope boundary, which is reflected by morphology and magnetic data. In the area south of the Juan Fernandez Ridge, the continental crust is undeformed whereas near the ridge it is deformed as seen in the Valparaiso Basin. Here, according to magnetic modeling (von Huene et al., 1997), a seamount (Papudo) is being subducted (Fig. 3). The Juan Fernandez Ridge deforms the thin crust of the continental margin but landward of the coast its effects are deep seated (Flueh et al., 1998).

On the margin, the most pronounced morphological features are the wide Valparaiso Basin and the San Antonio Canyon (Fig. 2). The Valparaiso Basin is the northernmost forearc basin along the Chilean coast (Gonzalez, 1989). The basin is ~100 km long and ~50 km wide near the center. However, Valparaiso Basin is located in a midslope position, whereas the other forearc basins along the margin are all situated on the uppermost slope close to the coast. The San Antonio Canyon is the
Figure 3. Oceanic and continental magnetic anomalies for International Geomagnetic Reference Field and diurnal variation. Data on continent from Servicio National de Geologia y Minería. Wide-angle profiles are shown by solid black lines. North of Profile 2, there is a magnetic anomaly associated with subducted Papudo Seamount.
most deeply incised one in this region. The zig-zag geometry of the canyon suggests that it is tectonically controlled (Hagen et al., 1996; Thornburg et al., 1990). The presence of a terrace or basin is significant as this type of morphology has been associated with subducted seamounts (Lallemand et al., 1990; von Huene et al., 1995).

2.3 Existing 2D Models for Profiles 1 and 2

Zelt (1996) and Flueh et al. (1998) developed crustal models for the two lines of the same wide-angle OBH data used in this thesis. These profiles extend for approximately 160 km from the shore to 50 km seaward of the Chile trench. Profile 1 lies ~75 km to the south of profile 2, some distance from the Juan Fernandez Ridge, and is called the off-ridge profile (Fig. 2). Profile 2 runs obliquely over the subducting Juan Fernandez Ridge and is called the ridge profile.

The models of Zelt (1996) were derived during the CONDOR cruise using the inverse method of Zelt and Smith (1992). These models are based on a preliminary analysis of a subset of the data; only half of the OBH records were employed in the modeling due to the time limitation of the cruise. Zelt's preliminary models image the major structural elements to a depth of ~25 km, such as a forearc wedge, subducting oceanic crust and the overriding continental crust (Fig. 4). These models reveal sediments in the trench and on the continental slope for both profiles. The sediments are up to 3.5 km thick on profile 1, but only ~1 km on profile 2. The sediment velocities vary from ~1.7 to ~2.9 km/s. On the middle continental slope of profile 2 a synclinal structure known as the Valparaiso Basin is observed. The basin is
Figure 4. Preliminary crustal 2-D models for profile 1 (a) and profile 2 (b) of Zelt (1996). Seismic velocities in km/s. OBH locations used in Zelt’s analysis shown as circled numbers.
approximately 40 km wide and sediment thicknesses are as high as 4 km. A forearc wedge is defined on both profiles above the oceanic crust landward of the trench. The wedge is about 45 km wide and up to 7 km thick with velocities ranging between ~2.3-5.9 km/s. According to Zelt’s modeling, the wedge is terminated by a subvertical interface (at ~100 km model distance, the model “origin” is 0 km), marked by a rapid velocity increase to ~5.3-6.0 km/s. This zone indicates the transition from accretionary wedge material to upper continental crust material (Zelt, 1996). On both profiles an average velocity within the upper continental crust is ~5.6 km/s. The lower continental crust is separated from the upper crust by a sub-horizontal mid-crustal boundary at ~12 km depth on both profiles. Beneath the forearc wedge at model distances between ~80 and ~110 km within the lower continental crust, velocities are ~0.7 km/s higher along profile 2 as compared to profile 1. This difference between the two profiles causes the velocity discontinuity from 5.9 km/s to 5.7 km/s at the mid-crustal boundary at a model distance ~100 km on profile 1. The subducting oceanic crust landward of the trench is up to 2.5 km thicker on profile 2 than on profile 1. The thicker oceanic crust is characterized by lower velocities (5.9-6.4 km/s) on profile 2 as compared to the velocities on profile 1 (~6.2-6.5 km/s) at the same location. The dip of the subducting oceanic plate from the trench axis to a depth of ~25 km is 12° and 16° along profile 1 and profile 2, respectively (i.e. slab has shallower dip to the south). The uppermost mantle is characterized by an average velocity of ~7.9 km/s on both profiles.

Two-dimensional models for profiles 1 and 2 were also developed by Flueh et al. (1998) using the forward modeling techniques of Luetgert (1992). Data from 14
OBHs and 9 land stations were used for profile 1, while for profile 2 a total of 19 OBHs and 7 land stations were incorporated into the modeling (Fig. 5). Based on the land station records, Flueh's models extend ~40 km further east and ~10 km deeper than Zelt's models. Land station records are not available to Rice University, and therefore, are not used in the inverse modeling.

To the west, profile 1 shows a normal oceanic crust with a thickness of 6 km and a clear division into a 2-km-thick layer 2 and layer 3. Velocities increase from 4.5 to 6.0 km/s and from 6.4 to 6.8 km/s within oceanic layers 2 and 3 respectively. A thickness of ~6.5 km for the oceanic crust was constrained by Moho reflections about 2 s below the oceanic basement from multichannel seismic reflection data. A thin sedimentary cover less than 0.2 km thick is observed seaward of the trench. Within the trench, sediment thickness increases to 3.5 km. These sediments are subdivided into two units with velocities of 2.0 km/s in the upper layer and increasing to 3.1 km/s in the lower layer. Situated landward of the trench is a 35 km wide accretionary wedge which is also subdivided into two layers with velocities of 2.0 to 2.3 km/s and 3.0 to 4.0 km/s. At model distances between ~90 and ~120 km, a body with velocities in excess of 5.0 km/s is located at a depth of 2.0 to 3.5 km below the seafloor. This is interpreted as a backstop. It has an abrupt seaward termination at 95 km model distance. Another high velocity body (6.4 to 6.5 km/s) lies east of OBH 13 (~120 km model distance). The two high velocity bodies are underlain by a 45-km-wide and up to 5.0-km thick low velocity zone with a lateral velocity gradient from 4.4 to 5.4 km/s. The low velocities of layer 2 in the oceanic crust increase rapidly to more than 6.0 km/s at about 15 km depth below the seaward termination of the backstop. Further
Figure 5. Crustal models for profile 1 (a) and profile 2 (b) obtained by forward modeling by Flueh et al. (1998). Seismic velocities in km/s. OBHs are numbered.
landward, layers 2 and 3 of the oceanic crust cannot be distinguished, but Moho depth of the down-going plate is clearly defined.

Along profile 2, a thin sedimentary cover is also observed on top of the oceanic crust, which is similar in thickness and velocity to profile 1. Beneath the steep lower slope thin, low-velocity (2.1 to 2.3 km/s) sediment covers a 45 to 50 km wide body. This body has a pronounced vertical and lateral velocity gradient with velocities ranging from 3.0 to 4.0 km/s. At model distance ~100 km, west of the Valparaiso Basin, this accretionary body is terminated against a subvertical interface, where velocities between 5.5 and 6.3 km/s are found. Another subvertical velocity discontinuity is observed at model distance ~140 km, where the basement velocities increase to about 6.7 km/s. In contrast to the Zelt's model, there is no velocity inversion within the continental crust on profile 2. The Moho of the oceanic crust is imaged up to the coast, where it reaches a depth of 38 km. The velocities within the uppermost mantle are ~8.0-8.1 km/s on both profiles.

Comparison between the preliminary models of Zelt and the forward models of Flueh reveals similarities as well as the significant differences in their structures. Both models show the main structures such as the subducted oceanic plate, the layer of the slope and trench sediments, the Valparaiso Basin sediments, the accretionary margin wedge, and the continental crust. However, different velocities imply that the composition of these structures is different. The Zelt's models do not distinguish the two oceanic layers, whereas the Flueh's models show clear division within the oceanic plate. Also, on profile 2 in the Zelt's model, the thickness of the subducted plate varies significantly reaching the maximum of ~7.5 km beneath the margin wedge (~90-100
km model distances) and then sharply decreasing to less than 5 km at model distances greater than ∼115 km. In the Flueh’s model, the oceanic crust thickness is increased beneath the wedge as well, however changes of the thickness are not as significant as in the Zelt’s model. Other observed differences between the compared models are a) the presence of a low-velocity structure between the oceanic plate and the continental margin at the depth of more than ∼15 km on both profiles in the Flueh’s model, whereas this structure is absent in the Zelt’s models; b) existence of the subvertical velocity discontinuities within the continental crust in the forward models, whereas the Zelt’s models do not constrain these boundaries; and c) the absence of the mid-crustal boundary in the Flueh’s models, whereas the continental crust is divided into the upper and lower part by the subhorizontal boundary in the Zelt’s models.

In a summary, despite the subjective nature of the interpretation, the above comparison shows significant differences between the results of forward and inverse modeling, and demonstrating the importance of assessing model reliability.
3 Modeling Methods

Wide-angle seismic techniques are widely used to investigate large-scale velocity variations in the crust and upper mantle. The forward modeling or traveltime inversion methods are commonly used to interpret wide-angle data. The conventional forward-modeling technique is based on two-dimensional ray tracing. The theoretical traveltime response of an inhomogeneous medium is repeatedly compared with observed record sections to decrease the divergence between calculation and observation. This trial-and-error iterative forward-modeling approach can provide a geologically reasonable model in laterally varying media. Forward-modeling algorithms based on asymptotic ray theory (e.g. Cerveny et al., 1977) have been developed and are often used. These algorithms include those of McMechan and Mooney (1980), Cassell (1982), Spence et al. (1984), and Zelt and Ellis (1988).

In recent years, advances in seismic instrument technology and numerical methods have enabled the collection of much larger sets of data. The increase in data quantity has exceeded the capabilities of conventional forward-modeling methods. Using forward modeling it is impossible to estimate model parameter uncertainty, resolution, non-uniqueness and to get an assurance that the data have been fit to minimize a particular norm. Inversion does not have these limitations of forward modeling. In addition, the time required to interpret data is significantly reduced. Inversion methods using first arrivals, and possibly reflected arrivals separately, include Firbas (1987), White (1989), and Lutter et al. (1990). A combined inversion of
all types of data allows more rigorous model assessment and provides simpler final models than that obtained from inverting each type separately (Zelt and Smith 1992; McCaughey and Singh 1997). A joint inversion for both interface depths and velocities allows the complete range of parameter trade-offs to be assessed.

Examples of inversion applications for laterally varying structure using wide-angle travel-times from controlled sources include White and Clowes (1990), Lutter et al. (1990), Hole (1992), Zelt and Smith (1992), McCaughey and Singh (1997), Zelt and Barton (1998), Zhang et al. (1998), and Lutter et al. (1999). A number of refraction traveltime tomography methods have also been developed (e.g. White 1989; Hole 1992; Lutter et al. 1999). However, most commonly used inversion algorithms (all those mentioned above, except the Zelt and Smith (1992) method) have limited effectiveness since they require the velocity and depth model parameters to be uniformly spaced (uniform grid) and do not allow one to perform an inversion for selected parameters while the other parameters are fixed. Due to these limitations, reliable prior geological information cannot be incorporated into the model, and model space cannot be explored rigorously. As a result, a final model may include unnecessary structure, and therefore, be less reliable geologically.

A simultaneous inversion for velocities and interfaces using refracted and wide-angle reflected traveltimes, which is offered by the Zelt and Smith (1992) method, provides a more time efficient, reliable and stable inverse solution. This inverse method is applicable to any set of traveltime data for which forward modeling is possible, regardless of the data quality and shot-receiver geometry. Employing a flexible model parameterization with a minimum number of independent parameters
and time-efficient ray tracing, the method of Zelt and Smith derives a geologically reliable model that fits the observed data appropriately. The principle advantage of simultaneous inversion is that it allows an assessment of the velocity models in terms of spatial resolution, uncertainty and non-uniqueness. These assessment capabilities distinguish the Zelt and Smith inverse method from the rest. Examples of applying the Zelt and Smith simultaneous inversion to refraction and wide-angle reflection data are given by Holbrook et al. (1994, 1999), Hughes et al. (1998), and Christeson et al. (1999).

3.1 Inversion Fundamentals

As in other high-frequency methods, in the ray method the complex wave field is composed of elementary waves such as the refracted waves, waves reflected from individual interfaces, converted waves, multiply reflected waves, etc. These elementary waves can be evaluated along individual rays (Cerveny, 1985). Because the ray method allows direct computing of traveltimes, inversion can be applied to these quantities. This means that the traveltimes of selected phases must be picked from the observed seismic data to allow the appropriate comparison between the observed and calculated traveltimes.

The basic step in the evaluation of the seismic wave field of individual elementary waves is ray tracing. Many ray tracing algorithms are available now. The relationship between the traveltime from a fixed source to a receiver and the velocity and interface model parameters is non-linear. The non-linearity of traveltime inversion
is due to dependence of the ray path on the model. Therefore, for linearization purposes a starting model and an iterative approach is needed. The iterative approach allows for the use of new ray paths and solving for model perturbations at each iteration.

From ray theory, the traveltime \( t \) of a ray along a ray path \( l \) in two dimensions is the integral of a continuous velocity field \( v(x, z) \):

\[
t = \int \frac{1}{v(x, z)} dl ,
\]

(1)

In this way, traveltime can be presented as a linear combination of slowness (reciprocal of velocity), although traveltime inversion is still a non-linear problem because the ray path depends on velocity. Using a Taylor series expansion about a starting model and neglecting higher order terms, equation (1) can be linearized (Zelt and Smith 1992). The resulting linear system of equations is written

\[
\delta t = G \delta m ,
\]

(2)

where \( G \) is the partial derivative matrix, which contains the elements \( g_{ij} = \partial t_i / \partial m_j \) (\( t_i \) is the \( i \)th observed traveltime and \( m_j \) is the \( j \)th model parameter), \( \delta t \) is the traveltime residual vector, and \( \delta m \) is the model parameter adjustment vector. The partial derivatives can be calculated while ray tracing through the model using analytical expressions. The model parameters can be represented by a vector \( m \) that contains both the velocities and the interface-depth parameters. The model parameter adjustment vector \( \delta m \) is defined as the perturbation of a model from that used in the previous solution of the forward modeling \( m_0 \)

\[
\delta m = m - m_0 ,
\]

(3)
Similarly a traveltime residual vector $\delta t$ is defined such that

$$\delta t = t_o - t_p,$$

where $t_p$ is the model (predicted) traveltime vector, and $t_o$ is the data (observed) traveltime vector.

Given a starting model, both the traveltime residual vector and the partial derivative matrix are calculated while ray tracing through a model for a particular iteration. Equation (2) is solved for the parameter adjustment vector and in this way used to obtain a new model. If the updated model is better than the previous one, it can be used as an improved starting model. The procedure of computing traveltimes and solving equation (2) is repeated until a satisfactory fit to the data is obtained. Thus, the iterative process consists of two steps: 1) forward modeling to obtain the traveltimes and their partial derivatives, and 2) inverse modeling to solve the linear set of equations to derive a new model.

The generalized least squares solution to the inverse problem $Gm = d$ is provided by

$$m^{est} = [G^TG]^{-1}G^Td,$$

where $G$ is the data kernel matrix that equals the partial derivative matrix for a non-linear problem, $d$ is a vector of the observed data, and $m^{est}$ is a vector of the estimated model parameters (Menke 1984). For non-linear traveltime inversion, equation (5) is written as

$$\delta m = (G^TG)^{-1}G^T\delta t.$$

From inverse theory, based on the correlation between the data and the model parameters, the inverse problem can be classified as an underdetermined,
overdetermined or mixed-determined. The problem is underdetermined when there are more unknown model parameters than data and therefore, the inversion \( Gm = d \) does not provide enough information to determine the model uniquely. In contrast, when there is more information than needed to obtain an exact solution, the problem is overdetermined. Because the inverse problem in practice is usually mixed-determined, equation (6) is solved using a damped least-squares method (Aki and Richards 1980).

In this approach, additional constraints are applied to the inversion solution by minimizing both the misfit of the data and the size of the model parameter adjustments. This minimizing function \( \Phi \), called an objective function, is a combination of the prediction error (data misfit) and the solution length (model parameter adjustments) for the model parameters:

\[
\Phi(\mathbf{m}) = E + \lambda^2 L = \mathbf{e}^T \mathbf{e} + \lambda^2 \mathbf{m}^T \mathbf{m},
\]

where \( \mathbf{e} \) is a vector of the misfit of the data defined as \( \mathbf{e} = \mathbf{d}_{\text{obs}} - \mathbf{d}_{\text{pre}} \) (\( \mathbf{d}_{\text{obs}} \) is the observed data, and \( \mathbf{d}_{\text{pre}} \) is the predicted data), and \( \lambda \) is a weighting factor called the damping parameter that determines the relative importance given to the prediction error \( E \) and solution length \( L \). A large value of \( \lambda \) provides a small perturbation by minimization of the underdetermined part of the solution mostly. However, in this case the overdetermined part of the solution is minimized as well, so the resulting model will introduce errors in the estimation of the true model parameters. A value of \( \lambda \) equal to zero provides minimization of the prediction error \( E \), but does not allow the incorporation of any \textit{a priori} information into the solution to constrain the underdetermined model parameters. In this case the solution will be unstable. Thus, a non-zero value of the damping parameter \( \lambda \) controls the strength of regularization.
(Zelt and Smith 1992), or equivalently, the size of the step length in model space. The compromise value for $\lambda$ is determined by trial and error to yield a solution that suppresses the misfit and takes a reasonably small step in the model space.

The damped least-squares solution to the inverse problem $Gm=d$ is

$$m^{\text{est}} = [G^TG + \lambda^2 I]^{-1}G^Td,$$

where $I$ is an identity matrix. In this way the solution damps the underdeterminacy of the inverse problem by minimizing the overall error that include the prediction error and the solution error (solution length) (Menke 1984).

For the non-linear traveltime inverse problem the damped least-squares solution to equation (2) given by

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

where $C_t$ and $C_m$ are the data and model covariance matrices, which describe $a$ priori estimates of the errors of the traveltime data and model parameters. These matrices can be written as

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

where $\sigma_i$ and $\sigma_j$ are the $a$ priori standard deviations of the error of the $i$-th traveltime and $j$-th model parameter. These covariance matrices allow one to determine how error in the data is mapped into error in the model parameters.

The final model does not present the full solution to an inverse problem since a complete solution must estimate the reliability of each parameter in the final model and obtain a measure of the correlation between the model parameters. These correlation and error statistics are obtained from an analysis of the model resolution matrix. The generalized solution to the inverse problem $Gm=d$ indicates that the
estimate of the model parameters is controlled by some matrix $G^{*g}$ that solves the inverse problem. This matrix is called the generalized inverse. For the least squares problem, as follows from equation (5), the generalized inverse is

$$G^{*g} = [G^T G]^{-1} G^T. \quad (11)$$

Relationship between the estimated model parameters $m^{est} = G^{*g} d_{obs}$ and the true model parameters $m^{true}$ that solves $Gm^{est} = d_{obs}$ can be expressed as

$$m^{est} = G^{*g} d_{obs} = G^{*g} [G m^{true}] = [G^{*g} G] m^{true} = R m^{true}, \quad (12)$$

where $R$ is the model resolution matrix with $M \times M$ dimension ($M$ is the number of model parameters). The model resolution does not depend on the actual values of the data. It depends only on the data kernel matrix $G$ and the a priori information added to the inverse problem. The diagonal elements of the resolution matrix represent the degree of linear dependence of the true model as provided by the estimated model. The row elements of the resolution matrix represent a resolution kernel that characterizes how well the true model parameters can be resolved.

The damped least-squares solution of the non-linear traveltime inverse problem provides an expression for the model resolution matrix $R$ as

$$R = (G^T C_{i}^{-1} G + \lambda C_{m}^{-1})^{-1} G^T C_{i}^{-1} G. \quad (13)$$

The diagonal elements of $R$ are often related to the relative number of rays that sample each model parameter (Zelt and Smith, 1992).

By minimizing the size of the parameter adjustment, the damped least-squares solution provides a new model, which is as close to the previous model at each iteration as possible. When regularization takes the form of minimizing the size if the model perturbation at each iteration (as in the Zelt and Smith method 1992), the
damped least-squares method may produce a rough model. This is because the
smallest perturbation regularization does not include any constraint on the relative
values of adjacent model parameters as with flatness or smoothness constraints.

In contrast to the damped least-squares method, another modeling approach,
that is different from Zelt and Smith (1992), is used to obtain a smooth final model. In
this case, the minimization constraints are applied to the whole model rather than the
model perturbation and in order to minimize the roughness of the solution, the first or
second spatial derivatives are minimized. In this case the misfit (objective) function
\( \Phi(m) \) is given as

\[
\Phi(m) = \delta t^T C_d^{-1} \delta t + \lambda [m^T C_h^{-1} m + s_v m^T C_v^{-1} m],
\]

where \( \delta t \) is the data (traveltime) residual vector; \( m \) is the model vector, \( C_d \) is the data
covariance matrix; \( C_h \) and \( C_v \) are the horizontal and vertical roughening matrices
which contain the 2-D and 1-D second derivative finite difference operators that
estimate the model roughness in the horizontal and vertical directions; \( \lambda \) is the
damping parameter that determines the amount of regularization; and \( s_v \) determines
the relative weight of the vertical and horizontal spatial-derivative regularization. This
approach was developed by Zelt and Barton (1998) and was used for tomographic
model assessment in this thesis.

### 3.2 Traveltime Inversion Method

Zelt and Smith (1992) developed a method of seismic traveltime inversion for
simultaneous determination of 2-D velocity and interface structure. In recent years this
inverse algorithm is one of the most commonly used for the interpretation of the wide-angle reflection/refraction data (Christeson et al., 1999; Holbrook et al., 1994, 1999; Hughes et al., 1998). The strongest characteristics of the Zelt and Smith method are: 1) a flexible (uniform or irregular) model parameterization that allows the incorporation of certain forms of prior information into the model; 2) the ability to include any or all model parameters in the inversion; 3) simultaneous inversion of all types of travelt ime arrivals including reflection and refraction data; 4) final model assessment techniques that estimate model resolution, errors, and non-uniqueness.

Due to the non-linearity of travelt ime inversion, a starting model and iterative approach are necessary. Zelt and Smith’s method uses ray tracing as its forward calculation at each iteration. Based on asymptotic ray theory, rays are traced through the velocity model using a numerical solution of the ray tracing equations. The take-off angles of particular ray group are determined automatically. For a specific model layer, three types of ray groups are modeled: 1) turning rays that refract within the layer; 2) rays that reflect off the bottom of the layer, and 3) head waves from the bottom of the layer.

Effectiveness of a ray tracing algorithm for inverse modeling strongly depends on the model parameterization. Zelt and Smith’ iterative inverse method uses a modification of the model parameterization of Zelt and Ellis (1988) that was proposed for forward modeling. Zelt and Smith use a layered 2-D isotropic velocity model parameterization (Fig. 6). Each layer is divided into variable-sized blocks for the purpose of ray tracing. The layered nature of the model allows the selection of specific layers within which, or layer boundaries at which, turning, reflection and head waves
Figure 6. Example of model parameterization used by the inverse method of Zelt and Smith (1992). The three-layer model is constrained by 20 independent model parameters: 8 boundary nodes (red squares) and 12 velocity nodes (blue circles). The model is automatically divided into 7 trapezoidal blocks for ray tracing purpose.
can occur. Each layer is specified by an arbitrary number and spacing of velocity and boundary nodes (model parameters). Layer thickness may be reduced to zero to model pinch-outs or isolated bodies and thus is an important feature for modeling data from a convergent margin. However, layer boundaries must extend across the whole model from left to right without crossing each other. The complete velocity field distribution is obtained by linear interpolation of the values at the velocity nodes along the boundaries and in vertical directions such that there are no velocity discontinuities within a layer.

Employing the non-uniform parameterization, Zelt and Smith's method is capable of an inverse approach that uses a minimum number of model parameters and seeks a prior-structure model (Zelt 1999). A number of the model parameters depend on the form of the prior information. If prior information consists of relatively simple features such as vertical velocity gradient or discontinuities, the starting model is simple also. In case of laterally homogeneous media, the starting model may be characterized by one node specifying each boundary and one node specifying the velocity at the top and bottom of each layer. In the case of more complex features representing the prior information such as the position and dip of a subducting plate at a convergent margin, the starting model is relatively complex as well.

Since the reliability of the final model depends on the correctness of the traveltime picks, the analysis of the wide-angle traveltime data is a very important pre-modeling step. In Zelt and Smith's algorithm, each traveltime pick is identified by a numerical code that characterizes the phase of the arrival: reflection or refraction for a specific layer. The mis-identification of a phase may lead to gross errors in the final
model. Each ray group traced is identified by the same numerical code to compare the calculated traveltimes with the proper observed traveltimes.

The calculated traveltimes and partial derivatives corresponding to a particular observed receiver location are linearly interpolated across the two closest ray endpoints. During ray tracing, the elements of the matrix of partial derivatives \( G \) that has dimension \( N \times M \) (\( N \) is the number of data, \( M \) is the number of model parameters) are calculated analytically for the model parameters selected for inversion. Zelt and Smith's approach uses damped least-squares inversion (equation 9) for the determination of the updated model parameters of those selected for adjustment, both velocities and boundary nodes simultaneously. After ray tracing, the smallest model perturbation is solved for according to the value of the damping parameter \( \lambda \). The solved parameter adjustment vector \( \delta m \) is applied to the current model. Then, rays are traced through the updated model. This process is iterated until a satisfactory fit to the observed data, based on the chi-squared \( \chi^2 \) measure, is achieved.

As stated before, Zelt and Smith's method uses the form of regularization that minimizes the size of the model perturbation at each iteration. This type of inversion allows one to honor the prior information and include reasonable additional geological information during the modeling. The strength of regularization is controlled by a value of the damping parameter \( \lambda \). Based on the non-uniform parameterization and parameter-selective characteristic of Zelt and Smith's algorithm, the model may be developed by using different approaches such as an 'across-and-down', 'whole-model' or combination both of them (Zelt, 1999). When the starting model is quite simple, new model parameters are incorporated into modeling across the model and then
downward in a layer-stripping way to fit the data better. This ‘across-and-down’
approach constrains the shallow structure first, and then defines the deeper layers
progressively. When the starting model is complex, inversion of all model parameters
simultaneously is preferred. In this case, the starting model contains the essential
structure of the whole model. Therefore, only few if any model parameters are added
during the modeling, and thus, the ‘whole-model’ inversion solves for all model
parameters at once.

Although a preferred final model may provide an appropriate fit to the data and
prior information, the non-uniqueness of the inverse problem yields a non-unique final
model. The inverse method of Zelt and Smith (1992) offers a wide range of
capabilities to assess the final model reliability in terms of resolution, uncertainties,
and misfit errors. Zelt and Smith's model assessment techniques include indirect and
direct methods. Indirect methods assess the final model in terms of resolution and
uncertainty. These assessment methods do not require the derivation of additional
models that satisfy the observed data. In contrast, direct methods assess the final
model by obtaining alternative models that fit the real data equally well. The non-
linearity of traveltime inversion can be addressed using direct assessment by
estimating the bounds for particular model features and associated model parameter
trade-offs (Zelt, 1999).
3.3 Minimum-Structure Inversion Method

The method of first-arrival refraction tomography presented by Zelt and Barton (1998) seeks a minimum-structure model by minimizing the roughness and size of the total slowness perturbation with respect to the starting model. This type of wide-angle data inversion uses an uniform/fine-grid parameterization. The starting model is laterally homogeneous and constructed on the basis of a 1-D analysis of all or few shots. Prior information is not included, except that which determines the starting model. Since only first-arrivals are used, the more subjective contribution of later reflections and refractions is removed. Regularization that contains flatness or smoothness constraints (equation 14) provides a stable inverse solution by avoiding unnecessary structure in the final model.

Thus, this method of refraction tomography is objective by comparison to the Zelt and Smith approach. In this thesis the tomography method is used only to objectively assess the large-scale features of the models obtained by the Zelt and Smith method.

3.4 Program Descriptions

The wide-angle OBH data were modeled using a package of programs developed by C.A. Zelt. Existing programs include those based on the Zelt and Smith inversion algorithms. They are ZPLOT, RAYINVR, VMODEL, and DMPLSTSQR. Figure 7 presents a flowchart of the main programs that were used for thesis work.
Figure 7. Program flowchart. Seismic data input is in SEGY format. SGY2Z program converts SEGY data to z-format for ZPLOT program. ZPLOT creates a traveltime pick file. HEADUP program updates the header file using traveltime pick file. Z2TX program creates a pick file in text format for RAYINVR program. RAYINVR is a main program of the software package which provides ray tracing in 2D isotropic media for forward and inverse modeling, and calculates traveltime residuals and partial derivatives for inversion. DMPLSTSQRI program applies the method of damped least-squares to the linearized inverse problem using a RAYINVR output file of traveltime residuals and partial derivatives. VMODEL program allows one to check, edit and plot a velocity model used by RAYINVR.
ZPLOT is a seismic data visualization program that allows interactive plotting, picking, filtering and data editing. It is specifically designed for wide-angle seismic data. The ZPLOT program requires the data to be in z-format, a special data format unique to this program package. The OBH data were transformed from SEGY format (as supplied by GEOMAR) using the program SGY2Z. The ZPLOT program does not modify the actual trace headers. After a picking session a program HEADUP updates the active header file changing the picks and dead trace flags only. The input for the ZPLOT program includes the data file, a parameter input file, a record file, a predicted traveltime file output from RAYINVR, and the header file obtained from the data file using SGY2Z. The output of ZPLOT is a log file that contains the pick times and dead traces flags. The log file is used to update the active header file by running HEADUP. A program Z2TX is used to create a pick file in text format for RAYINVR from the header file. This data file contains the observed traveltime-distance pairs and is input for RAYINVR.

RAYINVR (Zelt and Smith, 1992) is a program to trace rays in 2-D isotropic media for fast forward modeling and inversion of refraction and reflection traveltimes. It also calculates traveltime residuals and partial derivatives for inversion. A 2-D numerical solution of the ray tracing equations is used to trace rays through the model. The ray tracing provides an analytical calculation of the partial derivatives of traveltime with respect to the model parameters selected for adjustment. These parameters include velocities and the depth of boundary nodes. The input for RAYINVR consists of a program parameter file, observed traveltime data file, and velocity model file. The parameter file defines program input parameters related to
plotting, ray tracing, and inversion. The data file contains the observed traveltime-
distance pairs for all receivers and phases, including the estimated uncertainty of the
traveltime picks, and the type of arrival. The velocity file specifies the velocity model
parameterization. Each run of RAYINVR produces a plot of the model and all traced
rays as well as a diagram of reduced traveltime versus distance for the observed and
calculated data. This allows the user to verify seismic phase identifications and pick
accuracy.

The interactive nature of RAYINVR allows one to explore model space in a
variety of ways such as tracing the whole model, tracing a selected model layer or
layers, or tracing only in the vicinity of the selected model parameter(s). The partial
derivatives and traveltime residuals are output from RAYINVR and used as input to
the DMPLSTSQR program.

DMPLSTSQR is a program to apply the method of damped least-squares to the
linearized inverse problem using the traveltime residuals and the partial derivatives of
traveltimes with respect to the model parameters calculated by RAYINVR (Zelt and
Smith, 1992). The input for DMPLSTSQR is a file containing the matrix of partial
derivatives and vector of traveltime residuals that was generated by RAYINVR using
the current velocity model file and a parameter file that includes the damping factor $\lambda$
in equation (9). The output of DMPLSTSQR consists of the updated velocity model as
well as the corresponding resolution and covariance estimates (equations 13).

VMODEL is a program to check, edit and plot the velocity model used by the
RAYINVR program. This program checks the velocity model after each iteration for
errors and unusual features. These features include unrealistic velocity values,
negative vertical velocity gradients, large vertical or lateral velocity gradients, low-
velocity zones, crossing layer boundaries, and boundaries with unexpected large dip.

For data and model plotting purposes the Generic Mapping Tool (GMT)
(Wessel and Smith, 1998), XV, IslandDraw, and CorelDraw applications were used to
create and/or modify postscript files generated by RAYINVR, VMODEL, and
ZPLOT.
4 Data and Modeling Preparation

4.1 Data Acquisition

4.1.1 Wide-Angle Reflection/Refraction Data

As part of the CONDOR project, the wide-angle seismic data were collected along two profiles using OBHs. Each of the profiles is about 150 km long and roughly perpendicular to the Chilean margin (Fig. 2). The southern profile 1 is ~50 km south of the Valparaiso Basin and is a comparative example of supposedly “normal” subduction. The northern profile 2 is ~75 km north of profile 1, and crosses and extends oblique to the axis of the Juan Fernandez Ridge. The full length of each 2-D profile was shot twice with a total of 36 OBHs deployed (OBH 1 to 15 along profile1, OBH 31 to 51 along profile 2) (Fig. 2). Two OBHs (1 and 45) failed. The average OBH spacing is about 7.5 km. Seismic signals were generated using up to three airguns of 32 liter volume each, fired at 60 s intervals at a depth of 10-12 m. Average shot spacing is about 100 m (Flueh, 1995). The dominant frequency of the data is 5-7 Hz and was sufficient to trace signals on the OBH record sections to distances up to ~160 km.

4.1.2 Other Data

Deep multi-channel seismic reflection data were collected using a 3 km streamer with 120 channels along the wide-angle OBH lines. A tuned airgun array
with a total capacity of 52 liters was used as a source. Deep reflection data coincident with the OBH profiles provided images to 12 s.

About 2,000 km of high-resolution seismic reflection profiles were shot using a four-sleeve airgun source and a 24-channel, 150-m-long streamer. Some of the lines are roughly coincident with OBH profiles 1 and 2. The high-resolution data generally imaged the upper 2-4 s of the sediments and basement (Flueh, 1998).

In addition to seismic data, several other types of data were acquired during the CONDOR experiment. These include bathymetric and magnetic data, and core and dredge samples. Bathymetry data (Fig. 2) and magnetic anomaly data (Fig. 3) were acquired along 15,380 km. A sampling rate of 6 s of the magnetic data was considered sufficient to process the data and record the long wavelength variation of the magnetic field (typically 5 to 20 km) because of relatively low speed marine recording (10 to 12 knots).

In this thesis only wide-angle OBH seismic data and bathymetry data were used for modeling the velocity structure along the two profiles, since digital copies of reflection data were not available for Rice University. However, the high-resolution reflection data were available and used during the cruise to develop a starting model (Zelt, 1996).

### 4.2 Starting Models and Model Parameterization

The 2-D models of Zelt (1996) from preliminary modeling of the wide-angle data along both profiles were used as the starting models. These preliminary models
(Fig. 4) have been developed using only one half of the OBH records from the first deployment – the eight OBHs of profile 1 (2, 3, 4, 6, 7, 8, 9, 11) and eleven OBHs of profile 2 (33, 34, 36, 37, 38, 40, 42, 44, 46, 48, 50). Because of limited time during the cruise, the arrivals were picked at 2-5 km spacing on both profiles, resulting in a total of 858 and 1075 picks for all OBHs and phases for profile 1 and profile 2, respectively.

These arrival picks were modeled using the Zelt and Smith (1992) technique. First, it was necessary to develop an initial model. The observed OBH data were modeled first for profile 1. The starting model was built using the following information. The shallow features were constrained by the bathymetry measurements along profile 1 and an analysis of high-resolution reflection data recorded along the two OBH lines (von Huene et al., 1995). The crustal velocity model from about 1000 km to the north of the study area (between 23°S and 24°S) presented by Comte et al. (1994) defined the large-scale structural elements of Zelt’s starting model. The dip of ~15° of the subducting oceanic plate to 60 km was based on earthquake locations for the Valparaiso region (Tichelaar and Ruff, 1991). Figure 8 presents the starting model for profile 1. The final 2-D model of profile 1 obtained by Zelt (1996), and the corresponding bathymetry and high-resolution reflection data were used to construct the starting model for profile 2.

Six seismic phases were identified and inverted during the preliminary modeling of both profiles. These phases are the water wave, refracted waves within the sediments, refracted waves within the forearc wedge and continental upper crust, refracted waves within the oceanic crust, reflected waves from the oceanic Moho, and
Figure 8. Starting model for profile 1 from Zelt (1996) with velocities in km/s. Velocity and depth nodes are indicated by blue circles and red squares, respectively; layer boundaries are indicated by dashed lines. This model is based on 4 sources of prior information (see chapter 4.2).
refracted waves within the uppermost mantle. The water wave phase was used to control the OBH locations only. The traveltime data were fit with a RMS traveltime residual of \( \sim 450 \) ms with respect to the predicted data in the starting models for both profiles. The final models provide a traveltime fit with the RMS residual of \( \sim 185 \) ms and \( \sim 190 \) ms for profile 1 and profile 2, respectively. During several inversion iterations, more model parameters were added to constrain the lateral variation suggested by the data. Both preliminary models contain about 50 independent parameters.

The starting models and their parameterization based on Zelt's (1996) preliminary models are presented in Figure 9 for both profiles. The models are parameterized in accordance with the Zelt and Smith (1992) method requirements. The starting models are composed of a sequence of six layers that represent the main structural elements: 1 – water, 2 – shallow sediments, 3 – forearc wedge and upper continental crust, 4 – lower continental crust, 5 – oceanic crust, and 6 – uppermost mantle. Well-defined and well-constrained boundaries such as the water bottom (layer 1) and base of the sediments (layer 2) required more depth parameters to follow the boundary relief correctly. The models contain layers reduced to zero thickness to model pinch-outs for the two layers of the continental crust (layer 3 and 4). These pinch-outs arise because of the convergence between the oceanic plate and continental margin in a subduction zone.

The starting models of Zelt were modified as the modeling of the OBH data revealed additional structural elements. Figure 10 shows the final model parameterization for both profiles. During the modeling process, new layers were
Figure 9. Final model parameterization for preliminary models of Zelt (1996) for profile 1(a) and profile 2(b). Velocities in km/s. Velocity and depth nodes are indicated by blue circles and red squares, respectively; layer boundaries are indicated by solid lines. Model layers are numbered (bold numbers).
Figure 10. Model parameterization for the final models obtained by the inverse modeling in this thesis for profile 1(a) and profile 2(b). Velocity and depth nodes are indicated by blue circles and red squares, respectively; layer boundaries are indicated by solid lines; model layers are numbered (bold numbers).
incorporated into the models such as the subducting sediment layer on both profiles (layer 5 and 6 on profiles 1 and 2, respectively) and a layer of Valparaiso Basin sediments on profile 2 (layer 3). The velocity and depth nodes were added into the models so as to provide a better fit to the observed data. For profile 1 the final model consists of eight layers and contains 119 independent model parameters. The final model of profile 2 includes nine layers and 130 parameters. In the shallow part of the models (less than 10 km), the OBH data constrains the structural features with a higher degree of reliability. Therefore, model parameters are distributed more densely compared to the deeper part. The dense parameter distribution allows a better fit to the data. However, taking into account that a dense distribution of the parameters generally decreases the model resolution, it is important to find the appropriate trade-off between the number of model parameters and model resolution. On both profiles at model distances more than ~50 km, the base of the sediment layers are sampled by depth nodes with ~10 km spacing. For the deeper layers the depth nodes are spaced at ~30-40 km along the layer boundaries. The velocity node spacing ranges between ~5-15 km for the shallow portion of the models and more than 20 km for the deep part of the model.

4.3 Data Preparation

For both profiles the seismic wide-angle data consist of 34 records from the OBH instruments: 14 OBH records for profile 1 and 20 OHB records for profile 2. The quality of the data was evaluated at this stage. Some problems such as noise level,
dead traces, errors due to navigation and inaccuracy of the instrument locations, negatively affected the data. Therefore, the overall data quality is not considered good, but satisfactory, relative to other similar datasets (e.g. Christeson et al., 1999; Holbrook et al., 1994).

The main procedure in the data preparation stage is the picking of arrivals of the wide-angle data and classifying these picks. The reliability of a model derived by inverse modeling of the traveltime data depends on the accuracy of the picks. As mentioned before, about one half of the OBH records from the first deployment – the eight OBHs of profile 1 and eleven OBHs of profile 2 – were modeled by Zelt (1996), and the preliminary models were developed for both profiles. Using these models, first and clear later arrivals from the other non-modeled OBH records were identified by overlaying calculated times on the observed data. To improve the picking, the data were processed using amplitude scaling, filtering and mute operations, using capabilities of the ZPLOT program. A Butterworth filter with zero-phase and minimum-phase options was applied with varying low cut and high cut values to enhance arrival before picking. Where possible the data were picked from the unfiltered data. In the beginning, a semi-automated picking approach was used. However, due to spatial aliasing and noise in the data, all picks were made manually. The picks from the previously modeled OBH records were evaluated and repicked more densely and accurately. For all identified phases and shots, the picks have an average spacing of ~1 km. Table 1 presents the description of all identified seismic phases and statistical information for both profiles related to the modeling.
### Table 1. Seismic Phase Terminology and Traveltime Misfits

<table>
<thead>
<tr>
<th>Phase Symbol</th>
<th>Phase Description</th>
<th>No. Picks (Uncertainty) (s)</th>
<th>RMS Residual (s)</th>
<th>Chi-Squared ($\chi^2$)</th>
<th>No. Picks Used *</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Profile1</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Pw</td>
<td>Direct arrivals-water waves</td>
<td>338 (0.078)</td>
<td>0.059</td>
<td>0.567</td>
<td>338</td>
</tr>
<tr>
<td>PsP</td>
<td>Reflection, base of sediments</td>
<td>134 (0.106)</td>
<td>0.086</td>
<td>0.832</td>
<td>134</td>
</tr>
<tr>
<td>Ps</td>
<td>Refraction, sediments</td>
<td>273 (0.087)</td>
<td>0.082</td>
<td>1.046</td>
<td>269</td>
</tr>
<tr>
<td>PssP</td>
<td>Reflection, top of subducting sediments (model distance &lt;90 km)</td>
<td>85 (0.115)</td>
<td>0.094</td>
<td>0.702</td>
<td>83</td>
</tr>
<tr>
<td>PcP</td>
<td>Mid-continent crust boundary (model distance &gt;90km)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Pmw &amp; Pg</td>
<td>Refraction, Margin wedge (model distance &lt;100 km), upper continental crust (model distance &gt;100 km)</td>
<td>386 (0.095)</td>
<td>0.085</td>
<td>1.074</td>
<td>383</td>
</tr>
<tr>
<td>PocP</td>
<td>Reflection, top of slab (model distance &gt;50 km)</td>
<td>91 (0.115)</td>
<td>0.088</td>
<td>0.610</td>
<td>91</td>
</tr>
<tr>
<td>Poc2</td>
<td>Refraction, oceanic crust layer 2</td>
<td>24 (0.115)</td>
<td>0.084</td>
<td>0.607</td>
<td>24</td>
</tr>
<tr>
<td>Poc3</td>
<td>Refraction, oceanic crust layer 3</td>
<td>125 (0.098)</td>
<td>0.081</td>
<td>0.771</td>
<td>125</td>
</tr>
<tr>
<td>PmP</td>
<td>Reflection, base of slab</td>
<td>61 (0.120)</td>
<td>0.102</td>
<td>0.837</td>
<td>61</td>
</tr>
<tr>
<td>Pn</td>
<td>Refraction, oceanic Moho</td>
<td>331 (0.116)</td>
<td>0.147</td>
<td>1.711</td>
<td>331</td>
</tr>
<tr>
<td><strong>All data</strong></td>
<td></td>
<td>1848 (0.098)</td>
<td>0.096</td>
<td>0.995</td>
<td>1839</td>
</tr>
<tr>
<td><strong>Profile2</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Pw</td>
<td>Direct arrivals-water waves</td>
<td>570 (0.078)</td>
<td>0.081</td>
<td>0.975</td>
<td>570</td>
</tr>
<tr>
<td>PsP</td>
<td>Reflection, base of slope sediments</td>
<td>124 (0.111)</td>
<td>0.060</td>
<td>0.304</td>
<td>124</td>
</tr>
<tr>
<td>PsvP</td>
<td>Reflection, base of Valparaiso Basin sediments</td>
<td>57 (0.109)</td>
<td>0.096</td>
<td>0.773</td>
<td>57</td>
</tr>
<tr>
<td>Ps/Pvs</td>
<td>Refraction, sediments (slope and Valparaiso Basin)</td>
<td>257 (0.087)</td>
<td>0.079</td>
<td>0.898</td>
<td>257</td>
</tr>
<tr>
<td>PssP</td>
<td>Reflection, top of subducting sediments (model distance &lt;90 km)</td>
<td>186 (0.117)</td>
<td>0.110</td>
<td>0.820</td>
<td>186</td>
</tr>
<tr>
<td>PcP</td>
<td>Mid-continent crust boundary (model distance &gt;90km)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Pmw &amp; Pg</td>
<td>Refraction, margin wedge (model distance &lt;100 km), upper continental crust (model distance &gt;100 km)</td>
<td>566 (0.093)</td>
<td>0.090</td>
<td>1.022</td>
<td>565</td>
</tr>
<tr>
<td>PocP</td>
<td>Reflection, top of slab (model distance &gt;55 km)</td>
<td>117 (0.122)</td>
<td>0.092</td>
<td>0.610</td>
<td>117</td>
</tr>
<tr>
<td>Poc2</td>
<td>Refraction, oceanic crust layer 2</td>
<td>61 (0.112)</td>
<td>0.091</td>
<td>0.678</td>
<td>61</td>
</tr>
<tr>
<td>Poc3</td>
<td>Refraction, oceanic crust layer 3</td>
<td>209 (0.098)</td>
<td>0.067</td>
<td>0.478</td>
<td>207</td>
</tr>
<tr>
<td>PmP &amp; Pn</td>
<td>Refraction, oceanic Moho, or reflection, base of slab</td>
<td>502 (0.114)</td>
<td>0.112</td>
<td>1.012</td>
<td>502</td>
</tr>
<tr>
<td><strong>All data</strong></td>
<td></td>
<td>2649 (0.098)</td>
<td>0.093</td>
<td>0.897</td>
<td>2646</td>
</tr>
</tbody>
</table>
(* - the actual number of traveltime picks used in the modeling;

** - On profile 2, PmP and Pn arrivals are not clear distinguishable between each other. Therefore, the two seismic phases (PmP and Pn) were traced as one group for profile 2.)

12 and 14 seismic phases are used for the inverse modeling of profile 1 and 2, respectively. Example of ray paths for all modeled seismic phases is shown in Figure 11.

On both profiles, the comparison between the observed data and calculated traveltime curves reveals discrepancies due to the instrument depth location errors. These errors were more obvious for the direct arrivals and near-offset (< ~20 km) arrivals for all OBHs. Changing the depth locations of the OBHs in the trial-and-error way, several forward modeling RAYINVR iterations were performed to estimate the errors and get a more appropriate fit to the observed data. Average instrument location errors are 200 m, excluding OBH 36 on profile 2 in which the instrument was intentionally kept 1000 m above the sea floor for test purpose during the cruise.

For inverse modeling, it is essential to assign uncertainties to the arrival picks to avoid under- or over-fitting the traveltime data. The pick uncertainty bounds provide the suitable up- and down-weighting of the contribution to the inverse solution from relatively clear and unclear picks, respectively. Using the automated scheme of Zelt and Forsyth (1994), the pick uncertainties were assigned on the basis an empirical relationship between signal-to-noise ratio (SNR) and pick uncertainty. This scheme calculates the SNR in a 250-ms window directly before and after each pick and
Figure 11. a) Scheme of ray paths and b) travel times for all modeled seismic phases. For phase description refer to Table 1. Red lines indicate reflections, and blue lines indicate refractions.
correlates the SNR value to pick uncertainty. Table 2 lists the average SNR value for each of the 34 OBHs.

Taking into account the data noise, frequency content of the data, and the navigation and timing errors, picking errors were established to be in a range of 70 - 150 ms. Higher picking errors were assigned to the arrivals with lower SNR. Based on the ratio of trace energy in a 250-ms window immediately before and after a pick, all picks were divided into two types. First type characterizes first and clear later arrivals. Uncertainties for this type of arrival picks are assigned in range of 70 - 120 ms. Second type of pick characterizes the less clear later arrivals. Later arrivals have a greater potential to introduce interpretation errors into the modeling, and therefore, the later arrival picks have higher uncertainties according to the scheme of Zelt and Forsyth (1994). The second type of arrival picks are assigned in the range of 110 – 150 ms. Table 3 shows the relationship between SNR and uncertainty for both types of picks.
**Table 2.** Signal-to-Noise Ratio for each OBH.

<table>
<thead>
<tr>
<th>Profile 1</th>
<th>Profile 2</th>
</tr>
</thead>
<tbody>
<tr>
<td>OBH</td>
<td>No. Picks</td>
</tr>
<tr>
<td>2</td>
<td>169</td>
</tr>
<tr>
<td>3</td>
<td>188</td>
</tr>
<tr>
<td>4</td>
<td>195</td>
</tr>
<tr>
<td>5</td>
<td>171</td>
</tr>
<tr>
<td>6</td>
<td>176</td>
</tr>
<tr>
<td>7</td>
<td>99</td>
</tr>
<tr>
<td>8</td>
<td>130</td>
</tr>
<tr>
<td>9</td>
<td>87</td>
</tr>
<tr>
<td>10</td>
<td>79</td>
</tr>
<tr>
<td>11</td>
<td>77</td>
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<tr>
<td>12</td>
<td>77</td>
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<tr>
<td>13</td>
<td>115</td>
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<tr>
<td>14</td>
<td>144</td>
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<tr>
<td>15</td>
<td>141</td>
</tr>
<tr>
<td>46</td>
<td>90</td>
</tr>
<tr>
<td>48</td>
<td>141</td>
</tr>
<tr>
<td>50</td>
<td>90</td>
</tr>
</tbody>
</table>

Total 1848 3.378  Total 2649 4.456
Table 3. Pick Uncertainties

<table>
<thead>
<tr>
<th>Type 1</th>
<th>Type 2</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>SNR</strong></td>
<td><strong>σ, ms</strong></td>
</tr>
<tr>
<td>&gt; 2.0</td>
<td>70</td>
</tr>
<tr>
<td>&gt; 1.5</td>
<td>100</td>
</tr>
<tr>
<td>≥ 1.0</td>
<td>110</td>
</tr>
<tr>
<td>&lt; 1.0</td>
<td>120</td>
</tr>
</tbody>
</table>
5 Modeling and Model Assessment

5.1 Traveltime Inversion

The picked arrivals from 34 OBH records were modeled using the 2-D ray tracing inverse method of Zelt and Smith (1992). As previously mentioned, the first iteration of forward modeling was used to update the starting model and classify the arrival types. There are 12 and 14 modeled seismic phases for profile 1 and 2, respectively (Table 1). During the subsequent iterations of traveltime inversion, the values of model parameters were adjusted to obtain an appropriate fit between the observed and calculated data. A combination of both the across-and-down (layer-by-layer) and whole-model approaches (Zelt, 1999) was employed during the modeling. At the beginning, due to the difficulties in correlating many of the later and far offset arrivals with a particular phase from the initial model, the layer-stripping method was used. The layer-by-layer inverse modeling constrains the shallow features of the model first and then progressively constrains the deeper layers. Using the layer-stripping approach, the models were developed by adding several model parameters where required to fit the OBH data. Thus, based on the across-and-down inverse modeling, two new structural layers were incorporated into the model. These are the layer of the Valparaiso Basin sediments on profile 2 and the layer of subducting sediments on both profiles (Fig. 10). As the inversion proceeded, more seismic phases were incorporated into the modeling to develop the deeper model layers. However, based on the selective character of Zelt's algorithm, only velocity and depth nodes that
constrain a specific layer were adjusted at each stage of the layer-by-layer modeling. In spite of the fact that the layer-stripping approach can provide a reliable result, it is was necessary to apply the whole-model approach because of the complex starting model for both lines and to account for possible tradeoffs between the shallow and the deep structures. Also, in this case, the whole-model method allows one to incorporate and maintain prior information in the final model more easily than the layer-by-layer method. During several iterations of the whole-model inversion all model parameters were solved for simultaneously. For a few inverse iterations, the partly whole-model approach was performed for all model layers except the sediment layer that was considered well constrained by the high-resolution reflection data in the starting model and only slight changes were made during the layer-stripping modeling for this layer on both profiles. The final velocity models are presented in Figures 12(c) and 13(c) for profile 1 and 2, respectively. Figures 14 and 15 show the ray coverage provided by each of the modeled phases for all OBHs for profile 1 and 2, respectively. The comparison of the observed and calculated traveltimes for each OBH and phase is presented in Figure 16 for profile 1 and Figure 17 for profile 2. The interpreted crustal models are presented in Figures 18 and 19 for profile 1 and 2, respectively. The detailed modeling of the OBH data is described below with an indication of the source and degree of constraint for each structural element of the model.
Figure 12. Velocity models for profile 1: a) "1-D" starting model (follows bathymetry to <15 km depth) for tomographic modeling; b) final minimum-structure model obtained using the tomographic method of Zelt and Barton (1998) of first arrivals from all 14 OBHs; c) final minimum-parameter model obtained by traveltime inversion using the Zelt and Smith (1992) algorithm. Velocities in km/s are contoured. Contour interval is 1 km/s: black lines contour velocities of ≤ 6 km/s; white lines contour velocities of ≥ 7 km/s. Unsampled regions of the model are omitted (white space).
Figure 13. Velocity models for profile 2: a) "1-D" starting model (follows bathymetry to <15 km depth) for tomographic modeling; b) final minimum-structure model obtained using the tomographic method of Zelt and Barton (1998) of first arrivals from all 20 OBHs; c) final minimum-parameter model obtained by traveltime inversion using the Zelt and Smith (1992) algorithm. Velocities in km/s are contoured. Contours as in Figure 12.
Figure 14. Ray diagrams for each of the 12 modeled phase for all OBHs for profile 1. Two-point (source-to-receiver) ray tracing is used. Dashed lines indicate layer boundaries. For phase descriptions refer to Table 1.
Figure 14. (continued)
Figure 15. Ray diagrams for each of the 14 modeled phase for all OBHs for profile 2. Two-point (source-to-receiver) ray tracing is used. Dashed lines indicate layer boundaries. For phase descriptions refer to Table 1.
Figure 15. (continued)
Figure 16. Comparison of observed and calculated traveltimes for each of 14 OBHs used in the inverse 2D modeling of profile 1. The observed times are indicated by colored vertical bars with heights corresponding to their pick uncertainty and colors corresponding to the ray paths in Fig. 14; the calculated times are black lines. A reducing velocity of 6 km/s is applied. Horizontal scale is model position in km. For phase descriptions refer to Table 1.
Figure 17. Comparison of observed and calculated traveltimes for each of 20 OBHs used in the inverse 2D modeling of profile 2. The observed times are indicated by colored vertical bars with heights corresponding to their pick uncertainty and colors corresponding to the ray paths in Fig.15; the calculated times are black lines. A reducing velocity of 6 km/s is applied. Horizontal scale is model position in km. For phase descriptions refer to Table 1.
Figure 18. Structure of Chilean margin along profile 1. Annotated with a geological interpretation. Seismic velocities in km/s. Numbered OBH locations used in the inverse modeling are indicated. Dashed lines represent uncertain layer boundaries.
Figure 19. Structure of Chilean margin along profile 2. Annotated with a geological interpretation. Seismic velocities in km/s. Numbered OBH locations used in the inverse modeling are indicated. Dashed lines represent uncertain layer boundaries.
5.1.1 Sediments

The sediment layer extends across the whole model on both profiles. This layer is characterized by near offset (< 20 km) arrivals of the refracted waves within the sediments (Ps) from all OBH records of profile 1 (Fig. 14c). The base of the sediments is constrained by reflection (PsP) ray paths from all OBHs, except OBH-11 and OBH-13 (Fig. 14b), for profile 1 (Fig. 16). To provide a better fit to the observed data for profile 1, four velocity nodes were added to the starting model at 47 km and 115 km distance at the top and base of the sediment layer. The number of depth nodes remains the same, whereas horizontal locations of the depth nodes are slightly changed (Figs. 9a and 10a). On profile 1, Ps and PsP arrivals constrain an average sediment thickness of ~2.5 km in the model distance interval of ~55 – 120 km. At the model edges, the sediment layer thins approximately to ~0.7 km. Traveltimes of Ps constrain the lateral velocity variations in a range from ~1.9 to ~2.7 km/s. Higher apparent velocities are indicated by Ps branches for OBH-03 (Fig. 20) and OBH-04 (Fig. 21) at ~60-70 km model distance in the vicinity of the trench (Figs. 18 and 19, respectively). The vertical velocity gradient varies from ~0.1 s\(^{-1}\) to ~0.5 s\(^{-1}\) within the sediments of profile 1. However, neither Ps nor PsP arrivals constrain the vertical velocity gradient reliably. The vertical velocity gradients within each of the model layers are determined on the basis of ray bending, prior geological information and consistency with the existing models.
Figure 20. Record section from profile 1 OBH-03: a) observed record section; b) record section overlain with calculated traveltimes; c) comparison of observed and calculated traveltimes (see Fig. 16 for details). A reducing velocity of 6 km/s is applied. Butterworth bandpass filter with a low cut of 4 Hz and a high cut of 14 Hz is applied in (a) and (b). Phases used in the modeling are labeled in (c); refer to Table 1 for terminology.
Figure 21. Record section from profile 1 OBH-04: a) observed record section; b) record section overlain with calculated traveltimes; c) comparison of observed and calculated traveltimes (see Fig. 16 for details). A reducing velocity of 6 km/s is applied. Butterworth bandpass filter with a low cut of 4 Hz and a high cut of 14 Hz is applied in (a) and (b). Phases used in the modeling are labeled in (c); refer to Table 1 for terminology.
On profile 2, Ps arrivals are detected at the near offsets (<20 km) for all 20 OBHs, and reflections from base of the sediment layer PsP are observed at offsets of less than 10 km (Fig. 17). During the modeling, previously unidentified later arrivals at small offsets (<20 km) from five OBHs (43, 44, 46, 47, 48) were identified as reflections from the base of the Valparaiso Basin (PvsP) (Fig. 15 c). Turning rays within the Valparaiso Basin (Pvs) and within the slope sediments Ps are presented in Figure 15(d). Figures 22 and 23 show the records for OBH-46 and OBH-48, respectively. Thus, based on the modeling of these arrivals, the Valparaiso Basin sediments are divided into two layers with velocities of ~2.2 km/s and ~2.5 km/s and each of ~1.8 km thick. The base of the Valparaiso Basin is constrained by the first arrivals of turning waves within the upper continental crust Pg (OBH 43,44,48-51) (Fig. 17).

Landward from the trench, the slope sediment layer thickens up to ~1.8 km, whereas in the vicinity of the trench and in the seaward direction (model distance of < 70 km) an average sediment thickness is only ~0.8 km. Velocities within the slope sediments range from ~1.9 to ~2.6 km/s. For both profiles, the sediment layers were constrained by other deeper seismic phases such as first-arrival refractions through the underlying margin wedge (Pmw) (model distance ~60-100 km) and upper continental crust (Pg) (model distance > 100 km).
Figure 22. Record section from profile 2 OBH-46: a) observed record section; b) record section overlain with calculated traveltimes; c) comparison of observed and calculated traveltimes (see Fig. 17 for details). A reducing velocity of 6 km/s is applied. Butterworth bandpass filter with a low cut of 4 Hz and a high cut of 14 Hz is applied in (a) and (b). Phases used in the modeling are labeled in (c); refer to Table 1 for terminology.
Figure 23. Record section from profile 2 OBH-48: a) observed record section; b) record section overlain with calculated traveltimes; c) comparison of observed and calculated traveltimes (see Fig. 17 for details). A reducing velocity of 6 km/s is applied. Butterworth bandpass filter with a low cut of 4 Hz and a high cut of 14 Hz is applied in (a) and (b). Phases used in the modeling are labeled in (c); refer to Table 1 for terminology.
5.1.2 Accretionary Wedge and Upper Continental Crust

The accretionary wedge and upper continental crust were modeled from clear Pmw and Pg arrivals that occurred on many OBH records on both profiles (Figs. 16, 17, 20, 21, 24). According to the model parameterization, these two structural elements were part of one layer (layer 3 and 4 on profile 1 and 2, respectively) (Fig. 10). Lateral velocity variations are relatively well constrained by refractions within the wedge for both profiles due to the dense Pmw ray coverage. However, Pg ray paths are deeper and denser for profile 2 (Fig. 15 f) than for profile 1 (Fig. 14 e) in sampling the upper continental crust. Within the margin wedge velocities vary in the range from ~3.0 to ~4.0 km/s on profile 1, and from ~2.8 to 3.9 km/s on profile 2. The modeling of Pmw/Pg arrivals distinctively reveals a relatively sharp transition zone between the margin wedge and the upper continental crust at model distances between ~90-110 km for both profiles. This zone is characterized by an abrupt increase in lateral velocity gradient (Figs. 12c, and 13c). On profile 1, the transition zone is constrained by two velocity nodes at the top of the layer at model distance of 100 km and 110 km with velocities of ~4.0 km/s and ~5.5 km/s, and two nodes at the layer base at a distance of 80 km and 100 km with velocities of ~3.7 km/s and ~5.7 km, respectively. For profile 2, the velocity parameters that define the transition zone at the top of the layer are located at 100 km and 110 km model distance with velocities of ~4.0 km/s and ~5.3 km/s, and at the base of the layer at a distance of 60 km and 100 km with velocities of ~2.6 km/s and ~5.2 km/s, respectively. The transition zone dips approximately 30° in the seaward direction for both profiles.
**Figure 24.** Record section from profile 2 OBH-36: a) observed record section; b) record section overlain with calculated traveltimes; c) comparison of observed and calculated traveltimes (see Fig. 17 for details). A reducing velocity of 6 km/s is applied. Butterworth bandpass filter with a low cut of 4 Hz and a high cut of 14 Hz is applied in (a) and (b). Phases used in the modeling are labeled in (c); refer to Table 1 for terminology.
Velocities within the upper continental crust vary from ~5.5 to ~5.9 km/s on profile 1 and from ~5.5 to 6.6 km/s on profile 2, increasing in the landward direction. An average vertical gradient of 0.1 s\(^{-1}\) is derived in the upper continental crust for profile 1. For profile 2, the upper continental crust has a vertical velocity gradient of ~0.5 s\(^{-1}\) at a model distance of ~120 km and ~1.0 s\(^{-1}\) at a distance of ~145 km. The base of the upper continental crust is constrained by mid-crustal reflections (PcP) (Figs. 14 d, 15 e) that are observed for several OBH instruments such as OBHs 7, 8, 13-15 on profile 1 (Fig. 16), and OBHs 42, 44, 46, 47, 48, 50 on profile 2 (Fig. 17). PcP arrivals constrain the depth of a mid-crustal continental velocity discontinuity for both profiles at a model distance of 100-140 km. For both profiles the mid-crustal continental boundary lies sub-horizontal at a depth of ~10-13 km.

5.1.3 Subducting Sediments

A layer of subducting sediments was incorporated into the models for both profiles based on the modeling of later reflection arrivals from OBH records 3-6, 15 for profile 1 (Figs. 25, 26) and OBH records 38-43 for profile 2 (Figs. 27, 28, 29). These arrivals were identified as reflections from the top of the subducting sediments (PssP) (Figs. 14d, and 15e). PssP arrivals are not clear and therefore the evidence for this layer is weak. The data from the OBHs of profile 2 provide more reliable evidence of the existence of this layer. For this profile, the layer lies on top of the subducting slab at depths of ~7 km to ~13 km with velocities from ~3.4 to ~4.3 km/s.
Figure 25. Record section from profile 1 OBH-06: a) observed record section; b) record section overlain with calculated traveltimes; c) comparison of observed and calculated traveltimes (see Fig. 16 for details). A reducing velocity of 6 km/s is applied. Butterworth bandpass filter with a low cut of 4 Hz and a high cut of 14 Hz is applied in (a) and (b). Phases used in the modeling are labeled in (c); refer to Table 1 for terminology.
Figure 26. Record section from profile 1 OBH-15: a) observed record section; b) record section overlain with calculated traveltimes; c) comparison of observed and calculated traveltimes (see Fig. 16 for details). A reducing velocity of 6 km/s is applied. Butterworth bandpass filter with a low cut of 4 Hz and a high cut of 14 Hz is applied in (a) and (b). Phases used in the modeling are labeled in (c); refer to Table 1 for terminology.
Figure 27. Record section from profile 2 OBH-39: a) observed record section; b) record section overlain with calculated traveltimes; c) comparison of observed and calculated traveltimes (see Fig. 17 for details). A reducing velocity of 6 km/s is applied. Butterworth bandpass filter with a low cut of 4 Hz and a high cut of 14 Hz is applied in (a) and (b). Phases used in the modeling are labeled in (c); refer to Table 1 for terminology.
Figure 28. Record section from profile 2 OBH-40: a) observed record section; b) record section overlain with calculated traveltimes; c) comparison of observed and calculated traveltimes (see Fig. 17 for details). A reducing velocity of 6 km/s is applied. Butterworth bandpass filter with a low cut of 4 Hz and a high cut of 14 Hz is applied in (a) and (b). Phases used in the modeling are labeled in (c); refer to Table 1 for terminology.
Figure 29. Record section from profile 2 OBH-43: a) observed record section; b) record section overlain with calculated traveltimes; c) comparison of observed and calculated traveltimes (see Fig. 17 for details). A reducing velocity of 6 km/s is applied. Butterworth bandpass filter with a low cut of 4 Hz and a high cut of 14 Hz is applied in (a) and (b). Phases used in the modeling are labeled in (c); refer to Table 1 for terminology.
On profile 1, the layer of the subducting sediments is constrained with less confidence, because the reflection arrivals are weaker or unclear and observed only on a few OBHs. The low-velocity subducting sediment layer was modeled on profile 1 using the layer from profile 2 as a guide. The modeling of PssP arrivals together with arrivals of deeper phases that sample this part of the model reveals that the velocities range from ~3.0 km/s at the depth of ~7 km to 4.7 km/s up to the depth of ~17 km for profile 1.

5.1.4 Lower Continental Crust

Ray paths of reflections from the oceanic Moho (PmP) and refractions within the uppermost oceanic mantle (Pn) sample the lower continental crust for both profiles (Figs. 14i,j, and 15j). PmP/Pn arrivals are observed at far offsets of ~90-160 km on practically all OBH records. However, PmP/Pn ray paths cover this region of the model only at model distance of less than 120 km. Thus, velocities and vertical velocity gradients within the most part of the lower continental crust are not well constrained. For both profiles these velocities vary from ~5.7 to ~ 6.8 km/s increasing with depth in the landward direction.

5.1.5 Oceanic Crust

The slab of the subducting oceanic crust is constrained by reflection arrivals from the base of the sediments (PsP) at model distances ~0-55 km (Figs. 14b, and 15b) and reflections from the top of the oceanic crust (PocP) at model distances ~55-110
km (Figs. 14f, and 15g) on both profiles. Worldwide studies of oceanic plates confirm the layered nature of the oceanic crust (White et. al., 1992; Carlson 1998; Nakanishi et. al., 1998; Christeson et. al., 1999). In general, the oceanic crust is divided into two primary layers: a lower-velocity, high-vertical-gradient upper layer (layer 2) and a higher-velocity, low-vertical-gradient lower layer (layer 3). According to White et al. (1992), the thickness of layer 2 is an average 1/3 of the total oceanic crustal thickness. Average velocities of ~5.1 km/s and ~6.7 km/s with average velocity gradients of ~1.8 s\(^{-1}\) and 0.3 s\(^{-1}\) characterize layer 2 and layer 3, respectively. There is evidence for the existence of the oceanic layer 2 from the Chilean OBH data as well. Even though no reflections are observed to constrain the 2/3 layer boundary, an analysis of arrivals of turning waves provides a support for a two-layer oceanic crust. Arrivals of turning waves that are observed at offsets less than ~20 km on three instruments for profile 1 (OBHs 2, 3, 4) and six for profile 2 (OBHs 31-35, 40) were identified as refractions within layer 2 (Poc\(_2\)) (Figs. 14g, and 15h). Turning rays within the lower oceanic layer 3 (Poc\(_3\)) (Figs. 14h, and 15i) are indicated on many OBH records on both profiles (Figs.16, 17). Poc\(_2\), PocP, and Poc\(_3\) traveltimes provide relatively good constraint for the oceanic crustal velocities at model distances of less than 80 km (Fig. 30). Seaward of the trench (at model distance <~60 km), the final models have a velocity of ~4.1-4.2 km/s at the top of layer 2 and a velocity of ~6.6 km/s at the top of layer 3 on both profiles. The thickness of the oceanic layer 2 is about 2 km. Seaward of the trench, the velocity at the base of layer 3 is ~7.0 km/s on both profiles. Landward of the trench, the oceanic crust is modeled with less confidence since refracted arrivals from layer 2 are observed only at model distances less than 100 km.
Figure 30. Record section from profile 2 OBH-32: a) observed record section; b) record section overlain with calculated traveltimes; c) comparison of observed and calculated traveltimes (see Fig. 17 for details). A reducing velocity of 6 km/s is applied. Butterworth bandpass filter with a low cut of 4 Hz and a high cut of 14 Hz is applied in (a) and (b). Phases used in the modeling are labeled in (c); refer to Table 1 for terminology.
At model distances ~70-110 km, velocities at the top of layer 2 increase to ~4.5 km/s on both profiles. On profile 1, velocities do not significantly vary laterally within layer 3. However, on profile 2 at a model distance ~100 km, velocities are slightly decreased to ~6.5 and ~6.7 km/s at the top and the base of layer 3, respectively. This decreasing of velocity coincides with a ~1 km increase of slab crustal thickness beneath the margin wedge (~70-110 km model distances). On profile 2, an average velocity at the base of layer 3 of ~6.9 km/s is significantly less than a typical velocity value for the oceanic crust 7.2 km/s. The cause of this could be explained by the close proximity of profile 2 to the Juan Fernandez subducting ridge (Christeson et al., 1999).

5.1.6 Moho Depth and Upper Mantle

The depth of the oceanic Moho and velocities in the oceanic upper mantle were constrained by reflections from the base of the oceanic crust (PmP) and refractions turning within the uppermost mantle (Pn). On most of the OBH records, PmP arrivals are often difficult to separate and distinguish from the Pn arrivals. However, for a few OBHs the PmP phase was identified at relatively near offsets of ~30 km (Figs. 16, 17). As previously mentioned, on profile 2, the Moho depth is ~1 km deeper than on profile 1 at model distance between 70-110 km. The refracted Pn arrivals are clearly observed at far offsets for most OBHs on both profiles. For both final models the Pn arrivals constrain the upper mantle velocity to be ~7.9 km/s. Velocities in the overlying model layers affect the Moho depth and therefore, provide some constraints on Moho relief as well.
5.2 Model Assessment

The reliability of the final models for both profiles was judged using indirect and direct assessment techniques that were proposed by Zelt (1999).

5.2.1 Indirect Assessment

Indirect model assessment provides a measurement of the resolution and uncertainty of a final model in a time-efficient way without deriving additional models that satisfy the data. Since indirect methods involve measures that do not depend on the non-linearity of the traveltime inverse problem, these methods may be considered as linear methods of model assessment. Therefore, strictly speaking, this provides results that are valid only in a neighborhood about the final model, which behaves linearly.

Modeling statistics are presented in Table 1 that lists all identified seismic phases for both profiles (12 phases for profile 1 and 14 phases for profile 2). Statistics include the number of picks for each phase, the final value of the standard deviation (RMS) and chi-squared ($\chi^2$) measure of the traveltime misfit for each phase, and the total for all phases. The RMS traveltime misfit is defined as:

$$\sigma_i = \sqrt{\frac{1}{N} \sum_{j=1}^{N} (t_i^o - t_i^p)^2},$$ \hspace{1cm} (15)

where $t_i^o$ is the traveltime of the $i$-th pick, $t_i^p$ is the corresponding predicted traveltime from the model, and $N$ is the number of traveltime picks.
The chi-squared $\chi^2$ measure characterizes the agreement between the observed and predicted values of the traveltime measurements normalized by the picking error:

$$\chi^2 = \sum_{i=1}^{N} \left( \frac{t_i^o - t_i^p}{\sigma_i} \right)^2,$$

where $t_i^o$ is the $i$-th observed traveltime, $t_i^p$ is the $i$-th predicted (calculated) traveltime, and $\sigma_i$ is the estimated standard deviation of the $i$-th pick.

The value of the normalized misfit parameter $\chi^2$ is one of the criteria used to select the final model. If the value of $\chi^2$ is of the order of one, the data are considered to be fit to within their assigned uncertainties without overfitting (Zelt, 1999). The final models have the overall $\chi^2$ value of 0.995 and 0.897 for profile 1 and profile 2, respectively, and thus, these models provide an appropriate fit to the data. However, the final $\chi^2$ values for each phase vary from the value of one in some cases. The $\chi^2$ values for each phase vary from 0.567 (Pw) to 1.711 (Pn) for profile 1 and from 0.304 (PsP) to 1.025 (Pn) for profile 2. The values of $\chi^2$ of order less than one have been obtained for a few phases such as Pw ($\chi^2 = 0.567$), Poc$_2$ ($\chi^2 = 0.607$) and PocP ($\chi^2 = 0.610$) for profile 1, and PsP ($\chi^2 = 0.304$), PocP ($\chi^2 = 0.610$) and Poc$_3$ ($\chi^2 = 0.478$) for profile 2. However, those values of $\chi^2$ less than one are acceptable for some phases as long as the overall measure of $\chi^2$ is approximately one for all phases and OBHs. On profile 1, one seismic phase that is identified as refracions within the uppermost mantle (Pn) stands out with a large value of $\chi^2$ of 1.711. The modeling of the first arrivals of the Pn phase from OBH-6 caused such a misfit error (Figs. 16e and 25). It is not clear why this one phase was impossible to fit more closely.
Table 4 shows how many model parameters, both velocity and depth nodes, were involved to constrain each model structure for two profiles.

Graphical methods of indirect assessment include plots comparing the observed and calculated traveltimes for each phase and OBHs (Figs. 16, 17), a plot of the model parameterization (Fig. 10), ray diagrams (Figs. 14, 15), and a plot of the model (Figs. 18, 19). Since the distribution of the all velocity and depth nodes comprising the model parameterization is controlled by the prior information incorporated into the models, the model parameterization plot gives some insight into the geological constraint on the model. Plots of the decimated ray coverage for all phases and OBHs together are presented in Figure 31 for both profiles. These plots are important for qualitatively understanding how well the model can be resolved in different regions. Figures 14 and 15 show the ray diagrams of each separate phase for all OBHs for profile 1 and 2, respectively. These diagrams allow the contribution of each phase to the model constraint to be evaluated.
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Figure 31. Ray coverage from all phases and OBHs for profile 1(a) and profile 2(b). Two-point ray tracing is used. Rays are decimated by 5 for clarity.
Another important criterion to evaluate the final model on the basis of indirect assessment is model parameter resolution. The analysis of the resolution matrix ($R$; equations 12 and 13), both the diagonal and row values, provides a quantitative assessment of model resolution. However, the resolution values are generally important in a relative sense only. Figure 32 presents the diagonal values of the model resolution matrix for both profiles. The values for the velocity nodes are contoured in color and circles whose size is scaled to the value in the diagonal represent the depth nodes. Resolution matrix diagonal values greater than 0.5 typically indicate reasonably well resolved model parameters (Zelt, 1999). Maximum values are 1.0; negative values indicate a negative correlation of parameter values. Due to the non-uniform model parameterization that is used to represent the strong lateral variation of a convergent margin some portions of the model are constrained by a larger number of parameters compared to other parts. As a result, in such parts of the model a trade-off in the resolution values among the adjacent model parameter is observed, and overall values are decreased. In model regions in which adjacent parameters have small diagonal values, less then 0.5, reasonably good resolution is achieved over a length scale equivalent to a series of adjacent parameters in which the sum of their diagonal values is about 1.0. For both profiles the resolution matrices were calculated with a value of the damping parameter ($\lambda$) of 10.0 that was used during the inverse modeling. The model parameters are relatively well resolved up to a depth of approximately 10-12 km for both profiles. For profile 1 (Fig. 32 a), the final model is well resolved for the sediment layer (except for the model distance interval of 0-25 km), the margin wedge, the upper continental crust, the subducting oceanic crust and Moho boundary
Figure 32. Diagonal values of resolution matrix for profile 1(a) and 2(b) in contour format with shading for velocity parameters and circle size for depth parameters. In lower left corner of the plots the circle size that corresponds to the value of 1.0 is presented; layer boundaries are indicated by white solid lines. Contour interval is 0.1. Color scale is nonlinear.
within a model distance of \( \sim 60 \text{ km} \) to \( \sim 95 \text{ km} \). The final model of profile 2 (Fig. 32b) resolves the sediment layer, Valparaiso Basin, the margin wedge, the upper continental crust up to a model distance of \( \sim 130 \text{ km} \). Resolution of the subducting oceanic crust, Moho boundary, and the uppermost mantle is better on profile 2 compared to profile 1. Since the wide-angle data were modeled from 20 OBHs for profile 2 compared to 14 OBHs for profile 1, the more dense ray coverage of profile 2 results in higher resolution than for profile 1. However, the mid-continental crustal boundary, the lower continental crust, and the subducting oceanic crust at far model distance (\( > \sim 130 \text{ km} \)) are not resolved well on either profile due to the low density of rays in these parts of the model (Fig. 31).

A resolution kernel is a row of the resolution matrix, and it characterizes how a selected model parameter represents a weighted average of the true Earth structure, i.e. "true" model parameters. For a model with a non-uniform parameterization, resolution kernels provide more insight into the spatial resolution of the model compared to the diagonals of the resolution matrix. Figures 33 and 34 show representative resolution kernels corresponding to the velocity and depth nodes for profiles 1 and 2, respectively. The plot of the resolution kernel may be used to evaluate how well a particular parameter is resolved in the final model. The smearing that involves parameters adjacent to the chosen model parameter indicates the degree of the particular parameter resolution. Such display of model resolution for a mixed-parameter non-uniformly parameterized inverse problem is presented in this thesis for the first time.
Resolution kernels for velocity and depth nodes for profile 1 are displayed in contour format with shading for velocity parameters and circle size for depth parameters; blue color characterizes negative resolution values, red color – most positive; maximum value for depth resolution is given in lower left corner; black square contains the position of the selected parameter; layer boundaries are indicated by white solid lines. Type of parameter (velocity or depth) is indicated above each plot.
For profile 1 the velocity nodes at ~80 km model distance at the base of the sediment layer (Fig. 33 a) and at ~110 km distance at the mid-crustal boundary (Fig. 33 b) are relatively well resolved. Even though large negative correlation occurs between these velocity nodes and neighboring depth nodes, resolution values are high (~0.8 and ~0.5 respectively). For the depth node at ~135 km model distance at the mid-crustal boundary, the resolution kernel (Fig. 33 c) shows averaging that involves the nearest depth node at ~110 km at the crustal boundary. However, resolution for this depth parameter of ~0.15 is an order of magnitude higher than an average velocity resolution of ~0.01 in that model region. Therefore, the depth of the mid-crustal boundary is relatively well resolved around that node. Figure 33(d) presents the resolution kernel for the velocity parameter at ~100 km model distance at the top of the subducting sediment layer. For this node, velocity smearing as well as smearing into the nearest depth nodes indicates the linear dependency of the parameters that constrain the upper and lower continental crust at the model distance interval ~90-130 km, and the oceanic crust, Moho boundary and the uppermost mantle at the distance interval of ~60-120 km. There is not much velocity and depth averaging (less than 0.05) for the depth node at the top of the oceanic crust at ~120 km model distance (Fig. 33 e), and hence, this depth parameter is well resolved. In spite of the low degree of smearing of the kernel in Figure 33(f), the small resolution values indicates that the corresponding velocity parameter within the oceanic layer 2 at ~40 km model distance is poorly resolved because of many close-positioned velocity nodes in this model region (Fig. 10 a).
Figure 34. Resolution kernels for velocity and depth nodes for profile 2 are displayed in contour format with shading for velocity parameters and circle size for depth parameters; blue color characterizes negative resolution values, red color – most positive; maximum value for depth resolution is given in lower left corner; black square contains the position of the selected parameter; layer boundaries are indicated by white solid lines. Type of parameter (velocity or depth) is indicated above each plot.
a) Velocity Parameter

\[
\text{max } |Z_{rk}| = -0.128
\]

b) Velocity Parameter

\[
\text{max } |Z_{rk}| = -0.366
\]

c) Depth Parameter

\[
\text{max } |Z_{rk}| = 0.920
\]

d) Velocity Parameter

\[
\text{max } |Z_{rk}| = -0.143
\]

e) Depth Parameter

\[
\text{max } |Z_{rk}| = 0.749
\]

f) Velocity Parameter

\[
\text{max } |Z_{rk}| = -0.207
\]
Figures 33(g) and 33(h) show the resolution kernels for the velocity and depth parameters respectively at ~75 km horizontal distance at the base of the oceanic layer. Although some smearing occurs into adjacent velocity as well as depth nodes, the model is well resolved for both velocity and depth parameters. Figures 33(i, j, k, l) present the resolution kernels corresponding to the model parameters that constrain the Moho boundary. Figures 33(i) and 33(j) plot the kernels for velocity and depth nodes at ~35 km model distance where Moho depth is about 12 km. A significant negative tradeoff occurs between velocity and depth resolution for the velocity node, whereas for the depth node correlation is weak. Even though a wide range of model parameters is smeared for these two kernels, the corresponding velocity and depth nodes are relatively well resolved. The resolution kernels for the velocity and depth parameter at Moho depth of ~19 km are presented in Figure 33(k) and 33(l) respectively. These resolution kernels, as well as the plot of the diagonal values of the resolution matrix (Fig. 32(a), provide the evidence that the final model is resolved very well around these model parameters. This can be explained by the high density of rays in this part of the model.

For profile 2, Figures 34(a), (b), and (c) show the resolution kernels for the model parameters that constrain the sediments of the Valparaiso Basin. For the velocity nodes at the top (Fig. 34(a)) and base (Fig. 34(b)) of the sediments, the kernels show big negative correlation with neighboring depth parameters. Velocities around the node in the upper part of the basin are resolved relatively better than velocities near the node at the base of the basin. For the depth node at the base of the Valparaiso Basin (Fig. 34(c)), the kernel does not indicate much averaging with neighboring
velocity and depth parameters, and so, depth resolution is sufficient in this model region. Based on an analysis of these resolution kernels, the Valparaiso Basin is considered to be well resolved. The resolution kernel for the velocity node at ~100 km model distance at the mid-crustal boundary (Fig. 34 d) provides evidence for the high degree of resolution for velocities around this node, even though smearing involves a range of the adjacent depth parameters. Figures 34(e) and 34(f) present the resolution kernels for the depth and velocity nodes respectively at ~90 km model distance at the top of the subducting sediments. The depth parameter is well resolved in this part of the model. The kernel for the velocity node involves some smearing into the neighboring velocity parameters, and shows large negative correlation with the adjacent depth parameter. According to the analysis of the diagonal values of the resolution matrix (Fig. 32), the final model is best resolved at depths of 0 to 10 km. The resolution kernel for the depth node at ~60 km model distance at a depth of ~7 km (at the base of the sediments near the trench axis) confirms the high degree of resolution for model structures at depths of 10 km and less (Fig. 34 g). The kernel in Figure 34(h) shows that the depth node at ~20 km model distance and a depth of ~8 km (at the base of the oceanic layer 2) is also well resolved, even though the resolution value is relatively small due to limited ray coverage in this part of the model. Figures 34(i) and 34(j) represent the resolution kernels for the velocity parameters that constrain velocities within the upper part of the oceanic layer 3 at model distances of ~35 km and ~75 km respectively. For both kernels, smearing that occurs with a range of model parameters (both velocity and depth nodes) indicates relatively lower velocity resolution around these nodes. The velocity parameter at ~35 km model
distance at the base of the oceanic layer 3 (Fig. 34 k) is resolved more poor compared
to the velocity node at the same distance in the upper part of this layer (Fig. 34 i). The
resolution kernel in Figure 34(k) involves significant smearing into the neighboring
velocity as well as depth parameters. Figures 34(l), (m), and (n) show the resolution
kernels for the depth parameters that constrain Moho depth at ~30 km, ~75 km and
~100 km model distance, respectively. As follows from the analysis of these kernels,
resolution of Moho depth decreases with the model depth (analogous to the model
distance, since the Moho is being subducted). However, in general, resolution kernels
for these three depth nodes indicate Moho depth is relatively well resolved. The
resolution kernel for the velocity parameter at ~85 km model distance within the
uppermost mantle is presented in Figure 34(o). Even though a range of adjacent model
parameters is smeared for this kernel, the uppermost mantle velocity is well resolved
around this node.

5.2.2 Direct Assessment

In spite of the fact that indirect methods provide important information
concerning the reliability of a final model, direct assessment allows one to examine
the uncertainties and non-uniqueness of the model in a detailed way taking the
nonlinearity of the inverse problem into account. Direct methods are more time-
consuming since they require deriving alternate models that fit the observed data.
However, these methods allow one to evaluate the reliability of a particular model
feature required by the data and define the absolute bounds on the model parameter values (Zelt and White, 1995).

Two types of direct assessment are applied. First, a single-parameter test provides an estimation of absolute uncertainty for a particular model parameter. The following scheme is used. The parameter's value is slightly perturbed from its value in the final model and fixed. Then, using the final model as the starting model the observed data are inverted involving all model parameters except the fixed one. For subsequent iterations the anomaly of the perturbation is gradually increased until the recovered model is unable to fit the data as well as the preferred final model. Positive and negative parameter perturbations are tested for each model node. The maximum positive and negative perturbation that guarantees an appropriate fit to the data is an absolute uncertainty estimate for the particular parameter.

A multi-parameter uncertainty test is used to assess the constraint on some particular model features such as the depth and configuration of layer boundaries, and the velocity distributions within interesting parts of the model. Additional inversions were run for each selected structural element. The procedure is similar to the single-parameter test with the exception that several model parameters that represent a particular feature are held fixed.

Table 5 presents the results of all the single- and multi-parameter tests for both profiles. Due to the model’s irregular/non-uniform parameterization and ray coverage, the absolute uncertainties vary significantly for different regions of the models. For both velocity and depth nodes the absolute uncertainties increase with depth. The depth of the base of the relatively shallow structures (depth < 9 km), such as the slope
sediment layer, Valparaiso Basin, the subducting sediment layer, and the oceanic crust layer 2 at model distances of 0 km to ~80 km, has an absolute uncertainty on the order of ~0.2 km. Large values of the absolute bounds on the depth parameters are estimated for the mid-crustal boundary and Moho boundary (~ ± 1.0 km). The absolute uncertainties of the velocity parameters increase from ± 0.05 km/s within the sediments to ± 0.4 km/s within the oceanic crust. However, velocities within the uppermost mantle have an absolute uncertainty of ~ ± 0.15 km/s.

Two structural elements of both final models were examined with special interest. These are the layer of subducting sediments and the upper oceanic layer 2. The observed data do not provide clear evidence for the existence of these features since only a few OBHs contain the weak arrivals that constrained these layers (Figs. 16, 17). Using the direct methods the robustness of these two model structures was tested. On profile 1 the low-velocity layer extends along the slab up to ~17 km depth where absolute velocity uncertainties are ~±0.3 km/s. On profile 2, this layer is subducted up to ~12 km with velocity uncertainties of ~±0.2 km/s. The top of the subducting sediment layer on profile 2 is resolved reasonably well with an absolute uncertainty of ~ ± 0.15 km, whereas on profile 1 the depth of the top of this layer is poorly resolved with an absolute uncertainty of ~ ± 0.4 km. The oceanic layer 2 is relatively well resolved for model distance less than 80 km, where the ray coverage is good. Here, the velocities within this layer, as well as the oceanic layer 3, have an absolute uncertainty of ~ ± 0.4 km/s for both profiles. The depth of the oceanic 2/3 boundary is constrained with an absolute uncertainty of ~ ± 0.45 km for both profiles.
In spite of the fact that the direct assessment method has limitations for exploring all possibilities in model space, the estimation of the absolute bounds on the model parameters provides the range of models that fit the data appropriately and include the desired prior information.
Table 5. Absolute Uncertainties of Velocity and Depth Nodes

<table>
<thead>
<tr>
<th>Model Structural Element</th>
<th>Model Distance (km)</th>
<th>Velocity Uncertainty (±, km/s)</th>
<th>Depth Uncertainty (±, km)</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Profile 1</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Slope Sediments</td>
<td>0 – 60</td>
<td>0.02 – 0.03</td>
<td>0.25 (layer base)</td>
</tr>
<tr>
<td></td>
<td>&gt; 60</td>
<td>0.02 – 0.03</td>
<td>0.15 (layer base)</td>
</tr>
<tr>
<td>Forearc Wedge</td>
<td>~65 – 100</td>
<td>0.1</td>
<td></td>
</tr>
<tr>
<td>Subducting Sediments</td>
<td>~55 – 95</td>
<td>0.1</td>
<td>0.1 (layer top)</td>
</tr>
<tr>
<td></td>
<td>~95 – 110</td>
<td>0.3</td>
<td>0.4 (layer top)</td>
</tr>
<tr>
<td>Upper Continental Crust</td>
<td>&gt; 100</td>
<td>0.1</td>
<td></td>
</tr>
<tr>
<td>Lower Continental Crust</td>
<td>~90 – 125</td>
<td>0.15</td>
<td></td>
</tr>
<tr>
<td>Mid-Crustal Boundary</td>
<td>~90 – 100</td>
<td>0.3</td>
<td></td>
</tr>
<tr>
<td></td>
<td>&gt; 100</td>
<td>1.0</td>
<td></td>
</tr>
<tr>
<td>Oceanic Crust, Layer 2</td>
<td>0 – 80</td>
<td>0.4</td>
<td>0.25 (layer top)</td>
</tr>
<tr>
<td></td>
<td>&gt; 80</td>
<td>0.4</td>
<td>0.4 (layer top)</td>
</tr>
<tr>
<td>Oceanic Crust, Layer 3</td>
<td>0 – 120</td>
<td>0.3 – 0.4</td>
<td></td>
</tr>
<tr>
<td>Oceanic 2/3 Boundary</td>
<td>0 – 120</td>
<td>0.45</td>
<td></td>
</tr>
<tr>
<td>Moho Boundary</td>
<td>0 – 160</td>
<td>0.9 – 1.3</td>
<td></td>
</tr>
<tr>
<td>Uppermost Mantle</td>
<td>0 – 160</td>
<td>0.1</td>
<td></td>
</tr>
<tr>
<td><strong>Profile 2</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Slope Sediments</td>
<td>0 – 100</td>
<td>0.02 – 0.03</td>
<td>0.25 (layer base)</td>
</tr>
<tr>
<td></td>
<td>100 – 145</td>
<td>0.02 – 0.03</td>
<td>0.35 (layer base)</td>
</tr>
<tr>
<td>Valparaiso Basin Sediments</td>
<td>~100 – 145</td>
<td>0.05</td>
<td>0.25 (layer base)</td>
</tr>
<tr>
<td>Forearc Wedge</td>
<td>~55 – 100</td>
<td>0.15</td>
<td></td>
</tr>
<tr>
<td>Subducting Sediments</td>
<td>~55 – 90</td>
<td>0.2</td>
<td>0.15 (layer top)</td>
</tr>
<tr>
<td>Upper Continental Crust</td>
<td>&gt; 100</td>
<td>0.15</td>
<td></td>
</tr>
<tr>
<td>Lower Continental Crust</td>
<td>~100 – 125</td>
<td>0.2 – 0.3</td>
<td>0.25</td>
</tr>
<tr>
<td>Mid-Crustal Boundary</td>
<td>~90 – 100</td>
<td>0.3</td>
<td>1.0</td>
</tr>
<tr>
<td></td>
<td>&gt; 100</td>
<td>1.0</td>
<td></td>
</tr>
<tr>
<td>Oceanic Crust, Layer 2</td>
<td>0 – 80</td>
<td>0.4</td>
<td>0.25 (layer top)</td>
</tr>
<tr>
<td></td>
<td>&gt; 80</td>
<td>0.4</td>
<td>0.5 (layer top)</td>
</tr>
<tr>
<td>Oceanic Crust, Layer 3</td>
<td>0 – 120</td>
<td>0.4</td>
<td></td>
</tr>
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<td>0.45</td>
<td></td>
</tr>
<tr>
<td>Moho Boundary</td>
<td>0 – 160</td>
<td>1.0</td>
<td></td>
</tr>
<tr>
<td>Uppermost Mantle</td>
<td>0 – 160</td>
<td>0.15</td>
<td></td>
</tr>
</tbody>
</table>
5.2.3 Tomographic Assessment

The traveltime inversion method of Zelt and Smith (1992) is considered a subjective approach since it incorporates prior information and geological concepts into the modeling and the irregular model parameterization is subjective. As a result, the final model is consistent with the observed traveltime data as well as known geological structures. The minimum-structure tomographic method of Zelt and Barton (1998) allows for the objective assessment of the geologically constrained, minimum-parameter model obtained by the inverse approach of Zelt and Smith (1992). The tomographic method objectively seeks the model with the minimum structure required by the data according to uniform/fine-grid model parameterization, an objectively determined starting model, and a form of regularization that uses smoothness constraint. This method is applied only to the first arrivals since their identification and picks are the least subjective and more certain than for later arrivals.

The first arrivals are calculated using a finite-difference technique (Vidale, 1988; Hole and Zelt, 1995). Using the uniform/fine-sampled model parameterization with a horizontal and vertical node spacing of 0.5 km, the first arrivals from each OBH were involved. A linear-gradient velocity model was used as a starting model. The starting model does not include any prior information, except bathymetry and the shallow sediment layers that mimic the water-bottom horizon (Figs. 12a, 13a). The minimum-structure model is derived by minimization of the roughness (second spatial derivatives). Regularization was controlled by the relative weight of the spatial smoothing parameter $s_z$ of 0.15 (equation 14). Figures 12 and 13 show the starting models (Figs. 12a, 13a), final minimum-structure (Figs. 12b, 13b) and minimum-
parameter models (Figs. 12c, 13c) for profiles 1 and 2, respectively. For both profiles, the minimum-structure models determine the model features that are required by the OBH data without considering the subjective prior information. Such model features include the sediment layer that extends across the whole model, margin wedge, continental crust, and subducting oceanic crust. For profile 1, the tomographic model shows the thin sediment cover on the continental slope and thick sediment filling at the trench axis. For profile 2, the minimum-structure model objectively reveals such structure as the thin sediments in the vicinity of the trench axis, the Valparaiso Basin located at the middle continental slope, the uplifted forearc wedge. A comparison between the minimum-structure and minimum-parameter models confirms position of the transition zone between the margin wedge and continental crust, and a two-layer nature of the subducting oceanic crust for both profiles. Some model structures, such as the subducting sediment layer, mid-continental and Moho boundaries at far horizontal model distances, are not supported by the minimum-structure test and, therefore are considered less objective. Due to the poor ray coverage (Fig. 31), the tomographic test is unable to constrain any structures within the continental crust at model distances more than ~120 km. This is consistent with the result of model resolution tests (Fig. 32).
6 Discussion

6.1 Crustal models

The seismic wide-angle OBH data were interpreted using the iterative damped least square inversion method of Zelt and Smith (1992). As a result, two separate compressional velocity models were constructed along two wide-angle reflection/refraction lines. The principal structures of the interpreted final models are shown in Figures 18 and 19 for profile 1 and 2, respectively. These crustal velocity models reveal large-scale structural elements of the subsurface of the Chilean convergent margin in the vicinity of the subducting Juan Fernandez Ridge. The main structural elements for both profiles include sediments of the trench and the continental slope, the subducting sediments, the forearc wedge, the upper and lower continental crust, the two-layer oceanic crust, and the uppermost mantle.

On both profiles, the uppermost mantle has a velocity of $\sim 7.9 \pm 0.1$ km/s. However, the position of the Moho boundary is different on the two lines. On profile 1, at model distances from 0 to $\sim 55$ km, the depth of the Moho gradually changes from $\sim 10$ to $\sim 15$ km (the dip angle is about $5^\circ$). Within the same distance interval on profile 2, the Moho deepens from $\sim 10.5$ to $12.4 \pm 1.0$ km (the dip angle is about $2^\circ$). At model distances from 55 to 120 km, the dip angle of the Moho is about $11^\circ$ for both profiles.

A two-layer subducting oceanic crust is modeled beneath the Chilean continental margin. Even though strong reflections or large velocity discontinuities are not observed within the oceanic crust on the OBH records, an analysis of turning
waves provides constraints for the two-layer oceanic crust at model distances ~0-100 km on both profiles. There is no strong evidence supporting such division of the oceanic plate from the OBH data at far model distances (~>100 km). However, the two oceanic layers are assumed to exist across the entire inverse models that are consistent with results of convergent margin crustal seismic studies worldwide (Carlson, 1998; Nakanishi et al., 1998; Christeson et al., 1999; Holbrook et al., 1999).

On both profiles, the position of the subducting slab is relatively well constrained by the OBH data up to model distances of ~120 km with the plate boundary at ~18-19 km depth.

There are differences in the structure of the oceanic crust between the two lines. On profile 1, at model distances from 0 km to ~55 km (location of the trench), the top of the oceanic crust dips with an angle of 3.3° reaching a depth of 7.7 ± 0.25 km in the vicinity of the trench, whereas at the same distance on profile 2, the oceanic crust extends towards the trench slightly inclined (a dip angle of ~1.8°), and reaches a depth of ~6.5 ± 0.25 km. Landward of the trench, the oceanic crust is subducted with a slab dip of ~10.5° at a depth of ~20 km on both profiles. The thickness of the oceanic crust also varies across the model on both lines. Seaward of the trench (at model distance <=55 km), the oceanic crust is up to ~1.0 km thicker along profile 2 (6.3 ± 0.5 km) compared to profile 1 (5.3 ± 0.7 km). In the vicinity of the trench (at model distances 50-60 km), the thickness of the oceanic crust is 5.5 ± 0.7 km on profile 1 and 6.0 ± 0.5 km on profile 2. Landward of the trench, an average thickness of the oceanic crust on profile 1 is 5.5 ± 0.8 km, whereas on profile 2 it is 6.5 ± 0.7 km thick.
The velocity distribution within the oceanic crust is different for the two profiles as well. Seaward of the trench (at model distance \(<\)55 km), the upper part of the oceanic crust (layer 2) has velocities varying from 4.1 ± 0.4 km/s at the top to 6.6 ± 0.4 km/s at the base of the layer 2 on both profiles. The thickness of the oceanic layer 2 is about 2 km. Landward of the trench, the oceanic crust is modeled with less confidence since refracted arrivals from layer 2 are observed only at model distances less than 100 km. At model distances of \(~70-110\) km, velocities at the top of layer 2 increase to 4.5 ± 0.4 km/s on both profiles. On profile 1, velocities do not vary significantly laterally, ranging from 6.7 ± 0.4 km/s to 7.0 ± 0.4 km/s within the lower part of the oceanic crust (layer 3). The average thickness of layer 3 is 4.0 ± 0.5 km on profile 1. On profile 2, lateral velocity variations occur within layer 3; seaward of the trench, they vary from 6.7 to 6.9 ± 0.4 km/s. At a model distance of \(~80\) km (beneath the forearc wedge), velocities slightly decrease to 6.5 ± 0.4 km/s and 6.7 ± 0.4 km/s at the top and the base of layer 3 respectively. This decrease in velocity coincides with a \(~1\) km increase of the oceanic crustal thickness beneath the margin wedge (distance \(~70-110\) km), making oceanic layer 3 up to \(~1.0\) km thicker along profile 2 in comparison to profile 1. At model distances greater than 110 km, velocities increase to 6.7 ± 0.4 km/s at the top and 6.9 ± 0.4 km/s at the base of the layer 3. On profile 2, an average velocity of 6.85 ± 0.4 km/s at the base of layer 3 is significantly less than a typical velocity value for the oceanic crust, which is \(~7.2\) km/s. The cause of this as well as the thicker oceanic crust on profile 2 could be explained by the close proximity of profile 2 to the Juan Fernandez subducting ridge. The Juan Fernandez Ridge is
associated with the present hot spot, and according to Christeson et al. (1999), hotspot-modified oceanic crust could account for increase in thickness of the oceanic plate on profile 2.

The continental lower crust is not well resolved by the OBH data due to limited ray coverage. Although the modeling reveals a typical average velocity of $6.6 \pm 0.2$ km/s within this structural unit, velocities vary from $5.7 \text{ to } 6.8 \pm 0.2$ km/s laterally and have an average vertical gradient of $-0.2 ~s^{-1}$ on both profiles. Relatively low velocities of $\sim 5.7 ~\text{km/s}$ are observed near the slab, and hence, may be associated with the subduction of the oceanic plate.

The layer combining the continental margin wedge and the upper continental crust overlies the oceanic crust and the lower continental crust on both profiles. Development of the wedge occurs as a result of deformation of the continental margin due to the subduction of the oceanic plate. The forearc wedge is formed between the trench axis and the normal upper continental crust. The wedge on both profiles is about $\sim 40 ~\text{km}$ wide. The forearc wedge has been delineated with a thickness reaching $\sim 7.4 ~\text{km}$ and $\sim 6.6 ~\text{km}$ at $\sim 30 ~\text{km}$ landward from the trench axis on profiles 1 and 2, respectively. On profile 1, the wedge lies about $2 ~\text{km}$ deeper than on profile 2. At $\sim 20 ~\text{km}$ landward of the trench axis, velocities vary from $2.9 \text{ to } 3.4 \pm 0.1 ~\text{km/s}$, and from $3.0 \text{ to } 3.6 \pm 0.15 ~\text{km/s}$ on profiles 1 and 2, respectively. At $\sim 40 ~\text{km}$ landward of the trench, velocities range from $3.6 \text{ to } 4.9 \pm 0.1 ~\text{km/s}$ on profile 1 and from $3.7 \text{ to } 4.7 \pm 0.15 ~\text{km/s}$ on profile 2. The transition zone between the margin wedge and the upper continental margin occurs at $\sim 45 ~\text{km}$ landward from the trench axis. The OBH data constrain the lateral velocity gradient of $\sim 1.4 ~s^{-1}$ within this zone at model distances of
~90-110 km on both profiles. The transition zone is inclined seaward with a dip angle of ~30°, constrained by the inverse models for both lines. The tomographic assessment corroborates the reliability of the inclined transition zone for both profiles (Figs. 12b, and 13b). The upper continental crust is characterized by an average velocity of 5.7 ± 0.1 km/s and 5.8 ± 0.15 km/s on profiles 1 and profile 2, respectively. The subhorizontal mid-crustal boundary lies at a depth of 12 ± 1.0 km on both lines.

Due to the tectonic deformation caused by the imminent subduction of the Juan Fernandez Ridge, the upper continental crust (the middle and lower slopes) is much thinner on profile 2 than on profile 1. The upper crustal thickness on profile 2 is reduced up to ~5 km by the presence of the Valparaiso Basin on the middle slope of the margin (~ 120 km model distance). Due to the absence of the forearc basin, the upper crust on profile 1 at the same model distance is ~10 km thick. Based on the analysis of the coastal outcrops near Valparaiso, the upper continental margin is composed of metamorphic rocks of Paleozoic and Mesozoic age (Gonzalez, 1989; von Huene, 1989).

The accretionary wedge on both lines is underlain by a 3.4 – 4.5 ± 0.15 km/s layer. This layer is the most subjective feature on both profiles, since only a few OBH records provide evidence for its existence. The subducting sediments near the trench are not distinctive from the margin wedge because the velocities are practically the same within the sediment layer and the wedge. These velocities are too low to correspond to crystalline rocks, and thus are more likely to indicate sedimentary material that could be consolidated, dewatered, and possibly cemented and partially metamorphosed. Therefore, this layer can be interpreted as a layer of subducting
sediments and it extends to ~17 km depth on profile 1 and only to ~12 km on profile 2. The extension of the subducting sediments on profile 2 is likely terminated by the subduction of the Juan Fernandez Ridge that acts as a barrier for sediment supply into the trench (von Huene et al., 1997).

Tertiary sediments overlie the oceanic crust and the continental margin across the whole model on both profiles. On profile 1, the thickness of the hemipelagic sedimentary layer varies from 0.9 ± 0.15 km at the western and eastern edges of the model to 2.4 ± 0.15 km in the vicinity of the trench axis, reaching the maximum of ~2.9 km at ~80 km model distance (above the wedge). Within this layer, velocities range from ~2.0 to ~2.7 km/s. In contrast, the sediment cover is much thinner on profile 2. Seaward of the trench and in the vicinity of the trench axis, the sediment thickness is 0.9 ± 0.2 km with an average velocity of ~2.2 km/s. The sediments are ~1.1 km thick above the margin wedge (~60-100 km model distances). At model distances of ~100-140 km on profile 2, the Valparaiso Basin is a distinctive tectonic structure that was formed as a result of the subduction of the Juan Fernandez Ridge (von Huene et al., 1997). Within the Valparaiso Basin, the accumulated sediments are divided into two ~2-km thick layers with velocities of ~2.2 and ~2.5 km/s. The deformed basin sediments, as shown on the multichannel seismic reflection data (Flueh et al., 1998), and the middle slope position of the basin make the Valparaiso Basin a tectonic anomaly. According to von Huene et al. (1997), this tectonism is related to the southward migration of the subducted Juan Fernandez Ridge.

The above comparison between the two final models shows that their structures differ significantly, although the two lines are only ~70 km apart. Such differences are
explained by their location. Profile 1 lies relatively far from the Juan Fernandez Ridge and therefore is not affected by the ridge subduction. Conversely, the structures on profile 2 are deformed by the subduction of the Juan Fernandez Ridge. The main differences are, on profile 2: 1) thicker oceanic crust with lower velocities; 2) less steep dip of the oceanic plate toward the trench and therefore, the more shallow position of the plate boundary in the vicinity of the trench axis; more sharp bending of the slab landward of the trench; 3) shorter extent of the subducting sediment along the slab; 4) shallower margin wedge; 5) more tectonic deformation within the continental margin (presence of the forearc Valparaiso Basin on the middle slope); 6) thinner slope and trench sediments.

6.2 Comparison with results of forward modeling

The final crustal models that were derived using the method of minimum-parameter traveltime inversion of Zelt and Smith (1992) can be compared with the results of the forward modeling of Flueh (1998). Flueh et al. developed their crustal models for the same two seismic lines using the forward modeling techniques of Luetgert (1992). The structural image revealed by the multichannel seismic reflection data along these profiles was used as a starting model. 14 OBH and 9 land station records were modeled for profile 1 while 19 OBH and 7 land station records were modeled for profile 2. Based on the land station data, the forward model extends 80 km more landward and 10 km deeper as compared to the inverse model for both lines. Flueh et al. (1998) state that the average fit is better than 0.1 s for their models, and the
maximum misfits are 0.2 s. In the inverse models, an average fit is 0.096 s and 0.093 s for profiles 1 and 2, respectively; the maximum misfits are 0.15 s (Table 1). Flueh's models (Fig. 5) show the same main structural elements of the study area that were revealed by the inverse modeling such as the sediments of the trench and the continental slope, the forearc wedge, the continental crust, the two-layer oceanic crust, and the uppermost mantle. However, there are significant differences in the model structure between the results of the forward and the inverse modeling.

The uppermost mantle in Flueh's models have a velocity of ~8.0 km/s on profile 1, and ~8.1 km/s on profile 2, whereas the inverse models have a velocity of 7.9 ± 0.1 km/s within this layer for both lines. In the forward as well as in the inverse model, the oceanic crust is divided into two layers. The thickness of the oceanic layers and the velocities within the upper oceanic layer 2 in Flueh's models are comparable with those of the inverse models. However, the range of the velocities within the oceanic layer 3 of Flueh's models (from ~6.2 to 7.1 km/s) differs from that of the inverse models (from 6.7 to 7.0 ± 0.4 km/s). Also, on profile 2, the forward modeling does not reveal the decrease of the velocities within the lower oceanic layer 3 beneath the margin wedge (~80 km model distance) that occurs in the inverse model.

At model distances of ~55-100 km on both profiles, the forearc wedge with the velocities of ~3.0-4.4 km/s overlies the oceanic crust in the forward models as well as in the inverse models. Strong vertical and lateral gradients characterize the velocity field within this structure. In Flueh's models, this structure is defined as an accretionary prism. In the inverse models, this structure corresponds to the combination of the accretionary wedge and the layer of the subducting sediments.
Based on my detailed analysis of the OBH data (chapter 5.1.3), the subducting sediment layer becomes a separate structural element in the inverse models. Even though the velocities within the subducting sediments are similar in both Flueh’s and the inverse models, the sediment thickness and the extension along the slab differ. On profile 2, in both models, the subducted sediments extend to ~100-km model distance (the middle slope). However, in the forward model this sediments subduct to a depth of ~17 km, whereas in the inverse models they extend to only 13 ± 1.0 km depth. On profile 1 in Flueh’s models, the presence of the ~5-km thick sediment material is observed between the oceanic plate and the upper continental crust extending to ~140 km (the upper slope) model distance and ~19 km depth. In contrast, the inverse modeling constrains the extension of the ~2-km layer of subducting sediments to model distance of ~120 km and 17 ± 1.0 km depth.

At ~95-105 km model distances within the upper continental margin, the transition from the 3.0-4.5 km/s accretionary wedge to the 5.0-6.0 km/s structure of the upper continental crust occurs on both profiles and both models. However, Flueh et al. position a structure with velocities in excess of 5.0 km/s by a subvertical boundary and interpret it as a backstop, whereas the results of the inverse modeling do not support a sharp vertical boundary between the two model structures. According to the inverse model interpretation, at ~95-105 km model distances, the transition zone between the forearc wedge and the upper continental crust is inclined ~30° landward, and evidently constrained by both Zelt and Smith inverse technique and the tomographic assessment (Figs. 12a, 12b and 13a, 13b). Other significant discrepancies between the forward and inverse models are observed in the structure of the
continental crust on both lines. In the inverse models, the continental crust is divided into the upper and lower layers by the subhorizontal mid-crustal boundary at 12.0 ± 1.0 km depth on both profiles. The lower continental crust with an average velocity of 6.7 ± 0.1 km/s is terminated seaward by a pinch-out, and the upper continental crust with an average velocity of 5.9 ± 0.1 km/s is gradually terminated by the transition zone, as mentioned above. In Flueh’s models, the mid-crustal boundary is observed only on profile 2, where velocities increase from ~6.4 to ~6.6 km/s at depth of ~12 km. However, Flueh et al. do not divide the continental crust into two separated layers interpreting a structure with velocities of ~6.4-6.8 km/s as a one-layer continental crust instead. In the forward models, the continental crust is terminated seaward against a subvertical interface at ~120 km and ~140 km model distances on profiles 1 and 2, respectively. In the Flueh’s models, an extended material of the accretionary backstop overlies the continental crust at a depth of ~5 km, whereas the inverse modeling does not support this model feature.

Through the analysis of both Flueh’s models and the inverse models, I arrive at the conclusion that two different technical approaches (the forward and inverse) may lead to different results. Therefore, an assessment of the model reliability is an essential part of any modeling technique. The forward method of the wide-angle data modeling does not have the capabilities to evaluate how well the model space is resolved, how reliable and necessary the model structures are. In contrast, the traveltime inverse method of Zelt and Smith (1992) provides model assessment in terms of model resolution and uncertainty. In addition, the minimum-structure tomographic method of Zelt and Barton (1998) objectively assess the geologically
constrained inverse model obtained by the Zelt and Smith approach. The tomographic models are consistent with both inverse and forward models in revealing such structures as the slope sediments, the Valparaiso Basin, a margin wedge, the continental crust, and a layered oceanic plate. However, the tomographic assessment provides more support for the position of the transition zone between the wedge and the continental crust in the inverse models. Thus, using the combination of an inverse modeling technique and tomographic method to model the OBH data and assess the models, the inverse models for the two lines are considered more reliable and geologically reasonable compared to the forward models of Flueh et al.

6.3 Comparison with results of seismic wide-angle studies of other convergent margins

Several wide-angle seismic studies of convergent margins have been carried out worldwide. The analysis of the results of these studies gives essential insight into the complexity of the subduction process, helps establish model reliability, and evaluates the potential of the different modeling techniques such as forward and inverse methods.

Holbrook et al. (1999) conducted a seismic reflection and refraction survey of the Aleutian Island arc. Using OBH data and the traveltime inversion method of Zelt and Smith (1992), Holbrook et al. obtained a velocity model across the Aleutian arc, where the Pacific plate is subducting beneath the North American plate. This model reveals some similar structural features with the models across the Chilean margin.
These features are the upper mantle with an average velocity of ~7.8 km/s, a two-layered oceanic plate, ~2-km layer of subducting material up to ~15 km depth with velocities of ~4.3 ~5.0 km/s.

A wide-angle seismic survey across the eastern Nankai Trough (along the southeast coast of Japan) yields some results similar to this study (Nakanishi et al., 1998). Convergence near the Nankai Trough involves the subducting ridge as well. The model was constructed by forward modeling. The model shows an uppermost mantle velocity of ~7.85 km/s, and the two-layer oceanic crust. The upper oceanic crustal layer is related to the high-gradient ~3-km thick part of the crust with velocities ranging from ~4.0 to ~6.5 km/s. The lower layer of the oceanic crust with a normal average thickness of ~4.0 km is characterized by a range of velocities from ~6.7 km/s to ~6.9 km/s.

Studying the Cascadia subduction zone, Flueh et al. (1997) obtained a model that shows the subduction of the Juan de Fuca plate beneath the Olympic Peninsula. Based on forward modeling of the OBH data, a velocity of ~7.9 km/s was constrained within the uppermost mantle. Also, a two-layered oceanic crust was revealed. The ~2-km thick oceanic layer 2 has velocities ranging from ~5.0 to ~6.3 km/s, and the ~4-km thick layer 3 shows velocities of ~6.8 to 7.0 km/s.

Based on the modeling of wide-angle seismic data for the Costa Rica convergent margin offshore Nicoya Peninsula, using the inverse method of Zelt and Smith (1992), Christeson et al. (1999) modeled a two-layer structure of the oceanic crust as well. In this, region the oceanic layer 2 is up to 2.4 km thick with velocities ranging from ~4.0 to ~6.6 km/s. The lower layer 3 has a range of velocities from ~6.6 to ~7.2 km/s. Also,
the results of the Christeson et al. inverse modeling supports the existence a layer of subducting sediments that could extend to a depth more than 10km along the slab. According to authors, in the model across the Costa Rica convergent margin offshore Nicoya Peninsula, thickness of a layer of the subducting sediments varies between \( \sim 0.2 \) km and \( \sim 1 \) km. Velocities within this layer range from \( \sim 3.0 \) to \( \sim 4.0 \) km/s. Also, in their study, Christeson et al. summarized the worldwide velocity distribution within margin wedges worldwide (Table 6). In Table 6, worldwide margin wedge velocities are constrained by wide-angle seismic surveys that are relatively comparable in quality (data acquisition) (Christeson et al., 1999). According to these results, the Chilean margin offshore Valparaíso has velocities relatively similar to the Cascadia margin offshore Oregon (Fig. 35). However, for comparison it is important to consider geometry of the accretionary wedges, because the velocity (and density) distribution depends on size of the accretionary prizm.

In summary, the comparison of the results of this thesis with the velocity models across different convergent margins provides support for the existence of the subjective structures of the inverse models across the Chilean margin, such as the two-layered oceanic crust and the subducting sediments.
Table 6. Worldwide Margin Wedge Velocities from Wide-Angle Seismic Surveys

(from Christeson et al. (1999), Flueh et al. (1998), and results from this thesis)

<table>
<thead>
<tr>
<th>Model Number</th>
<th>Region</th>
<th>Velocity Range, km/s</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>20 km from trench</td>
</tr>
<tr>
<td>1</td>
<td>Costa Rica, NW of Osa</td>
<td>4.0 – 4.0</td>
</tr>
<tr>
<td></td>
<td>Peninsula</td>
<td></td>
</tr>
<tr>
<td>2</td>
<td>Costa Rica, Nicoya Peninsula</td>
<td>4.4 – 4.9</td>
</tr>
<tr>
<td>3</td>
<td>Cascadia, Oregon</td>
<td>3.0 – 4.0</td>
</tr>
<tr>
<td>4</td>
<td>Cascadia, Washington</td>
<td>3.5 – 4.2</td>
</tr>
<tr>
<td>5</td>
<td>Alaska, Kodiak</td>
<td>3.0 – 3.3</td>
</tr>
<tr>
<td>6</td>
<td>Alaska, Shumagin</td>
<td>3.2 – 4.7</td>
</tr>
<tr>
<td>7</td>
<td>Taiwan, Hengchun Peninsula</td>
<td>3.3 – 3.6</td>
</tr>
<tr>
<td>8</td>
<td>Taiwan, Ryukyu Trench</td>
<td>3.5 – 3.7</td>
</tr>
<tr>
<td>9</td>
<td>Eastern Nankai Trough</td>
<td>3.0 – 4.2</td>
</tr>
<tr>
<td>10</td>
<td>Western Nankai Trough</td>
<td>3.5 – 4.4</td>
</tr>
<tr>
<td>11</td>
<td>Chile, Valparaiso, Profile 1</td>
<td>3.3 – 3.7</td>
</tr>
<tr>
<td></td>
<td>(Flueh et al., 1998)</td>
<td></td>
</tr>
<tr>
<td>12</td>
<td>Chile, Valparaiso, Profile 2</td>
<td>3.1 – 4.0</td>
</tr>
<tr>
<td></td>
<td>(Flueh et al., 1998)</td>
<td></td>
</tr>
<tr>
<td>13</td>
<td>Chile, Valparaiso, Profile 1</td>
<td>2.9 – 3.4</td>
</tr>
<tr>
<td></td>
<td>(this study)</td>
<td></td>
</tr>
<tr>
<td>14</td>
<td>Chile, Valparaiso, Profile 2</td>
<td>3.0 – 3.6</td>
</tr>
<tr>
<td></td>
<td>(this study)</td>
<td></td>
</tr>
</tbody>
</table>
Figure 35. Comparison of worldwide margin wedge velocities from wide-angle seismic surveys: a) velocities at a ~20 km distance from a trench; b) velocities at a ~40 km distance from a trench. Velocity in km/s. Pink line indicates velocities at a top of a wedge; blue line indicates velocities at a base of a wedge. Model descriptions are given in Table 6. Red squares single out velocities that were constrained in this thesis for the two inverse models. Red dashed lines show inverse model velocities for a purpose of comparison.
7 Conclusions

2D compressional velocity models were obtained along two wide-angle reflection/refraction profiles. The lines cross the strongly varying media of the Central Chilean margin. The iterative damped square inversion method of Zelt and Smith (1992) was applied. Based on the OBH records, 12 and 14 seismic phases from 14 and 20 OBHs were identified and modeled for profile 1 and profile 2, respectively.

The results show the effectiveness of a minimum-parameter inversion to derive crustal models, generally consistent with known or expected geological features, and to assess the final models in terms of resolution and uncertainty. The reliability of the minimum-parameter models was proved by a minimum-structure tomographic assessment. The final models yield overall normalized misfit values of $\chi^2 = 0.995$ and $\chi^2 = 0.897$ for profile 1 and profile 2, respectively.

An effective strategy for the analysis of wide-angle data is the complimentary nature of minimum-parameter and minimum-structure methods for modeling and model assessment. Minimum-parameter inversion is effective for obtaining the geological model because it is designed to allow the inclusion of as much prior information as desired. Minimum-structure tomography is able to objectively determine the minimum model structure required by the data.

The final crustal models include the following structural elements: slope sediments, the Valparaíso forearc basin, subducting sediments, a margin wedge, upper and lower continental crust, a two-layer oceanic crust, and uppermost mantle. In general, the model layer boundaries are well resolved except the mid-crustal and
oceanic layer 2/3 boundaries. For the shallow structures (< 10 km) the average velocity uncertainty is ~0.2 km/s and the average depth uncertainty is ~0.3 km. Deeper, the velocity uncertainties increase to ~0.45 km/s and depth uncertainties to ~1 km.

The final models reveal significant differences between the two seismic lines. These differences are likely due to the influence of the subducting Juan Fernandez Ridge. The ridge profile 2 extends obliquely over the Juan Fernandez Ridge and is affected by ridge subduction. The off-ridge profile lies ~75 km to the south away from the subducting ridge. The Juan Fernandez Ridge terminates the sediment transport in the North-South direction along the Chilean trench and causes deformation of the subsurface structures. In comparison with profile 1, profile 2 reveals thicker oceanic crust with lower velocities; a less steeply dipping oceanic plate toward the trench and therefore, a more shallow position of the plate boundary in the vicinity of the trench axis; a more sharply bending slab landward of the trench; less extent of the subducting sediment along the slab; less shallow position of the margin wedge; more tectonic deformation within the continental margin (presence of the forearc Valparaiso Basin on the middle slope); thinner slope and trench sediments.

The results of the inversion method were compared with that of the forward modeling method (Flueh et al., 1998). This comparison shows similarities as well as significant discrepancies between the forward and inverse models. The similarities are expected since the same wide-angle data are modeled. Both forward and inverse models reveal the main structures such as the slope and trench sediments, the accretionary wedge, the continental crust, the subducted sediments, the two-layered
oceanic crust. The differences between the inverse and Flueh's models are: slightly lower upper mantle velocities in the inverse models (7.9 ± 0.1 km/s) than in Flueh's models (~8.0 km/s); decrease of the velocities within the oceanic layer 3 under the margin wedge on profile 2 in the inverse model and absence of this fact in the forward model; different interpretation of structure of the margin wedge: existence of a ~30° inclined landward transition zone between the wedge and the upper continental crust in the inverse models, and in contrast, position a subvertical boundary at same location in Flueh's models; differences in thickness and extension of the subducting sediments and structure of the continental crust. These differences could be explained by a higher degree of dependence of the results of the forward modeling from a subjective opinion of an interpreter compared to the inverse modeling technique.

It has been shown that the inversion approach has the following advantages: 1) additional information can be incorporated into the model during the modeling process through the model parameterization and a parameter-selective inversion; 2) final model resolution and uncertainty can be assessed; 3) parameter trade-offs can be assessed in such a way that a simpler model (without unnecessary structures) can be obtained.

The comparison between the final inverse models with the results of the analogous seismic crustal studies helps to understand more complexity of convergent margins, and serves to check the reliability of the models in my study area.
7.1 Future Research

The results of traveltime inversion of the wide-angle data presented in this thesis provide some knowledge of the subduction process of the Chilean margin offshore Valparaiso based on the velocity models along the two seismic lines. Future work should concentrate on the inverse modeling of all types of data together, including the vertical incidence reflection and land data, as well as more geologic interpretation that takes into account experience from convergent margin studies worldwide. It is essential to compare different convergent margins using key parameters such as the age of the subducting plate, the rate and angle of convergence, presence or absence of a subducting ridge, and amount of sediment in the trench axis.
References

Akt, K. and Richards, P. G., 1980, Quantitative Seismology, Theory and Methods, Freeman, W.H., San Francisco


Cerveny, V., Molotkov, I., and Psencik, I., 1977, Ray Method in Seismology, University of Karlova, Prague, Czechoslovakia.

Cerveny, V., 1985, Gaussian beam synthetic seismograms: J. Geophys. 58, 44-72


Seismic investigation of the continental margin off- and onshore Valparaiso, Chile, Tectonophysic, 288, 251-263.


Zelt, C. A., 1996, Seismic velocity structure of the central Chilean margin near the subducting Juan Fernandez ridge: effective inversion of traveltime data across complex, laterally varying structure, Presentation at the 7-th International Symposium on Deep Seismic Profiling of the Continents, Asilomar, CA.


