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Crustal Velocity Structure of
the Superior and Grenville Provinces of the Southeastern Canadian
Shield

by

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Submitted in partial fulfillment
of the requirements for the degree of
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ABSTRACT

In the summer of 1992, the Lithoprobe Abitibi-Grenville Seismic Refraction Experiment was conducted across the Grenville and Superior Provinces of the Southeastern Canadian Shield. More than 17,000 seismograms were collected during the survey from 44 shot points at an average shot spacing of 30 km and station spacing of 1 - 1.5 km. Each shot was recorded by 417 instruments, 47 of which were three-component systems which were placed strategically across the Grenville Front. The dense ray coverage in the area allowed us to make a detailed study of the variation of crustal velocities using seismic refraction tomography. The shallow velocity structure under the profiles was mapped using a delay-time analysis of the P_g phases. A study of wide-angle reflected waves from the Moho (P_mP) was carried out to constrain the thickness of the crust and the nature of the crust-mantle boundary. Three-component analysis was also employed to explore the anisotropic nature of the upper crust in the vicinity of the Grenville Front Tectonic Zone.

The 2-D tomographic inversion was performed using an iterative conjugate gradient technique, whereas its corresponding uncertainty in the velocity determinations were estimated using a combination of the very fast simulated annealing and calculation of the ray-path density. From the three-component data set, polarization and propagation vectors were extracted. Angular deviations between these two directions (5 - 15 degrees) were then inverted to give parameters describing the anisotropic nature of the uppermost crust in the vicinity of the receivers. The

latter was performed using a non-linear inversion technique utilizing the very fast simulated annealing.

The delay-time analysis and tomographic inversion of the data set demonstrate significant lateral and vertical variations in crustal velocities from one terrane to another, with the largest velocity values occurring underneath the Central Gneiss and the Central Metasedimentary Belts south of the Grenville Front. The Grenville Front Tectonic Zone is imaged as a southeast dipping region of anomalous velocity gradients extending to the Moho. The velocity anomaly maps suggest that Archean crust may extend, horizontally, 140 km beneath the northern Grenville Province. Near surface velocity anomalies correlate well with the known geology. The most prominent of these are the Sudbury Structure which is well mapped as a low velocity basinal structure. The tomography images also suggest underthrusting of the Pontiac and Quetico subprovinces beneath the Abitibi Greenstone Belt. Wide-angle P_mP signals indicate that the Moho varies from a sharp discontinuity south of the Grenville Front to a rather diffuse and flat boundary under the Abitibi Greenstone Belt north of the Grenville Front. A significant crustal thinning near the Grenville Front may indicate post-Grenvillian rebound and/or the extensional structure of the Ottawa-Bonnechere graben. Crustal thickening resulting from continental collision may explain the tomographic images which show that the Moho is 4 – 5 km deeper south of the Grenville Front. The inversion of the anisotropic parameters, on the other hand, indicates that the anisotropic nature of the uppermost crust in the vicinity of the Grenville Front may be caused by an aligned dry-cracks or fractures oriented subparallel to the predominant rock foliation in this region.

*To My Wife, Cicilia
My Daughter, Cindy
and My Son, Chandra*

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Chapter 1

INTRODUCTION

1.1 A General Overview

The Abitibi-Grenville project, carried out in Quebec and Ontario, is part of the LITHOPROBE program. The characterization of the crust in the Abitibi region of the Superior Province and in the Grenville Province is its major objective. LITHOPROBE is Canada's national, collaborative, multidisciplinary earth science research project created to answer fundamental questions on the nature, structure, and evolution of the continental lithosphere. The program integrates modern geophysical, geological, and geochemical concepts, methods, and technology to extend knowledge of the lithosphere, in various key areas across Canada, into the third dimension - depth (Clowes, 1993).

Why study the Abitibi-Grenville crust ? Based on geological studies, the Abitibi region of the Superior Province represents new crust that was formed and accreted rapidly towards the end of the Archean era. The Archean, the period from the formation of the Earth to approximately 2.5 Ga ago, records almost 45% of the Earth's total history. However, there is still no consensus on how the Archean Earth op-

erated. The study of the Abitibi region is obviously of great scientific interest. Studying the geometrical relationship between terranes in the Abitibi region and its associated tectonic processes, for example, may be useful in clarifying the fundamental argument as to whether the tectonic processes operated on the Archean crust resemble those that are active today.

The Grenville province, on the other hand, is largely composed of reworked crust of pre-Grenvillian age, some of which was once buried to as much as 30 km, thrust northward, and juxtaposed against the Superior Province. Based on metamorphic grade, other geological studies, and largely by analogy with other high grade metamorphic regions (e.g. The Ivrea Zone of the southern Alps, the Limpopo belt in southern Africa, and the Dharwar craton in the southern India (Fountain and Salisbury 1981)), the Grenville Province is considered to represent a relic of a Proterozoic mountain-building process which involved stacking of crustal slices comparable to that of modern continental mountain chains. These two segments of crust, the Abitibi and the Grenville, thus reflect fundamentally different processes in the evolution of the Earth. Fundamental questions concerning how the continental crust was finally stabilized, the roles of magmatic intraplate and underplating, and tectonic thickening in continental crustal growth may then be addressed by studying the Abitibi-Grenville crust.

Although many structural features in the Abitibi-Grenville region have been recognized based on surface observation, none of these is conclusive. To fully understand the origin and evolution of the crustal structures, a complete three-dimensional picture is needed. We must then rely on geophysical techniques which are capable of mapping the subsurface formations of the structures. Among other geophysical techniques, the seismic reflection technique is probably the one that provides the

highest-resolution measurements of the structure of the crust. However, aside from the fact that the seismic reflection profiling is relatively very costly, this technique cannot provide an unambiguous determination of depth, due to the lack of velocity information. Seismic refraction profiling, on the other hand, provides both information on the crustal thickness and velocity structure which are important to the characterization of crustal tectonic history. In addition, this technique also provides us with the ability to infer crustal composition as a function of depth (for example, by comparing the velocity results with laboratory measurements of seismic velocity for a wide range of rocks commonly found in the crust). The seismic refraction profiling, thus, allows us not only to reveal the crustal velocity structures but also develop a model for the composition of the continental crust.

1.2 Objectives of the thesis

Because of its tectonic significance in understanding the evolution of the continental crust, the Abitibi-Grenville region was selected as one of the main LITHOPROBE targets (Clowes, 1993). A long range seismic refraction profiling was conducted over the Abitibi-Grenville region in 1992 to obtain information on its velocity structure. In this thesis, the seismic experiment is described and the results of the analysis on the data are presented. The analysis involved seismic refraction tomography, to image the overall seismic velocity structure, combined with delay-time analysis of crustal refracted waves (P_g) to map near surface velocities better. A study of wide-angle reflected waves from the Moho ($P_m P$) was also carried out to give constraints on the thickness of the crust and the nature of the crust-mantle boundary underneath the study area. Three-component polarization analysis was also exploited to reveal the anisotropic nature of the upper crust in the vicinity of the Grenville Front

Tectonic Zone. The overall aim is to provide a detailed picture of the crustal velocity structure and of the structural development in the area.

1.3 Outline of the Thesis

In essence, the contents of the thesis may be described as follows. In chapter 1, a rationale for why this study is pursued is put forward. In the same chapter, the importance of the Abitibi-Grenville seismic refraction experiment to further constrain the model of crustal evolution in the Archean and Proterozoic is described. In chapter 2, a quick review of the geological and geophysical settings in the area under study is given. The 1992 Lithoprobe Abitibi-Grenville Seismic Refraction Experiment is fully described in chapter 3. Chapter 4 describes the strategy on how to handle the analysis and interpretation of the seismic refraction data. Descriptions of all the techniques and tools developed and used to reveal the nature of the crust and upper mantle velocity structures beneath the study area are also given in chapter 4. That chapter, thus, serves as a technical introduction for the subsequent chapters in this thesis. Chapter 5 elaborates the delay-time analysis in revealing the near-surface velocity structures. Chapter 6 describes the tomographic technique developed, tested and used in this study. Chapter 7 is dedicated to describe the application of the tomographic inversion to elucidate the deep structure of the Abitibi region and the Grenville Province of the Southeastern Canadian Shield. Analysis of the uncertainty involved in the final velocity model is also given in that chapter. Results of interpretation and its corresponding tectonic implications on the crustal evolution in the Archean and Proterozoic are given in chapter 8. Chapter 9 is entirely devoted to the analysis of the three-component data set. The anisotropic nature of the upper crust in the vicinity of the Grenville Front Tectonic Zone is

explored and discussed. Concluding remarks and recommendations for future work are given in chapter 10.

Chapter 2

GEOLOGICAL AND GEOPHYSICAL SETTINGS

2.1 Introduction

In this chapter, a brief review of the geology and geophysics of the study area is given. The discussion is organized into two main subdivisions, terranes to the north and to the south of the Grenville Front. Since this study covers a huge survey area involving many geological terranes/subdivisions, no attempt is made to give a detailed geological discussion. Readers who are interested in gaining a more detailed geological/geophysical understanding are referred to the references cited herein.

2.2 Terranes north of the Grenville Front

2.2.1 The Abitibi Subprovince

The Abitibi subprovince is located in the southeastern part of the Superior province (Fig. 2.1). This geological subprovince is the world's largest Archean greenstone terrane, which is a belt of greenschist to lower amphibolite facies metamorphosed volcano-plutonic and sedimentary rocks (Card 1990). The Abitibi greenstone belt (AGB) is bounded on the north by the Opatika and on the south by the Pontiac subprovinces. On the east and southeast, the belt is bordered by the Grenville Province. To the west, the Abitibi belt is separated from the Quetico and Wawa

subprovinces by the high-grade metamorphic rocks of the Kapuskasing structural zone (Percival et al. 1989).

The Abitibi subprovince is dominated by a young granite-greenstone belt with 80% volcanics and related intrusions and 20 % metasediments (Card 1990). The age for the volcanic rocks ranges from 2735 to 2698 Ma and 2740 to 2640 Ma for the plutonic rocks (Corfu et al. 1989; Chown et al. 1992; Kimura et al. 1993). The Abitibi and adjacent subprovinces have been variously interpreted as resulting from arc-arc collision, accreted oceanic and arc fragments and rifted arc sequences (Dimroth et al. 1983; Ludden et al. 1986; Chown et al. 1992). However, a recent view based on the Lithoprobe seismic reflection results (Ludden et al. 1993) suggested models of oblique convergence of ocean-crust relative to an accretionary complex in an island-arc sequence (see also Hoffman 1991a; Desrochers et al. 1993; Kimura et al. 1993). The validity of this hypothesis will be further tested with advent of the 1992 Lithoprobe Abitibi-Grenville Refraction data presented in this study.

2.2.2 The Pontiac Subprovince

The Pontiac Subprovince is an Archean metasedimentary and plutonic terrane (Card and Cieselski 1986) located along the southeastern margin of the Superior Province. The Pontiac is dominated in the north by an east-west band of metasediments, in the south by volcanics and metagreywackes and in the east by granitoids and gneisses (Ludden et al. 1986; Rive et al. 1990). In the north, the Pontiac subprovince is separated from the AGB by the Cadillac Larder-Lake Fault Zone (CLLFZ), while in the south, its boundary to the Grenville Province is marked by the Grenville Front Tectonic Zone (GFTZ).

The supracrustal rocks of the Pontiac subprovince are dominated by turbiditic

metasediments of the Pontiac group. U/Pb ages of detrital zircons from the Pontiac group varied from 2980 to 2683 Ma (Davis 1992; Mortensen and Card 1993). This led to an interpretation that the metasediments and metagreywackes of the Pontiac group were probably part of an accretionary prism formed after 2683 Ma (Davis 1992). This interpretation is consistent with the seismic reflection survey (Green et al. 1990; Clowes et al. 1992; Ludden et al. 1993) as well as gravity and magnetic modeling in the area (Keating 1992; Benn et al. 1994) which showed a model of underthrusting of Pontiac subprovince beneath the Abitibi subprovince. The study presented in this thesis will thus add some constraints on the nature of the crustal velocity underneath this subprovince.

2.2.3 The Sudbury structure and the Southern Province

The Sudbury structure lies in a geologically complex region of the Canadian Shield. It is located at the boundary between the Archean Superior Province, the Paleoproterozoic Southern Province, and the Mesoproterozoic Grenville Province (Fig. 2.1). The Sudbury structure, based on the generally accepted model, represents the remnants of a large meteorite impact that occurred during the Paleoproterozoic (1850 Ma) (Grieve et al. 1991). The most obvious feature of the Sudbury structure is the oval-shaped outcrop pattern of the Sudbury Igneous Complex (SIC) (Fig. 2.1). This has been interpreted as an impact-generated melt sheet that differentiated into a layered body of gabbro-norite and granophyre (Grieve et al 1991). The complex is overlaid by a layer of fall-back breccia of Onaping formation, and by sediments of the Onwatin and Chelmsford formations (Rousel 1984).

In the north, the SIC is underlain by the Archean Levack gneiss complex (Card 1994). North of the Levack gneiss, the rocks are late Archean granites and metavol-

canics of the Abitibi greenstone belt (Card and Innes 1981). South of the SIC, the rocks are Paleoproterozoic metasediments of the Huronian supergroup. The rocks are highly brecciated and contain shatter cones, providing some of the strongest evidence for an impact origin of the Sudbury structure (Dressler 1984). The southern part of the Sudbury structure may have been deformed during the Penokean orogeny and also affected by the Mesoproterozoic Grenvillian orogeny (Davidson 1992). The northern part, on the other hand, has been used to study possible impact ring structures (Butler 1994; Spray and Thompson 1995) because of the apparent lack of post-impact deformation. Line AB and its cross-line XY of the 1992 Lithoprobe Abitibi-Grenville seismic refraction experiment are dedicated to study the deep structure of this enigmatic structure. The geometrical relationship between the Sudbury structure and its adjacent structures, based on seismic velocity information, will be elucidated in this thesis.

2.3 Terranes south of the Grenville Front

2.3.1 The Grenville Province in Central Ontario

The Geology of the Grenville Province (GP) in Central Ontario has been described by several authors (e.g. Wynne-Edwards 1972; Davidson 1986; Bethune and Davidson 1988; Rivers et al. 1989; Haggart et al. 1993). In Ontario, the GP can be subdivided into two major terranes: the Central Gneiss Belt (CGB) and the Central Metasedimentary Belt (CMB). These terranes are separated from one another by major shear zones. The CGB is separated from the Paleoproterozoic rocks of the Southern Province and Archean rocks of the Superior Province by the GFTZ, while the CMB is separated from the CGB by the CMB Boundary Zone (CMBBZ) (Davidson 1986; Hanmer 1988). No attempt is made to describe the detailed geol-

ogy of the area. Only the salient features of each terrane in the vicinity of the study area will be highlighted here.

2.3.2 The Grenville Front

The Grenville Front is one of the most prominent tectonic features of the Canadian Shield (Wynne - Edwards 1972; Rivers et al. 1989). It constitutes the NE trending limit of the Grenville Province which extends approximately 1900 km from the northern shore of Lake Huron to the coast of Labrador. This trend is shown as a prominent NE trending gradient in gravity and magnetic anomalies (Rivers et al. 1989). The Front marks the northwestern limit of the Grenvillian deformation. It separates rocks deformed and metamorphosed during the Grenvillian orogeny (ca. 1000 Ma) from Archean and Proterozoic rocks to the northwest. The Front is characterized by moderately to steeply dipping NW-verging thrust faults and mylonite zone over most of its length (Davidson 1986; Owen et al. 1988). In the western part of the orogen near Georgian Bay, however, the Front is marked by a sharp increase in metamorphic grade. It increases from greenschist/lower amphibolite facies in the foreland northwest of the Front to upper amphibolite/granulite facies just southeast of the Front (Rivers et al. 1993; Haggart et al. 1993). In this part, the Front is characterized by a 15-30 km wide zone of highly strained gneiss and anastomosing shear zones with dip-parallel lineations (Rivers et al. 1993). This zone is known to be the Grenville Front Tectonic Zone (GFTZ).

Previous seismic refraction studies across the Front revealed evidence for a zone of thickening just south of the Front (Mereu and Jobidon 1971; Berry and Fuchs 1973; Mereu et al. 1986). A near vertical seismic reflection study (Green et al. 1988) and a wide-angle reflection study (Epili and Mereu 1991) across the Lake Huron/Georgian

Bay (see Fig. 2.2) imaged the GFTZ as a package of southeast dipping reflectors extending down to the lower crust. These images led the GFTZ to be interpreted as a major shear zone along which high-grade lower crustal rocks were transported to the surface (Rivers et al. 1993). These images have also added credence to a continental collision hypothesis which considered the high-grade terranes south of the Front as the deep crustal equivalents of the modern Himalayan collision belt (Windley 1986; Rivers et al. 1989). However, Lithoprobe line 15, a seismic reflection profile from Western Quebec that crossed the Grenville Front as shown in Fig. 2.2 (see also Kellet et al. 1994) does not show a similar image as seen on GLIMPCE Line-J of Green et al. (1988). One of the seismic refraction profiles (line MG) analyzed in this study partly coincides with Line 15 mentioned above. The tomographic image of the crustal velocity structure underneath line MG is presented in this thesis to address that question.

2.3.3 The Central Gneiss and Central Metasedimentary belts

The Central Gneiss Belt (CGB) of the Grenville Orogen in Ontario is bounded to the northwest by the GFTZ and to the south by the CMBBZ. The CGB is characterized by layered felsic gneisses and amphibolites with upper amphibolite to granulite facies grade of metamorphism. Easton (1986) reported that a variety of Archean to Mesoproterozoic crustal assemblages within the CGB have been affected by the Grenvillian orogeny. Evidence for high-grade regional metamorphism at 1400 Ma with the onset of synmetamorphic thrusting at 1600 - 1120 Ma has also been reported (Tucillo et al. 1992). A number of ductile shear zones that separate distinctive lithotectonic domains and subdomains in the CGB showed an indication of a northwest directed low angle ductile thrusting (Davidson 1986). This has led

some authors to interpret the CGB as a series of imbricated crustal slices assembled during the orogeny (e.g. Davidson et al. 1984).

The Central Metasedimentary Belt lies to the southeast of the CGB. The CMB consists primarily of marbles, quartzites and mafic to volcanic rocks metamorphosed in the greenschist to granulite facies during the Grenvillian orogeny. These rocks were intruded by a variety of gabbroic, tonalitic, granitic and syenitic plutons (Davidson 1986). The CMB is separated from the CGB by the CMB boundary zone. Shear indicators in this zone indicated thrusting of the CMB northwestwardly against the CGB (Davidson et al. 1991; Easton 1992).

Previous seismic refraction surveys in the CGB and CMB (Mereu et al. 1986) revealed evidence for a crustal thickening underneath the CMB. Those studies also suggested that the boundary between the CGB and CMB appeared to be a deep-seated feature. Seismic reflection data (Milkereit and Forsyth 1992; Forsyth et al. 1994a, 1994b; White et al. 1994) also showed a band of southeast dipping reflectors that can be traced to the surface. Apparently, both seismic refraction and reflection interpretations are generally consistent with models of northwest-directed thrusting and crustal shortening for the Grenville orogen in Ontario (e.g. Rivers et al. 1989). The validity of this hypothesis will be further tested with the present analysis of this seismic refraction data.

Figure 2.1: Location of the 1992 Lithoprobe Abitibi-Grenville seismic refraction survey, with simplified regional geology. CMBBZ, Central Metasedimentary Belt Boundary Zone; GFTZ, Grenville Front Tectonic Zone; CLLFZ, Cadillac Larder-Lake Fault Zone; KSZ, Kapuskasing Structural Zone; SIS, Sudbury Impact Structure (see text for descriptions).

Figure 2.2: Location of the 1992 Lithoprobe Abitibi-Grenville seismic refraction survey with respect to several previous reflection and refraction studies conducted in this area (see text for descriptions).

Chapter 3

THE SEISMIC REFRACTION EXPERIMENT

3.1 Introduction

The 1992 Lithoprobe Abitibi-Grenville Seismic Refraction Experiment was a large-scale seismic refraction and wide-angle reflection experiment which was conducted across the Grenville and Superior Provinces of the southeastern part of the Canadian Shield. This experiment was a multi-agency scientific project involving the Geological Survey of Canada, the United States Geological Survey, Geophysics GPR International and five Canadian Universities (Western Ontario, Manitoba, Ecole Polytechnique, Laval and Dalhousie). Primary tectonic targets of the experiment were the Archean Abitibi greenstone terranes of the Superior Province, the Sudbury Impact Structure (SIS), the Grenville Front (GF) and its associated Grenville Front Tectonic Zone (GFTZ), the Ottawa Graben (OG), the Central Gneiss Belt (CGB), the Central Metasedimentary Belt (CMB) and the boundary zones between them.

The experiment consisted of four seismic refraction lines (AB, XY, EF and MG). They were layed out in the form of two cross-arm arrays as shown in Fig. 2.1. The lines crossed the Grenville Front in two places, south of Sudbury and southeast of Lake Temiscaming, providing an opportunity to study along-strike variations in tectonic style of the Front (see Green et al. 1988; Rivers et al. 1993 and Kellet

et al. 1994). Several hundred kilometres of these refraction lines were also either coincident or parallel with the near vertical reflection lines of the 1991 Lithoprobe reflection experiment (see Clowes et al. 1992; Ludden et al. 1993). The southern end of line-MG also intersects the western end of the 1988 wide-angle seismic profile which runs across the CMB of the southern GP in Ontario and New York State (Hughes and Luetgert 1992; Zelt et al. 1994) (see Fig. 2.2). These allow one to correlate between the reflectivity patterns and the inferred model of the velocity structure. Considering that the Superior Province was constructed largely by the addition of mantle-derived material between 3100 and 2600 Ma (Card 1990) and that the Grenville Province is comprised of Archean and Proterozoic rock assemblages that were tectonically stacked and reworked between 1300 and 1000 Ma (Rivers et al. 1989), these seismic refraction profiles provide critical information concerning the tectonic characteristics of these two important parts of Canada's Precambrian Shield.

3.2 The Refraction Experiment

The seismic refraction experiment was conducted during the month of July 1992 across the southeastern Canadian shield in Ontario and Quebec. Four high resolution seismic refraction profiles were acquired. The experiment was coordinated from a field headquarters which was located at the Canadore College in North Bay, Ontario. Figure 2.1 shows the location of the profiles (assigned by letter) and the associated shot points (numbered). Further information on the shot points is presented in Table 3.1. The experiment was accomplished in three major deployments, with each deployment completed in about a week. The northern half of line GD and its cross-array EF were done in the first deployment. The southern half of line GD

was combined with line MD to form the long refraction profile, MG. This was done in the second deployment. The third deployment was dedicated to lines AB and XY. It is important to note that shots from EF were not only recorded on EF but also on GD receivers and vice-versa. Similarly, shots from AB were also recorded on both AB and XY and vice-versa. This configuration, thus, provides both in-line and fan shot refraction profiles. The details of the data acquisition parameters for this survey was given in an internal report by the Geological Survey of Canada (Irving et al. 1993). Only general information on the experiment will be described here.

Line EF transects from west to east, almost parallel to the general structural trend in the Abitibi Greenstone belt, for approximately 210 km in length. It begins from Cochrane, Ontario and ends in the AGB 50 km northeast of La Sarre, Quebec. This line provides information on along strike variation in crustal velocity structure underneath the AGB. Line MG, the longest line in the ABGR'92 experiment, is 620 km in length. This line runs from Belleville, Ontario and extends northwest across the CMB, the CGB, the Ottawa Graben, traverses perpendicularly the GFTZ, the Pontiac subprovince and ends in the AGB about 170 km north of Rouyn-Noranda, Quebec. The other two lines, AB and its cross-array XY, were designed to study the structure underneath the SIS and its relationship to the surrounding subprovinces including the Southern subprovince, the northern part of the Grenville Province (GP) and the GFTZ. Line AB, 280 km in length, runs almost parallel to line MG. It begins from Parry Sound, perpendicularly transects the GF, crosses the Sudbury Impact Structure (SIS) and extends further northwest into the Superior Province. Line XY, the shortest line in the experiment (170 km in length), was dedicated mainly to study the SIS.

During the experiment, more than 17,000 seismograms were collected from 44

shot points which were placed at an average shot spacing of 30 km. Each shot was recorded by 417 instruments, 47 of which were three-component systems. These three-component seismometers were deployed strategically across the Grenville Front Tectonic Zone to provide detailed information on the seismic nature of this major tectonic boundary. The average station spacing in this experiment was 1 - 1.5 km. The large number of traces per shot allowing us to use tomographic imaging techniques to analyze the data set.

Four types of recording instruments were employed during the survey. The Geological Survey of Canada provided 185 Portable Refraction Seismograph of vertical components (PRS1) and 47 three-component systems (PRS4). All geophones, vertical and horizontal components, were Mark Products L4A with a natural frequency of 2 Hz except for the PRS4-300 series which used 4.5 Hz geophone of L28LBH system. The United States Geological Survey provided 144 Seismic Group Recorder (SGR) with 2 Hz vertical geophones and 41 systems of 8 Hz geophone strings.

The locations of each shot and receiver site for lines EF, GD and MD were determined by a Global Positioning System (GPS) with a horizontal accuracy of 100 metres. An accuracy of 10 metres was achieved, however, for lines AB and XY by using the GPS in a differential mode. A total of 46 shotholes were drilled with 33 holes drilled to 35 metres and 13 holes to 43 metres. However, only 44 shotholes were actually fired. The deeper holes were loaded with 800 Kg of Hydromex T3 and shallower holes were loaded with 400 Kg. The complete list of shots for the whole experiment is shown in Table 1. All geophones were placed on rock outcrops, wherever possible. After completion of the field work, all data were then checked for time corrections and synchronized to the same sample rate of 120 sample per second, and stored on tape in SEG Y format.

Table 1: List of shots for the 1992 Lithoprobe Abitibi-Grenville Seismic Refraction Experiment.

No	Deployment	Shot ID	Shot Depth (m)	Shot Size (kg)
1	Deployment 1	GD12	43	800
2		GD3	35	400
3		GD2	35	400
4		GD5	35	400
5		EF2	35	400
6		EF3	35	400
7		GD1	43	800
8		GD4	35	400
9		EF1	43	800
10		EF5	35	400
11		EF4	35	400
12		EF6	35	350
13	Deployment 2	GD6	43	1600
14		GD8	35	400
15		MD1	35	400
16		MD6	35	400
17		GD14	35	400
18		MD8	35	300
19		GD9	35	400
20		MD2	35	400
21		MD7	35	400
22		GD7	35	400
23		GD13	35	400
24		GD11	35	400
25		MD3	35	400
26		MD5	35	400
27		GD10	35	400
28		GD12	43	800
29		MD4	43	1600
30	Deployment 3	AB0	43	800
31		AB2	35	400
32		XY0	43	800
33		XY4	35	200
34		XY5	43	800
35		AB1	35	400
36		AB3	35	400
37		XY1	35	400
38		XY2	35	400
39		AB9	35	400
40		AB8	35	400
41		AB6	43	800
42		AB10	43	1600
43		AB7	35	400
44		AB4	35	400
Data Acquisition Instrumentation		Number		
GSC - PRS1 Vertical component		185		
GSC - PRS4 Three component		47		
USGS - SGR Vertical Component		185		

Table 3.1: List of shots for the 1992 Lithoprobe Abitibi-Grenville Seismic Refraction Survey

Chapter 4

ANALYSIS AND INTERPRETATION STRATEGY

4.1 Introduction

A top-down approach of analysis and interpretation was made to handle effectively a huge seismic refraction data from the 1992 Lithoprobe Abitibi-Grenville experiment. The analysis begins with a rough and global estimate of the velocities of the crust underneath the study area, then moves to more sophisticated techniques such as a seismic refraction tomography. The analysis consists of five main steps:

- A rough estimate of the average crustal velocity gradient under the study area is the first attempt. This average velocity model, which represents the average background velocity of the crust of the Canadian Shield, is then used to obtain a standard traveltimes curve for comparison with all traveltimes obtained for each shot. This first approach, thus, attempts to obtain a rough idea on the lateral variation of velocities from one terrane to another.
- Secondly, a rough estimate of the thickness of the crust and nature of the crust-mantle boundary is carried out by studying the P-waves that have been reflected back from the Moho (PmP). This second approach was accomplished by using a PmP-composite display technique (Mereu et al, 1986) described later in this chapter.

- By employing a delay-time technique which is applied in a sliding window mode, an estimate of the lateral variation of near surface velocities was made. The fact that seismic traveltimes are highly sensitive to the near surface velocity makes this third step essential to take before one attempts to do further/deeper analysis using a tomographic technique.
- Fourth, a tomographic imaging technique is developed and employed to map the detailed velocity structure of the deeper portion of the crust and upper mantle.
- Fifth, a polarization analysis is carried out specifically to handle the three-component array data.

Two out of five steps listed above are described in this chapter. The other three methods, however, are discussed in chapters 5, 6, and 7.

4.2 Tools/ Softwares employed to aid the analysis

A series of tools which designed specifically for seismic refraction data was used to handle efficiently the huge amount of data that resulted from this experiment. The printer-plotter program (Mereu et al. 1990) was revised and updated, enabling one to easily send and plot large refraction record sections from a SUN Workstation to an ink-jet and/or laser printer. The program allows one to plot the refraction data using a variable area method similar to seismic reflection record sections. Such a type of display sometimes enables one to see better a large-scale refraction and wide-angle reflection patterns which cannot be seen easily on the conventional record section. Examples of typical record sections for lines AB, XY, MD, GD, and EF, created using this program, are shown in Figs. 4.1 – 4.9.

An interactive program, written by Prof. R. Mereu, which enables a user to easily pick first and later arrivals from the refraction data sets on the screen display was also employed. The program used a combination of automatic and manual methods in picking the onset times. The picks resulted from the automatic picking are displayed on the computer screen and the user can then check and modify them (if necessary) using the arrows keys. The program was made for the ease of use in that it enables a user to change distance, time and amplitude scales with only a touch of a key. This feature allowing us to zoom-out and zoom-in on the screen display to check the consistency of the picks. The final results of picks were then saved into a file which can easily be accessed by other programs, such as a delay-time routine or a tomographic routine, for further analysis.

An interactive ray-tracing program developed by Mereu (1990) was also used to create several good starting models which are necessary for the tomographic inversion. This program enables one to create a crustal model and trace seismic rays through the model. In fact, this ray-tracing routine served as the core of the forward modeller for the tomographic inversion routine developed in this thesis. This particular topic will be discussed further in chapters 6 and 7.

A set of polarization routines (Winardhi 1994; Winardhi and Mereu 1994), designed specifically to handle the three-component data, was employed to extract the direction of polarization of seismic phases and other three-component attributes such as degree of polarization and rectilinearity. This is of particular importance for the study of the upper crust in the vicinity of the Grenville Front Tectonic Zone. This particular study will be discussed further in chapter 9. These polarization routines, which make use of the polarization property of the seismic signals, can also be used to enhance the signal to noise ratio and aid in phase/arrival identification.

Readers interested in the polarization filtering are referred to Winardhi (1994).

4.3 A least squares travelttime curve

It is often desirable to understand a general/average feature first before going into a more detailed analysis. This scheme was used in this study to show the existence of lateral variations in crustal velocities as observed from the travel time data. For this purpose, the average crustal model was derived by performing a least squares analysis on the complete first arrival observations. A smooth curve, which is the least squares travelttime curve, as shown in Fig. 4.10, was then used as a standard curve for comparing with the travelttime observations from several profiles. This average travelttime curve actually represents the theoretical travelttime expected for a crustal velocity model with constant vertical velocity gradient but without lateral velocity variation. This average crustal model was later used in seismic tomography to obtain the velocity anomaly maps. The average curve obtained in this study yielded a crustal model within which the velocity varies linearly with depth from 6.13 km/sec at the surface to 7.10 km/sec at a depth of 40 km. A similar study was also performed on the S wave arrivals. It produced a model with the velocity varied from 3.56 km/sec at the surface to 4.11 km/sec at the 40 km depth. The average Poisson's ratio derived from this study was 0.245 ± 0.005 . A plot of the complete S wave arrivals is shown in Fig. 4.11.

The existence of a significant lateral velocity variation from one terrane to another is illustrated in Fig. 4.12. It can be seen that the arrival times on the profile shot from MD4 (located in the Grenville Province) are generally earlier than the standard curve. On the other hand, the reversed profile from shot GD6 (located in the Archean Abitibi Greenstone terrane) shows arrivals that come later than the

average curve. Figures 4.13 – 4.15 show similar observations for line AB, XY and EF, respectively. It is seen that most of the arrivals which were shot within the Grenville Province are generally earlier than the average curve, whereas, the reversed profiles from shots situated in the Abitibi Greenstone terrane show arrivals that come later than the standard curve. These observations clearly demonstrate the significant lateral variations in crustal velocities from one terrane to another. Several observations from other shots also support this finding.

4.4 The nature of the Moho from P_mP waves

A portion of the record sections shown in Figs. 4.16 and 4.17 show the nature of the Moho underneath shot MD8 and GD6. The Moho appears as a strong reflector, as indicated by a strong and clear P_mP waves, on the observed record section from shot MD8. However, the Moho tends to become transparent or diffuse on the section from shot GD6. To get a general idea of the roughness/topography of the crust-mantle boundary and on the variation in its nature from one terrane to another, a special method of displaying the P_mP waves was employed (see Mereu et al. 1986).

First, normal moveout (NMO) correction was applied to all the traces. The resulting traces were then plotted at the half-way between the shot and recording stations. This is justified as the approximate locations where the P_mP waves would sample the Moho, assuming the average crustal velocity to be constant or only a function of depth. A P_mP composite section was then produced by using all the traces from all the shots to a profile. These traces were also plotted with the time axis increasing downward, mimicking the way the seismic reflection data are displayed. The NMO corrections used for these plots were computed using the root mean squares (rms) velocity based on the average crustal model obtained from the least

squares analysis described previously. Figures 4.18 – 4.20 and 4.21 show the $P_m P$ composite section underneath line MG and AB, respectively, obtained using this technique. The $P_m P$ composite sections for lines XY and EF were not created due to the short profile. Analysis of the information seen on these $P_m P$ composite plots was later incorporated as an additional information for the tomographic inversion described later in this thesis.

This simple analysis, as it is seen in Figs. 4.18 – 4.21, reveals a significant variation in the nature and topography of the Moho. Underneath the Archean Abitibi Greenstone terrane, the Moho is generally transparent (see Fig. 4.20) compared to areas beneath the Grenville Province which tend to show the Moho as a strong reflector (see Fig. 4.18). The very strong $P_m P$ was observed to occur right underneath the surface expression of the Grenville Front in which the depth to the Moho appears to be very shallow (see Fig. 4.19). The relatively deeper Moho underneath the Grenville Province provides evidence for a thickening of the crust south of the Front. The south-east dipping features of the crust are in agreement with those observed by Forsyth et al. (1994a, 1994b), White et al. (1994) and Roy and Mereu (1996).

Figure 4.1: Record section for line AB from shot AB0.

Figure 4.2: Record section for line AB from shot AB10.

Figure 4.3: Record section for line XY from shot XY0.

Figure 4.4: Record section for line XY from shot XY5.

Figure 4.5: Record section for line MD from shot MD1.

Figure 4.6: Record section for line MD from shot MD4.

Figure 4.7: Record section for Line GD from shot GD14.

Figure 4.8: Record section for Line EF from shot EF1.

Figure 4.9: Record section for Line EF from shot EF6.

Figure 4.10: Composite plot of complete P_g arrival time observations. Smooth curve is the least squares travel time curve. It represents the travel time curve for a crust in which the velocity increases linearly from 6.13 km/s at the surface to 7.1 km/s at a depth of 40 km.

Figure 4.11: Composite plot of complete S_g arrival time observations. Smooth curve is the least squares travel time curve. It represents the S-wave travel time curve for a crust in which the velocity increases linearly from 3.56 km/s at the surface to 4.11 km/s at a depth of 40 km.

Figure 4.12: Arrival time observations for P_g signals for shots GD6 and GD10 of line GD and shots MD4 and MD8 of line MD.

Figure 4.4: Arrival time observations for P_g signals for shots AB0, AB4, AB9 and AB10 of line AB.

Figure 4.5: Arrival time observations for P_g signals for shots XY0, XY2, XY4 and XY5 of line XY.

Figure 4.15: Arrival time observations for P_g signals for shots EF1, EF0, EF5 and EF6 of line EF.

Figure 4.16: Record section for line MD from shot MD8.

Figure 4.17: Record section for line GD from shot GD6.

Figure 4.18: The normal moveout (NMO) corrected $P_m P$ composite section for the south portion (40 – 240 km) of line MG.

Figure 4.19: The normal moveout (NMO) corrected $P_m P$ composite section for the south portion (240 – 440 km) of line MG.

Figure 4.20: The normal moveout (NMO) corrected $P_m P$ composite section for the south portion (440 – 640 km) of line MG.

Figure 4.21: The normal moveout (NMO) corrected $P_m P$ composite section for line AB.

Chapter 5

DELAY-TIME ANALYSIS

5.1 Introduction

It has been stated in the previous chapter that seismic travel time is highly sensitive to near-surface velocity variations. A sudden jump in the near surface velocity creates a big bump in the observed travel time, whereas a similar jump in the deep crust can hardly be noticed from a single travel time observation. Separating the near-surface effect from the deeper crust contribution to the observed travel times is therefore suggested to produce a better map of the overall velocity structure of the crust.

Figure 5.1 shows examples of the picked P_g arrivals/travel time data for line AB. Two distinct fluctuations of the travel time curve are observed. Velocity variation in the deeper portion of the crust are responsible for the long wavelength/low frequency fluctuation, whereas the near-surface velocity structure has more contributions to the high frequency part. A closer examination of the scatter of points in the individual travel time data also indicated that much of the small scale scatter could be attributed to significant lateral variations in velocity in the near surface rocks. It is these rocks which are easiest to correlate with the observed geology. Since in general the near surface can be modelled as a low velocity layer overriding

a much higher velocity layer, a simple least squares delay-time method to estimate these velocity variations is therefore employed.

5.2 Least squares Delay-time Analysis of the P_g Arrivals

The delay-time method, first proposed by Scheidegger and Willmore (1957), is actually a tomographic method which simultaneously makes use of all the ray paths which pass through any given region to estimate the velocities. The fundamental principle of the delay-time (sometimes called time-term) analysis is very straightforward. The delay-time associated with the path SMNG in Fig. 5.2 is the observed refraction time at G, t_g , minus the time required for the wave to travel from P to Q (the projection of the path on the refractor) at the velocity V . Writing δ for the delay time, we have

$$\begin{aligned}
 \delta &= t_g - \frac{PQ}{V} \\
 &= \left(\frac{SM + NG}{V_0} + \frac{MN}{V} \right) - \frac{PQ}{V} \\
 &= \left(\frac{SM + NG}{V_0} \right) - \left(\frac{PM + NQ}{V} \right) \\
 &= \left(\frac{SM}{V_0} - \frac{PM}{V} \right) + \left(\frac{NG}{V_0} - \frac{NQ}{V} \right) \\
 &= b + a
 \end{aligned} \tag{5.1}$$

where b and a are known as the source (or shot) delay time and the geophone (or station) delay time because they are associated with the portions of the path down from the shot and up to the station.

An approximate value of δ can be found by assuming that the dip is small enough that PQ is approximately equal to the geophone offset Δ . Provided the dip

is less than 10 degrees, this approximation is sufficiently accurate for most purposes (Sheriff and Geldart, 1995). In this case, for a set of refraction data, the travel time equation can be assumed to be of the form

$$t_{ij} = \frac{\Delta_{ij}}{V} + a_i + b_j \quad (5.2)$$

where t_{ij} and Δ_{ij} are the travel-time and distance from shot j to station i , V is the refractor velocity and a_i and b_j are the corresponding station and shot delay times respectively. Since the number of shot-station pairs is usually more than the number of unknowns (i.e. a_i and b_j), the problem of determining these unknowns becomes a least squares problem of solving the normal equations. In this study the normal equations were solved using the Gauss-Seidel iterative method similar to the one used by Mereu (1966).

In this analysis, a window of length 130 km was chosen. Tests showed that a smaller window yielded unstable results. For the method to work, rays must bottom within the higher velocity layer. Tests also showed that larger distances resulted in rays penetrating the model to depths which were regarded as being too deep for a shallow analysis. The 130 km windows were slid at 10 km intervals along the model to obtain values of a_i and b_j and V for each window. Figures 5.3 – 5.6 show a composite plot of a_i values for the full length of lines AB, XY, EF and MG. The consistency of the delay-time features from window to window indicated the stability of the method in revealing the near surface structures. Major physical discontinuities such as a fault typically manifest themselves as abrupt discontinuities/anomalies in the a_i values. The final a_i s and b_j s were obtained by averaging the overlapping values. The procedure also showed how V varied laterally along each of the profiles.

To translate the delay times into apparent depths d_i , the simple refraction formula

$$d_i = a_i \frac{VV_0}{\sqrt{V^2 - V_0^2}} \quad (5.3)$$

was used. Here V , the refractor velocity, was obtained from the delay-time analysis and V_0 the velocity at the top of the model was estimated from a least squares analysis of the P_g arrivals within 20-30 km from the shot point. The observed times were checked to ensure that a proper distance range was used in calculating the values of V_0 .

Figures 5.7 and 5.8 show the results of the delay-time analysis on all 4 lines AB, XY, MG and EF along with the geology of the area. These diagrams demonstrate the clear correlation of some of the low and high velocity structures with the known geology. Further comments are given in the next section.

5.3 Discussion of the Results

5.3.1 Delay-time analysis of lines AB and XY

The delay time analysis shown in Figs. 5.7 and 5.8 reveals the topography of a higher velocity layer beneath 0 - 3 km of lower velocity surficial layer. Figure 5.7, for example, shows the basinal structure of the Sudbury Impact Structure, where the low velocity layer corresponds to the sedimentary rocks of the Whitewater Group. Beneath line XY, the high velocity layer corresponds well to the Gabbro Norite and Granophyre sublayer of the Sudbury Igneous Complex (SIC) (see Fig. 5.7b), whereas, beneath line AB, this is not entirely the case. The image also shows that the south part of the Sudbury Impact Structure (SIS in Fig. 5.7a) must have undergone significant deformation resulting in an unsymmetrical shape of the SIS. This interpretation is in general agreement with the interpretation based on seismic

reflection study of Wu et al. (1994). However, in contrast to their model which showed that the SIC is well elongated to the south, we interpret that the SIC, more or less, mimics the basinal structure shown in Fig. 5.7. This is based on the fact that at surface, the outcrop of the SIC is a continuous roughly oval structure (G.M. Young and W.R. Church, personal communication, 1995; Card and Jackson 1995).

Refractor velocity profiles, obtained from the analysis, demonstrate the existence of other lateral velocity variations in the uppermost crust of the survey area. Figure 5.7a, for example, clearly shows that the Central Gneiss Belt rocks of the Grenville Province has, generally, higher velocities than the Archean Superior Province to the north. This high velocity result is consistent with the high metamorphic grade rocks commonly observed in this area, supporting the idea of lower crustal exhumation.

Figure 5.7b, on the other hand, shows that the uppermost crust of the Southern Province has relatively lower velocities compared to the Archean Superior crust to the west. This low velocity region correlates with the metasedimentary rocks of the Huronian Supergroup of the Southern Province.

5.3.2 Delay-time analysis of lines MG and EF

The near surface structure beneath lines MG (Fig. 5.8a) also augments several known geological structures. The paleozoic cover near the south end of line MG is shown to decrease gradually to the north. The Central Metasedimentary Belt is shown to decrease gradually to the north. The Central Metasedimentary Belt is dominated by high velocity rocks. This is indicated by the refractor beneath it which occurs at a very shallow depth. The Central Metasedimentary Belt Boundary Zone (CMBBZ) is marked by a refractor discontinuity in the vicinity of the shot points MD4 and MD5. A somewhat discontinuous features in the vicinity of shot point GD14 may correlate with the Ottawa-Bonnechere Graben (Kay 1942; Kumarapelli

and Saul 1966). Three main tectonic zones, the Grenville Front Tectonic Zone (GFTZ), the Cadillac - Larder Lake Fault Zone (CLLFZ) and the Casa-Berardi Tectonic Zone (CBTZ), are also well mapped as three refractor discontinuities in the vicinity of shot points GD11, GD6-GD7 and GD1-GD2, respectively.

Refractor velocity profiles beneath line MG, again show that the Grenville Province has generally higher velocities compared to the Archean Superior to the north. Beneath the Paleozoic cover, a relatively high velocity upper crust is observed, providing evidence for the extension of the CMB crust beneath it. The upper crust of the CGB in the vicinity of the Ottawa River, on the other hand, is dominated by the Ottawa-Bonnechere graben which has a relatively lower velocity as compared to the upper crust of the CMB. A gradual decrease in the uppermost crustal velocities from south to north is also observed. The upper crust of the Pontiac subprovince, which is dominated by the metasedimentary rocks, is also shown to have a relatively lower velocity as compared to the Archean volcanic rocks of the Abitibi subprovince.

In addition, refractor topography beneath line EF also reveals the known surface geology in this study area. The low velocity layer of the Quetico metasedimentary rocks is shown to thicken over the upper boundary/refractor beneath line EF in the vicinity of the shotpoints EF4 and EF3 (see Fig. 5.8b). This, in fact, marks the juxtaposition between the Archean Abitibi and the Quetico subprovince.

The refractor velocity of this east-west profile (line EF), moreover, shows a generally higher velocities (+ 0.07 km/s) than the one obtained from its cross line (MG). This suggests anisotropy of the uppermost Archean Abitibi crust with the fast-velocity direction being east-west, subparallel to the general structural trend in this subprovince. This result is consistent with the surface geology in this area showing two sets of transcurrent faults in the crust, trending N 90 E and N 120 E

(see Ji et al., 1996), and also consistent with the shear direction (N 103 E) of the upper mantle shear zone beneath the Archean Abitibi as inferred from the teleseismic shear-wave splitting studies (Ji et al., 1996).

5.3.3 Correlation with the gravity data

Additional support for the existence of the near surface structures as revealed by the delay-time analysis is provided by gravity data. The data was made available by the Geological Survey of Canada. Figures 5.9 and 5.10 show the free-air and Bouguer gravity maps in the vicinity of the study area, whereas Figs. 5.11 – 5.14 show the four particular profiles which coincide with our four refraction lines, AB, XY, MG and EF, respectively.

It is observed in Figs. 5.11 – 5.14, that the gravity data basically consists of two distinct short and long wavelength fluctuations. The low frequency/ long wavelength component may usually be regarded as due to the regional field, whereas the high frequency/short wavelength fluctuation is usually caused by the residual or anomalous field due to lateral density variations in the uppermost part of the crust. In fact, the observed similarities in the high frequency fluctuations pattern between the free-air and the Bouguer gravity further indicates that this high frequency component is mainly caused by the near surface disturbances and not due to the effect of terrane's topography. This is justified since the free-air correction only accounted for the variation in elevation of the measurements, but not accounted for the additional mass of the topographic edifice, whereas the Bouguer correction has accounted for both elevation and the additional mass. If we wish to correlate between the gravity data and the near surface lateral velocity variations, the residual field has thus to be first separated from the regional one. This is done straightforwardly by fitting the

regional trend using polynomial function and then subtracting it from the Bouguer gravity to get the residual gravity anomaly.

It is interesting to note, from visual inspection of the gravity data, that most features revealed by the delay-time analysis are also observed to correspond well with the gravity data. It is seen from Fig. 5.11, for example, that a gradual decrease, from south to north, in the uppermost crustal velocities beneath profile AB may be explained in terms of a gradual decrease in crustal densities (as represented by a gradual north dipping trend in the Bouguer gravity profile). The refractor topography (as seen in Fig. 5.7a), on the other hand, was found to be more or less mimicked the residual gravity data (see the lower panel of Fig. 5.11).

Figure 5.12 shows similar observation for profile XY. The low velocity basinal structure of the SIS is seen to correspond well to the low residual gravity anomaly in the middle part of the profile. The high velocity layer of the Gabbro Norite sublayer, on the other hand, nicely corresponds to the high residual anomalies at the two edges of the basinal structure. It is interesting to note here that the basinal structure of the SIS is shown better in the east-west (XY) than the south-north (AB) profiles. This is obviously due to its dimension which is wider (about 60 km) in the east-west cross-section as compared to its south-north section (which is only about 20 km wide).

Figure 5.13 shows the Bouguer and free-air gravity anomaly profiles for line MG. A general correlation between the delay-time features (see Fig. 5.8) and the gravity data is again observed. Several near surface features described previously such as the Paleozoic cover, the CMBBZ, the Ottawa-Bonechere graben, the GFTZ and the CLLFZ are all observed to be well correlated with the residual gravity anomaly.

Figure 5.14 also shows nice correlation between the delay-time results and the

gravity data for line EF. It is also seen here that the pattern of the residual gravity anomaly more or less mimics the pattern of the refractor topography, whereas the gradual decrease in the refractor velocities corresponds to the gradual decrease in the observed Bouguer gravity. It is also interesting to note here that the boundary between the Archean Abitibi and the Quetico subprovince, which was previously shown by the delay-time analysis as the near surface depression, was also marked by a distinct depression in the gravity data.

Figure 5.1: Examples of the picked P_g arrivals/travel-time data for line AB. Much of the small scatter can be attributed to significant lateral variations in the velocity of near-surface rocks.

Figure 5.2: Diagram showing the ray that travels from the source (S) down to the refracting horizon (MN) and gets refracted back to the the recording station/geophone (G)..

Figure 5.3: A composite plot of a_i values for the full length of line AB.

Figure 5.4: A composite plot of a_i values for the full length of line XY.

Figure 5.5: A composite plot of a_i values for the full length of line EF.

Figure 5.6: A composite plot of a_i values for the full length of line MG.

Figure 5.7: Upper refractor topography and its corresponding upper and lower layer velocities obtained from the delay-time analysis of *Pg* waves for Lines AB (a) and XY (b). Numbers inside models indicate P-wave velocity in kilometres per second.

Figure 5.8: Upper refractor topography and its corresponding upper and lower layer velocities obtained from the delay-time analysis of P_g waves for Lines MG (a) and EF (b). Numbers inside models indicate P-wave velocity in kilometres per second.

Figure 5.9: Free-air gravity map of the Superior and Grenville Provinces of the southeastern Canadian shield.

Figure 5.10: Bouguer gravity map of the Superior and Grenville Provinces of the southeastern Canadian shield.

Figure 5.11: Free-air, Bouguer (upper panel) and residual (middle panel) gravity profiles which coincide with the 1992 Lithoprobe Abitibi-Grenville refraction line AB (lower panel).

Figure 5.12: Free-air, Bouguer (upper panel) and residual (middle panel) gravity profiles which coincide with the 1992 Lithoprobe Abitibi-Grenville refraction line XY (lower panel).

Figure 5.13: Free-air, Bouguer (upper panel) and residual (middle panel) gravity profiles which coincide with the 1992 Lithoprobe Abitibi-Grenville refraction line MG (lower panel).

Figure 5.14: Free-air, Bouguer (upper panel) and residual (middle panel) gravity profiles which coincide with the 1992 Lithoprobe Abitibi-Grenville refraction line EF (lower panel).

Chapter 6

TOMOGRAPHIC INVERSION TECHNIQUE

6.1 Introduction

This chapter describes various stages involved in the development of the tomographic inversion technique designed and employed in this study. Seismic tomography basically consists of the following steps:

- Creating a reasonable starting model. This step, of course, deals with how to parameterize the model.
- Shooting rays through this model to create a set of theoretical travel times. This step involves how to simulate the nature of the seismic waves in travelling down from the source through the medium and back again to the surface (recording station).
- Carrying out an optimization process which tries to fine tune the model such that the misfit errors between the observed travel times and theoretical travel times are minimized. This step is the core of the inversion technique in the seismic tomography.

6.2 Model parameterization and Forward Modelling

The effectiveness of a ray tracing method for use in tomographic inversion depends on how a velocity model was parameterized. Wesson (1971), for example, used a smooth functional representation of the medium. Lutter and Nowack (1990) applied spline interpolation of velocities as well as interfaces which were defined at nodes (see also Cervený and Psencik, 1984). Zelt and Ellis (1988) and Zelt and Smith (1992), parameterized their models using trapezoidal blocks. In this study, in order to handle laterally varying structures within the whole crust, the model was parameterized using a set of triangular blocks as described by Mereu (1990) (see also White, 1989). Each block in this model was assigned its own 2-D linear velocity function

$$V(x, z) = ax + bz + c \quad (6.1)$$

where $V(x, z)$ is the seismic velocity at any point (x, z) in the model. The constants a , b and c are determined from velocity values assigned at each node of every triangle. These three constants are different for each triangle. A constant velocity gradient of

$$\nabla V = a\hat{i} + b\hat{k} \quad (6.2)$$

may therefore be computed in each triangle where $\hat{i}$ and $\hat{k}$ are unit vectors in the x and z directions, respectively. The constant gradient in the velocity field ensures that all the ray paths traced through each of the triangular blocks will be arcs of circles. This enables one to rapidly and accurately trace rays through the velocity model.

Furthermore, this constant gradient approach allows us to model all first arrivals easily without specifically identifying individual refractions. This offers a more flex-

ible modelling approach as compared to the conventional one. The rationale is as follows: in the conventional refraction analysis, first arrivals for a given refraction segment are always assumed to increase linearly with distances. As a consequence, prior to modelling the data, one must distinguish each individual refraction segment on the first arrival observations. However, as seen in previous chapters, first arrivals rarely appear to be composed of a series of distinct linear segments. Clearly, this shows the flexibility of the model parameterization employed in this study compared to the conventional approach.

The ray tracing is accomplished by solving ray arc equations with triangular boundary line equations and applying Snell's law at each intersection point. The triangle is chosen for the model building block because it is very effective in modelling lateral structures and dipping faults. The manner in which the model is divided into triangles, however, is not unique. An example of theoretical traveltime curves for a simple two-layer crustal scale model obtained using the method described above is shown in Fig. 6.1. Further details on this forward method are given by Mereu (1990).

6.3 Tomographic Inversion

"Tomography can be defined as the reconstruction of a field from the knowledge of linear path integrals through the field. In seismology, the analysis of lateral velocity variations fits this definition, if the travel time equation is perturbed about a reference velocity model. The field in this case is slowness perturbations, and the observations are travel time deviations"

(Clayton, 1984).

6.3.1 Fundamental concept

To clarify the concept of seismic tomography quoted above, consider a very simple case in which a wave travels in a straight line in a medium with a constant velocity V . The travel time is then simply obtained by multiplying the slowness s by the distance traveled d , where the slowness is defined as the reciprocal velocity ($1/V$). If the wave travels a total distance l , a formula for the travel time can then be expressed as:

$$\begin{aligned} t &= \int_0^l s \, dl \\ &= \int_0^l \frac{1}{V} \, dl \end{aligned} \quad (6.3)$$

If we now proceed further by assuming that the total distance l is in fact a ray path along which the wave travels in the Earth, and the slowness value varies along the path, a generalized version of the above formula can then be given as:

$$t_{ray} = \int_{raypath} \frac{1}{V(x, y, z)} \, dl \quad (6.4)$$

where x , y and z represent the horizontal and the vertical coordinates, respectively. The core of the tomographic problem is to derive $V(x, y, z)$ from a number of travel time measurements at the surface. It seems easy. Equation (6.4) is, however, a very complicated problem, because the unknown $V(x, y, z)$ is also implicitly present in the raypath l . This makes the tomographic inversion a highly nonlinear inverse problem and nonlinear equations are difficult to solve.

Fortunately, we usually have a fairly good idea on what $V(x, y, z)$ is like. Spherically symmetric Earth models such as those developed by Jeffreys and Bullen in the 1930's, for example, are able to predict travel times with quite high accuracy (Nolet, 1987). This indicates that the symmetric Earth can actually be used as a starting model which can then be improved further by allowing small deviations from this symmetry. This approach serves as the fundamental concept of seismic tomography. This is why, equation (6.4) is usually solved by first linearizing it in the vicinity of a reference model (i.e. a model which could approximately produces the observed travel times).

6.3.2 Formulation of the Tomographic Problem

As it has been mentioned in the previous section, for crustal study, seismic velocities are usually assumed to increase with depth. As a consequence, curved ray paths are normally taken as a better approximation than straight lines (or paths). Consider now, a wave travelling in a medium with velocity $V_0(\mathbf{x})$, along a curved ray path. Its travel time, following White's (1989) notation, can then be given as:

$$t(V_0(\mathbf{x})) = \int_{L_0} \frac{dl}{V_0(x)} \quad (6.5)$$

where L_0 represents the curved ray path. If the velocity of the medium is slightly perturbed by a value $\delta V(\mathbf{x})$ such that $V(\mathbf{x}) = V_0(\mathbf{x}) + \delta V(\mathbf{x})$, the travel time and the ray path are also perturbed such that

$$t(V(\mathbf{x})) = \int_L \frac{dl}{V(\mathbf{x})} \quad (6.6)$$

where L is now the new ray path. The travelttime through the perturbed medium can, however, be expressed as an expansion about the reference medium $V_0(\mathbf{x})$ as

$$\begin{aligned} t(V(\mathbf{x})) = & t[V_0(\mathbf{x})] + \frac{\partial t}{\partial V}|_{V=V_0} \delta V(\mathbf{x}) + \\ & \frac{\partial t}{\partial L}|_{V=V_0} \frac{\partial L}{\partial V}|_{V=V_0} \delta V(\mathbf{x}) + \\ & O[\delta^2 V(\mathbf{x})] \end{aligned} \quad (6.7)$$

Using Fermat's principle which states that the travelttime for the ray path is stationary (i.e. $\frac{\partial t}{\partial L} = 0$), and discarding terms of $O[\delta^2 V(\mathbf{x})]$, we then obtain

$$\delta t = t[V(\mathbf{x})] - t[V_0(\mathbf{x})] = \frac{\partial t}{\partial V}|_{V=V_0} \delta V(\mathbf{x}) \quad (6.8)$$

Since we use a discrete (as opposed to continuous) model formulation, in which the velocity model is specified at the nodes of a triangular grid by N discrete values V_j , $j = 1, 2, \dots, N$, equation (6.8) can then be written as

$$\delta t = \sum_{j=1}^N \frac{\partial t}{\partial V_j} \delta V_j \quad (6.9)$$

For a set of M rays (observations), a linear system of equations can then be formed from this equation to yield

$$\Delta \mathbf{T} = \mathbf{A} \Delta \mathbf{V} \quad (6.10)$$

where $\Delta \mathbf{T}$ is the vector of traveltime residuals (i.e. $T_{observed} - T_{theoretical}$), $\mathbf{A}$ is the Jacobian matrix of the partial travel time derivatives $\partial T_i / \partial V_j$, and $\Delta \mathbf{V}$ is the velocity perturbation vector. It can be seen from equation (6.10) that, based on $\Delta \mathbf{T}$, tomographic inversion attempts to solve for $\Delta \mathbf{V}$. However, in order to solve for $\Delta \mathbf{V}$, one must be able not only to calculate the traveltime residuals, but also to construct the matrix of partial traveltime derivatives.

6.3.3 Construction of the System of Linear Equations

The most important step for the construction of the matrix element A_{ij} is the calculation of the ray-path. This is accomplished using the ray-tracing program developed by Mereu (1990) as discussed in the previous section. As is shown in Fig. 6.2, each node in the velocity model is labelled/numbered. A set of rays is then traced through the model with the number of rays is chosen to be, at least, as great as the number of observations (or stations). Then, one of the nodal velocities is perturbed, and another set of rays is traced again through the model.

Compared to the initial unperturbed set, the perturbed velocity will eventually yield slightly different travel times. With a given ray and the nodal velocity perturbation, it is then possible to begin constructing the matrix element A_{ij} using the deviations in the measured travel times observed at a particular receiver location. If, for example, the travel time deviation for ray $i = 1$ is ∂T_1 , and the third nodal velocity is perturbed by ∂V_3 , then $\partial T_1 / \partial V_3$ is the value for $A(1, 3)$. If one has M number of observations, and N numbers of velocity nodes in the model, then matrix A will be of size M by N .

It can be seen from the process described above, that element of matrix A_{ij}

actually represents the sensitivity of the travel time to a particular model parameter. This is why matrix A is sometimes called the sensitivity matrix. With the matrix A completed, the differences between the observed and the calculated (or theoretical) travel times can then be used to solve for new nodal velocity perturbations. These values are added to the initial nodal velocity values to yield a new/updated model. The process iterates until a satisfactory updated model is achieved.

6.3.4 Solving the System of Linear Equations

The solution of the system of linear equations $\Delta T = A\Delta V$ is however not straightforward. There are a couple of things to note, before one tries to solve this set of equations.

- First, in a real problem, the equations are expected to be inconsistent since data usually contain small or large errors. Thus, $A\Delta V = \Delta T$ cannot be fit exactly. Instead, one must aim for a minimization of the misfit in some particular norm.
- Secondly, despite the fact that M is usually greater than N , the system of equations are almost certainly ill-conditioned. This is because the velocity in at least one nodal point, can usually be freely varied without significantly affecting the fit to the the system of equations. One must then damp the solution to keep the propagation of errors under control. However, damping always affects the fit to the equations. Choosing a preferred solution, therefore, involves a trade-off between the goodness of fit and the reliability of the solution.

- Finally, one must realize the importance of scaling the solution of the system of equations for different data reliabilities or errors. This is simply because data or observations cannot all be treated equally. Data near the shot point should be regarded as the more reliable information compared to the far away data. As such, the former should have a much more influence in the solution of equations than the latter.

To address the first item, the least square norm is used in this study. As such, problem stated in equation (6.10) is replaced by the minimization problem

$$\min \|\mathbf{A}\Delta\mathbf{V} - \Delta\mathbf{T}\|^2 \quad (6.11)$$

This is done, equivalently, by simply premultiplying equation (6.10) with the transpose of the $\mathbf{A}$ matrix to form

$$\mathbf{A}^{tr}\mathbf{A}\Delta\mathbf{V} = \mathbf{A}^{tr}\Delta\mathbf{T} \quad (6.12)$$

Equation (6.12) completely characterizes the least-squares solution of equation (6.10) (Aki and Richards, 1980; Menke, 1989). If matrix $\mathbf{A}$ has as many rows as columns and is of full column rank, then the least-squares solution of equation (6.10) is unique. On the other hand, if equation (6.10) is underdetermined or rank deficient, the least-squares solution is not unique.

To address the second item, the minimization problem of (6.11) is further replaced with

$$\min \|\mathbf{A}\Delta\mathbf{V} - \Delta\mathbf{T}\|^2 + \lambda^2 \|\Delta\mathbf{V}\|^2 \quad (6.13)$$

where λ is the damping constant. The normal equations associated with this new damped objective function are given as (see chapter 3 of Menke, 1989)

$$(\mathbf{A}^{tr}\mathbf{A} + \lambda^2\mathbf{I})\Delta\mathbf{V} = \mathbf{A}^{tr}\Delta\mathbf{T} \quad (6.14)$$

This sort of damping is straightforwardly done by simply augmenting the original matrix A with the damping factor λ to form

$$A' = \begin{pmatrix} A \\ \lambda \mathbf{I} \end{pmatrix} \quad (6.15)$$

The augmented right hand side is

$$\Delta\mathbf{T}' = \begin{pmatrix} \Delta\mathbf{T} \\ 0 \end{pmatrix} \quad (6.16)$$

Since $A'^{tr}\Delta\mathbf{T}' = A^{tr}\Delta\mathbf{T}$ and $A'^{tr}A' = A^{tr}A + \lambda^2 \mathbf{I}$, the least squares solutions of $A'^{tr}\Delta\mathbf{V} = \Delta\mathbf{T}'$ satisfy equation (6.14).

The third consideration is addressed in this study using the weighting scheme. As such, an uncertainty σ is assigned to each observation or data point. This uncertainty can be regarded as a combination of noise, errors in picking the travel time,

misidentification of the travel time branch, and any other factors that affect the reliability in the observed data (Zelt and Smith, 1992; Hamilton and Mereu, 1993; Zelt and Forsyth, 1994). This is implemented in this study by simply premultiplying equations (6.15) and (6.16) by a diagonal matrix C_d^{-1} to form

$$\begin{bmatrix} C_d^{-1} \mathbf{A} \\ \lambda \mathbf{I} \end{bmatrix} \Delta \mathbf{V} = \begin{bmatrix} C_d^{-1} \Delta \mathbf{T} \\ 0 \end{bmatrix} \quad (6.17)$$

where the diagonal entries of the matrix C_d are the uncertainties (or standard deviations) of the measured data σ_i . Equation (6.17) is the final form of the system of linear equations to be solved in this tomographic inversion.

6.3.5 Iterative Solution Using Conjugate Gradient

Introduction

To solve equation (6.17), a direct inversion method such as Singular Value Decomposition (SVD) (Golub and Van Loan, 1983) is available to be used. However, this system of equations are usually too large to be inverted directly. Van der Sluis and Van der Vorst (1987) reported that SVD requires computing time and memory space of orders (MN^2) and MN , respectively. This method is therefore not preferred for a large problem. Moreover, despite the fact that tomographic data are always noisy, the linear system (6.17) is by itself only an approximation. Iterative solution techniques are therefore commonly employed to solve this tomographic problem.

The iterative methods such as Simultaneous Iterative Reconstruction Technique (SIRT) and Algebraic Reconstruction Technique (ART) which are usually referred

to as *row action methods* are examples (Dines and Lyttle, 1979; Censor, 1981). Conjugate Gradient (CG) of Hestenes and Stiefel (1952), on the other hand, provides an alternative to those methods. In fact, CG is by far the most widely used iterative method for solving large linear systems (Scales, 1987; Scales and Smith, 1994). Ivansson (1986) compared ART, SIRT and CG and found that the CG method converge more rapidly. Scales (1987) showed that CG is a fast and accurate method for solving large tomographic problem. Scales et al (1988) derived a robust algorithm to solve noisy and ill-conditioned systems using preconditioned CG approach by simply iteratively reweighting the residual vectors. Excellent books on seismic tomography edited by Iyer and Hirahara (1993) and Nolet (1987) provide the basic foundation for the technique developed in this study.

Conjugate Gradient for Least Squares Problems (CGLS)

The classic Conjugate Gradient algorithm solves

$$\mathbf{A} \mathbf{x} = \mathbf{b} \quad (6.18)$$

only in the case that $\mathbf{A}$ is an N by N symmetric and positive definite matrix. The algorithm is actually based on the idea of minimizing the function

$$f(\mathbf{x}) = \frac{1}{2} \mathbf{x} \cdot \mathbf{A} \cdot \mathbf{x} - \mathbf{b} \cdot \mathbf{x} \quad (6.19)$$

This function is minimized when its gradient

$$\nabla f = \mathbf{A} \cdot \mathbf{x} - \mathbf{b} \quad (6.20)$$

is zero, which is equivalent to equation (6.18). A simple gradient based minimization algorithm, such as the steepest descent algorithm, minimizes along successive directions of the steepest descent, which generally are not conjugate to each other. Each iteration of the algorithm, thus, spoils some of the work done in the previous iterations. The conjugate gradient algorithm, on the other hand, makes use of conjugate directions or non-interfering directions. That is, they are directions for which minimization along one direction does not undo minimization achieved along other directions (Press et. al., 1992).

The CG minimization, therefore, is carried out by generating a succession of search directions $\mathbf{p}_k$ and improved minimizer $\mathbf{x}_k$. At each stage a quantity α_k is found that minimizes $f(\mathbf{x}_k + \alpha_k \mathbf{p}_k)$, and $\mathbf{x}_{k+1}$ is set equal to the new point $\mathbf{x}_k + \alpha_k \mathbf{p}_k$. The $\mathbf{p}_k$ and $\mathbf{x}_k$ are built up in such a way that $\mathbf{x}_{k+1}$ is also the minimizer of f over the whole vector space of directions already taken, $\mathbf{p}_1, \mathbf{p}_2, \dots, \mathbf{p}_k$. After N iterations, one eventually arrives at the minimizer over the entire vector space, i.e., the solution to (6.18) (Press et al., 1992; Kincaid and Cheney, 1996).

Conjugate gradient can also be extended to the least squares solution of arbitrary linear systems. It has been shown in the previous section that the solutions of the normal equations $\mathbf{A}^{\text{tr}} \mathbf{A} \mathbf{x} = \mathbf{A}^{\text{tr}} \mathbf{b}$ are critical points of the function $\|\mathbf{A} \mathbf{x} - \mathbf{b}\|^2$. Note that $\mathbf{A}^{\text{tr}} \mathbf{A}$ is always symmetric and positive definite. However, applying CG to the normal equations by explicitly forming the product $\mathbf{A}^{\text{tr}} \mathbf{A}$ should be avoided (Paige and Saunders, 1982). This is because the matrix $\mathbf{A}^{\text{tr}} \mathbf{A}$ is usually dense even when $\mathbf{A}$ is sparse. Furthermore, if $\mathbf{A}$ is ill-conditioned, $\mathbf{A}^{\text{tr}} \mathbf{A}$ will be "double" ill-

conditioned. This results in slowly convergent iteration (Paige and Saunders, 1982; Scales, 1987). To avoid this problem CG is then applied to the normal equations but by, first, factoring terms of the form $A^{tr} b - A^{tr} A x$ into $A^{tr} (b - A x)$, which is simply doing the subtraction before the final matrix multiplication.

Algorithmically, this is done as follow: Choose x_0 . Put $s_0 = b - A x_0$, $r_0 = p_0 = A^{tr} (b - A x_0) = A^{tr} s_0$, $q_0 = A p_0$. Then for $k = 0, 1, 2, \dots$

$$\begin{aligned}
 \alpha_{k+1} &= (r_k, r_k) / (q_k, q_k) \\
 x_{k+1} &= x_k + \alpha_{k+1} p_k \\
 s_{k+1} &= s_k - \alpha_{k+1} q_k \\
 r_{k+1} &= A^{tr} s_{k+1} \\
 \beta_{k+1} &= (r_{k+1}, r_{k+1}) / (r_k, r_k) \\
 p_{k+1} &= r_{k+1} + \beta_{k+1} p_k \\
 q_{k+1} &= A p_{k+1}
 \end{aligned} \tag{6.21}$$

Scales (1987) showed that factoring the matrix multiplications in this way results in improved rounding behaviour. This is the CG algorithm which is employed in this study to solve equation (6.17). Paige and Saunders (1982) presented a variant of CGLS called LSQR which is very popular because it is publicly available through the *Transactions on Mathematical Software*. However, since the CGLS code is easy to write, the former is preferred. This is also chosen in order to incorporate easily the regularization and weighting schemes discussed previously in this chapter.

6.4 Some Synthetic Examples

Before a technique is applied to a real data, it is always important to test that technique using synthetic data. This is needed in order to gain insight into the behaviour and performance of the technique. This section is dedicated to discuss this issue. The CGLS technique of inversion will be tested in parallel with the SVD. Readers interested specifically in the SVD method are however referred to Hamilton (1991) and/or Hamilton and Mereu (1993).

6.4.1 Test Model 1 - Simple crustal model

A model with a length of 270 km and a depth of 35 km was considered for the first test of the tomographic inversion procedure. Figure 6.3 shows this very simple, laterally homogeneous, crustal scale model with a velocity of 6.0 km/s at the surface and 7.0 km/s at a depth of 35 km. Two shot points were established at the two edges of the model, and rays were traced through it. The resulting travel times were then taken to be the observed synthetic data.

The model was divided into 6 triangular blocks of equal area as is shown in Fig. 6.4a. An initial estimate of the velocity structure, 6.5 km/s at the surface and 7.5 km/s at a depth of 35 km, was then made to provide a starting model for the inversion. In this first test, it is assumed that very little is known about the structure of the true model, based mainly on the travel time observations. As a result, due to this wild guess in the starting model, a large RMS travel time deviation as shown in Table 6.1, was observed. However, it is evident from the rapidly decreasing RMS values within only a couple of iterations, that both inversion techniques, CGLS and SVD, do work.

The final model resulted from the tomographic inversion is shown in Fig. 6.4b.

It is interesting to note here, that even though the inversion converges to a quite low RMS value, the velocity value at the 8th node located at ($x = 270.0$ km; $z = 35.0$ km), was observed to be too far off from its true value. This is obviously due to a poor ray coverage caused by the geometrical arrangement of the 6th triangular block as is seen in Fig. 6.4. To prove this hypothesis, the 5th and 6th blocks of the original arrangement were flipped in such a manner, shown in Fig. 6.5a, that evenly distributed ray coverage is now achieved. The new arrangement and the tomographic result using this arrangement are given in Fig. 6.5b and Table 6.2, respectively. It is evident that the solutions given by both techniques, now converges to the true values. It is also observed that both techniques, basically, converge to the same solution.

Recalling that tomographic inversion seeks the solution iteratively, the quality of the final solution, consequently, depends on the initial estimate/starting model. Of course, the closer the initial guess is to the true model, the better the inversion will be. However, there are many ways of choosing a starting model: a model with a low initial velocity and a model with a high one. Which one should be chosen? The experiment with test model 1 was then extended to illuminate this point. Two inversions were performed with two different starting models. A model with velocities of (5.75 km/s at the surface; 6.75 km/s at 35 km depth) serves as a 'low' initial model and a model with initial estimate of (6.25 km/s at the surface; 7.25 km/s at 35 km depth) as a 'high' initial model. Tables 6.3 and 6.4 as well as Fig. 6.6 show clearly that the 'high' initial model achieved better solution/final model than the 'low' initial model. This is understandable, because with a given model length, a high velocity gradient implies that more rays will dive and bend back to the surface, resulting in a denser ray illuminations of the area than the

'low' initial model does. This test has, thus, shown us the importance of how the triangular blocks should be arranged, and how to choose the starting model wisely to achieve a better tomographic solution.

6.4.2 Test Model 2 - Shallow depth model

A simple shallow depth and relatively short model with a high velocity gradient, shown in Fig. 6.7a, was taken as a second example. In this case, a velocity of 5 km/s at the surface and 6 km/s at a depth of 3 km was used as a 'true' model. Two shot points were set up on the surface at the coordinates ($x=0$ km, $z=0$ km) and ($x=20$ km, $z=0$ km). Two initial estimates of the velocity structure were made in order to study the non-uniqueness of the inversion: (i) a model with a low guessing velocity (4.75 km/s at the surface and 5.75 km/s at a depth of 3 km) and (ii) a model with a high initial velocity (5.25 km/s and 6.25 km/s at the surface and at a depth of 3 km, respectively).

The results of the tomographic inversion using both CGLS and SVD are shown in tables 6.5 and 6.6 as well as Figs. 6.7b and c. It is observed that the high initial estimate yielded only a slightly better final solution. However, unlike the crustal scale example discussed in test model 1 which showed convergence to the true model, the final solution in this shallow model appears to be non-unique and contains artifacts of the initial model. As it is seen in Fig. 6.7c, the low velocity starting model shows a tendency of giving a better solution for the near surface structure, whereas the high velocity initial model shows better illumination of the deeper structure (see Fig. 6.7b). This may, in fact, be due to the high velocity gradient and the shortness of the model length. The high velocity gradient was probably too high such that the linear approximation used in this tomographic

inversion is not sufficient. The short length effect, on the other hand, is noticeable from the significant difference in velocities (compared with the true model in Fig. 6.7a) at the two bottom edges of the model. Obviously, this effect may be handled by expanding the model length and, consequently, incorporating more shotpoints to gain more information.

It is also important to note here that both techniques, in principle, give almost identical final solutions. The difference is that the SVD is generally more memory intensive compared to the CGLS, even though the former uses a little bit less processing time due to its non-iterative /direct nature. Perhaps, in this computer era, speed and/or memory is no longer a matter of concern. However, with the increasing power and popularity of a Personal Computer (PC), a tomographic method which requires less memory but retains its ability to give a reliable solution in a reasonable time is still preferred. The CGLS method naturally fits this requirement. This is one of the reasons why the author preferred the CGLS than SVD.

There is another interesting fact about the CGLS algorithm that deserves to be mentioned in that its low-frequency components of its solution tend to converge faster than its high-frequency components (Kincaid and Cheney, 1996). Hence, the CG method has some regularization effect where the number of CG iterations plays the role of the regularization parameter. This effect is needed, especially in handling an ill-conditioned inversion problem. Since it has been shown that CGLS is just as effective as SVD, from now on, all examples will be solved using CGLS.

6.4.3 Test Model 3 - Laterally heterogeneous model using exact parameterization

The two previous examples deal with models which are rarely observed in a real situation. A real tomographic problem usually involves both vertical and lateral

velocity changes. Test model 3 is designed to test the ability of the tomographic technique to resolve this problem.

Figure 6.8a shows this laterally heterogeneous crustal model used to create the synthetic data. Three sets of travel time observations were obtained through ray tracing from the three shot points. A flat layered crustal model with velocities of 6 km/s at the surface, 6.429 km/s at a depth of 15 km and 7 km/s at a depth of 35 km was created as an initial model. A plot between the RMS values versus number of iterations, shown in Fig. 6.9, demonstrates the high rate of convergence of the CGLS technique for the tomographic inversion. The final velocity solution for this test model, which is the solution after 10 iterations, is presented in Fig. 6.8b. It can be seen that the lateral variation in velocities have apparently been recovered.

It is important to note here that, in all examples described above, exact parameterization was used. This means that the model parameterization used in the true model to create the synthetic data, is exactly the same with the model parameterization employed to model that data. In a real world problem, however, very little (if not at all) is known about the nature of the true model. Hence, there is no way of knowing how to exactly model the real data. After all, the earth's crust is not built out of the triangular blocks! It is then crucial to understand the effect of under and over parameterization to the solution of the tomographic inversion. The next example is dedicated to illuminate this issue.

6.4.4 Test Model 4 - Laterally heterogeneous model using under and over parameterization

In this example, the synthetic data of test model 3 was used. However, in creating a model to explain the synthetic data, two cases of parameterizing a model were

considered. In an under parameterized case, fewer triangular blocks, hence fewer nodes (compared with the true model), were employed. On the other hand, a model is said to be over parameterized if it uses more model parameters than the true model does.

Figure 6.10a shows the triangular blocks used to test an under parameterized case of inversion. This model comprises of 12 triangular blocks, as opposed to 16 blocks in the exact parameterized model. The convergence results are shown in Fig. 6.11 and the best fit model (based on minimum RMS value) is that obtained after 6 iterations. The final solution for this problem is shown in Fig. 6.10b. It can be seen that the lateral variation can still generally be recovered. However, the plot of RMS misfits versus iterations for this model shows a completely different picture compared with the similar plot that was obtained using exact parameterization (compare Figs 6.9 and 6.11). Under parameterization seems to lead to a relatively large RMS misfits.

The experiment is continued with an over parameterized problem. Figure 6.12a shows a model with 20 triangular blocks, as opposed to 16 blocks in the exact model. The convergence results for this model is shown in Fig. 6.13 and the best fit (based on minimum RMS misfit) is that obtained after 20 iterations. As is shown in Fig. 6.12b, the final solution for this over parameterized model appears to recover the main feature of the true model. The plot of RMS misfits versus iterations also shows that, generally, low misfits are achieved (compared with the under parameterized case). Fig. 6.14 shows a summary of all the case: under, exact, and over parameterization to better illustrate the overall effect of model parameterization. It is important to note that it is generally safer to be over parameterized rather than being under parameterized. A smoothing constraint could then be used to reduce

the roughness of the over parameterized solution. This may either be applied inside the inversion routine (as in the case of this study) or after the solution is obtained (White, 1989; White and Boland, 1992).

6.4.5 Test Model 5 - Crustal scale model with random noise

At this point in the development of the tomographic routine, several test models have been discussed. They range from shallow to crustal scale models with exact, under and over parameterizations. However, all of them deal with a clean, noise-free, synthetic data set. This is, of course, unrealistic, since noise will always present in real data. To simulate a real tomographic problem, random noise was generated and added to the synthetic data. The true model used in this example is exactly the same as test model 3, except that the three sets of observed travel times have now been contaminated with random noise.

An initial estimate with velocities of 6 km/s at the surface, 6.429 km/s at a depth of 15 km and 7 km/s at a depth of 35 km was again used. The convergence result is shown in Fig. 6.15. A relatively high level of RMS misfits compared to the noise-free case in test model 3 (see Fig. 6.9) is noticed. However, convergence can still be reached only in a few iterations. The plot of observed versus model times in Fig. 6.16a and the final velocity solution in Fig. 6.16b, clearly demonstrate the reliability and robustness of the technique in handling the noisy data. It remains to show the application of the CGLS technique to the 1992 Lithoprobe Abitibi-Grenville Seismic Refraction Data.

Iteration	RMS Misfit (msec)	
	CGLS	SVD
0	1542.27	1542.27
1	115.56	115.56
2	16.73	14.79
3	2.94	4.94
4	3.48	4.30
5	3.30	4.27

Table 6.1: RMS misfits for test model 1 resulting from the same initial model for both inversion techniques, Conjugate Gradient for Least Squares (CGLS) and Singular Value Decomposition (SVD) using $\lambda = 0.5$.

Iteration	RMS Misfit (msec)	
	CGLS	SVD
0	1542.27	1542.27
1	120.88	120.88
2	2.74	2.76
3	1.33	1.49
4	0.88	1.17
5	0.86	0.94
6	0.52	0.78

Table 6.2: RMS misfits for test model 1, using new block arrangement, resulting from the same initial model for both inversion techniques, Conjugate Gradient for Least Squares (CGLS) and Singular Value Decomposition (SVD) using $\lambda = 0.5$.

Iteration	RMS Misfit (msec)	
	CGLS	SVD
0	799.96	799.96
1	28.33	28.33
2	1.02	1.03
3	0.58	0.38
4	0.27	0.25
5	0.25	0.24
6	0.24	0.23

Table 6.3: RMS misfits for test model 1, resulting from an initial model which has higher velocities than the true model - a 'high' initial model ($\lambda = 0.5$).

Iteration	RMS Misfit (msec)	
	CGLS	SVD
0	860.08	860.08
1	32.52	32.52
2	5.96	5.98
3	3.11	3.99
4	1.83	3.19
5	1.09	1.92
6	1.14	1.47
7	0.70	1.13
8	0.45	0.93
9	0.43	0.84
10	0.42	0.67

Table 6.4: RMS misfits for test model 1, resulting from an initial model which has lower velocities than the true model - a 'low' initial model ($\lambda = 0.5$).

Iteration	RMS Misfit (msec)	
	CGLS	SVD
0	95.38	95.38
1	15.92	15.92
2	5.62	5.72
3	5.43	5.51
4	5.25	5.66
5	5.15	5.70
6	5.09	5.74
7	5.06	5.64
8	5.26	5.71
9	5.14	5.58
10	5.12	5.77

Table 6.5: RMS misfits for test model 2 (at 1 – 10 iterations), resulting from an initial model which has lower velocities than the true model - a 'low' initial model ($\lambda = 0.8$).

Iteration	RMS Misfit (msec)	
	CGLS	SVD
11	4.77	5.59
12	5.21	5.43
13	4.96	5.58
14	4.94	5.13
15	4.89	5.13
16	5.08	5.14
17	4.98	5.13
18	4.92	5.09
19	4.90	5.10
20	4.85	5.18

Table 6.6: RMS misfits for test model 2 (at 11 – 20 iterations), resulting from an initial model which has lower velocities than the true model - a 'low' initial model ($\lambda = 0.8$).

Iteration	RMS Misfit (msec)	
	CGLS	SVD
0	111.65	111.65
1	24.54	24.54
2	7.25	7.14
3	6.51	6.01
4	6.04	5.79
5	5.68	5.65
6	5.59	5.55
7	5.46	5.58
8	5.36	5.50
9	5.33	5.43
10	5.28	5.37

Table 6.7: RMS misfits for test model 2 (at 1 – 10 iterations), resulting from an initial model which has higher velocities than the true model - a 'high' initial model ($\lambda = 0.8$).

Iteration	RMS Misfit (msec)	
	CGLS	SVD
11	5.23	5.39
12	5.30	5.41
13	5.15	5.36
14	5.13	5.40
15	5.03	5.30
16	5.15	5.34
17	5.15	5.47
18	4.41	5.37
19	5.32	5.36
20	5.19	5.27

Table 6.8: RMS misfits for test model 2 (at 11 – 20 iterations), resulting from an initial model which has higher velocities than the true model - a 'high' initial model ($\lambda = 0.8$).

Figure 6.1: Theoretical travel time curves for a simple two layer model, and corresponding ray paths.

Figure 6.2: Constant velocity gradient triangular block model. Each node has a given velocity value such that a linear gradient within each block is established.

Figure 6.3: 'True' velocity model for test model 1: a very simple, laterally homogeneous, crustal scale model with a velocity of 6 km/s at the surface and 7 km/s at a depth of 35 km.

Figure 6.4: Initial (a) and final (b) models for the tomographic inversion of test model 1. Numbers at each velocity node show CG/SVD comparison. Note the significant difference in velocities at the 8th node (the bottom-right corner of the final model) compared to the 'true' model of Fig. 6.3. This is due to uneven distribution of rays caused by block arrangement in the model.

TEST MODEL 1 - SIMPLE CRUSTAL MODEL A NEW BLOCK ARRANGEMENT

Figure 6.5: Initial (a) and final (b) models for the tomographic inversion of test model 1 after triangular block rearrangement. Note the significant improvement in velocities at the 8th node (the bottom-right corner of the final model) compared to the final model of Fig. 6.4b. This is achieved because rays are now evenly distributed throughout the model.

Figure 6.6: Final velocity model of test model 1 using 'high' (upper panel) and 'low' (lower panel) initial models (see text for detail). Note the better solution achieved by the 'high' initial model compared with the 'low' one.

TEST MODEL 2 - A SHALLOW DEPTH MODEL

Figure 6.7: 'True' velocity (a) and final (b and c) models for test model 2. Note the better solution achieved by the 'high' (b) initial model compared with the 'low' (c) one. Also note the significant difference in velocities (compared with the true (a) model) for nodes located at the bottom-left and bottom-right corners of the model.

TEST MODEL 3 - A LATERALLY HETEROGENEOUS MODEL

Figure 6.8: 'True' (a) and final (b) velocity models for test model 3: A laterally heterogeneous crustal scale model. Note that the lateral variation in velocities have been reasonably well recovered.

RMS Travel Time Deviations for Test Model 3

Figure 6.9: RMS misfit versus number of iterations for test model 3.

TEST MODEL 4 - A Laterally Heterogeneous Model Under Parameterization

Figure 6.10: Triangular blocks arrangement used to test a tomographic inversion with under parameterization: initial (a) and final (b) velocity models. Note this model comprises of 12 blocks as opposed to 16 blocks (in exact parameterization). Also note that the lateral variation can still generally be recovered.

Figure 6.11: RMS misfit versus number of iterations of test model 4 in the case of under parameterization.

TEST MODEL 4 - A LATERALLY HETEROGENEOUS MODEL OVER PARAMETERIZATION

Figure 6.12: Triangular blocks arrangement used to test a tomographic inversion with over parameterization: initial (a) and final (b) velocity models. Note this model comprises of 20 blocks as opposed to 16 blocks (in exact parameterization). Also note that the overall features of the 'true' model can generally be recovered.

RMS Travel Time Deviations for Test Model 4

Under, Over and Exact Parameterized Models

Figure 6.13: RMS misfit versus number of iterations of test model 4 in the case of over parameterization.

Figure 6.14: Contour plots of final models for all cases used in test model 4: under, over, and exact parameterized models compared with the true model.

Figure 6.15: RMS misfit versus number of iterations of test model 5: a laterally heterogeneous crustal scale model with noise.

TEST MODEL 5: A LATERALLY HETEROGENEOUS MODEL WITH NOISE

Figure 6.16: Plot of observed versus model times (a) and final velocity model (b) for test model 5: a laterally heterogeneous crustal scale model with noise. Note the robustness of the tomographic technique in recovering the 'true' model even in the presence of noise.

Chapter 7

TOMOGRAPHIC ANALYSIS OF THE DATA

7.1 Introduction

It has been shown in the previous chapter that, for a given synthetic data, the inversion technique developed was able to arrive at a velocity model which is reasonably close to the 'true' model. The robustness of the technique has also been tested on the data in the presence of lateral velocity variations and noises. The technique is then ready to be applied to a real data. The four seismic refraction profiles from the 1992 Lithoprobe Abitibi-Grenville experiment serve as the problems to be solved. This chapter is dedicated to describe the various stages in the tomographic analysis/modelling of the data. Modelling strategy, analysis of the final velocity models and its corresponding velocity anomaly images, as well as uncertainties involved in the models will be thoroughly discussed in this chapter.

7.2 Modelling strategy

A modelling strategy known as layer stripping (Zelt and Smith, 1992) or expanding cell (Hamilton and Mereu, 1993) was employed in this study to model the data. The inversion is started at the top, imaging the shallower crust using near source observations and gradually working to deeper crust in the model with the more

distant observations. Using this approach, the overall crust is modelled into two or three subdivisions: upper crust and mid- to lower crust. Hence, two or three major modelling steps are generally needed. The length of the profile dictates the number of major modelling steps involved. For a relatively short length profile, as in the case of lines XY and EF, the modelling was, basically done using two major steps. However, for lines AB and MG, a three step modelling/inversion was needed.

It has also been previously shown that the quality of the final solution of the tomographic inversion depends on the initial estimate/starting model. The closer the initial model is to the true model, the better the inversion will be. For this reason, a one dimensional velocity model (or a two dimensional model without lateral velocity variations) based on the best fitted average crustal model was used as the initial model. Similarly, in parameterizing the model, the triangular blocks were constructed in such a way that the presumed 'true' model is closely mimicked. It is difficult, of course, to find the best guess of the 'true' velocity structure. However, a rough idea of the 'true' structure could reasonably be obtained from:

- the geology of the area,
- travel time observations,
- near surface structures based on the delay-time results and
- nature and topography of the Moho based on the PmP composite sections.

This is, basically, the modelling strategy/approach which is used in this study to model the overall velocity structure of the Superior and Grenville Provinces beneath the four refraction profiles.

7.2.1 Upper crust tomography

For the upper crust (0-10 km), the initial model was designed to resemble the image of the near surface structure resulted from the delay-time analysis. The triangular blocks were constructed in such a way that each shotpoint was located very near to the triangular node. This is chosen in order to easily model the near source observations.

The tomographic inversion was performed using paired shot points to determine velocity structure in the vicinity of the shotpoints better. The inversion proceeded iteratively until a model of the upper crust which satisfies all near source observations was found. The velocity model for the upper crust was then fixed and the inversion process continued for the deeper crust. This was done in order to reduce the number of parameters involved and to, consequently, speed up the convergence of the iteration. Moreover, by carefully imaging the upper crust first and fixing it, the deeper crust is then better constrained and therefore less prone to unreliable results.

7.2.2 Mid- to lower crust and upper mantle tomography

In mapping velocity structure at the mid- and lower crustal level, more distant observations and hence more shot points were used. Models with depth ranges from 10 to 35 km and length ranges from 100 to 400 km were involved. The thickness of the crust underneath each refraction profile was constrained by the depth to the Moho obtained from the P_mP observations. The enhanced P_mP composite sections, described previously in this thesis, which had assumed a constant lateral velocity, were used as guiding models.

The nature of the crust-mantle boundary, on the other hand, was determined by

the appearance of the Moho as a reflector. The Moho was modelled as a sharp first order boundary only if strong P_mP waves were observed. Otherwise, the Moho was modelled as a diffuse boundary/transition zone. The distance from the shotpoint to the location where the high amplitude P_mP waves were first observed was used to determine roughly the thickness of the Moho transition zone. The validity of this procedure is justified by the fact that thick transition zone tends to modify the ray paths of the seismic waves that were reflected back to the surface, resulting in the shortening of the length of the PmP branch being observed at the surface (as shown in Figs 7.1 and 7.2). The mantle refracted waves P_n were also used in combination with the PmP waves to constrain the shape of the crust-mantle boundary. The velocity structure of the upper mantle, on the other hand, were determined entirely from the P_n observations.

The modelling approach for mapping the deeper crust was actually rather unique and differs from the modelling approach applied previously for mapping the shallower crust. This is because both velocities and depth of the crustal-mantle boundary are unknown. One could try, of course, to invert both parameters at the same time. This approach is, however, not suggested. This is because of the fact that the deeper crusts are less sampled by the rays compared to the shallower crusts resulting in poor resolution at the deeper crust. Since, at this stage, the mid-crust velocities have already been determined, the deep blocks were therefore easily interpolated using the knowledge of the average velocity gradient based on the least squares study described previously. Interactive forward modelling was then performed prior to the tomographic inversion to adjust the velocities further at these depths. This is done to improve the fit between the model and the observed times for P_mP and P_n while maintaining the nature and topography of the Moho as seen on the P_mP composite

section. From here, the P_mP and P_n were then inverted to better image the Moho.

7.2.3 Selecting the final model

The final (important) stage in a tomographic inversion is the selection of a final velocity model. Why is it important? The rationale is as follows. Often, it is not possible to trace rays to all observation points using a model obtained from an inversion. A common cause of this problem is that the model consists of too many parameters, such that the data are unable to adequately constrain all model parameters, causing an unrealistic lateral variation/oscillation of the velocity values and/or a velocity reversal with depth at certain locations in the model. These features cause rays to focus and/or scatter which result in shadow zones at the surface. The final model selected is, therefore, the one which:

- provides optimum RMS (root-mean-square) traveltimes residual, and
- allows rays to be traced to all observation points.

This is, once again, the place where the interactive modelling comes into play. Velocity adjustment/tweaking for a number of nodes is sometimes needed to ensure that the rays can be smoothly traced to most observation points.

The triangular blocks used to parameterize the whole crust and upper mantle for all the four refraction profiles are shown in Figs. 7.3 – 7.6. A general increase in size of the triangular cells with depth is observed. The rationale behind this design are:

- To model the overall crust realistically. Since it is expected that the shallower crust has greater ray coverage than the deeper crust, it is then wise not to

treat them equally. The shallower crust should be better resolved than the deeper crust.

- To balance between the number of data points, hence the number of rays, and the number of model parameters involved, such that an overdetermined system of linear equations is established. This is done in order to give a robust tomographic solution (i.e., a stable and reliable final solution).

The diagrams showing the typical fit between the observed and the model times for the final models obtained from the tomographic inversion for lines MG, EF, AB and XY are shown in Figs. 7.7 – 7.26, whereas their final RMS misfits are shown in Tables 7.1 – 7.5. It is evident from these figures and tables that the tomographic routine developed in this study was able to create images which best explain the observed data. The question that may arise is then: How reliable are these images in representing the actual crustal structure? This particular question will be further addressed in a later section on non-uniqueness and uncertainty involved in the final model.

7.3 The Final Velocity and Velocity Anomaly Images

7.3.1 The final velocity model

The final velocity model of the crust and upper mantle beneath profiles MG, AB, XY and EF are shown in figures 7.27a, 7.28a and 7.29a, respectively. These colour images were created by taking advantage of the high performance graphic visualization environment of MATLAB (the MATLAB code or m-file is included in the appendix for the interested readers). This is done in order to easily visualize any lateral and/or vertical changes in the velocity field.

It is seen that the tomographic images demonstrate a large lateral variation in the velocity values from one terrane to another, as was expected from the visual examination of the travel times. An obvious difference between a relatively low velocities Archean rocks and high velocity crust underneath the GP with the Grenville Front marked the transition from a relatively higher velocity terranes to the south to a relatively lower velocities to the north was also revealed. However, the boundaries between terranes cannot be distinctively resolved. This was due to the fact that refraction tomography tends to smooth over the detailed structures. It is then difficult to compare an observed seismic section (from the coincident near vertical reflection survey) with the image resulted from the refraction tomography. This problem inspired us to design a better way to present the tomographic results: a velocity anomaly map/image.

7.3.2 The velocity anomaly image

The velocity anomaly image was created by simply subtracting the average crustal velocity field (based on the standard crust with constant velocity gradient discussed in the previous chapter) from the final velocity model. Results of using this procedure to obtain the velocity anomaly maps are shown in figures 7.27b, 7.28b and 7.29b for lines MG, AB, XY and EF, respectively. It can be seen in figure 7.27b, for example, that several distinct boundaries can now be easily recognized. The boundary between the Middle Ordovician rocks south of the CMB and the CMB was marked by a sharp transition from low to high velocity anomalies. A distinct transition from the Central Metasedimentary Belt to the Central Gneiss Belt was also clearly observed. This zone, the CMBBZ, was marked by the southeast dipping boundary from high to low anomalies. Similarly, the Grenville Front Tectonic Zone

which was imaged as a southeast dipping anomalous velocity gradient in the conventional tomographic representation (as shown in Fig. 7.27a) can now be clearly seen as a juxtaposition between a relatively high velocity anomaly beneath the Grenville Province and low velocity anomaly of the Superior crust (see Fig. 7.27b). The boundary between Pontiac subprovince and the Abitibi Greenstone Belt, the CLLFZ, was also resolved as a northeast dipping low velocity anomaly.

The velocity anomaly images also help revealing the deep structure of the GFTZ underneath Line AB (see Fig. 7.28b). The Front is imaged as a southeast dipping juxtaposition between the Britt domain of the CGB and the Abitibi subprovince of the Superior province which have a relatively lower velocity. This image appears mimicking the southeast dipping reflectors of GLIMPCE line-J (Green et al. 1988). The low velocity region underneath the SIS was also clearly imaged as a low velocity basinal structure (see Fig. 7.28b). The nature of the Grenville Front in two localities, south of Sudbury and southeast of Temiscaming, as crossed by these two refraction profiles can now be contrasted by simply comparing their velocity anomaly images side by side. Further comments on these tomographic results will be given later in the discussion chapter.

7.4 Non-uniqueness and uncertainty analysis

It has been shown that the tomographic inversion, like many other geophysical inverse problems, seeks a solution that best explains the observed data. Whenever an earth model is determined, by whatever methods, it is crucial to know whether it is unique. If it is not, then it is possible that the model obtained only represents one of many different models that satisfy the observed data. However, if the acceptable models all fall within some narrow bounds, then the determination of these

bounds/uncertainties on the model may provide some useful geological information. A solution of the inverse problem is therefore usually described as complete if in addition to estimating the best fitted model, the uncertainty involved in the result is also quantified. Wouldn't it be nice to have a complete tomographic solution? In this section, basic foundation, methods and results of uncertainty analysis in the tomographic inversion will be discussed.

7.4.1 Basic foundation: Rationale

A formal method of estimating parameter resolution and uncertainty for the solution of a linear (Gaussian) inverse problem exists in the literature (see for example Aki And Richards, 1980; Menke, 1989). This is achieved through the use of a generalized matrix inversion approach which can then be solved by the virtue of the singular value decomposition (SVD) (see for example Press et al. 1992). In refraction tomography, White and Boland (1992) and Zelt and Smith (1992) , to name a few, have applied this method to estimate the uncertainty involved in their tomographic solutions. However, resolution and uncertainty analysis carried out by this method gives only a relative measure of the quantities. Zelt and Smith (1992) also proposed that absolute parameter uncertainty may be obtained by perturbing each model parameter and hold it fixed while reinverting the observed data. The maximum perturbation of the parameter that allows a comparable fit to the observed data is then an estimate of its absolute uncertainty. It can easily be seen that applying this type of analysis to all velocity and boundary nodes could take longer than the time required to obtain the final model. A simpler approach which avoids reinversion but is applicable to a nonlinear problem should then be pursued.

One other approach of describing general inverse problem is to use a statistical

framework (Tarantola and Vallete, 1982; Tarantola, 1987) and to make an attempt to describe or characterize the nonuniqueness of the solution by describing the solution in terms of probability density (distribution) in model space. The simplest way of deducing model parameters that satisfy some observed dataset, based on this approach, is to use an enumerative (or grid) search. A systematic exploration of the discretized model space is performed, and all models within the considered model subspace are visited. Although this method may be ideal for problem with low dimensionality (i.e., with few parameters), the task is computationally infeasible when problems with many model parameters are considered.

When analysing a nonlinear inverse problem of high dimensionality, as in the case of this study, it is necessary to restrict the number of models to be visited, hence the number of forward model calculations, as compared to the grid search. A question that may arise is then: how to draw samples restrictively from the model space, such that only a region which contains plausible solutions are visited. One way to do this is to use a guided random walk (i.e., a variant of a well known *Monte Carlo* search) to select samples from the model space such that the probability of selecting a model/sample is approximately proportional to the probability distribution of the solution sought.

This approach can, in fact, be implemented with the help of Bayes's theorem. This theorem relates a 'posterior' probability distribution (i.e., a distribution which describes the solution of an inverse problem) to a likelihood (i.e., a misfit/objective function) and a 'prior' distribution (i.e., a prior information about the model without the knowledge of the data) (see Tarantola, 1987; Duijndam, 1988). A fundamental problem in Bayesian statistic is however the need for explicit evaluation of integrals. This means that realistic choices of likelihood and prior will necessitate the use of

sophisticated numerical integration or analytic approximation techniques. This can pose problems when analyzing nonlinear system involving many parameters. Fortunately, Smith and Gelfand (1992) (both are statisticians) addressed this problem by taking a new look at Bayes's theorem from a sampling-resampling perspective (readers interested are urged to read this excellent paper). Their approach opened the way of obtaining the posterior probability density (distribution) with, essentially, calculus-free insight. But what is Bayes's theorem, anyway?

7.4.2 Bayesian formulation - made easy

Bayes' rule can be described as a mathematical tool to update a current knowledge, when new measurements/data become available. A detailed description of Bayesian inference models can be found in the text by Box and Tiao (1973) and a version of the same for geophysical applications is given in Jackson and Matura (1985) and Cary and Chapman (1988). In the following, only a quick summary of Bayes formulation will be given.

Let a model be represented by a vector $\mathbf{m}$ and data by a vector $\mathbf{d}$ given by

$$\begin{aligned} \mathbf{m} &= m_1, m_2, m_3, \dots, m_M \\ \mathbf{d} &= d_1, d_2, d_3, \dots, d_N, \end{aligned} \tag{7.1}$$

with elements m_i and d_i are considered to be a random variable. The quantities M and N are the number of model parameters and data points, respectively. Assume further that there exists a theory which relates the (parameterized) model $\mathbf{m}$ with the predicted/theoretical data:

$$d^{theo} = g(m) \quad (7.2)$$

Based on the *Bayesian approach* to inverse problems, a solution vector is fully described by a *a posteriori* probability density function $\sigma(m)$ which is a combination of *a priori* information, $\rho(m)$ on the model with information provided by the measurement of the data and the information given by the theory. A formal approach, based on probability theory, can then be made to derive the relation between them (see Tarantola and Valette, 1982; Tarantola, 1987; Sen and Stoffa, 1995). Whatever the particular approach may be, a final *simple* relation can usually be expressed as

$$\sigma(m) = k\rho(m)L(m), \quad (7.3)$$

where k is a normalization constant and $L(m)$ is a *likelihood function* which measures the fit between the observed data and data predicted/calculated from the model. As an example, consider an experimental result d^{obs} which has Gaussian experimental uncertainties described by a covariance matrix C . Using the l_2 -norm, the likelihood function can be expressed as:

$$L(m) = \exp\left[-\frac{1}{2}(g(m) - d^{obs})^T C^{-1}(g(m) - d^{obs})\right] \quad (7.4)$$

If we proceed further by contemplating the essence of a sampling process and probability density function in the probability theory, there exists, actually an essen-

tial duality between a sample and the density (distribution) from which that sample is generated (Smith and Gelfand, 1992). It is obvious that the density generates the sample; conversely, given a sample we can approximately recreate the density (as a histogram or whatever!). Based on this duality, according to Smith and Gelfand (1992), Bayes's theorem can alternatively be viewed as a process of updating of a sample from prior $\rho(m)$ to a sample from posterior $\sigma(m)$ through the likelihood function $L(m)$ (see again equation (7.3)). But, how does this new insight help us in solving the problem? This question will be addressed in the next section.

7.4.3 Sampling the posterior probability

At this stage, since we are dealing with a discrete quantity/sample (as opposed to a continuous probability density function), it is necessary to use discrete formulation of equation (7.3). The discrete version of this equation, following Mosegaard and Tarantola (1995), can be expressed as

$$\sigma_i = \frac{\rho_i L_i}{\sum_j \rho_j L_j}, \quad (7.5)$$

where σ_i and ρ_i represent the posterior and prior probability of a point m_i in the model space, respectively, whereas L_i is the likelihood value of a point m_i . Using this formulation, the sampling problem can then be stated as follows: *given a set of points in space with a probability ρ_i attached to every point (i.e., the probability of selecting point i using a random rules is ρ_i), how can the rules be modified so that the new random walk converges at the probability σ_i expressed by equation (7.5).*

This can actually be done as follows: Assume that some rules define a random walk which samples the prior probability ρ_i . At each step, the random walker is

at point i , an application of the rules would lead to a transition to point j . If the proposed transition is always accepted, then the random walker would sample the prior probability ρ_i . Consider now a new rule of accepting the proposed transition which sometimes thwarts the proposed move, as opposed to always accept it. This new rule is stated as follows:

1. If $L_j \geq L_i$ (i.e., if the 'new' point has higher (or equal) likelihood than the 'old' point), then accept the proposed transition to j .
2. If $L_j < L_i$, then make a random decision to either move to j , or to stay at i , with the probability of moving to point j to be equal to $\frac{L_j}{L_i}$.

Mosegaard and Tarantola (1995) proved that, by using this rule, the random walker will eventually sample the posterior probability σ_i defined by equation (7.5).

As an example, consider the case of independent, identically distributed Gaussian uncertainties. The likelihood function for this case (see equation (7.4)), then, degenerates into

$$L(m) = k \exp\left(-\frac{S(m)}{s^2}\right), \quad (7.6)$$

where

$$S(m) = \frac{1}{2} \sum_{i=1}^N (d_i^{theo} - d_i^{obs})^2 \quad (7.7)$$

is the misfit function and s^2 is the total noise variance which is the same for all N observations. The acceptance probability for a perturbed model then becomes

$$P_{accept} = \begin{cases} 1, & \text{if } S(m_{new}) \leq S(m_{old}); \\ \exp(-\frac{\Delta S}{s^2}), & \text{if } S(m_{new}) > S(m_{old}); \end{cases} \quad (7.8)$$

where

$$\Delta S = S(m_{new}) - S(m_{old}) \quad (7.9)$$

This means that the perturbation is always accepted if it improves the data fit, but it is accepted with probability P_{accept} if it degrades the data fit.

It is interesting to note that the acceptance rule given by equation (7.8) mimics the classical Metropolis rule (Metropolis et al., 1953) which was used in the early days of digital scientific computing for simulating the evolution of a solid in a heat bath to thermal equilibrium. This algorithm was later adapted by Kirkpatrick et al. (1983) as an optimization tool known then as a *Simulated Annealing* (SA). In the simulated annealing world, the misfit function is usually called the thermodynamic energy E , whereas the denominator term is expressed as the temperature T :

$$P_{accept} = \exp(-\frac{\Delta E}{T}), \quad (7.10)$$

It is seen here that, in the case of a uniform prior, the algorithm of *modifying the prior into the posterior* becomes exactly identical to the Metropolis algorithm at a constant temperature of s^2 .

7.4.4 Simulated Annealing

It has just been shown how a random walk sampling some prior probability ρ_i can be modified by the Metropolis rule to sample the posterior probability σ_i . The Simulated Annealing (SA) algorithm introduced above actually uses this Metropolis rule as its acceptance rule. It is obvious, however, that if a constant temperature is used, this algorithm requires a very long time to converge. This is simply because the ratio between the accepted and the proposed move is very low. In order to flexibly explore the model space, the probability of accepting a new model in the SA algorithm was then designed as a function of the temperature T . This algorithm usually starts at a sufficiently high temperature such that the perturbed model could be far away from the current model. Then, the temperature is gradually lowered following a cooling schedule, such that in the limit (i.e., as the temperature approaches zero), the global minimum energy state can be reached.

Geman and Geman (1984) showed that a *logarithmic* cooling schedule for SA is necessarily required in order to achieve convergence to the global minimum. This means that the algorithm requires a very long time to converge to the globally optimal solution. Szu and Hartley (1987) addressed this issue and proposed a new algorithm which they called Fast Simulated Annealing (FSA). The cooling schedule required for convergence, using the FSA, is now reduced to be *inversely proportional to iteration number*. Ingber (1989, 1993) proposed several modifications of FSA and created a new algorithm called Very Fast Simulated Annealing (VFSA) which further speeded the cooling process but retained its ability to converge to the globally optimum solution. The latter algorithm is the one which is employed in this study to estimate the uncertainty in the tomographic solution.

7.4.5 Very Fast Simulated Annealing

It has been shown that the VFSA algorithm is actually a variant of a global optimization technique called simulated annealing (SA) (Metropolis et al. 1953; Kirkpatrick et al. 1983; Press et al. 1990). This optimization process simulates the evolution of a physical system as it cools and anneals into a state of minimum energy. A physical annealing process occurs when a solid is initially heated by increasing the temperature such that all the particles are randomly distributed in a melt. This process is followed by slow cooling such that all the particles arrange themselves and crystallizes in the low energy state. In the VFSA technique, a model is drawn from a Cauchy-like distribution (as opposed to uniform distribution) which is a function of a control parameter called temperature. Following this, a decision of whether rejecting or accepting this model is then made, based on statistical criterion described later in this section. The general aim of using this technique is to find the globally optimum solution as represented by the model which give the least minimum error analogous to the low energy state in a real physical annealing process. However, in this study, the objective of using this technique is not to promote a method of the finding global optimum, but rather to cast the inverse problem into the statistical framework and then devise a method based on this technique to estimate uncertainties. Excellent reviews of the various SA techniques are given by Ingber (1993) and Sen and Stoffa (1995).

The algorithm employed in this study essentially comprises the following steps:

1. Compute theoretical travel times through an initial model and determine the least-squares error, E_0 . For each iteration i , the least-squares error (which is

the objective function to be minimized) is defined as:

$$E_i = \frac{1}{n} \sum_{j=1}^n (t_j^{obs} - t_j^{mod})^2 \quad (7.11)$$

where n is the number of observations, j denotes each observation, while t^{obs} and t^{mod} are the observed and theoretical travel times, respectively.

2. The second step is to perturb the model. The velocity field is perturbed by making small random variations, within a given bound, to the velocity values of the nodes of the triangular blocks. The new least-squares error, E_1 based on this perturbed model is then computed.

3. The third step is to determine the probability P of accepting the new model:

$$\text{if } E_1 \leq E_0; \quad P = 1 \quad (7.12)$$

$$\text{if } E_1 > E_0; \quad P_a = P_{accept} = \exp\left(\frac{-(E_0 - E_1)}{T}\right) \quad (7.13)$$

where P_c is the probability of conditional acceptance, T is called the acceptance temperature.

4. Repeat the process of perturbation and acceptance explained in steps 2 and 3 at a fixed temperature. Then the temperature is lowered following a cooling schedule and the process is repeated.

The convergence is determined when the difference in least-squares error between successive models and the probability of accepting new models become very small. Ingber (1989) showed that a global minimum can be statistically obtained by using the following cooling schedule

$$T(k) = T_0 \exp(-Ck^{1/NM}) \quad (7.14)$$

where T_0 is the initial temperature, C is called a decay factor, a parameter to be used to control the temperature schedule and NM is the number of model parameters involved. It can be seen that, compared to the SA and FSA, this cooling schedule is much faster.

7.4.6 Uncertainty in the velocity determination

Sampling based approach

A typical RMS error versus iteration curve in the VFSA run of the final velocity model beneath line MG is shown in Fig. 7.30. The starting temperature and the number of moves per temperature were chosen to be 1.0 and 8 respectively. Small improvement, as represented by the RMS error in Fig. 7.30, was achieved from this inversion. This perhaps indicates that the final model of the tomographic inversion was quite close to the global optimum model.

In calculating the estimates of the uncertainties in the velocity determinations, two different approaches were employed. A sampling based approach described above was used in the first approach. The uncertainties in the model, using this method, were intrinsically calculated during the process of finding the global optimum. The histogram of a frequency of occurrence of a model parameter being sampled within a given bound was then used to estimate the probability distribution (PD) of that particular model parameter. Rather than viewing each and every histogram in analyzing the uncertainty, the estimate of mean value and its corresponding standard deviation were then made by assuming that the PD (i.e., the histogram) has a shape of a Gaussian distribution.

Ray density approach

The second approach is a novel approach which attempts to obtain a fast rough estimate of the velocity uncertainties. This method is based on the number of times a triangular block is being sampled by the rays. The rays were traced from all the shotpoints through the final velocity model. The ray density at each node was then calculated by averaging the value of the ray densities from all the surrounding triangular blocks. The value of the ray density within each triangle itself was calculated from the total time of rays spent in that particular triangle divided by the area of that triangle. This method, however, cannot provide an estimate of the absolute parameter uncertainty as given by the sampling method. A simple plot between the value of the ray density versus the standard deviation obtained from the sampling method was created (see Fig. 7.31) to calibrate the ray-density method for estimating velocity uncertainty. As expected, the uncertainties in the velocity determination in each node was found to be inversely proportional to the value of the ray density at that particular node.

Velocity uncertainty image

The velocity uncertainty images beneath lines MG and AB are shown in Figs. 7.27c and 7.28c, respectively. It is seen that the velocity uncertainties in the top 2 km of the upper crust calculated using the method described above ranging from 0.02 to 0.04 km/sec. The lower part of the upper crust, depth of 10 km, is characterized by the velocity uncertainties of 0.04 to 0.1 km/sec. The value of the velocity uncertainties in the upper part (20 km depth) and lower part (depth of 37 km) of the lower crust are in the range of 0.05 to 0.12 km/sec and 0.06 to 0.16 km/sec, respectively. In all cases, the higher values are located at the edges of the model.

Shot	RMS Misfit (sec)
AB10	0.077
AB9	0.079
AB8	0.059
AB7	0.083
AB6	0.079
AB4	0.071
AB3	0.051
AB2	0.057
AB1	0.058
AB0	0.074

Table 7.1: RMS misfits for the final inversion result for line AB.

Shot	RMS Misfit (sec)
XY0	0.075
XY1	0.077
XY2	0.055
XY3	0.090
XY4	0.049
XY5	0.057

Table 7.2: RMS misfits for the final inversion result for line XY.

Shot	RMS Misfit (sec)
EF6	0.068
EF5	0.069
EF3	0.073
EF0	0.087
EF2	0.084
EF1	0.085

Table 7.3: RMS misfits for the final inversion result for line EF.

Shot	RMS Misfit (sec)
MD1	0.085
MD2	0.081
MD3	0.074
MD4	0.076
MD5	0.075
MD6	0.094
MD7	0.095
MD8	0.058
GD14	0.084
GD13	0.087
GD12	0.082

Table 7.4: RMS misfits for the final inversion result for line MG (southern portion).

Shot	RMS Misfit (sec)
GD11	0.082
GD10	0.088
GD9	0.084
GD8	0.068
GD7	0.056
GD6	0.078
GD5	0.059
GD4	0.046
GD3	0.073
GD2	0.088
GD1	0.092

Table 7.5: RMS misfits for the final inversion result for line MG (northern portion).

Figure 7.1: Diagram showing rays traced through the model for line MG shot GD09. Note the effect of the Moho transition zone in turning the rays to shorten the length of the P_mP branch.

Figure 7.2: Diagram showing rays traced through the model for line MG shot MD04. Note the effect of the Moho as a sharp, first order boundary which tends to extend the length of the observed P_mP branch.

ABITIBI-GRENVILLE LINE MG

Figure 7.3: Diagram showing a full crustal and upper mantle model underneath line MG parameterized using triangular cells.

Figure 7.4: Diagram showing a full crustal and upper mantle model underneath line EF parameterized using triangular cells.

ABITIBI-GRENVILLE LINE AB

Figure 7.5: Diagram showing a full crustal and upper mantle model underneath line AB parameterized using triangular cells.

ABITIBI-GRENVILLE LINE XY

Figure 7.6: Diagram showing a full crustal and upper mantle model underneath line XY parameterized using triangular cells.

Figure 7.7: Observed versus model times for line MG shots MD1 (upper panel) and MD2 (lower panel).

Figure 7.8: Observed versus model times for line MG shots MD3 (upper panel) and MD5 (lower panel).

Figure 7.9: Observed versus model times for line MG shots MD8 (upper panel) and GD14 (lower panel).

Figure 7.10: Observed versus model times for line MG shots GD13 (upper panel) and GD12 (lower panel).

Figure 7.11: Observed versus model times for line MG shots GD11 (upper panel) and GD10 (lower panel).

Figure 7.12: Observed versus model times for line MG shots GD08 (upper panel) and GD07 (lower panel).

Figure 7.13: Observed versus model times for line MG shots GD6 (upper panel) and GD5 (lower panel).

Figure 7.14: Observed versus model times for line MG shots GD2 (upper panel) and GD1 (lower panel).

Figure 7.15: Observed versus model times for line EF shots EF6 (upper panel) and EF5 (lower panel).

Figure 7.16: Observed versus model times for line EF shots EF3 (upper panel) and EF0 (lower panel).

Figure 7.17: Observed versus model times for line EF shots EF2 (upper panel) and EF1 (lower panel).

Figure 7.18: Observed versus model times for line AB shot AB0.

Figure 7.19: Observed versus model times for line AB shots AB2 (upper panel) and AB1 (lower panel).

Figure 7.20: Observed versus model times for line AB shots AB4 (upper panel) and AB3 (lower panel).

Figure 7.21: Observed versus model times for line AB shots AB7 (upper panel) and AB6 (lower panel).

Figure 7.22: Observed versus model times for line AB shots AB9 (upper panel) and AB8 (lower panel).

Figure 7.23: Observed versus model times for line AB shots AB10.

Figure 7.24: Observed versus model times for line XY shots XY0 (upper panel) and XY1 (lower panel).

Figure 7.25: Observed versus model times for line XY shots XY2 (upper panel) and XY3 (lower panel).

Figure 7.26: Observed versus model times for line XY shots XY4 (upper panel) and XY5 (lower panel).

Figure 7.27: Final velocity model (a), velocity anomaly map (b) and contour plot of velocity uncertainties (c) for line MG.

Figure 7.28: Final velocity model (a), velocity anomaly map (b) and contour plot of velocity uncertainties (c) for line AB.

Figure 7.29: Final velocity model (a and c) and velocity anomaly models (b and d)

for line XY and EF.

Figure 7.30: A typical RMS error versus iteration curve in the Very Fast Simulated Annealing run of the final velocity model beneath line MG.

Figure 7.31: Uncertainty in the velocity determination versus ray density plot for the velocity nodes of the crustal model beneath line MG.

Chapter 8

RESULTS AND INTERPRETATIONS

The delay-time and tomographic analysis have brought us complete images of the overall velocity structure of the crust and upper mantle beneath the four refraction profiles. The data sets, as demonstrated by the final velocity models, show large lateral variations in crustal velocities from one terrane to another (see Figs. 7.27a, 7.28a, 7.29a and 7.29c). The velocity anomaly maps shown in Figs. 7.27b, 7.28b, 7.29b and 7.29d, on the other hand, clearly accentuate most features shown in the final crustal velocity models. It is now time to discuss the geological significance of these results.

8.1 Crustal structure in the Grenville Province

The paleozoic cover near the south end of line MG (Figs. 7.27a and 7.27b) is shown to decrease gradually to the north. Near Belleville, the total thickness of the Paleozoic rocks is 2 – 3 km. This is sufficient to explain the observed significant negative velocity anomaly as well as pronounced negative Bouguer gravity anomaly in this area. The southern part of the velocity model for line MG (Fig. 7.27a) shows apparent south dips of the CMBBZ penetrating to the lower crust. The velocity anomaly map for line MG (Fig. 7.27b) shows that this boundary is a deep-seated

feature. The CMBBZ is imaged on the seismic reflection sections as a distinct package of reflectors with apparent south-east dip (Forsyth et al. 1994b; White et al. 1994). The well defined Moho (40 km depth) underneath this structure is consistent with a Moho reflection observed on the north end of Lithoprobe line 32 (see Fig. 8 in White et al. 1994). The Moho of 43 km depth shown by the wide-angle study (Hughes and Luetgert 1992; Zelt et al. 1994) crossing the south end of line MG further supports the south-east dipping crustal structure in this vicinity. The largest positive velocity anomaly values are associated with the Central Metasedimentary Belt and the Central Gneiss Belt south of the Grenville Front. The velocity structure agrees with results from the previous seismic refraction study in this area (Mereu et al. 1986). The high seismic velocities are consistent with the high metamorphic grade of rocks exhumed from lower crustal depths within the CMB and CGB, resulting from the collision of the Central Metasedimentary Belt with the Central Gneiss Belt. These observations, support the theory of the thrusting of the Central Metasedimentary Belt northwestwardly over the Central Gneiss Belt with the CMBBZ as the main thrust zone (Davidson et al. 1984; 1991; Windley 1986; Easton 1992).

The upper crust of the CGB in the vicinity of the Ottawa River is dominated by the Ottawa-Bonnechere graben (Kay 1942) which has a relatively lower velocity as compared to the upper crust of the Central Metasedimentary Belt (CMB in Fig. 7.18b). Line MG crosses this graben obliquely which appears as a shallow structure (5 – 8 km deep) underlain by a relatively high velocity mid crust.

The velocity anomaly images also help define the Grenville Front Tectonic Zone underneath line MG (Fig. 7.27b) and line AB (Fig. 7.28b). The GFTZ south of Sudbury (line AB) and southeast of Temiscaming (line MG), is imaged as a

remarkable south-east dipping juxtaposition of a relatively high velocity crust south of the Grenville Front and a relatively low velocity crust to the north. Images 7.27b and 7.28b show how the GFTZ penetrates the crust to the Moho. These images suggest an Archean crust may extend 140 km beneath the northern Grenville Province. The tomographic images also reveal some differences in the structure of the GFTZ. The deep structure of the GFTZ underneath line AB (Fig. 7.28b) is very similar to that implied by the GLIMPCE line J (see Fig. 2 in Green et al. 1988). A south-east dipping anomaly as shown in Fig. 7.28b, appeared to closely resemble the south-east dipping velocity structure of the GFTZ in the Georgian Bay/Lake Huron as modelled by Epili and Mereu (1991) (see their Fig. 10). This image is also in general agreement with models proposed previously for this area (Jiao and Moon 1995; Long and Salisbury 1996). However, the lower crustal signature of the GFTZ beneath line MG is distinctly different from line AB. The rise of the Moho (Fig. 7.27b) beneath the surface expression of the GFTZ, appears to propagate upward, modifying the earlier structure at lower and mid crustal levels. This feature is consistent with the results of the Abitibi-Grenville Lithoprobe seismic reflection profile 15 (see Figs. 4 and 5 in Kellet et al. 1994). The Moho updoming and the subsequent crustal thinning observed underneath line MG could be associated with either the extensional structure along the Ottawa-Bonnechere (OB) graben of the St. Lawrence Ottawa river rift system (Kumarapelli and Saul 1966) or a representation of post-Grenvillian rebound (Hynes, 1994). However, the latter interpretation is preferred due to the fact that the OB graben is evidently shown as a shallow feature (about 8 km maximum depth extent).

8.2 The Sudbury Structure and The Southern Province

The delay time analysis (Figs. 5.7 and 5.8) clearly reveals the topography of a higher velocity layer beneath 0 - 2 km of lower velocity surficial layer. Figure 5.7, for example, clearly show the basinal structure of the Sudbury Impact Structure, where the low velocity layer corresponds to the sedimentary rocks of the Whitewater Group. The high velocity layer seems to corresponds well to the Norite and Granophyre sublayer of the Sudbury Igneous Complex (SIC). The image also shows that the south part of the SIS (Fig. 5.7a) must have undergone significant deformation resulting in an unsymmetrical shape of the SIS, in general agreement with the interpretation based on seismic reflection study of Wu et al. (1994). However, in contrast to their model which showed that the SIC is well elongated to the south, we interpret that the SIC to more or less mimic the basinal structure shown in Figs. 5.8a and 5.8b. This is based on the fact that at surface, the outcrop of the SIC is a roughly oval structure (G.M. Young and W.R. Church, personal communication, 1995; Card and Jackson 1995). The SIS, shown in the tomographic image (Fig. 7.28b), is underlain by rocks of a relatively lower velocity anomaly than the surrounding rocks. We interpret that these rocks belong to the Southern province which is dominated by volcanic and sedimentary rocks of the Huronian Supergroup. The unsymmetrical nature of the SIS is also shown by the elongated low velocity anomaly of basinal structure in the middle of line XY (Fig. 7.29b). The difference in the velocity anomaly values between the upper crust beneath line XY (Fig. 7.29b) and the one beneath line AB (Fig. 7.28b) could be attributed to an anisotropy of the upper crust of the Southern province with fast-velocity direction being east-west, subparallel to the general structural trend in this province.

8.3 Crustal structure in the Superior Province

The Pontiac subprovince, north of GFTZ (Fig. 7.27a), is marked by the dramatic changes in velocity from the relatively higher velocities in the Grenville Province to lower velocities beneath the Pontiac subprovince. The south dipping velocity anomaly contour beneath the southern part of the Pontiac subprovince suggests that this subprovince possibly underlies the GFTZ. This interpretation is consistent with the seismic reflection data crossing this area (Kellet et al., 1994) and also in agreement with the previous interpretation of the area based on the seismic refraction data of Grandjean et al. (1995). The north dipping velocity anomaly contour beneath the Pontiac subprovince is consistent with the Lithoprobe seismic reflection line 16A (see Fig. 4 in Ludden et al., 1993) which showed that the Pontiac subprovince was dominated by the north dipping, highly reflective mid- to lower-crust. The boundary between the Pontiac and the AGB (see the northern part of line MG in Fig. 7.27b) is marked by steeply north dipping velocity anomaly which we interpreted to be an image of the Cadillac-Larder Lake Fault Zone. The low velocity anomaly overlaying the CLLFZ is consistent with the metasedimentary rocks of Pontiac Group in the surface geology. The low velocity anomaly juxtaposed with the higher velocity rocks to the south, in a very good agreement with the juxtaposition between the metasedimentary assemblages and the granitic plutons. This north dipping feature is interpreted to either be a consequence of the Moho updoming which occurred right underneath the transition from the GFTZ to the Pontiac subprovince or a representation of a series of thick southwest-verging imbricated thrusts (Hubert et al. 1992). If the latter is the case, the images have thus shown the underthrusting of the Pontiac subprovince beneath the Abitibi Greenstone Belt.

The Abitibi subprovince beneath line MG (Fig. 7.27b) is imaged as a relatively

higher velocity anomaly upper crust (8 – 10 km thick) which is underlain by a north dipping low velocity anomaly mid- to lower-crust. In contrast to the earlier refraction studies of this area (Grandjean et al., 1996), which interpreted the Abitibi belt as a simple layered crustal velocity structure, the velocity anomaly contour presented in this study (see the northern part of Fig. 7.27b) shows a remarkable resemblance to the seismic reflectivity patterns of the Lithoprobe reflection line 28 (see Fig.3 in Ludden et al., 1993). Consistent with the tectonic architecture in this area as inferred from the reflection data (Ludden et al., 1993), the north dipping low velocity anomaly is interpreted as turbidite sequences that would constitute the upper crust of the Pontiac subprovince, whereas the higher velocity anomaly upper crust comprised an allochthonous veneer of volcanic and plutonic rocks.

The middle and lower crusts beneath line EF (Fig. 7.29d) is shown to be dominated by rocks which have relatively lower velocities than the average crust. Figure 7.29d also shows the detailed variations in rocks lithology crossed by this west-east profile which runs subparallel to the general structural trend in the Abitibi greenstone belt. The high velocity anomaly in the western end of the line, for example, represents mafic to intermediate volcanic rocks that belong to the upper crust of the Abitibi greenstone belt. As we move eastward, the profile crosses the Quetico metasedimentary rocks. This is indicated by surficial low velocity anomaly and eastward thinning of the high velocity anomaly. The surficial high velocity anomaly in the middle of the section represents the massive granodiorite to granite rocks that belong to the Opatica subprovince. The low velocity anomaly located in between the high anomalies corresponds to the Quetico metasedimentary rocks in the vicinity of shotpoints EF3 and EF4 (see also Fig. 5.8b). The high velocity anomaly upper crust in the eastern part of the section again represents rocks of the Archean

Abitibi Greenstone Belt. This high velocity anomaly upper crust is interpreted to be associated with the volcanic and plutonic rocks of the Abitibi subprovince. The relatively lower velocity anomaly middle crust is associated with an imbricated sequences of metaplutonic and metasedimentary rocks which could possibly belong to the Quetico and Opatika subprovinces. If this is the case, the velocity anomaly images have thus shown a picture of the imbrication of the Quetico and possibly Opatika rocks under the Abitibi Greenstone Belt. The overall images support the model involving oblique convergence of ocean-crust relative to an accretionary complex in an island-arc sequence (Hoffman, 1991; Desrochers et al., 1993; Kimura et al., 1993).

8.4 Moho definition and crustal thickness

The variations in Moho depth can perhaps be best seen from the tomographic image beneath line MG as shown in Figs. 7.27a and 7.27b. It can be seen that the shallowest Moho (34 km) occurred just north of the surface representation of the Grenville Front. The very clear P_mP signals (see Fig. 4.19) indicated that the Moho is a sharp first order crust-mantle discontinuity, in this location. This is interpreted as an indication of mantle involvement in response to flexural loading (Hynes, 1994) during the Grenvillian orogeny.

The Moho just south of the Grenville Front is generally deeper (39 - 43 km) compared to the Moho underneath the Archean Pontiac and Abitibi greenstone belt which is 34 - 36 km deep. The thickening of the crust south of the front is consistent with the results of the previous seismic refraction studies and also consistent with the crustal shortening scenario associated with the compressional event during the Grenvillian orogeny. Beneath the Central Gneiss Belt, the Moho

appears as a relatively diffuse (see the southern part of Fig. 4.19 and the northern part of Fig. 4.18) boundary, compared with the Moho beneath the Grenville Front. This could be interpreted as due to a rough crust-mantle boundary or a gradual change in seismic properties between the lower crust and the mantle, as Dainty and Schultz (1995) found that the P_mP character resulted from the Moho as a rough boundary is similar to the one resulted as a response of a gradient layer/transitional zone. The former interpretation is, however, preferred due to the fact that this area is a segment of a major crustal collisional orogen that was considered to be responsible for joining many continental blocks into a possible Late Proterozoic supercontinent (Hoffman, 1991b).

The Moho is generally defined as a sharp boundary further south of the Front (beneath the Central Metasedimentary Belt), but tends to become a diffuse transition zone (about 5 - 7 km thick) beneath the Archean Abitibi subprovince. The very strong P_mP signals (see Fig. 4.18) beneath the surface expression of the CMBBZ suggested that Moho in this vicinity is a sharp first order crust-mantle discontinuity. This could be an indication of post-Grenvillian extensional event which is also confirmed by the seismic reflection data (Forsyth et al., 1994a) showing extensional troughs containing post-Grenville sediments. The diffuse nature of the Moho beneath the Archean Abitibi subprovince, on the other hand, is interpreted to be associated with underplating of the high-density komatiitic magma which is predominant in the Archean, since magma of komatiitic composition would be seismically indistinguishable from the mantle (Durrheim and Mooney, 1994). The relatively thin crust and low velocity lower crust beneath the Archean Abitibi greenstone terrane of the Superior Province is consistent with the global pattern of Archean crusts as reported by Durrheim and Mooney (1991, 1994).

Chapter 9

THREE-COMPONENT/POLARIZATION ANALYSIS

9.1 Formulation of the problem

During the 1992 Lithoprobe Abitibi-Grenville Seismic Refraction Experiment, an array of three-component (3-C) seismometers was deployed strategically in the vicinity of the Grenville Front Tectonic Zone (GFTZ) (Fig. 9.1). In an attempt to study the nature of the crust in this particular area, polarization and propagation directions/vectors for P_g phases from a number of shotpoints were measured using these 3-C arrays. These vectors can be specified in terms of azimuth and incidence angle. The polarization direction, the dominant linear particle motion of the P_g arrival, was extracted from each 3-C station for each shot using the covariance matrix technique (Winardhi, 1994; Winardhi and Mereu, 1994). The propagation direction, on the other hand, was obtained from the array of vertical seismometers using the conventional beam-forming technique. Apparent angles of incidence were then computed from these measurements using knowledge of the local velocity obtained from the tomographic study described previously in this thesis. Interestingly, a systematic deviation (angular difference) between the polarization and propagation directions was observed. The difference between these two vectors often approaches 15 degrees, rendering them easily measurable. What causes the observed anomalies?

Apparently, in anisotropic media, the direction of P-wave polarization computed from particle motion does not generally coincide with the propagation direction (Crampin et al, 1982; Li et al, 1987). The magnitude and pattern of the polarization anomalies depend on the spatial orientation and elastic properties of the anisotropic medium (Babuska and Cara, 1991). The fundamental questions that may arise are then: Which part of the crust is anisotropic? Can we localize it? If yes, how? What mechanism possibly creates this observed anisotropic phenomenon? These are some questions for which we seek answers. It is then worthwhile to study the effect of anisotropic medium on the polarization of seismic waves in order to possibly explain and/or constrain, to some extent, the observed anomalies. Basic foundation on wave propagation in anisotropic media, methods of extracting information from the three-component array data as well as inversion analysis to reveal the nature of the crust in the vicinity of the Grenville Front Tectonic Zone will be discussed in this chapter to address the aforementioned questions.

9.2 The propagation of plane waves in anisotropic media

Comprehensive treatment of the propagation of plane waves in anisotropic media has been given, for example, by Crampin (1981) and Babuska and Cara (1991). Only features of propagation in anisotropic media which are of particular importance to this study will be briefly reviewed here.

Let us now consider the simplest type of seismic waves in anisotropic media: plane waves propagating in a homogeneous anisotropic medium. If $u(\mathbf{r}, t)$ denotes the motion of a particle located at coordinates $\mathbf{r}$ at time t , one can write for a plane wave propagating in the direction $\mathbf{n}$ at phase velocity v :

$$\mathbf{u}(\mathbf{r}, t) = \mathbf{a} f(t - \mathbf{n} \cdot \mathbf{r}/v) \quad (9.1)$$

where the function $f(\)$ gives the time dependence of the particle motion at a fixed point in space, $\mathbf{a}$ is a vector describing amplitude and polarization of the wave, and $\mathbf{n}$ is a unit vector perpendicular to the phase surface.

The conditions to be satisfied by the vector $\mathbf{a}$ and the phase velocity v can easily be obtained from the equation of motion applied to a homogeneous medium. In Cartesian coordinates, one can write:

$$\partial_j \sigma_{ij} = c_{ijkl} \partial_j (\varepsilon_{kl}) = \rho \partial_t^2 (u_i) \quad (9.2)$$

where ∂_j denotes the space derivative relative to the coordinate x_j , ∂_t denotes the time derivative and ρ is the density. For small strain, we have

$$\varepsilon_{kl} = (\partial u_k / \partial x_l + \partial u_l / \partial x_k) / 2 \quad (9.3)$$

and taking into account the symmetry of both the strain tensor and the elastic tensor, one can also write:

$$c_{ijkl} \partial_j \partial_k (u_l) = \rho \partial_t^2 (u_i) \quad (9.4)$$

Finally, using the relation of the displacement field $\mathbf{u}(\mathbf{r}, t)$, one gets:

$$\Gamma_{il} a_l = \rho v^2 a_i \quad (9.5)$$

where the Christoffel matrix elements Γ_{il} are

$$\Gamma_{il} = c_{ijkl} n_j n_k \quad (9.6)$$

The density times the squared phase velocity ρv^2 appear as eigenvalues of the Christoffel matrix (Γ_{il}) , and the vector $\mathbf{a}$, the polarization vectors, are the associated eigenvectors. Due to the symmetry conditions, the fourth order elasticity tensor (or the elastic constant matrix) c_{ijkl} , is given by only 21 independent constants for a general anisotropic elastic medium. This number, however, reduces to 5 in the case of transversely isotropic media.

9.2.1 Transversely Isotropic Media

This study restricts the analysis to the case of anisotropy with hexagonal symmetry or a transversely isotropic (TI) media, since this is the simplest and viable model for most of the possible causes of effective anisotropy (Babuska and Cara 1991). A transversely isotropic medium is fully described by five independent elastic constants. In this case, the elastic constant matrix has the simple shape

$$c_{\alpha\beta} = \begin{pmatrix} c_{11} & c_{11} - 2c_{66} & c_{13} & & & \\ c_{11} - 2c_{66} & c_{11} & c_{13} & & & \\ c_{13} & c_{13} & c_{33} & & & \\ & & & c_{44} & & \\ & & & & c_{44} & \\ & & & & & c_{66} \end{pmatrix} \quad (9.7)$$

with five unknowns, assuming that the symmetry axis is in x_3 direction. Note that we have adopted the Voigt convention where α and β are two indices which represent, in a more compact way, the double indices (ij) and (kl) of the elastic constant c_{ijkl} such that:

$$\begin{array}{cccccccc} ij \text{ or } kl & : & 11 & 22 & 33 & 32 = 23 & 31 = 13 & 12 = 21 \\ \downarrow & & \downarrow & \downarrow & \downarrow & \downarrow & \downarrow & \downarrow \\ \alpha & \beta & 1 & 2 & 3 & 4 & 5 & 6 \end{array} \quad (9.8)$$

An isotropic medium, for example, can then just be represented as a special case of a transversely isotropic medium with $c_{33} = c_{11} = \lambda + 2\mu$, $c_{66} = c_{44} = \mu$, and $c_{13} = c_{33} - 2c_{44} = \lambda$ where λ , and μ are the Lamé parameters.

9.3 The polarization of P-waves in a Transversely Isotropic Media

Let us now move to discuss the polarization nature of the P-waves. The polarization of plane waves propagating in transversely isotropic media can be expressed in an analytically *exact* form. It has been shown previously that the propagation

of plane wave in anisotropic media is governed by the Christoffel equation. The Christoffel equation is obtained by substituting the plane wave expression into the equation of motion. For P- and SV- waves propagating in the (x_1, x_3) plane of a transversely isotropic medium with the x_3 direction coincides with the symmetry axis, the Christoffel equation reduces to (Musgrave 1970)

$$\Gamma_{11} a_1 + \Gamma_{13} a_3 = 0 \quad (9.9)$$

$$\Gamma_{13} a_1 + \Gamma_{33} a_3 = 0 \quad (9.10)$$

where a_1 and a_3 are the components of the polarization vector, and Γ is the Christoffel matrix with

$$\Gamma_{11} = c_{11} \sin^2 \theta + c_{44} \cos^2 \theta - \rho v^2 \quad (9.11)$$

$$\Gamma_{33} = c_{44} \sin^2 \theta + c_{33} \cos^2 \theta - \rho v^2 \quad (9.12)$$

$$\Gamma_{13} = (c_{13} + c_{44}) \frac{\sin \theta \cos \theta}{v^2} \quad (9.13)$$

where ρ is the density, and phase angle, θ , is the angle between the direction of propagation (or the wavefront normal) and the symmetry axis x_3 (see Fig. 9.2).

The P- and SV- wave phase velocities can be obtained by combining these equations and the quadratic equation $\Gamma_{11} \Gamma_{33} - \Gamma_{13}^2 = 0$ formed from equations (9.9) and (9.10) (Daley and Hron 1977). They can then be written as

$$2\rho v_P^2 = c_{33} + c_{44} + (c_{11} - c_{33}) \sin^2 \theta + D(\theta) \quad (9.14)$$

$$2\rho v_{SV}^2 = c_{33} + c_{44} + (c_{11} - c_{33}) \sin^2\theta - D(\theta) \quad (9.15)$$

A compact notation for the quadratic combination $D(\theta)$ is given as

$$\begin{aligned} D(\theta) = & \left\{ (c_{33} - c_{44})^2 \right. \\ & + 2 \left[2(c_{13} + c_{44})^2 - (c_{33} - c_{44})(c_{11} + c_{33} - 2c_{44}) \right] \sin^2\theta \\ & \left. + \left[(c_{11} + c_{33} - 2c_{44})^2 - 4(c_{13} + c_{44})^2 \right] \sin^4\theta \right\}^{1/2} \end{aligned} \quad (9.16)$$

The polarization direction for the P- or SV-waves can be obtained by substituting Γ_{11} , Γ_{33} , and Γ_{13} into equation (9.9) or (9.10). The polarization angle, given by the angle γ , measured from the symmetry axis x_3 (Fig. 9.2), can then be defined as

$$\tan \gamma_{P,SV} = \frac{a_1}{a_3} = \frac{\sin\theta \cos\theta (c_{13} + c_{44})}{\rho v^2 - (-1)^i (c_{11} \sin^2\theta - c_{44} \cos^2\theta)} \quad (9.17)$$

where $i = 1$ for SV-waves and $i = 2$ for P-waves. Equation (9.17) is valid for any strength of the anisotropy, since this equation is an exact expression of the polarization angle, γ , as a function of the propagation direction, θ .

The practical use of equations (9.14) and (9.15) is, however, limited (if not difficult) because θ cannot be measured directly from three-component data. It is this data which is available to this study. Moreover, these equations are too complicated to be readily used for inversion of elastic parameters. To resolve this problem an expression which relates the direction of propagation to the direction of polarization is then needed. This will be shown below.

9.3.1 Definition of two "de Parscau" anisotropy parameters

To express the P- and SV-wave phase velocities as a function of the polarization angle γ , two dimensionless parameters (η and τ) proposed by de Parscau (1991b) are employed. These parameters are defined as

$$\eta = \frac{c_{13} + c_{44}}{c_{33} - c_{44}} \quad (9.18)$$

and

$$\tau = \frac{c_{11} - c_{44}}{c_{33} - c_{44}} \quad (9.19)$$

Rewriting equations (9.14) and (9.15) using equations (9.18) and (9.19) gives

$$\begin{aligned} \rho v_{P,SV}^2 = & \frac{(c_{33} - c_{44})}{2} \left[\frac{(c_{33} + c_{44})}{(c_{33} - c_{44})} \right. \\ & + (\tau - 1) \sin^2 \theta + (-1)^i \{ (1 - (1 \\ & \left. + \tau) \sin^2 \theta)^2 + (\eta \sin 2\theta)^2 \}^{1/2} \right] \end{aligned} \quad (9.20)$$

where $i = 1$ for SV-waves and $i = 2$ for P-waves. This equation shows that the P-velocity is sensitive to τ variations for θ higher than 45 degrees because of the $\sin^2 \theta$ term, and to η variations for θ lower than 45 degrees because of the $\sin 2\theta$ term. Note that for an isotropic medium we have $\eta = \tau = 1$.

Equation (9.17) can also be rewritten using the two parameters η and τ to give

$$\tan \gamma_{P,SV} = \frac{A}{B + (-1)^i \sqrt{B^2 + A^2}} \quad (9.21)$$

with A and B defined as follows

$$A = \eta \sin 2\theta \quad (9.22)$$

$$B = 1 - (1 + \tau) \sin^2 \theta \quad (9.23)$$

where, again, $i = 1$ for SV-waves and $i = 2$ for P-waves. After some algebra, it is easily found that

$$\begin{aligned} \frac{A}{B} &= \frac{2 \tan \gamma_{P,SV}}{1 - \tan^2 \gamma_{P,SV}} \\ &= \tan \gamma_{P,SV} \end{aligned} \quad (9.24)$$

and a simple relation is obtained

$$\tan 2\gamma_{P,SV} = \frac{\eta \sin 2\theta}{1 - (1 + \tau) \sin^2 \theta} \quad (9.25)$$

This equation (9.25) is valid for P- and SV- wave polarizations (which are orthogonal to one another) due to the periodicity of the tangent ($\tan \alpha = \tan(\alpha + \pi)$). It is interesting to note here that the polarization angle depends only on the two parameters η and τ and not on the two elastic constants c_{33} and c_{44} (which are related to the vertical P- and SV- wave phase velocities). This means that the pattern of the angular difference between polarization angle and the propagation angle (which

herein is referred as polarization anomaly) is solely governed by these two dimensionless parameters and not by the velocity of the anisotropic medium. This equation also shows, however, that the four elastic constants c_{11} , c_{13} , c_{33} , and c_{44} cannot be independently determined from P-wave polarization data. Instead, constraints on these constants must be derived from η and τ using the non-linear relations (9.18) and (9.19). In addition to η and τ , the orientation of the symmetry axis is also unknown. Therefore, if we were to invert the nature of the anisotropic medium from P-wave polarization data, we will have two additional unknowns representing strike and dip of the symmetry axis of the transversely isotropic media.

9.4 Angular deviation between polarization and propagation vectors

A plane P-wave propagating in an anisotropic (or isotropic) media advances at the phase velocity in the direction of propagation vector, normal to the surface of constant phase. It is clear from the previous discussion that, in general, the propagation and polarization vectors do not coincide for waves propagating in anisotropic media. They do, however, coincide in the isotropic case. The propagation vector may be specified by azimuth θ_s and horizontal apparent velocity $\bar{V}$. Alternatively, $\bar{V}$ can be represented in terms of incidence angle $\phi_s = \sin^{-1}(v/\bar{V})$ using the knowledge of local velocity v . This propagation vector can be estimated from an array data either from traveltimes or from F-K analysis such as a beam-forming technique (see for example Beauchamp, 1975). The polarization vector, $\mathbf{a}$, which is the dominant direction of linear particle motion, on the other hand, can be computed directly from the three-component data by polarization analysis (see for example Kanasewich, 1981; Winardhi, 1994). This vector can also be specified by back azimuth θ_a and incidence angle ϕ_a .

It is interesting to note that the beam-forming technique is well suited for signal enhancement, event detection and earthquake localization (see for example Beauchamp, 1975). However, so far, this particular method has apparently not been used in anisotropic studies. The angular deviation between phase and group velocity received quite some interest in the literature (see for example Crampin 1981; Crampin et al 1982). However, the deviation between polarization and phase velocity vectors has merely been discussed for model crystals (e.g Crampin 1978, 1981; Crampin et al 1982). So far, there are only a few applications to real data. White et al (1983) successfully applied this concept to infer effective anisotropy in Pierre Shale. Li et al (1987) studied the effect of fault-zone fracture on the polarization of seismic P-wave using VSP data. de Parscau (1991a) analyzed the phase velocities of P- and SV- waves also for VSP borehole data. Recently, the angular deviation between these two vectors has been studied to help define the anisotropy of the upper crust using regional to teleseismic earthquake data (Bokelmann 1995). It is therefore worthwhile to analyse the polarization-propagation deviation of seismic P-wave based on array of three-component refraction data in order to gain knowledge of its usefulness and limitations in the context of anisotropy study.

It is important to note, at this stage, that neither of these two azimuth estimates, θ_s and θ_a , necessarily corresponds to the azimuth of the shot-receiver, since heterogeneity, anisotropy and other effects may cause deviations of both propagation and polarization. However, it can be seen that the angular deviation between polarization and propagation directions, as a single relative quantity, depends only on local anisotropy in the vicinity of the receivers. This is because all remote effects, such as distant lateral heterogeneity and/or distant anisotropy, affect both polarization and propagation simultaneously. Near-receiver anisotropy should then explain the

observed polarization anomalies. But how about near-receiver heterogeneity? Small scale heterogeneity near the receivers may also cause distortions on the incoming wavefront. Hence, polarization data from a single three-component station has to be regarded with caution. However, the effect of near-receiver small-scale heterogeneity is expected to be uncorrelated from station to station. Averaging over a number of three-component stations within the array is therefore expected to remove or average out most of its effect on the difference pattern effectively. The effect of larger scale heterogeneity is, on the other hand, likely to be correlated from station to station. A net effect which give the observed pattern of the angular difference between polarization and propagation directions may then be explained as a manifestation of local effective macroscopic anisotropy of the upper crust in the vicinity of the receiver array. This treatment, in fact, allows us to simplify the analysis of the data in terms of an anisotropic but homogeneous model.

9.5 Forward modelling of P-wave polarization patterns

Equation (9.25) allows us to study analytically the behaviour of the P-wave polarization directions in transversely isotropic media for various values of η and τ . This equation is therefore the core for the forward modelling process. Unfortunately, equation (A.5) is limited only for a 2-dimensional problem. To understand the pattern and magnitude of the deviation of the polarization from propagation directions (polarization anomalies) completely, a full 3-dimensional modelling is needed. This equation has thus to be extended into a 3-D formulation. This is done quite straightforwardly using a method similar to that used by Niazi (1966) to calculate azimuthal deviation due to a dipping interface (see Appendix A).

To help visualize the 3-dimensional patterns of the polarization anomalies in

a 2-D paper, a lower-hemisphere polar diagram representation commonly used by geologist and earthquake seismologist was employed. This type of diagram is well suited to this analysis since both propagation and polarization vectors can be specified in terms of azimuth and incidence angle. Using this polar diagram, a vector is simply plotted as a point specified by its azimuth and sine of its incidence angle. A plane wave incident from vertically beneath a receiver, for example, can easily be illustrated as a point plotted at the center of the lower-hemisphere diagram. A horizontally travelling plane wave approaching the receiver, on the other hand, is simply visualized by a point located at the periphery of the diagram.

Figures 9.3 – 9.5 show the patterns of the P-wave polarization anomalies for a medium with $\eta = 1.43$ and $\tau = 2.13$ for three different orientations of the symmetry axis. This model actually describes a medium containing well aligned vapour-filled cracks or dry-crack of Hudson (1981). The effective anisotropic elastic constant for this particular model was calculated based on model HCD1 of Crampin (1984). The changes in strike and dip of the crack alignments can easily be simulated as the changes in the orientations of the symmetry axis. The variation in the density, size and type of the cracks (wet or dry), on the other hand, can simply be accommodated as the variation in η and τ values. Further discussion on this topic will be found later in the analysis section.

9.6 Analysis of the ABGR-92 Three-component data

9.6.1 The Three-component data

An array of three-component (3-C) seismometers which were deployed in the vicinity of the Grenville Front Tectonic Zone (GFTZ) during the 1992 Lithoprobe Abitibi-Grenville Seismic Refraction Experiment is shown in Fig. 9.1. There were 47 3-C

systems located strategically across the GFTZ at an average station spacing of 1 - 1.5 km. These 3-C systems were deployed in two localities: south of Sudbury (line AB) and southeast of Lake Temiscaming (line MG). This provides us an opportunity to study the similarities (and dissimilarities) on the anisotropic nature of the front in these two locations.

The differences (and perhaps the similarities) between the nature of the crust to the south and north of the GFTZ will be explored utilizing these 3-C arrays. For the purposes of this study, the 3-C array dataset were grouped into two regions, namely south and north of the GFTZ. The south group consists of the 3-C data which were recorded by the 3-C stations/seismometers situated south of the Grenville Front, whereas the north group consists of the dataset recorded by the stations which were located north of the Grenville Front. Since two profiles (AB and MG) were considered in this study, we have in total four groups of 3-C data to be analyzed. Two groups belong to line AB and the other two groups to line MG.

It can be seen in Fig. 9.1 that the 3-C data for line MG involves nine shot points (i.e., GD6 – MD6), whereas for line AB, only eight shot points (i.e., AB0 – AB10) are involved. Figures 9.6 – 9.8 show, for example, the nature of the 3-C seismograms which was shot from shot point MD8 (see Fig. 9.1). A high quality first arrivals and $P_m P$ are clearly demonstrated in that figure. However, in this study, only the first 0.5 – 1.0 second of the seismograms which were analyzed, since the polarization is measured from the direction of the particle motion of the first arrival.

9.6.2 Determination of Polarization direction

The procedures of extracting azimuth and incidence angle information from three-component data are relatively simple. Consider a three-component seismogram as

three time series, $x_z(t)$ for the vertical component, $x_n(t)$ for the North-South component and $x_e(t)$ for the East-West component. In a vector form this three-component data can then be expressed as (see for example Kanasewich, 1981)

$$\mathbf{X} = [x_z(t), x_n(t), x_e(t)]^{tr} \quad t = n\Delta t; n = 0, 1, \dots, N-1, \quad (9.26)$$

where Δt is the sampling interval, $N\Delta t$ is the length of the time series within a specified window and tr denotes the operation of vector transposition. The $(k, m)th$ component of the covariance matrix at zero-lag over a specified window is then evaluated as

$$\begin{aligned} S(k, m) &= \frac{1}{N} \sum_{i=1}^N x_{ik} x_{im} \\ &= \langle x_k x_m \rangle \end{aligned} \quad (9.27)$$

where the angle brackets represent time averages. In a matrix form, the 3×3 covariance matrix, which is real and symmetric, can then be rewritten as

$$S = \begin{bmatrix} \langle x_z x_z \rangle & \langle x_z x_n \rangle & \langle x_z x_e \rangle \\ \langle x_n x_z \rangle & \langle x_n x_n \rangle & \langle x_n x_e \rangle \\ \langle x_e x_z \rangle & \langle x_e x_n \rangle & \langle x_e x_e \rangle \end{bmatrix} \quad (9.28)$$

where $\langle x_z x_n \rangle$, for example, denotes the cross-variance of the vertical and north components, etc. Note that the index convention used here is $z = 1$, $n = 2$, and $e = 3$. The covariance matrix $\mathbf{S}$ is real and symmetrical, its eigenvalues are therefore

real and non-negative (some may be zero). $\mathbf{S}$ is actually the matrix of coefficients for a quadratic form which is an ellipsoid. This ellipsoid, termed the polarization ellipsoid, is the best fit to the data in a least-squares sense. Winardhi (1994) showed that the principal axes of the ellipsoid are found by simply solving the algebraic eigenproblem for $\mathbf{S}$. This involved finding the eigenvalues $(\lambda_1, \lambda_2, \lambda_3)$ and its corresponding eigenvectors $(\mathbf{a}_1, \mathbf{a}_2, \mathbf{a}_3)$ which are solutions to

$$(\mathbf{S} - \lambda^2 \mathbf{I}) \mathbf{a} = 0 \quad (9.29)$$

where $\mathbf{I}$ is the 3×3 identity matrix and 0 is a column vector of zeros. This eigenproblem is easily solved using standard numerical libraries such as *Numerical Recipes* of Press et al (1992). A set of three orthonormal eigenvectors $\mathbf{a}$ are chosen and $\lambda \mathbf{a}$ then give the axis of the polarization ellipsoid. The eigenvalues are ordered such that $\lambda_j \geq \lambda_k$ for $j < k$. If the particle motion is linearly polarized, then $\lambda_2 = \lambda_3 = 0$. For elliptical particle motion within a polarization plane, then $\lambda_3 = 0$. In general, due to the presence of noise, the particle motion is ellipsoidal.

The polarization direction in terms of the incidence angle ϕ_a and back azimuth θ_a were easily calculated by

$$\phi_a = \cos^{-1} |a_{11}| \quad (9.30)$$

$$\theta_a = \begin{cases} \tan^{-1} \left(\frac{a_{31}}{a_{21}} \right), & \text{for } a_{11} < 0 \\ \pi + \tan^{-1} \left(\frac{a_{31}}{a_{21}} \right), & \text{for } a_{11} \geq 0 \end{cases} \quad (9.31)$$

where a_{j1} ; $j = 1, 2, 3$ are the three direction cosines of eigenvector $\mathbf{a}_1$. A simple check of the quality of the measurement, in this study, is given by the degree of rectilinearity

$$R = 1 - \frac{\lambda_2 + \lambda_3}{2\lambda_1} \quad (9.32)$$

which is equal to 1 for the case of linearly polarized waves and 0 in the worst case of spherical polarization (i.e., when all eigenvalues have the same size).

Figures 9.9 – 9.11 show examples of this extraction procedure. An early portion of a band-passed (2 – 14 Hz) filtered seismograms from shot AB0 is shown in Fig. 9.9. A clear first arrival is demonstrated in this diagram. Fig. 9.10 shows its corresponding polarization attributes, namely degree of rectilinearity, angle of incidence and back azimuth, obtained from the analysis. Notice that as the value of the rectilinearity reaches a certain level, the angle of incidence and back azimuth values become relatively stable for sometime. It is then obvious that those quantities (ϕ_a and θ_a) can be extracted by simply defining a *cut off* level (for the degree of rectilinearity) of 0.8, for example, and simply reading its corresponding values of incidence angle and back azimuth. Of course, there will be many values of these quantities which fall above this specified level. The final value is therefore obtained by simple averaging.

The validity of this procedure can be checked by plotting a particle motion diagram for this particular arrival, as shown in Fig. 9.11. The very close similarities between both quantities, angle of incidence and back azimuth values as obtained using the covariance matrix technique and the one obtained from the visual in-

spection of the particle motion diagram, justifies the validity and reliability of the method. Clearly, a visualization method to extract the polarization attributes, such as the particle motion diagram, is highly inefficient as compared to the automatic covariance matrix procedure employed in this study.

9.6.3 Free-surface effect

It has been shown in the previous section that the direction of polarization, expressed in terms of angle of incidence and back azimuth can be extracted using the covariance matrix technique. However, this measured polarization direction is only an apparent incidence angle (i.e., not a true incidence angle). This is because the conversion at the free surface perturbs the true incidence angle of the polarization direction, as illustrated in Fig. 9.12 (upper panel). The true angle of incidence can still be obtained, however, by the virtue of the Wiechert's relation (see Aki and Richards, 1980)

$$2 \sin^2 \gamma = \left(\frac{V_p}{V_s} \right)^2 (1 - \cos \bar{\gamma}) \quad (9.33)$$

where γ denotes the true incidence angle, while $\bar{\gamma}$ is the apparent incidence angle. V_p and V_s are P- and S- waves velocity, respectively. To illustrate the dependence of $\bar{\gamma}$ on γ , equation (9.33) may simply be manipulated, using simple triginometrical relation, to give

$$\bar{\gamma} = \tan^{-1} \left[\frac{2 \sin(\gamma) \sqrt{A^2 - \sin^2(\gamma)}}{A^2 - 2 \sin^2(\gamma)} \right] \quad (9.34)$$

where A denotes the velocity ratio $A = \frac{V_p}{V_s}$. Figure 9.12 (lower panel) shows the dependence of $\bar{\gamma}$ on γ for a number of Poisson's ratios κ , which are related to A by

$$\kappa = \frac{(A^{-2} - 0.5)}{(A^{-2} - 1)} \quad (9.35)$$

The three values illustrated in that figure are: $\kappa = 0.2$ (i.e., $A = 1.63$), $\kappa = 0.25$ (i.e., $A = 1.73$) and $\kappa = 0.3$ (i.e., $A = 1.87$). It can be seen from this figure that the free-surface correction is particularly important for large incidence angles. Clearly, the free-surface effect is an important and observable effect, which must be taken into account. In this study, a crustal Poisson ratio of $\kappa = 0.245$ was used for the free-surface correction. This corresponds closely to the crustal mean value obtained from the least-squares study described previously in chapter 4.

9.6.4 Determination of Propagation direction

Propagation directions (θ_s, ϕ_s) can be extracted from array of vertical component data either from traveltimes or from F-K analysis. In this study, the former is employed. The method, which is well known for event detection and earthquake localization, is relatively simple to implement. It can basically be summarized as follows. Consider an array made up of N seismometers/stations scattered on the surface with a point O to be the center of the array (see Fig. 9.13). Consider now a wavefront of P-wave crossing the array with an azimuth θ and apparent velocity of $\bar{V}$ as shown in Fig. 9.13. Since coordinates of each stations (x_i, y_i) with respect to the center point O are all known or can easily be calculated based on their geometrical relationship, the arrival time of the P-wave at the center of the array, T_0 can then

be expressed as

$$\begin{aligned}
 T_0 &= T_i + T_{OM} + \epsilon_i \\
 &= T_i + \frac{OM}{\bar{V}} + \epsilon_i \\
 &= T_i + \frac{R_i(\cos \theta_i \cos \psi_i + \sin \theta_i \sin \psi_i)}{\bar{V}} + \epsilon_i
 \end{aligned} \tag{9.36}$$

where T_i denotes the arrival times at each station and ϵ_i is an error term for station i . By letting $P = \sin \theta / \bar{V}$ and $Q = \cos \theta / \bar{V}$ and noting that $x_i = R_i \sin \psi_i$ and $y_i = R_i \cos \psi_i$, the above relation can be simplified as

$$T_0 = T_i + x_i P + y_i Q + \epsilon_i \tag{9.37}$$

The error term ϵ_i can then be stated as

$$\epsilon_i = T_0 - T_i - (x_i P + y_i Q) \tag{9.38}$$

The best values of T_0 , P and Q can then be obtained by minimizing the error term with respect to these three unknown. This results in a 3 x 3 normal equation which can be solved easily using matrix inversion. The apparent horizontal velocity $\bar{V}$ and the azimuth θ of the incoming wavefront can then be obtained from

$$\bar{V} = \frac{1}{\sqrt{P^2 + Q^2}}, \quad \text{and} \quad \theta = \tan^{-1} \left(\frac{P}{Q} \right) \tag{9.39}$$

Conversion from the apparent horizontal velocity $\bar{V}$ to incidence angle $\phi_s = \sin^{-1}(v/\bar{V})$

was then done using the local velocity estimate v obtained from the tomographic analysis.

9.6.5 The polarization anomaly diagram

It has been mentioned in the previous section that the angular difference between the polarization and propagation directions (i.e., polarization anomalies/deviations) is the main subject of interest. It has also been shown that in order to easily visualize these patterns, the lower hemisphere polar diagram representation was employed. For each shot we have a number of observation of polarization direction (θ_a, ϕ_a) and propagation direction (θ_s, ϕ_s) . These two vectors can easily be plotted on the diagram as two points. A small circle was used here to represent the propagation vector, whereas a small star was used to represent the polarization vector. A straight line connecting these two points then represents the polarization vector deviation indicating the presence of anisotropy.

Figures 9.14 – 9.17 show the polarization anomaly diagram for line MG and AB, respectively. The diagram clearly shows the pattern of the polarization anomaly vectors (before and after averaging), for both receiver groups, namely south and north of the Grenville Front (GF) (please refer back to section 9.6.1). The size of the polarization anomalies ranges from 5 to 15 degrees. It can be seen from the raw anomaly diagram (i.e., before averaging) that scatter is invariably present. However, this scatter is dominated by random fluctuation and not by effects from discrete local scatterers. This is justified from the fact that the predominant feature of the anomaly pattern observed in the raw anomaly diagram remains practically the same after averaging. This is a very important observation for telling us that the cause of the observed anomalies is mainly due to the anisotropic nature of the

crust. It is also observed from those figures, the significant difference in magnitude and pattern of the polarization anomalies between these two localities, south and north of the GF. Distinct pattern of anomalies are observed mainly for the south of the GF groups (see Figs. 9.14 and 9.16), whereas the north group for line MG (shown in Fig. 9.15) shows a diffuse/noisy pattern. The north group for line AB (see Fig. 9.17), however, shows a somewhat similar pattern of anomalies to the one shown for the south part of line AB.

It has just been shown that the anisotropy does affect the relative data (propagation-polarization), as shown in Figs. 9.14 – 9.17. However, a question arises concerning to what depth extent are these data sensitive to? This is a crucial question as the answer to this question may guide us in choosing the most plausible *a priori* model which may be used to explain the observed anomalies. To answer this question we may have to recall to the fact that polarization is a local property (Hearty et al., 1977; Hu and Menke, 1992; Hu et al., 1994). This implies that the data are affected by anisotropy in the depth range equivalent to about a wavelength of the seismic phases employed. Since the typical dominant frequencies in this study ranges from 2 – 6 Hz, we have wavelengths of about 1 – 3 km. The data are therefore expected to represent *effectively* a spatial average of anisotropic properties in the top few kilometers of the upper crust. Also note that these are the depth ranges where the brittle (as opposed to ductile) deformation is expected. Moreover, since we average over a number of three-component stations within the array, it is then valid to assume that discrete heterogeneity within the array can be ignored. This allows us to, once again, analyse the data in terms of anisotropic but homogeneous model.

9.6.6 Inversion technique for recovering the anisotropic parameters

There are several phenomena which may cause effective anisotropy. Intrinsic anisotropy of minerals, as well as small-scale structure with some preferred symmetry may lead to directional dependence of macroscopic elastic parameters. This results in directional dependence of propagation velocity, hence anisotropy (Rabbel and Mooney, 1996). Sedimentary layering may also cause a potentially prominent effect (see for example White, 1983). Cracks, fractures (see for example Crampin, 1984) and rock foliation (Babuska and Cara, 1991), if certain orientations prevail, can also give rise to substantial effects.

The analysis is therefore restricted to the case of transversely isotropic media, since this is a viable model for most of the possible causes of effective anisotropy described above. In this case, the parameters involved in the inversion are η and τ , as well as two angles $\hat{\Theta}$ and $\hat{\Phi}$ describing azimuth and dip, respectively, of the symmetry axis. This is because of the fact that the four elastic constants cannot be determined independently. Constraints on these constants may instead be derived from η and τ .

Each event is described by polarization direction (θ_a, ϕ_a) and propagation direction (θ_s, ϕ_s) to give the observed angular differences (i.e., the observed anomalies)

$$\begin{pmatrix} \Delta\theta \\ \Delta\phi \end{pmatrix} = \begin{pmatrix} \theta_a - \theta_s \\ \phi_a - \phi_s \end{pmatrix} \quad (9.40)$$

The inversion problem can then be stated as: *find the parameters $(\hat{\Theta}, \hat{\Phi}, \eta, \tau)$ which give the best prediction $(\Delta\theta' \Delta\phi')$ to the observed anomalies $(\Delta\theta \Delta\phi)$.* Since the lower-hemisphere diagram was employed, the misfit on the sphere was therefore cho-

sen. This misfit function is defined as

$$S(\hat{\Theta}, \hat{\Phi}, \eta, \tau) = \frac{1}{N} \sum_{i=1}^N \{ \cos^{-1} [\cos(\Delta\theta_i - \Delta\theta'_i) \cos(\Delta\phi_i - \Delta\phi'_i)] \}^2 \quad (9.41)$$

which is analogous to the standard sum-of-squares misfit criterion.

Recalling that our data are sensitive to a spatial average over several kilometers, macroscopic anisotropic behaviour is therefore expected to predominate. Furthermore our P-wave polarization data are sensitive to the depth range where brittle deformation is expected. This means that the observed anomalies may then reasonably be explained as due to a spatially coherent orientation of cracks or fractures which predominate in the topmost part of the upper crust. This guess may be used as our a priori model.

The inversion was therefore started with fitting $\hat{\Theta}$ and $\hat{\Phi}$ using fixed values of η and τ . In this case, we use two existing models of cracks HCD1 and HCS1 of Crampin (1984). HCD1 represents a model for the dry-cracks, whereas HCS1 represents the saturated cracks (i.e., waterfilled cracks). The values of η and τ which correspond to these two models were calculated using the values of elastic constants for these two models (listed on Crampin, 1984). HCD1 is represented by ($\eta = 1.43$ $\tau = 2.13$), whereas HCS1 is represented by ($\eta = 0.82$ $\tau = 1.0$). A non-linear search utilizing the forward problem formulated in the previous section was then performed. Since the model parameters involved are only two ($\hat{\Theta}$ and $\hat{\Phi}$), we employed a combination of the Very Fast Simulated Annealing technique (described previously in chapter 7) and the conjugate gradient rather than using only gradient method. This is chosen

for the purpose of fast but fine exploration of any possible value of the symmetry axis. The VFSA was used to explore the model space, while the conjugate gradient was employed to seek the real minimum. This method is much faster compared to the grid search in the case of fine grids and, most importantly, allows us to graph pictures of prior and posterior distribution of the solutions.

The results of the inversion are shown in Figs 9.18 – 9.22. It can be seen that, in general, the dry-cracks explains the observed anomalies better than the saturated cracks as indicated by the histograms of the posterior distributions which have been considerably narrowed. It is also observed from the histogram plots of the posterior solutions that the parameter 'dip' is generally better resolved than the 'azimuth'. This is obviously due to the fact of lack in azimuthal coverage in the three-component datasets (see Fig. 9.1) as compared to the coverage in the incidence angles. By the virtue of constrained but global search utilizing the Very Fast Simulated Annealing, the inversion still however manage to give a reasonable solution.

It is observed that the polarization anomalies data for line MG south of the Grenville Front can be explained as a manifestation of anisotropic phenomenon due to the presence of dry-cracks. Its symmetry plane has a strike of (N 70 E) and a dip of 80 – 90 degrees (i.e. almost vertical) to the south. Similarly, the observed anomalies beneath line AB south of the GF can also be regarded as due to an aligned dry-cracks with the plane of symmetry oriented at a strike of (N 45 E) and a dip of 80 – 90 degrees to the south.

9.7 Discussion of the Results

It has been shown from the inversion results that in principle the macroscopic effect of dry-cracks was able to explain the pattern and size of the observed anomalies. To understand the nature of this anisotropy, the surface geology of the area in the vicinity of the 3-C arrays should be closely examined. The receivers within the array are mostly located on an outcrop of granite and gneiss. Apparently granite has no preferred orientation/alignment. However, the gneiss shows strong foliation with consistent orientations. The strike of this foliation (see Rivers et al., 1993) closely coincides with the orientation of the strike of the symmetry axis obtained from the best-fitted dry-cracks model. This suggests that the occurrence of cracks may be located along the plane of weakness given by the plane of foliation.

Long and Salisbury (1996) have reported, based on laboratory and refraction studies, that several shallow (< 8 km) layers in the vicinity of the Grenville Front Tectonic Zone were found to be anisotropic with fast-velocity direction dipping to the south. Wang et al. (1975) have demonstrated that, for small pressures in the laboratory, cracks opening along the foliation direction is very common to observe. Luschen et al. (1991) have also shown, based on the integrated P- and S- wave borehole experiments at the KTB (Continental Deep Drilling Program) site in the Oberpfalz area (South-East Germany), that rock foliation may also be the cause of effective anisotropy. Of course, the overall anisotropy could also be related to intrinsic properties of minerals. Gneisses, for example, typically contain large amounts of mica, a mineral which gives rise to very large anisotropy. It appears difficult, however, to distinguish between effects of intrinsic anisotropy and cracks or fractures based mainly on this limited amount of data. Perhaps additional insight can be gained if S- wave arrivals are also employed. However, unambiguous determination

of upper crustal anisotropy requires incidence angles within the shear-wave window (up to 35 degrees) (see Nuttli, 1961; 1964 and Evans, 1984). These conditions are difficult to meet for our refraction data. This is, in fact, the rationale of not employing the S-wave arrivals for this particular study of upper crustal anisotropy.

Figure 9.1: Location of the three-component seismometers array deployed in the vicinity of the GFTZ during the 1992 Lithoprobe Abitibi-Grenville Seismic Refraction Survey.

Figure 9.2: Definition of polarization and propagation directions for P-waves. The propagation vector is normal at the plane wave tangent to the wavefront and defines an angle θ with the axis of symmetry. The polarization vector represents the direction of particle motion and defines an angle γ with the symmetry axis.

Figure 9.3: Lower hemisphere representation of P-wave polarization anomalies for $(\eta \text{ and } \tau) = (1.43, 2.13)$ corresponding to the dry-crack model HCD1 of Crampin (1984). In this display (and the following ones), north is up. A small circle represents the propagation vector (θ), whereas a straight line points to the direction of polarization (γ).

Polarization Anomaly Diagram
 Dry-crack ($\eta = 1.43$, $\tau = 2.13$)
 Symmetry plane (strike = 60, dip = 20 to the South)

Figure 9.4: Lower hemisphere representation of P-wave polarization anomalies for the dry crack model (η and τ) = (1.43, 2.13). The symmetry plane has a strike of 60 degree and a dip of 20 degree to the south.

Polarization Anomaly Diagram
 Saturated-crack ($\eta = 0.82$, $\tau = 1.00$)
 Symmetry plane (strike = 60, dip = 60 to the South)

Figure 9.5: Lower hemisphere representation of P-wave polarization anomalies for the saturated crack model (η and τ) = (0.82, 1.00). The symmetry plane has a strike of 60 degree and a dip of 60 degree to the south. Note that the pattern and magnitude of the polarization anomalies differ substantially from the dry-crack model shown in Fig. 9.4.

Figure 9.6: Vertical component of the three-component record section for line MG from shot MD8.

Figure 9.7: North-South component of the three-component record section for line MG from shot MD8.

Figure 9.8: West-East component of the three-component record section for line MG from shot MD8.

Figure 9.9: An early portion of a band-passed (2 – 14 Hz) filtered three-component seismograms for line AB from shot AB0. A clear first arrival is demonstrated in this diagram.

Figure 9.10: Polarization attributes for the three-component data shown in Fig. 9.9, namely degree of rectilinearity (upper panel), angle of incidence (middle panel) and back azimuth (lower panel) extracted by utilizing the covariance matrix technique.

Figure 9.11: Particle motion diagram of the first arrival of the seismograms shown in Fig. 9.9. Note the very close similarities between angle of incidence (upper panel) and back azimuth (lower panel) as compared to the one obtained using the covariance matrix technique (shown in Fig. 9.10).

Figure 9.12: Free-surface effect (upper panel) on polarization incidence angles for a set of Poisson's ratios (lower panel).

Figure 9.13: Diagram illustrating a wavefront of P-wave crossing an array of seismometers with an azimuth θ .

Figure 9.14: Lower hemisphere representation of P-wave polarization anomalies under a portion of line MG south of the Grenville Front (GF) showing event polarization anomalies (upper panel) and its corresponding average pattern (lower panel). In this display (and the following ones), a small circle represents the propagation vector (θ), whereas a small star represents the polarization vector (γ). A straight line connecting these two points then represents the polarization vector deviation indicating the presence of anisotropy.

Figure 9.15: Lower hemisphere representation of P-wave polarization anomalies under a portion of line MG north of the Grenville Front (GF) showing event polarization anomalies (upper panel) and its corresponding average pattern (lower panel).

Polarization Anomaly Diagram

Line AB - South of GF

Figure 9.16: Lower hemisphere representation of P-wave polarization anomalies under a portion of line AB south of the Grenville Front (GF) showing event polarization anomalies (upper panel) and its corresponding average pattern (lower panel).

Polarization Anomaly Diagram

Line AB - North of GF

Figure 9.17: Lower hemisphere representation of P-wave polarization anomalies under a portion of line AB north of the Grenville Front (GF) showing event polarization anomalies (upper panel) and its corresponding average pattern (lower panel).

Figure 9.18: A scatterplot of uniformly distributed prior samples (upper panel) with histograms (lower panel) confirming the uniform shape of the priors.

Figure 9.19: A scatterplot of posterior samples (upper panel) with histograms (lower panel) confirming significant variance reductions of the posterior solutions (line MG in the case of the dry-cracks priors).

Figure 9.20: A scatterplot of posterior samples (upper panel) with histograms (lower panel) confirming no significant variance reductions of the posterior solutions (line MG in the case of the saturated-cracks priors).

Figure 9.21: A scatterplot of posterior samples (upper panel) with histograms (lower panel) confirming significant variance reductions of the posterior solutions (line AB in the case of the dry-cracks priors).

Figure 9.22: A scatterplot of posterior samples (upper panel) with histograms (lower panel) confirming no significant variance reductions of the posterior solutions (line AB in the case of the saturated-cracks priors).

Chapter 10

CONCLUSIONS

10.1 Conclusions

The main thrust of the study for developing a method of analyzing a huge refraction dataset as well as applying the method to elucidate the nature of the crust of the Superior and Grenville Province has been accomplished.

1. A simple and reliable tomographic technique using the conjugate gradient method has been successfully developed, tested and applied to analyze the real data.
2. A fast method of estimating uncertainty in the velocity determination has also been developed. This is accomplished through the use of a combination of sampling based technique using the Very Fast Simulated Annealing and ray density approach which utilized a fast ray-tracing technique.
3. A delay-time technique using sliding window has been demonstrated to successfully reveal the variation of the near-surface velocities as well as its topography. The results of the delay-time analysis have also been demonstrated to be nicely correlated with the pattern of the residual gravity anomalies indicating the existence of the near surface density and/or lithology variations.

4. A simple method of displaying the P_mP arrivals using an enhance P_mP composite section has also been demonstrated to be useful in revealing the nature and the topography of the Moho.
5. A special technique of extracting a wealth of information from the three-component dataset has also been developed and applied to reveal the anisotropic nature of the upper crust in the vicinity of the Grenville Front Tectonic Zone.
6. An overall approach of analysis to seismic refraction data based on the above mentioned techniques has been applied successfully (Winardhi and Mereu, 1997) in elucidating the nature of the crust of the Southeastern Canadian Shield. This will be summarized below.

The results of analysis on the 1992 Lithoprobe Abitibi-Grenville Refraction data sets as revealed by the delay times analysis, P_mP studies, tomographic analysis as well as three-component polarization studies, led us to draw several important features regarding with the crustal structure in the vicinity of the study area:

1. The average crustal velocity underneath the Southeastern Canadian Shield varies from 6.13 km/sec at the surface to 7.10 km/s at a depth of 40 km. The average value of Poisson's ratio was 0.245 ± 0.005 .
2. Delay-time analysis of small scale irregularities in the Pg travel-time curves reveals that these may be attributed to low velocity structures in the top 0 to 3 km which correlate well with the known geology. The most prominent of these is the Sudbury Structure which was imaged as a low velocity basinal structure underlain by higher velocity rocks of the Southern and Superior Provinces. This analysis also helps define the underthrusting of the Quetico subprovince beneath the Abitibi Greenstone terrane.

3. Tomographic analysis of the deeper crust shows large lateral and vertical variation in the seismic velocities with the largest values occurring under the uplifted rocks of the Central Metasedimentary belt. This image is consistent with the high metamorphic grade of rocks exhumed from mid to lower crustal depths within this area.
4. The Grenville Front Tectonic Zone is shown as a south dipping region of anomalous velocity gradients down to the Moho which marks the major crustal juxtaposition between the relatively higher velocity rocks of the Grenville Province and the lower velocity rocks of the Superior Province. The image also suggests that the Archean crust may extend laterally to the south as much as 140 km beneath the northern Grenville Province, in contrast to the previously reported result (only 35 km), based on seismic reflection data (Kellet et al., 1994). This interpretation is further supported by the absence of a change in either direction or magnitude of seismic and electrical anisotropies across the Grenville Front (Ji et al., 1996), indicating a possible lateral extension of the Archean upper mantle within the Grenville Province.
5. North of the Front, the images indicate thrusting of the Abitibi Greenstone terrane over the the Pontiac, Quetico and possibly Opatika subprovinces. Such an image is compatible with a plate tectonic scenario involving oblique convergence of sedimentary and volcanic terranes in an accretionary regime.
6. Analysis of the lateral variation in the character of the PmP reflections from the Moho indicate that the Moho generally varies from a sharp discontinuity south of the Grenville Front to a rather diffuse boundary under the Abitibi Greenstone terrane north of the Front. The crust to the south of the Front is 4

to 5 km thicker than that to the north suggesting crustal thickening/stacking during the Grenvillian orogeny. The Moho just south of the Front (beneath the Central Gneiss Belt) is a diffuse boundary, which suggests a rough crust-mantle boundary in which the crust is injected by mantle-derived mafic dikes and sills as opposed to a continuous sequence of mafic/ultramafic intrusions.

7. The subsurface velocity images, as revealed by the tomographic analysis, support collisional tectonic theory to the origin of the Grenville Province and the compatibility of plate tectonic theory of an accretionary regime for the Archean Abitibi Greenstone terrane.
8. Polarization analysis of the three-component array data in the vicinity of the Grenville Front Tectonic Zone has demonstrated the existence of anisotropy in the uppermost crust of the area. In principle, intrinsic anisotropy as well as macroscopic effects from oriented cracks, fractures or layering can cause anisotropy of the size as the observed polarization anomalies (5 – 15 degrees). However, it is evident from the inversion of the angular difference between propagation and polarization, obtained from the three-component dataset, that cracks or fractures aligned in the direction of the gneiss foliation may be the major cause of anisotropy observed. This study has thus demonstrated that the constraint on near-surface anisotropy in the vicinity of the receivers can in fact be obtained using the simple technique described in this thesis.

10.2 Recommendations for future work

It has been demonstrated that the delay-time analysis clearly revealed most of the near surface velocity variations. The fact that seismic traveltimes are highly sen-

sitive to these variations has made this particular analysis an important step to consider. However, experience based on this study has confirmed that, perhaps, it is easier to first subtract these delay-times from the traveltimes data and performing tomographic analysis based on the corrected traveltimes than attempt to model the near surface velocities using tomographic inversion directly. This is especially valid for complex near surface structures such as the Sudbury Impact Structure (lines AB and XY) and/or the juxtaposition between the Abitibi Greenstone Belt and the Quetico terrane (line EF).

The enhanced P_mP composite stacked sections have been demonstrated as a simple but useful method of displaying the character of the P_mP signals and topography of the Moho. The Moho relief obtained by this approach was used to help design the starting model for the tomographic inversion. However, no attempt was made to utilize amplitude information of the P_mP waves. P_mP waveform modelling/inversion is therefore suggested as a potential tool to characterize the nature and roughness of the Moho.

A non-linear, *constrained*, but global inversion for anisotropic parameters based on array of three-component data has been shown to be a useful technique in exploring the anisotropic nature of the uppermost crust. Perhaps, a more fruitful way of analyzing such data is to use an *unconstrained* approach of inversion. This is suggested to explore the many possible causes of anisotropy in the crust. The latter analysis could not be performed in this study due to the incompleteness of azimuthal coverage in the datasets. A combination of refraction and teleseismic studies to reveal the overall anisotropic nature of the crust and upper mantle is therefore suggested as a recommendation for future work.

Appendix A

3-D formulation of the polarization angle

Consider a dipping anisotropic medium with dip angle δ . Choose x-axis such that the dipping plane is parallel to the x-axis. The direction cosines of the plane's normal $\hat{N}$ can then be expressed as

$$\hat{A} = 0\hat{i} + \sin \delta \hat{j} + \cos \delta \hat{k} \quad (\text{A.1})$$

Consider now, a ray $\mathbf{R}$ approaching the above plane. The direction cosines of this ray can then be written as

$$\hat{R} = \sin \xi \cos \phi \hat{i} + \sin \xi \sin \phi \hat{j} + \cos \xi \hat{k} \quad (\text{A.2})$$

Consider further that the ray strikes the dipping plane at an angle θ with respect to the plane's normal $\hat{N}$. Since $\hat{R}$ and $\hat{N}$ lie on plane, relations between the direction cosines of both quantities exist

$$\hat{R} \cdot \hat{N} = \left| \hat{R} \right| \left| \hat{N} \right| \cos \theta \quad (\text{A.3})$$

and therefore

$$\cos \theta = \sin \xi \sin \phi \sin \delta + \cos \xi \cos \delta \quad (\text{A.4})$$

The magnitude of the polarization angle γ with respect to the normal can then be computed from θ using equation (9.25) stated previously in chapter 9.

$$\tan 2\gamma_{P,SV} = \frac{\eta \sin 2\theta}{1 - (1 + \tau) \sin^2 \theta} \quad (\text{A.5})$$

where η and τ are the two "de Parscau" anisotropic parameters described previously in chapter 9. An expression for the polarization vector still however need to be formulated . This will be proceeded as follows. Let $\hat{P}$ be the unit vector of the polarization angle

$$\hat{P} = l\hat{i} + m\hat{j} + n\hat{k} \quad (\text{A.6})$$

and let $\hat{B}$ be the unit vector which is normal to the plane of $\hat{R}$, $\hat{P}$, and $\hat{N}$

$$\hat{B} = \alpha\hat{i} + \beta\hat{j} + \zeta\hat{k} \quad (\text{A.7})$$

Since $\hat{B}$ is normal to $\hat{N}$ and $\hat{R}$, we then have

$$\hat{B} = \frac{\hat{N} \times \hat{R}}{|\hat{N} \times \hat{R}|} \quad (\text{A.8})$$

with a simple trigonometrical manipulation, α , β and ζ can then be further expressed as

$$\alpha = \frac{\tan \delta \cot \xi - \sin \phi}{G} \quad (\text{A.9})$$

$$\beta = \frac{\cos \phi}{G} \quad (\text{A.10})$$

$$\zeta = \frac{-\tan \delta \cos \phi}{G} \quad (\text{A.11})$$

where

$$G = \sqrt{1 + \tan^2 \delta \cot^2 \xi + \tan^2 \delta \cos^2 \phi - 2 \tan \delta \cot \xi \sin \phi} \quad (\text{A.12})$$

Since $\hat{P} \perp \hat{B}$, we then have

$$l\alpha + m\beta + n\zeta = 0 \quad (\text{A.13})$$

The polarization angle, which is the angle between $\hat{P}$ and $\hat{N}$ can then be found

using $\hat{P} \cdot \hat{N} = |\hat{P}| |\hat{N}| \cos \gamma$ which will then give

$$\cos \gamma = m \sin \delta + n \cos \delta \quad (\text{A.14})$$

Since $\hat{B}$ is parallel to $\hat{N} \times \hat{P}$, we can further express α as

$$\alpha = -\frac{m \cos \delta - n \sin \delta}{\sin \gamma} \quad (\text{A.15})$$

From equations (A.13), (A.14) and (A.15) we can then compute l , m and n as

$$n = \cos \gamma \cos \delta + \alpha \sin \gamma \sin \delta \quad (\text{A.16})$$

$$m = \frac{\cos \gamma - n \cos \delta}{\sin \delta} \quad (\text{A.17})$$

$$l = \frac{-(m\beta + n\zeta)}{\alpha} \quad (\text{A.18})$$

The vector of the polarization angle in terms of incidence angle and azimuth can then be simply computed from

$$\theta_a = \cos^{-1}(n) \quad (\text{A.19})$$

$$\phi_a = \tan^{-1}(m/l) \quad (\text{A.20})$$

It has thus been shown that the presence of the dipping anisotropic medium perturbs not only the incidence angle but also the azimuth of the polarization vector. This 3-D formulation is therefore needed if one wished to invert the angular difference between the polarization and the propagation vector to seek *not only* for η and τ *but also* for the dip and strike of the anisotropic medium.

Appendix B

Computer Programs

Several main computer programs used in this study will be included in a diskette in order to clarify the implementation of all the techniques described. Most of them were written in FORTRAN programming language but some (i.e., the postprocessing routines) were written in MATLAB. The latter is chosen due to its capability in creating high quality graphics in almost any format.

The codes listed here could be ported in various environments without major problem, since compliancy with ANSI FORTRAN-77 was maintained. However, some graphics may not work properly due to the need of specific graphic library. Most of data processing for the work of this study was performed on a Sun Sparc WorkStation running Solaris 5.1. The code has also been succesfully ported on a 486-33 PC using NDP-FORTRAN 386 Compiler and on a Pentium PC running LINUX. The latter is implemented by the virtue of a free GNU FORTRAN-77 Compiler, called g77.

The programs were grouped according to three headings:

- Tomographic inversion and delay-time analysis
- Three-component analysis and
- additional programs

B.1 Tomographic inversion and delay-time analysis

B.1.1 Tomographic Inversion using CGLS and SVD

The code for the tomographic inversion and uncertainty analysis were based on the main ray-tracing forward-modelling routine developed by Prof. R.F. Mereu. This code was then modified for this study to allow one to do tomographic inversion using conjugate gradient and/or singular value decomposition techniques. The name of this routine is TTCGSVD.F which stands for Travel Time Tomography using Conjugate Gradient and Singular Value Decomposition. This code has to be compiled and linked with several object files to use it. The object files could be requested from Dr. R.F. Mereu.

B.1.2 Delay-time analysis

The delay-time analysis was performed utilizing the routine called TTERM.F

B.2 Three-component analysis

Extraction of the polarization vector was performed using a routine called AGCOV3.F, whereas of the propagation vector was performed using a routine called XBMF.F. The 3-D Forward modelling of polarization anomalies due to an arbitrarily oriented transversely isotropic media was implemented through routine ANIFOR5.F, whereas the 3-D Inversion of parameters describing a transversely isotropic medium was performed using routine VFSACG.F

B.3 Additional Programs

Two main MATLAB routines were included here to clarify the usage of a lower-hemisphere diagram for presenting the polarization anomalies. Routine MPOLX.M was used to extract and plot the observed polarization anomaly data, while ANIVPLOT.M was used to plot the forward model resulting from routine ANIFOR5.F

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