

EVOLUTION AND DYNAMICS OF THE LITHOSPHERE IN NORTHWESTERN CANADA

CLÉMENT ESTÈVE

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SUMMARY

The lithosphere of northwestern Canada recorded more than 2.5 Gy of complex tectonic evolution, from the formation of the ancient cores of the continental lithosphere such as the Slave craton to the Phanerozoic Cordilleran orogeny with substantial variations in crust and upper mantle structures that led to the concentration of natural resources (i.e., diamonds in cratons). Present-day northwestern Canada juxtaposes a thin and hot Cordilleran lithosphere to the thick and cold cratonic lithosphere, which has important implications for regional geodynamics. Recently, seismic station coverage has drastically increased across northwestern Canada, allowing the development of seismic tomography models and other passive-source seismic methods at high resolution in order to investigate the tectonic evolution and dynamics of the lithosphere in this region. The P- and S-wave upper mantle structures of northwestern Canada reveal that the distribution of kimberlite fields in the Slave craton correlates with the margin of fast and slow seismic mantle anomalies, which could delineate weak zones in the lithosphere. Based on our tomographic models we identify two high-velocity seismic anomalies straddling the arcuate Cordillera Deformation Front that have controlled its regional deformation, including a newly identified Mackenzie craton characterized by high seismic velocities extending from the lower crust to the upper mantle to the north of the Mackenzie Mountains. Furthermore, our P-wave tomography model shows sharp velocity contrasts beneath the surface trace of the Tintina Fault. Estimates of seismic anisotropy show a progressive rotation of fast-axis directions when approaching the fault zone. Together, they provide seismic evidence for the trans-lithospheric nature of the Tintina Fault. We further propose that the Tintina Fault has chiseled off small pieces of the Laurentian craton between the Late Cretaceous and the Eocene, which would imply that large lithospheric-scale shear zones are able to cut through small pieces of refractory cratonic mantle and transport them over several hundred kilometers.

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LIST OF SYMBOLS

Following is a list of symbols used in this thesis. Their meanings (and typical units if appropriate) are indicated. If a symbol is used multiple times, its different meanings are separated by a semicolon.

Symbol	Description (Units)
C	Covariance matrix
d	data vector
E	Event matrix
F	Regularization matrix
G	Model operator
H	Station matrix
m	Model vector
W	Matrix containing the reciprocal of the travel time standard deviations determined from the MCCC analysis
t_0	Predicted arrival time based on IASP91 seismic reference model
$\bar{t}_{net}$	Predicted average phase arrival time for all stations recording the same earthquake
t_1	Calculated phase arrival time after running the ICCS algorithm
t_2	Input phase arrival time to the MCCC
t_3, t_i^{est}	Final phase arrival time for the i-th trace after running the MCCC
Δ	Epicentral Distance ($^{\circ}$) from a seismic station to an event
δt	Relative P- or S-wave delay time; accumulated delay time between the fast and slow shear waves (s)
ϕ	Fast direction of wave propagation ($^{\circ}$)
ε	Data errors
δV_P	P-wave velocity anomaly (%)

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Symbol	Description (units)
δV_S	S-wave velocity anomaly (%)
λ	Weight of either damping, flattening or smoothing
U	Group Velocity (km/s)

LIST OF ACRONYMS AND ABBREVIATIONS

Following is a list of acronyms and abbreviations used in this thesis. Their definition and explanations are indicated, as well as references where required.

Acronym	Description (Units)
AIMBAT	Automated and Interactive Measurement of Body-wave Arrival Times
AK135	1D seismic reference model of Kennett et al. (1995)
CANOE	Canadian Northwest Seismic Experiment
CHIS	Canadian Hazard Information Service
CMB	Core Mantle Boundary
CNSN	Canadian National Seismograph Network
CS	Canadian Shield
CSMC	Central Slave Mantle Conductor
DF	Denali Fault
GSLSZ	Great Slave Lake Shear Zone
Gy	Billion years
LAB	Lithosphere-Asthenosphere Boundary
IASP91	1D seismic reference model of Kennett & Engdahl (1991)
ICCS	Iterative Cross-Correlation and Stack
IM	the International Miscellaneous Stations
IRIS	Incorporated Research Institutions for Seismology
LPO	Lattice-preferred orientation
LTZ	Liard Transfer Zone
Ma	Million years ago
MCCC	Multi-Channel Cross-Correlation
MRA	Mackenzie River magnetic Anomaly

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Acronym	Description (units)
My	Million years
NCC	Northern Canadian Cordillera
POLARIS	Portable Observatories for Lithospheric Analysis and Research Investigating Seismicity
PPD	Posterior Probability Density
rjMCMC	reversible-jump Markov Chain Monte Carlo
RMS	Root mean square
SCLM	Sub-Continental Lithospheric Mantle
SNORCLE	Slave-Northern Cordillera Lithospheric Evolution
SNR	Signal-to-Noise Ratio
SPO	Shape-preferred orientation
STEEP	St-Elias Erosion/Tectonics Project
TA	Earthscope-USArray Transportable Array
TF	Tintina Fault
YNSN	Yukon-Northwest Seismic Network
YO	Yukon Observatory

PREFACE

Some of the chapters in this thesis have already been published or submitted for publication and written in collaboration with *Pascal Audet and Jeremy Gosselin at University of Ottawa, Andrew J. Schaeffer at Geological Survey of Canada, Pacific Division, Derek Schutt and Richard Aster at Colorado State University and Joel Cubley at Centre for Northern Innovation in Mining, Yukon College*. The content of these articles was unaltered to conform with Journal copyrights.

The results presented were generated using programs written mostly in Fortran and Python, as well as shell-scripts. All Fortran programs were compiled using the open source GNU Fortran compiler *gfortran*, running on Ubuntu and Centos Linux distributions. The figures were produced almost entirely using Generic Mapping Tools (GMT; [Wessel et al., 2013](#)) and Adobe Illustrator.

DECLARATION

I hereby certify that the submitted work is my own work, was completed while registered as a candidate for the degree of Doctor of Philosophy, and I have not obtained a degree elsewhere on the basis of the research presented in this submitted work.

Clément Estève
Ottawa, Ontario, Canada,
June 2020

STATEMENT OF CO-AUTHORSHIP

Chapter 3 was published with *Pascal Audet* at *University of Ottawa* and *Andrew J. Schaeffer* at *Geological Survey of Canada, Pacific Division*. I obtained and processed the data, generated the models, performed the analysis, and made the interpretation. I wrote the manuscript with input from *Pascal Audet* and *Andrew J. Schaeffer*.

Chapter 4 was published with *Pascal Audet* at *University of Ottawa*, *Andrew J. Schaeffer* at *Geological Survey of Canada, Pacific Division*, *Derek Schutt* and *Richard Aster* at *Colorado State University* and *Joel Cubley* at *Centre for Northern Innovation in Mining, Yukon College*. I obtained and processed the data, generated the models, performed the analysis, and made the interpretation. I wrote the manuscript with input from *Pascal Audet*, *Andrew J. Schaeffer*, *Derek Schutt*, *Richard Aster* and *Joel Cubley*.

Chapter 5 has been accepted for publication with *Pascal Audet* at *University of Ottawa*, *Andrew J. Schaeffer* at *Geological Survey of Canada, Pacific Division*, *Derek Schutt* and *Richard Aster* at *Colorado State University* and *Joel Cubley* at *Centre for Northern Innovation in Mining, Yukon College*. I obtained and processed the data, generated the models, performed the analysis, and made the interpretation. I wrote the manuscript with input from *Pascal Audet*, *Andrew J. Schaeffer*, *Derek Schutt*, *Richard Aster* and *Joel Cubley*.

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I would like to thank Andrew Schaeffer who has been (and still is) a fantastic mentor and friend since I started my PhD. I will never forget this road trip through the Rockies and the first landings on Banks Island.

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Je me permets d'emprunter ma langue maternelle afin de remercier toutes les personnes qui m'ont soutenu depuis la France et Ottawa. Tout d'abord, mes premières pensées vont à Manon, ma femme, qui n'a jamais cessé de me soutenir dans ce projet que j'entreprenais, par delà l'Océan Atlantique, loin de nos familles respectives et sans qui tout cela n'aurait jamais été possible. Ensuite, Maël, mon fils, qui a su illuminer mes journées avec sa joie dans les moments plus compliqués. Mes parents, Brigitte et André pour leur inexorable soutien et amour dans ma vie de tous les jours. Ma soeur, Justine, et mon frère, Damien, qui ont toujours été présents pour leur petit frère. Bien évidemment, ma famille de l'Hérault Benoît, Magali, Judith et Elisa pour le soutien qu'ils ont su me fournir tout au long de ce doctorat. Enfin, je remercie tous mes amis Hugo, Charlotte, Etienne, Nico, Marion, Virgile, Tristan, Romain, Laurent, et tous les autres.

Je dédie ce travail à mon fils, Maël.

Part I

INTRODUCTION AND THEORY

INTRODUCTION

Since its formation, the evolution of Earth has been characterized by the cyclic formation and break-up of continents due to the motion of tectonic plates, which has resulted in the juxtaposition of geologically young and tectonically active cordilleras (i.e., wide coastal mountain ranges) with old and stable continental interiors called cratons. One such example is located in northwestern Canada, where the northern Canadian Cordillera (NCC) comprises several exotic terranes that accreted to the passive margin of Laurentia. In this region, the lithosphere records ~750 My of Earth evolution and deformation processes, with substantial variations in crust and upper mantle structures. Understanding the complex tectonic evolution and the associated deformation processes in such tectonic settings is the focus of this thesis.

Archean cratons such as the Slave craton form the ancient cores of Earth's continental lithosphere and were preserved for billions of years sustaining entire Wilson cycles (e.g., [Currie & van Wijk, 2016](#)). They are characterized by a cold and thick lithospheric root and are chemically depleted in Fe-rich materials ([Jordan, 1978](#)). Some cratons are juxtaposed against Phanerozoic lithospheres such as those in Africa, South America, western Europe and western North America. Their longevity and stability have been attributed to the fact that Phanerozoic mobile belts are weaker than Archean cratons, and therefore are the preferred location for continental ruptures during supercontinent cycles ([Audet & Bürgmann, 2011](#)). Nonetheless, the timing and the mechanisms involved in their formation is still under debate. Proposed models involve mantle plumes (e.g., [Griffin et al., 2003](#); [Arndt et al., 2009](#)), buoyant subduction and imbrication by lithospheric-scale stacks (e.g., [Helmstaedt & Schulze, 1989](#); [Pearson & Wittig, 2008](#); [Snyder et al., 2015](#)) and successive accretion-collision episodes (e.g., [Jordan, 1988](#); [Gray & Pysklywec, 2010](#)) ([Figure 1.1](#)).

The western Laurentian cratonic margin has experienced several episodes of deformation such as rifting in the Neoproterozoic ([Lund, 2008](#)), subduction ([Johnston, 2008](#); [Hildebrand, 2009](#)), and the Cordillera orogeny ([Nelson & Colpron, 2007](#)). All of these deformation episodes may have important implications for regional geodynamics.

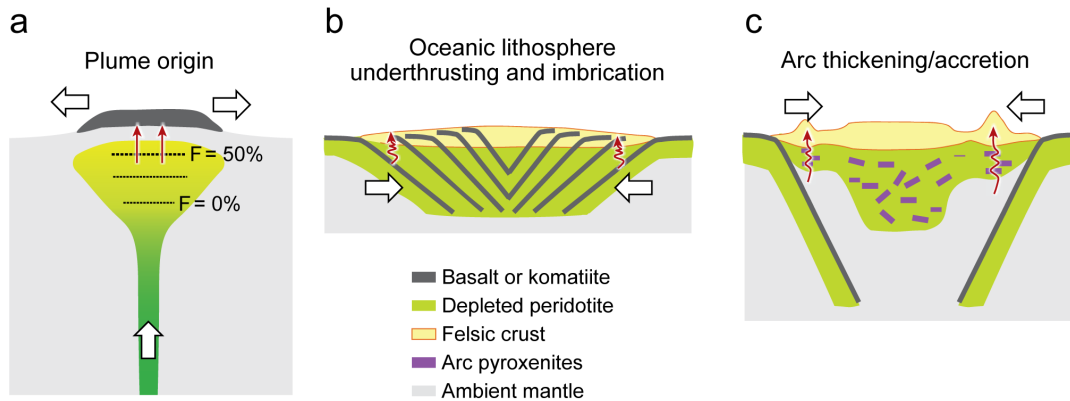


Figure 1.1.: Proposed mechanisms for the formation and growth of cratonic mantle: (a) plume origin, (b) oceanic lithosphere underthrusting and imbrication and (c) arc thickening/accretion. F average degree of melting (from Lee et al., 2011).

The NCC of northwestern Canada is an accretionary orogen consisting of a succession of terranes accreted to the Laurentian passive margin (Figure 1.2). The present-day Canadian Cordillera is characterized by a thin and hot lithosphere (e.g., Audet et al., 2019; Hyndman, 2017). Seismic tomography studies spanning various portions of the Canadian Cordillera show an abrupt change in seismic velocities at uppermost mantle depths between the Cordillera and the adjacent cratonic lithosphere (e.g., Mercier et al., 2009; Bao et al., 2014; Schaeffer & Lebedev, 2014; Chen et al., 2018). Despite decades of research, it remains unclear whether or not the lithospheric mantle roots of overlying terranes were preserved during the accretion to the continental margin. Based on results from a number of geophysical methods, it has been suggested that North American basement rocks in the lower crust extend hundreds of kilometers west of the Tintina Fault (Figure 1.2). The accreted tectonostratigraphic terranes are thought to represent thrust sheets separated from the original lower crust and mantle (e.g., Snyder et al., 2002; Cook et al., 2004; Cook & Erdmer, 2005; Hayward, 2015). An Osmium (Os) model age for the NCC mantle lithosphere derived from mantle xenoliths, 1.12 ± 0.26 Ga (Peslier et al., 2000), is younger than the age of the overlying basement and tectonostratigraphic terranes, suggesting no direct genetic link between the crust and lithospheric mantle (Peslier et al., 2000; Snyder et al., 2002; Cook et al., 2004).

In contrast to the proposed single age of lithospheric mantle formation by Peslier et al. (2000), Abraham et al. (2001) identify differences in isotopic data from alkaline basalts in the Insular and Intermontane superterrane and North American margin rocks from southern Yukon and northern British Columbia. The interpretation of these differences is that lithospheric mantle roots for each zone were preserved during accretion to the continental margin.

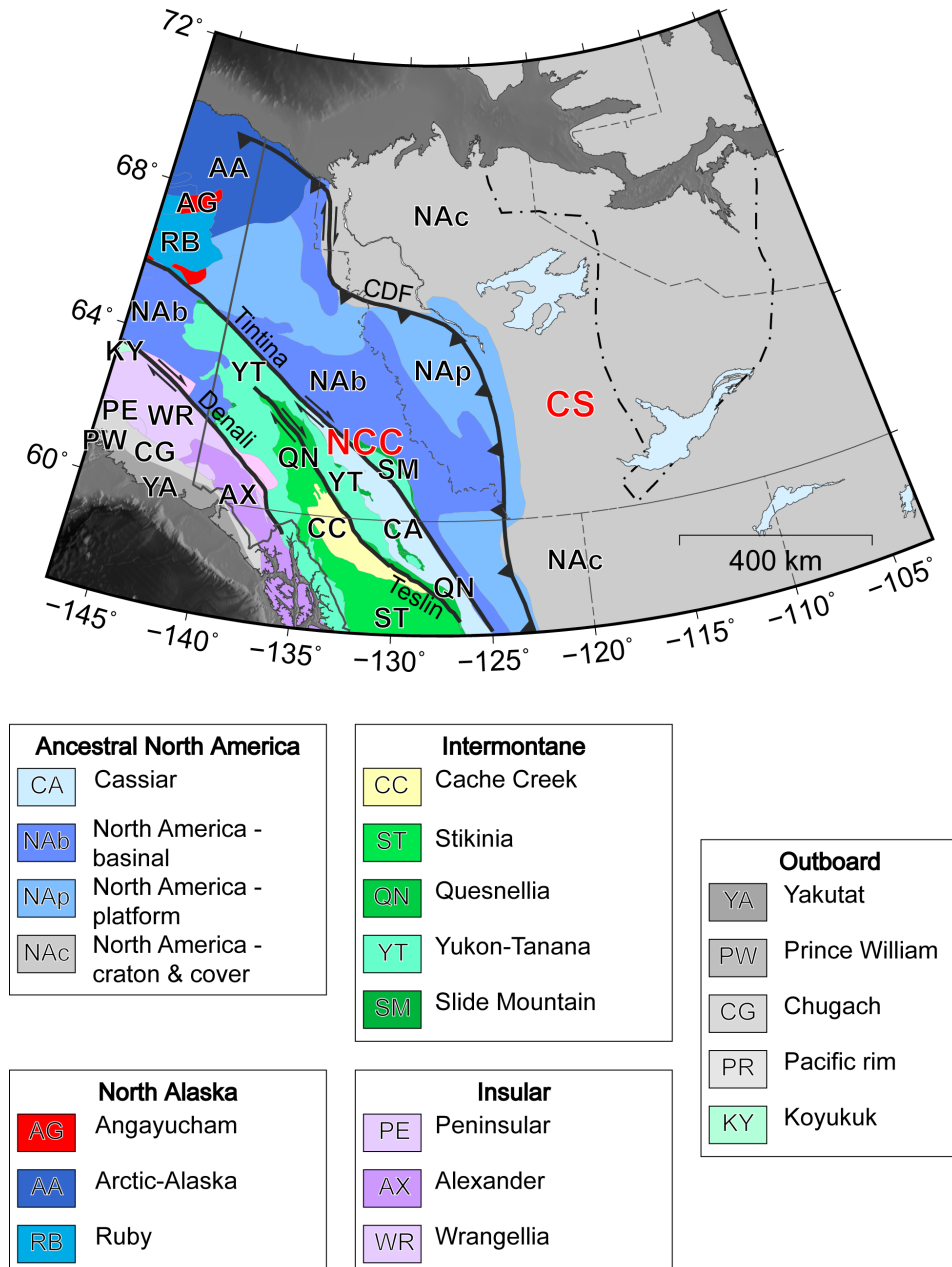


Figure 1.2.: Surface terrane map of the NCC modified from Nelson & Colpron (2007). Also shown are the main tectonic structures (black lines) and the outline of the Slave craton (dash-dotted line). Abbreviations: CDF: Cordillera Deformation Front; CS: Canadian Shield; NCC: Northern Canadian Cordillera.

Based on interpretations of active source seismic profiles acquired during the SNORCLE-Lithoprobe experiment, Cook & Erdmer (2005) propose that underlying lithospheric mantle of North American affinity was preserved during the accretion of overlying terranes in the Mesozoic, which is in agreement with magnetic anomaly models (Pilkington et al., 2006).

The accretionary model of Johnston (2008) and Hildebrand (2009) invokes west-dipping subduction in the Cretaceous – Early Cenozoic and collision of North America with a ribbon continent. The proposed ribbon continent includes the Intermontane and Insular superterrane (i.e., no pre-accretion of the Intermontane superterrane), with the implication that all Cordilleran lithosphere is exotic west of a suture along the present-day Tintina Fault. This would imply that the Tintina Fault marks a major geological boundary and demarcates fundamentally different mantle composition across it, consistent with variations in isotopic compositions of erupted alkaline basalts (Abraham et al., 2001).

The orogenic float model (Oldow et al., 1990), which stipulates that the high elevations, in the NCC, are supported by positive buoyancy in the upper mantle due to thermal expansion of hot mantle rocks, has been invoked in the NCC to explain thrust faulting in the Mackenzie Mountains and stress transfer to the Cordillera Deformation Front (CDF) (Mazzotti & Hyndman, 2002), 1000 km from the present-day plate, through a detachment within the deep crust that displaces the entire crust from the underlying lithosphere. Thermal and rheological properties of the Cordilleran lithosphere suggest that the lower crust is mechanically weak, allowing an eastward displacement of the upper crust above a lower crustal detachment zone (Mazzotti & Hyndman, 2002; Hyndman, 2017) and triggering shortening and thrust faulting at the boundary between that mobile belt and the North American craton. However, these models were devised at a time when detailed constraints on upper mantle structure were extremely sparse. The constraints were derived from heat flow measurements, low-resolution surface-wave models, sparse active source profiles, and potential field data. Imaging techniques that probe the upper mantle at high resolution require dense seismic station coverage.

During the past 25 years, northwestern Canada has witnessed a dramatic increase of broadband seismic station deployments (Figure 1.3). Recent seismic studies in the Southern Canadian Cordillera have revealed a complex tectonic evolution of the cratonic margin (Bao et al., 2014; Chen et al., 2018) but little has been done with the new available data set in the NCC. This data set allows the development of tomographic models, which illuminate the 3-dimensional seismic velocity structure of northwestern Canada from the crust to the upper mantle at a higher resolution compared to previous studies. The combined insights from this thesis provide a framework for a better understanding of the tectonic evolution of the western Canadian cratonic margin juxtaposed to the Canadian Cordillera, which concentrates the ongoing deformation.

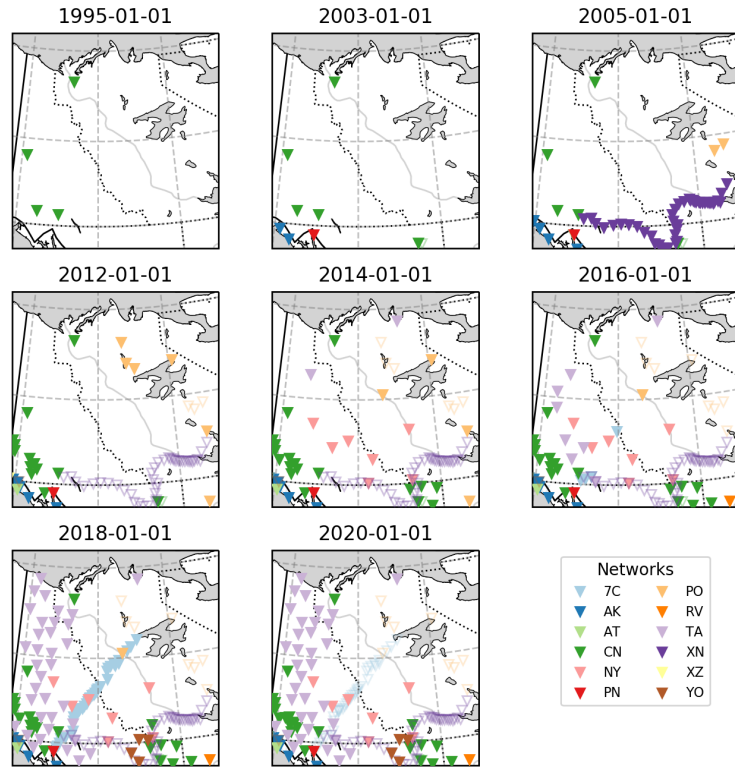


Figure 1.3.: Seismic station coverage in northwestern Canada between 1995 and 2020. Seismic stations (inverted triangles) are color coded by seismic networks. Empty triangles represent temporary deployments that are no longer active (Audet et al., 2020).

1.1 SCOPE OF THE THESIS

In this thesis, we develop seismic tomography models for the crust and the upper mantle underlying northwestern Canada. We also provide constraints on the upper mantle dynamics from estimates of seismic anisotropy. We summarize the fundamental questions that we address in this thesis following the chapter order.

- Can we attribute the propensity of diamondiferous kimberlites to a particular seismic upper mantle structure? Does the Slave craton have a similar upper mantle seismic structure compared to the other Archean cratons ?

- What is the nature and the origin of the lithospheric mantle underlying the NCC and past and present geodynamics of the region ? Does the upper mantle structure have implications to surficial geological features and the ongoing deformation throughout the NCC?
- What is the mantle seismic expression of the main fault zones across the NCC? How was the craton margin modified during the Cordillera orogeny ? What is the depth penetration of the major transcurrent fault and their implications for the tectonic evolution of the NCC ?
- Do we have evidence of the westward extent of the North American craton in the NCC at crustal depths ? What is the pattern of crustal fabric in relation with the major strike-slip faults ?

1.2 OUTLINE

In [Chapter 2](#), we introduce the concepts of teleseismic (i.e., far distant) body-wave travel time tomography, Bayesian (i.e., probabilistic) surface-wave tomography and seismic anisotropy. Body-wave travel time data are inverted to retrieve the 3-dimensional seismic velocity structure beneath the Slave craton ([Chapter 3](#)) and northwestern Canada ([Chapter 4](#)) using the method of [VanDecar \(1991\)](#). [Chapter 5](#) presents a seismic anisotropy analysis conducted across northwestern Canada. The results are used in conjunction with our teleseismic P-wave tomography model over northwestern Canada. In [Chapter 6](#), we invert surface wave group velocity dispersion measurements obtained from recordings of regional earthquakes within a Bayesian framework in order to estimate group velocity maps across northwestern Canada using the method of [Gosselin et al. \(2020\)](#). In [Chapter 7](#), we summarize the main contributions of this thesis and make some brief concluding remarks.

SEISMIC TOMOGRAPHY AND SEISMIC ANISOTROPY: THEORY

Tomography literally means slice picture (derived from Ancient Greek *tomos* meaning slice). This imaging technique is widely applied in various fields such as medicine, astrophysics, and Earth sciences. Seismic tomography is one of the most useful geophysical tools to map seismic parameters (velocity, attenuation, anisotropy, density) and derive the thermal state to constrain the structure and dynamics of the Earth's interior from the crust to the core. There exist a wide range of tomographic methods applied at different scales (local, regional or global), different types of seismic waves (body and surface waves), different types of data sets (derived data such as travel time or full waveform), different inversion methods (regularized linear approaches versus probabilistic tomography). The first 3-dimensional inversion of seismic body-wave (P-wave) travel times was successfully performed by [Aki et al. \(1977\)](#) in southern Norway and since then is known as the ACH method. This method was then applied worldwide in a variety of tectonic settings using either local or teleseismic earthquakes (i.e., far distant earthquakes) ([Aki, 1982](#)). Nowadays, the increasing number of operating seismic stations worldwide (e.g., Earthscope-Transportable Array, AlpArray, ChinArray) and, therefore, the increasing volume of seismic data, have motivated the development of new tomographic inversion techniques, which are growing ever more capable of producing high resolution Earth models that capture detailed structure and dynamics across a range of scale lengths.

In this chapter, we give a description of regularized teleseismic body-wave tomography, Bayesian surface-wave tomography and seismic anisotropy.

2.1 TELESEISMIC BODY-WAVE TOMOGRAPHY

In teleseismic tomography, earthquakes (sources), located outside the volume of interest, produce seismic waves, which propagate through the Earth's interior and are recorded at the surface by seismic stations (receivers) ([Figure 2.1](#)). The travel times are defined by

the station geometry and the distribution of seismic velocity anomalies along the ray path. Seismic parameters (i.e., seismic velocities) of the medium can be estimated through the formulation of an inverse problem (Equation 2.1) using travel times and assuming linearity (Aster et al., 2018),

$$\mathbf{d} + \varepsilon = \mathbf{G}\mathbf{m} \quad (2.1)$$

where $\mathbf{d}$ is the vector of observations (i.e., travel time data), ε represents errors in the observations, which can originate from measurement inaccuracies or deficiencies/assumptions in the underlying physics (i.e., theory errors), $\mathbf{G}$ is the model operator (i.e., matrix of partial derivatives) and $\mathbf{m}$ is the vector of unknown model parameters (i.e., slowness perturbations).

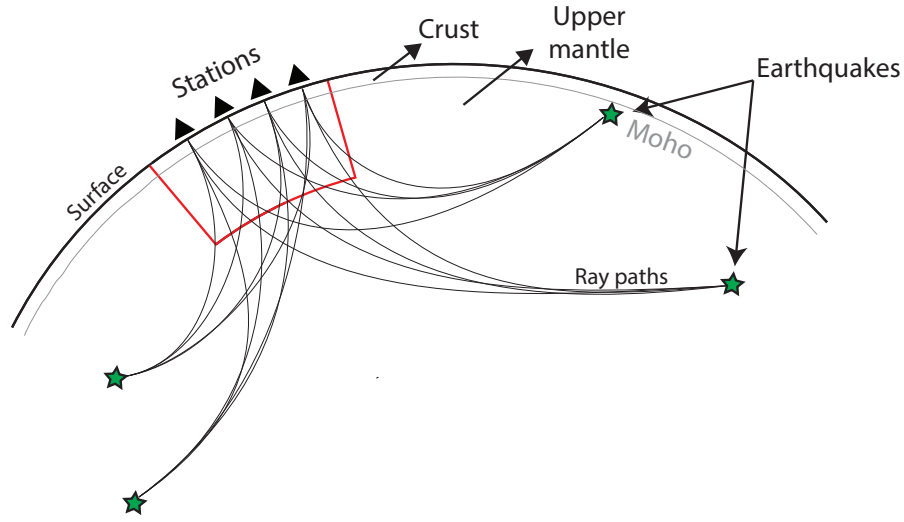


Figure 2.1.: Schematic diagram showing the principle of teleseismic tomography.

2.1.1 Setting up the tomography problem

The travel time, t_{AB} , of a seismic wave traveling from a point A to a point B through the Earth is given by

$$t_{AB} = \int_{L_0} \frac{1}{V_0(r)} dl \quad (2.2)$$

where L_0 is the ray path between A and B, dl is an element of path length, $V_0(r)$ is the intrinsic seismic wave velocity and r is a vector of spatial coordinates (r, ϕ, θ) . Ray paths are assumed to remain perpendicular to wavefronts in an isotropic medium (VanDecar, 1991). Equation 2.2 describes a non-linear problem due to the fact that the path of integration depends on the integral itself. Fermat's principle states that the travel time remains stationary under a small perturbation to the ray path, which suggests that, to first order, travel times are insensitive to small perturbations in the ray paths. We define the slowness $s(r)$, as the reciprocal of $V(r)$:

$$s(r) = \frac{1}{V(r)} \quad (2.3)$$

Let's assume that the slowness along the path is perturbed by an amount $\Delta s(r)$ small enough that the ray path remains unchanged, by considering the slowness $s + \Delta s$ (close to s) and the resulting travel time $t + \Delta t$ (close to t), one can substitute the perturbed values into Equation 2.2 without changing the path of integration L_0

$$t + \Delta t = \int_{L_0} (s(r) + \Delta s(r)) dl \quad (2.4)$$

Subtracting Equation 2.2 from Equation 2.4 gives the first order linear relation between travel time perturbations and slowness perturbations,

$$\Delta t = \int_{L_0} \Delta s(r) dl \quad (2.5)$$

Therefore, the slowness along a known ray path is defined as the sum of slowness perturbations along the path. In order to formulate Equation 2.5 in a matrix form, we must choose a suitable form to represent the slowness field, Δs , as a set of discrete parameters.

2.1.2 Parameterization

VanDecar (1991) represents the slowness field as splines under tension constrained at a series of regular knots. We describe the parameterizations we used for the models over the Slave craton and its surroundings in Section 3.4.4 and over northwestern Canada in Section 4.3.3. Figure B.1 shows the locations at which the splines are pinned in the region beneath northwestern Canada, where intersecting lines indicate node positions. The parameterization grid extends beyond the volume of interest, outlined by the red box in Figure B.1. Increasing

the size of the model space is designed to avoid mapping spurious structures from the outer region into the inner region of the model (VanDecar, 1991). Although this approach may remove some real structure from the inner region, the philosophy is to invert for the simplest model required by the data.

Furthermore, smooth structure in such a large region is required to perform 3-dimensional ray tracing (VanDecar, 1991). The outside knots are forced to represent zero anomaly such that the 3-dimensional model can merge smoothly into the radial Earth velocity model (based on IASP91; Kennett et al., 1995). This is achieved by heavily damping the perturbations at outside knots in the inversion procedure.

2.1.3 Inverse problem formulation and regularization

Relative delay times are inverted simultaneously for slowness perturbations, station static terms and source terms. Station static terms compensate for local shallow structure and avoid projecting crustal short wavelength anomalies into the upper mantle. Source terms account for first order uncertainties in event origin time and location, and large-scale velocity heterogeneities in the source region and lower mantle (VanDecar, 1991). The formulation of the unweighted travel time equation for the i th ray is, therefore, given by VanDecar (1991) as,

$$P_{ij}^{k-1} \Delta s_j^{(k)} + h_r^{(k)} + e_q^{(k)} = \Delta t_i^{(k-1)}, \quad (2.6)$$

where $P_{ij}^{k-1} = [\frac{\delta t_i}{\delta s_j}]_{s=s^{(k-1)}}$, $\Delta s_j^{(k)}$ is the perturbation to slowness $s^{(k-1)}$ at node j , $h_r^{(k)}$ is the time correction for station r at iteration k , $e_q^{(k)}$ is the time correction for event q at iteration k , and $\Delta t_i^{(k-1)}$ is the travel time residual for i^{th} ray relative to iteration $(k-1)$ of the model. The inclusion of the regularization term $\mathbf{F}$ in Equation 2.7 imposes additional constraints on the problem and creates a system of rank n (number of rays) allowing the computation of a least-squares solution.

The weighted system of linear equations at iteration k can now be written (VanDecar, 1991) as

$$\begin{pmatrix} \mathbf{WP} & \mathbf{WH} & \mathbf{WE} \\ \lambda \mathbf{F} & 0 & 0 \end{pmatrix} \begin{pmatrix} \Delta s^{(k)} \\ h^{(k)} \\ e^{(k)} \end{pmatrix} = \begin{pmatrix} \mathbf{W} \Delta t^{(k-1)} \\ \lambda \mathbf{F} \left\{ \sum_{i=0}^{k-1} \Delta s^{(i)} \right\} \end{pmatrix} \quad (2.7)$$

where,

$$\begin{aligned}
 W_{ij} &= \begin{cases} \frac{1}{\sigma_i^{res}}, & \text{if } i = j \\ 0, & \text{if } i \neq j \end{cases} \quad n \times n \text{ matrix} \\
 H_{ir} &= \begin{cases} 1, & \text{if ray } i \text{ was recorded at station } r \\ 0, & \text{otherwise} \end{cases} \quad n \times (\# \text{ station}) \text{ matrix} \\
 E_{iq} &= \begin{cases} 1, & \text{if ray } i \text{ originated from earthquake } q \\ 0, & \text{otherwise} \end{cases} \quad n \times (\# \text{ event}) \text{ matrix} \\
 F_{ij} &= \begin{cases} \delta_{ij}, & \text{for simple damping} \\ filter_{ij}, & \text{otherwise} \end{cases} \quad m \times m \text{ matrix}
 \end{aligned}$$

where the variable λ can represent the weight of damping ($\lambda = \lambda_{damping}$), flattening ($\lambda = \lambda_{flattening}$), or smoothing ($\lambda = \lambda_{smoothing}$); n = number of rays, and m = number of model nodes. Zero terms on the left-hand side of Equation 2.7 represent null matrices of appropriate size. The term σ_i^{res} in the weighting matrix W_{ij} is the error of the cross-correlation derived delay times (see Equation 4.2 in Section 4.3.2). The regularization term (**F**) may be defined as the sum of damping, smoothing and flattening.

The inversion problem posed by Equation 2.6 and Equation 2.7 is under-determined, which means there are more unknowns (model parameters) than equations (ray paths) for both tomography models presented in Chapter 3 and Chapter 4. For example, our body-wave tomography models over northwestern Canada have 211,356 slowness perturbation unknowns, 127,138 relative P-wave delay times and 59,107 relative S-wave delay times. Here, the matrix **G** is not full rank (Aster et al., 2018) and, therefore, this inverse problem requires some form of regularization to obtain a solution.

The regularization applied by VanDecar (1991) minimizes a 7-point finite element approximation to the Laplacian operator ($\nabla^2 s$) in order to penalize the roughness (second derivative) of the final slowness model and produce a smooth model that fits the data. The Laplacian operator minimizes the roughness of the final model by setting the right-hand side of Equation 2.7 equal to the negative roughness of the previous model's perturbations.

The constraint imposed in Equation 2.7 for regularizing the final total model at iteration k is obtained by setting the roughness of the model (weighted by λ) equal to zero. We therefore have (VanDecar, 1991)

$$\lambda \mathbf{F} s^{(k)} = 0 \quad (2.8)$$

or, since $s^{(k)} = \Delta s^{(k)} + s^{(k-1)}$,

$$\lambda \mathbf{F} (\Delta s^{(k)} + s^{(k-1)}) = 0 \quad (2.9)$$

Using the distributive property of the linear operator and rearranging yields

$$\lambda \mathbf{F} \Delta s^{(k)} = -\lambda \mathbf{F} s^{(k-1)} \quad (2.10)$$

To smooth the final perturbation from $s^{(0)}$ we similarly have

$$\lambda \mathbf{F} \sum_{i=0}^k \Delta s^{(i)} = 0 \quad (2.11)$$

Or, since $\sum_{i=0}^k \Delta s^{(i)} = \Delta s^{(k)} + \sum_{i=0}^{k-1} \Delta s^{(i)}$,

$$\lambda \mathbf{F} (\Delta s^{(k)} + \sum_{i=0}^{k-1} \Delta s^{(i)}) = 0 \quad (2.12)$$

Once again applying distributivity and rearranging Equation 2.12 yields

$$\lambda \mathbf{F} \Delta s^{(k)} = -\lambda \mathbf{F} \sum_{i=0}^{k-1} \Delta s^{(i)} \quad (2.13)$$

which is used in Equation 2.7. This sort of iterative inversion that smooths the total model through the addition of constraints as used here is known as a "jumping" procedure (Shaw, 1986) as opposed to a "creeping" procedure that simply constrains perturbations relative to the previous iteration's result.

Results to this inverse problem (Equation 2.7) are usually expressed in terms of a trade-off curve between regularization (i.e., smoothness, flattening and damping) and the data misfit. Such curve shows how the data fit varies when changing regularization levels. The optimal model solution lies in the corner of the curve representing a reasonable data fit and a fairly regularized model. Figures 4.3, A.4 and A.5 show the data fit (% root mean square RMS residual reduction) versus rms model roughness for the body-wave tomography models

over northwestern Canada and the Slave craton, respectively. Each point corresponds to an inversion with different regularization levels.

According to Equation 2.1, we will represent Equation 2.7 as,

$$\mathbf{d} = \mathbf{G}\mathbf{m}, \quad (2.14)$$

where

$$\mathbf{d} = \begin{pmatrix} \mathbf{W}\Delta t^{(k-1)} \\ \lambda \mathbf{F} \left\{ \sum_{i=0}^{k-1} \Delta s^{(i)} \right\} \end{pmatrix}; \quad \mathbf{G} = \begin{pmatrix} \mathbf{W}\mathbf{P} & \mathbf{W}\mathbf{H} & \mathbf{W}\mathbf{E} \\ \lambda \mathbf{F} & 0 & 0 \end{pmatrix}; \quad \mathbf{m} = \begin{pmatrix} \Delta s^{(k)} \\ h^{(k)} \\ e^{(k)} \end{pmatrix} \quad (2.15)$$

The least-square solution to Equation 2.14

$$\mathbf{m} = (\mathbf{G}^T \mathbf{G})^{-1} \mathbf{G}^T \mathbf{d} \quad (2.16)$$

The operator $(\mathbf{G}^T \mathbf{G})^{-1} \mathbf{G}^T$ is called the *generalized inverse of G*. It provides the best solution in a least-square sense, because it gives the smallest squared misfit. Obtaining the least-square solution to Equation 2.14 in a straightforward fashion was problematic when this tomographic software was developed due to the limitations in computer memory. Although the matrix $\mathbf{G}$ is sparse (most of its elements are zero) the matrix $(\mathbf{G}^T \mathbf{G})^{-1}$ is dense. Therefore, VanDecar (1991) takes advantage of the matrix sparseness and uses the conjugate gradient method, which does not require the storage of the entire matrix. Initially, conjugate gradient methods were developed to solve large sparse systems in geophysical problems (Hestenes & Stiefel, 1952; Scales, 1987). We found that the solution converges after 200-300 iterations in agreement with VanDecar (1991), however, we perform 2000 conjugate gradient iterations for both teleseismic tomography models as this step is no longer computationally intensive under modern computer architectures.

Least-square solutions are highly susceptible to the effects of data outliers. VanDecar (1991) applies Huber's method that iteratively down-weights residuals that are 1.5 standard deviations away from the mean (Huber, 1964). We perform 10 Huber down-weighting iterations to obtain a more robust solution for our teleseismic tomography models.

2.2 BAYESIAN SURFACE-WAVE TOMOGRAPHY

In this section, we provide a description of the Bayesian surface-wave tomography theory applied in [Chapter 6](#). Surface waves propagate near the Earth’s surface as opposed to body waves, which travel through Earth’s interior. An important property of surface waves is dispersion, meaning that waves of different periods travel at different velocities. Therefore, typical surface wave arrivals recorded on a seismogram are spread out in time. In [Chapter 6](#), we measure fundamental-mode Rayleigh-wave group velocities from regional earthquakes, which provide constraints on the seismic velocity structure between each source-station pair at different periods. These dispersion curves can be integrated to construct a 2-dimensional group velocity map at each considered period.

We saw in the previous section ([Section 2.1](#)) that, in seismic tomography, using a regular grid of model parameters combined with irregularly distributed observations (data) creates instabilities during the inversion. In the previous [Section 2.1](#), we saw that when matrix $\mathbf{G}$ is not full rank due to irregular sampling of the model space, we can apply regularizations that impose additional constraints on the problem ([Aster et al., 2018](#)).

In this section, we present the concept of trans-dimensional inversion of surface-wave data in a Bayesian (i.e., probabilistic) framework where the solution is a large ensemble of Earth models that constrain the seismic velocity structure in the region of interest. This approach presents the following advantages:

- The dimension of the model, or the number of model parameters, is treated as unknown during the inversion, avoiding potential subjective decisions (i.e., the trans-dimensional aspect).
- Data errors are also treated as unknown and are determined as part of the inversion.
- The data determine the complexity of the model during the inversion. Regularization of model parameters is not required.

2.2.1 *Model parameterization*

[Bodin & Sambridge \(2009\)](#) propose to parameterize the velocity field for seismic tomography with a discrete set of Voronoi cells ([Okabe et al., 1992](#)) in a trans-dimensional tomography approach (i.e., the number of model parameters becomes an unknown in the problem). The distribution of Voronoi cells is defined by a discrete set of points (Voronoi nuclei) with associated Earth model parameters. Each Voronoi cell encloses all the points in space

that are closer from its nucleus than to any other Voronoi nucleus (i.e., nearest-neighbour interpolation). This parameterization scheme presents the advantage of being adaptive, which means that the spatial distribution of nodes (size and location) that define the Voronoi cells varies during the inversion, based on the data information content. However, Voronoi cells have been found to be inconvenient when applied to seismic tomography. In particular, the discrete nature of nearest-neighbour interpolation leads to individual models containing spurious edges that appear non-physical (Gosselin et al., 2020).

Gosselin et al. (2020) uses an alternative model parameterization, the Delaunay triangulation with linear interpolation introduced by Hawkins et al. (2019). This parameterization scheme is slightly different to the Voronoi cells, since nodes no longer represent Voronoi nuclei but rather vertices of triangles. The Delaunay triangulation maximizes the smallest angle in the triangulation and tend to avoid sliver triangles. Each point in the model space is linearly interpolated using the velocity values assigned to the nodes of the confining triangle (Gosselin et al., 2020). The Delaunay triangulation parameterization scheme performs particularly well in the recovery of features in the inversion of smooth models such as tomographic inversions (Hawkins et al., 2019) (Figure 2.2).

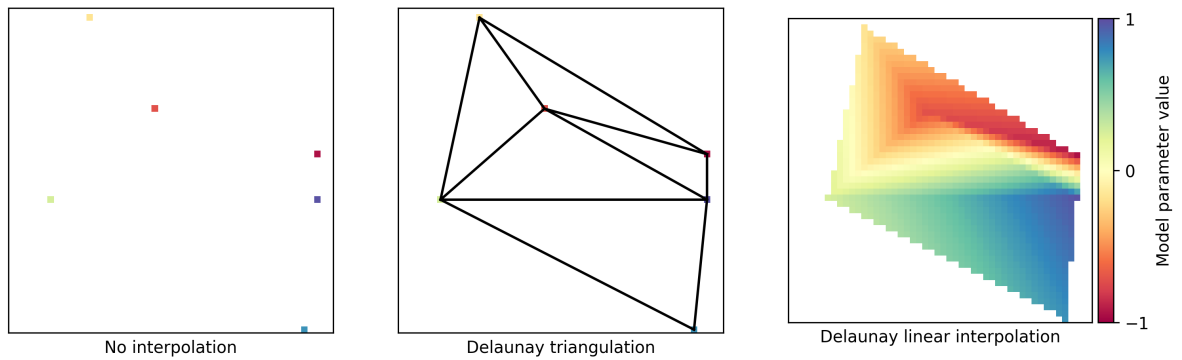


Figure 2.2.: (Left) A distribution of nodes with associated model parameter value. (Middle) Delaunay triangulation applied to the nodes. (Right) Delaunay triangulation with linear interpolation (from Gosselin et al., 2020).

2.2.2 Inversion procedure

Using the trans-dimensional tomography approach of Gosselin et al. (2020), surface-wave group velocity dispersion data are inverted simultaneously for isotropic group velocity and

azimuthal anisotropy. The velocity of a surface wave propagating through an anisotropic medium at a given period (T) and azimuth (ψ) is given by :

$$C(T, \psi) = C_0(T) + A1(T) \cos(2\psi) + B1(T) \sin(2\psi) + A2(T) \cos(4\psi) + B2(T) \sin(4\psi) \quad (2.17)$$

where $C_0(T)$ is the isotropic group velocity, 2ψ is the variation of the group velocity with π periodicity, and 4ψ is the variation with $\pi/2$ periodicity. In this work, the variation of group velocity with $\pi/2$ periodicity is neglected for computational simplifications (Gosselin et al., 2020). Furthermore, the number and the position of nodes and the uncertainty associated with the surface-wave dispersion measurements are treated as unknowns in the inversion, which are referred to as hyper-parameters in trans-dimensional tomography. Each node is defined by five parameters: the isotropic group velocity ($C_0(T)$), two parameters that describe the anisotropy with π periodicity ($A1$ and $B1$) and two coordinate values (Gosselin et al., 2020). The uncertainty σ (with respect to surface-wave travel time) associated with our data set is defined as a linear relationship between a source-to-station distance Δi and two hyper-parameters a and b for a given path i (Bodin et al., 2012).

$$\sigma_i = a\Delta_i + b \quad (2.18)$$

Figure 2.3 shows the hyper-parameters estimates obtained from the inversion of surface-wave group velocity dispersion measurements at 36 s.

2.2.3 Bayesian formulation

In a Bayesian framework, unknowns are typically treated as random variables and described with probability distributions. The posterior probability density (PPD), as defined by Bayes' theorem, is formulated as :

$$P(m|d, H) = \frac{P(d|m, H)P(m|H)}{P(d|H)} \quad PPD = \frac{Likelihood * Prior}{Evidence} \quad (2.19)$$

where $P(d|m, H)$ is the likelihood or the conditional probability of the observed data d (i.e., group velocity travel times at a given period) given a set of model parameters m (i.e., the number/position of nodes, and their associated properties) for a specific model H . H represents the physics/assumptions and parameterization choices used to produce the model. The likelihood describes how well a given model with a particular set of model parameters

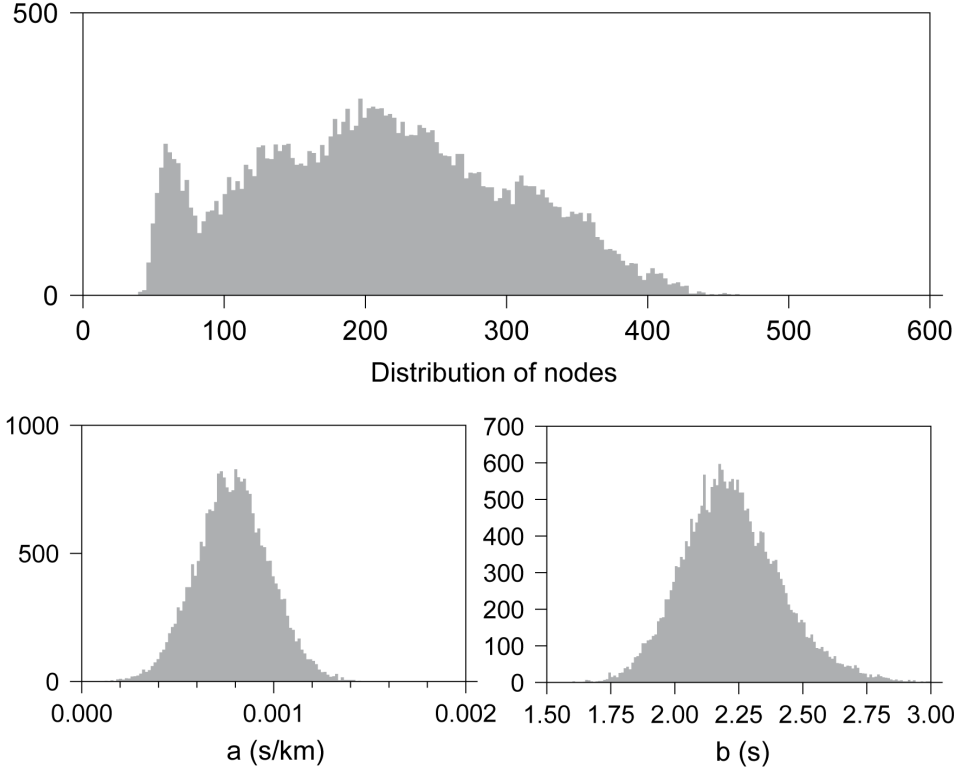


Figure 2.3.: Hyper-parameter estimates obtained from the inversion of group velocity dispersion data at 36 s. Top diagram shows the distribution of nodes. Bottom left diagram shows the linear coefficient between unknown data errors and travel path length (a). Bottom right diagram shows the constant term for unknown data error (b).

can reproduce the observed data. $P(m|H)$ is the probability of a set of model parameters m given a specific model H and is the prior probability distribution, which corresponds to the state of knowledge on the model before measuring the data. $P(d|H)$ is the probability of the observed data d given a specific model H and is referred to as the Bayesian evidence and provides normalization over the parameter space. Finally, $P(m|d, H)$ is the conditional probability of m given d for a specific model and is called the PPD.

We use the approach based on a generalization of the Metropolis-Hastings algorithm (Hastings, 1970; Metropolis et al., 1953), called reversible-jump Markov chain Monte Carlo (rjMCMC) method (Green, 1995), to generate a large ensemble of models in order to approximate the PPD of the model parameters. The rjMCMC algorithm randomly perturbs a set of model parameters m to a new state m' . This new model is accepted or rejected from the ensemble with a probability of acceptance $A(m|m')$, which is defined as

$$A(m|m') = \min \left[1, \frac{Q(m|m')}{Q(m'|m)} \frac{P(m')}{P(m)} \frac{\mathcal{L}(m')}{\mathcal{L}(m)} |J| \right] \quad (2.20)$$

where $\frac{Q(m|m')}{Q(m'|m)}$ is the proposal ratio, $\frac{P(m')}{P(m)}$ is the prior ratio, $\frac{\mathcal{L}(m')}{\mathcal{L}(m)}$ is the likelihood ratio and $|J|$ is the determinant of the Jacobian matrix describing the transition from one set of model parameters (state) to another. The prior probability density consists of a bounded uniform distribution. The bounds are chosen to be wide enough to include all geophysically-realistic models and allow the data set to predominantly constrain the solution.

At each iteration, 3 types of model perturbations are randomly applied. The first one, the "birth" perturbation, adds a node and its associated model parameters to the pre-existing Delaunay triangulation parameterization scheme. The second one, the "death" perturbation, removes a node. The third one, the "move" perturbation, moves the position of node. These model perturbations permit generation of a wide range of seismic velocity models by modifying the geometry of the Delaunay triangulation.

For all inversions, 20 rjMCMC chains ran in parallel and generated 5 million samples each. We discard the first half (i.e., 50 million burn-in samples) of the ensemble of samples as they are dependent on the starting model, which is randomly chosen. Therefore, our final group velocity maps are based on the post burn-in samples. We calculate the mean and standard deviation of the ensemble of models at every location on the map as a representative model with associated uncertainty

2.3 SEISMIC ANISOTROPY

Let's assume that seismic waves travel through an isotropic medium, in which elastic behaviour does not depend on direction. The equation (called *Hooke's law*) that relates the stress tensor σ_{ij} to the strain tensor e_{kl} via the elastic tensor c_{ijkl} can be written as,

$$\sigma_{ij} = c_{ijkl}e_{kl} \quad (2.21)$$

Or, in terms of Lamé parameters, [Equation 2.21](#) is written

$$\sigma_{ij} = \lambda\theta\delta_{ij} + 2\mu\varepsilon_{ij} \quad (2.22)$$

where θ is the cubic dilatation, λ and μ are the two independent elastic constants (Lamé parameters) and δ_{ij} is the Kronecker delta ($\delta_{ij} = 1$ for $i = j$, $\delta_{ij} = 0$ for $i \neq j$). When considering anisotropy, meaning that the material properties differ depending on the direction of propagation, *Hooke's law* involves at least three independent elastic constants and there can be up to 21 independent elastic constants depending on the symmetry of the elastic tensor. Seismic wave velocities depend on the elastic constants, therefore, seismic waves

traveling through anisotropic medium travel faster or slower depending on the direction of propagation. For example, a shear wave entering an anisotropic medium splits into two orthogonal shear waves, each with different polarity and traveling at different speed. This phenomenon is called shear-wave splitting.

The source of seismic anisotropy is attributed to several mechanisms from the crust to the upper mantle. For example, a stack of horizontally-stratified rocks with different isotropic velocities can, as a whole, behave anisotropically, where seismic waves traveling parallel or perpendicular to the layers will have different speeds. This is called shape-preferred orientation (SPO) anisotropy. Anisotropy can also arise from a preferred orientation of intrinsically anisotropic mineral crystals such as micas and/or amphiboles in the crust and olivine in the upper mantle. This is called lattice-preferred orientation (LPO). Olivine LPO is acquired during mantle flow, as the olivine fast axis aligns with the maximum principal strain axis (Nicolas & Christensen, 1987). Therefore, there is a clear link between deformation occurring at upper mantle depths and seismic anisotropy (Tommasi et al., 2000).

Because estimates of seismic anisotropy provide useful information on crust and upper mantle dynamics, interpretation of seismic anisotropy has become a major field in seismology for studying dynamic processes in the Earth's interior.

2.3.1 Radial anisotropy and azimuthal anisotropy

The symmetry of the stress and strain tensors reduces the number of independent elastic constants from 81 to 21. Therefore, the c_{ijkl} tensor can be written using the Voigt notation as a matrix C_{mn} ,

$$C_{mn} = \begin{pmatrix} C_{11} & C_{12} & C_{13} & C_{14} & C_{15} & C_{16} \\ C_{21} & C_{22} & C_{23} & C_{24} & C_{25} & C_{26} \\ C_{31} & C_{32} & C_{33} & C_{34} & C_{35} & C_{36} \\ C_{41} & C_{42} & C_{43} & C_{44} & C_{45} & C_{46} \\ C_{51} & C_{52} & C_{53} & C_{54} & C_{55} & C_{56} \\ C_{61} & C_{62} & C_{63} & C_{64} & C_{65} & C_{66} \end{pmatrix} \quad (2.23)$$

As we have seen in the previous section, an isotropic material has 2 independent elastic constants. Therefore, the c_{ijkl} tensor for an isotropic material can be written as,

$$c_{ijkl} = \lambda \delta_{ij} \delta_{kl} + \mu (\delta_{ik} \delta_{jl} + \delta_{il} \delta_{jk}) \quad (2.24)$$

so its matrix form is

$$C_{mn} = \begin{pmatrix} \lambda + 2\mu & \lambda & \lambda & 0 & 0 & 0 \\ \lambda & \lambda + 2\mu & \lambda & 0 & 0 & 0 \\ \lambda & \lambda & \lambda + 2\mu & 0 & 0 & 0 \\ 0 & 0 & 0 & \mu & 0 & 0 \\ 0 & 0 & 0 & 0 & \mu & 0 \\ 0 & 0 & 0 & 0 & 0 & \mu \end{pmatrix} \quad (2.25)$$

In seismology, one of the most important forms of anisotropy has one axis of symmetry about the vertical axis that is normal to a plane of isotropy and is known as radial anisotropy (or transverse isotropy). For example, radial anisotropy occurs in seismic reflection surveys of horizontal sediment layers in the crust. A material or medium exhibiting radial anisotropy is characterized by five independent elastic parameters defining hexagonal symmetry A , C , F , L and N (Stein & Wysession, 2003). Let's consider that the axis of symmetry is in the vertical direction, so properties in that direction differ from those in the horizontal plane, the elastic constant matrix C_{mn} becomes,

$$C_{mn} = \begin{pmatrix} A & A - 2N & F & 0 & 0 & 0 \\ A - 2N & A & F & 0 & 0 & 0 \\ F & F & C & 0 & 0 & 0 \\ 0 & 0 & 0 & C & 0 & 0 \\ 0 & 0 & 0 & 0 & L & 0 \\ 0 & 0 & 0 & 0 & 0 & L \end{pmatrix} \quad (2.26)$$

For azimuthal anisotropy, hexagonal crystal symmetry is assumed with symmetry axis in the horizontal plane yielding a fast, v_{sv1} , and a slow, v_{sv2} , propagation direction for vertically polarized shear waves (Becker et al., 2012). One can obtain azimuthal anisotropy by rotating the system of coordinates describing radial anisotropy with the vertical axis turned to horizontal.

2.3.2 *Seismic anisotropy in the lithosphere and asthenosphere*

Seismic anisotropy occurs within all the layers of the Earth and there exists a variety of sources of anisotropy in the Earth. A primary source of seismic anisotropy in the upper crust is the presence of vertically-aligned fluid-filled cracks induced by the maximum horizontal compressive stress parallel to the cracks (Crampin, 1994). Deeper in the crust, seismic anisotropy is attributed to the LPO of micas, amphiboles and/or orthopyroxenes in metamorphic rocks (Moschetti et al., 2010). Seismic anisotropy within the upper mantle is attributed to the LPO of olivine crystals (Tommasi et al., 2000).

Seismic anisotropy in the lithospheric mantle and in the asthenosphere is often determined from shear wave splitting analysis of teleseismic core-refracted shear waves (SKS and SKKS). Using SKS waves presents two advantages. First, the polarization of a SKS wave is known. SKS waves are fully polarized in the radial direction (SV) since the conversion from longitudinal (compression, or "K" phase in the core) to transverse (shear, or "S" phase back in the lower mantle) only generates SV, when the downgoing S-wave encountered the core mantle boundary (CMB). Second, the loss of SH energy in the core constrains the observed splitting to be on the receiver side of the path. Upon entering an anisotropic medium, a shear wave splits into two orthogonal shear waves traveling along the fast and slow directions of propagation (assuming radial anisotropy with a horizontal axis of symmetry) (Figure 2.4). Shear wave splitting is therefore defined by two parameters, the fast direction of wave propagation (or fast axis ϕ) and the accumulated delay time (δt) between the fast and slow components. These parameters reflect integrated effect along the receiver-side ray path and therefore have poor vertical resolution; however, it is typically assumed that the dominant contribution lies within the lithosphere and/or asthenosphere given the model of shear-aligned olivine being primarily responsible for the anisotropic signature (Nicolas & Christensen, 1987). We calculate SKS splitting parameters for stations in northwestern Canada using SplitPy (Audet & Schaeffer, 2019), a Python version of SplitLab (Wüstefeld et al., 2008). The method is described in detail in Appendix C.

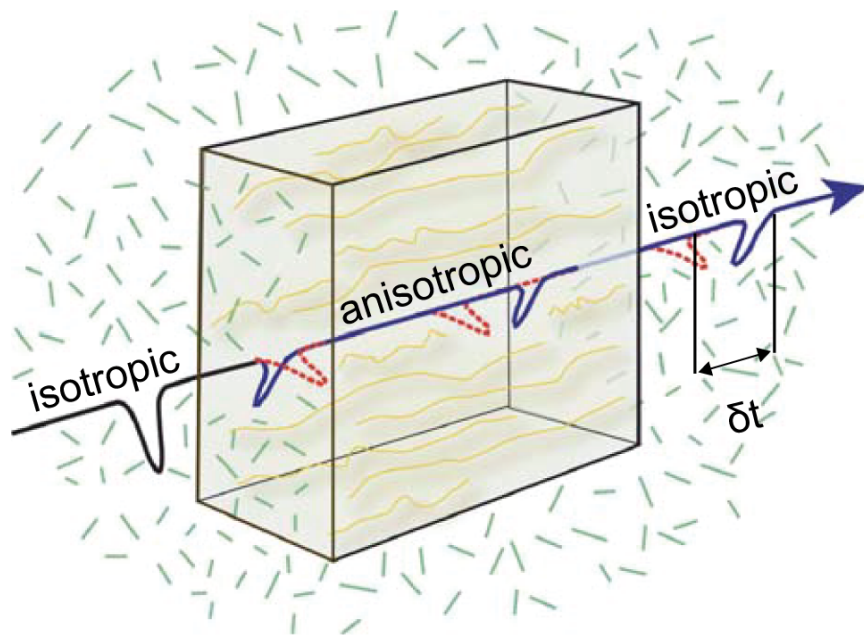


Figure 2.4.: Example of a shear wave traveling through an anisotropic medium. Upon entering an anisotropic medium, a shear wave splits into two orthogonal shear waves traveling along the fast (blue) and slow (red) directions of propagation. As the seismic wave travels through the anisotropic medium, a delay time accumulates (δt) between the fast and slow components (modified from [Maupin & Park, 2007](#)).

Part II

UPPER MANTLE STRUCTURE OF NORTHWESTERN CANADA

UPPER MANTLE STRUCTURE UNDERLYING THE DIAMONDIFEROUS SLAVE CRATON FROM TELESEISMIC BODY-WAVE TOMOGRAPHY

3.1 SUMMARY

Cratons are, by definition, the most tectonically stable and oldest parts of the continental lithosphere on Earth. The Archean Slave craton is located in the northwestern part of the Canadian Shield. The propensity of diamondiferous kimberlite pipes in the central Slave craton raises many questions regarding their structural environment and source. Here, we provide the most robust teleseismic P and S body wave tomography models over the Slave craton region based on 20,547 P-wave delay times, 6,140 direct S-wave delay times and 3,381 SKS delay times. The P-wave model reveals an alternating pattern of relative positive and negative anomalies over a fine scale region within the central Slave craton. Furthermore, the P-wave model revealed two fine structures located in the lithosphere beneath the Lac de Gras kimberlite field, with relatively slow anomalies (B - C) that extend from 75 km to 350 km depths with an apparent dip to the north. These relatively slow P-wave anomalies are associated with metasomatised regions within the lithosphere. The most recent kimberlite pipes (75-45 Ma) located in the Lac de Gras field are located on steep V_P and V_S gradients. The S-wave model displays a slow S-wave anomaly lying from 300 km depth to the transition zone beneath the central Slave craton. This anomaly is located beneath the Lac de Gras kimberlite field. We suggest that this anomaly is not the cause of the actual kimberlites at the surface since the last eruptions occurred 75-45 Ma ago but may be related to a potential kimberlite magma ascent in the asthenosphere.

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3.2 INTRODUCTION

Archean cratons are characterized by a thick sub-continental lithospheric mantle (SCLM) reaching 200 km depths on average, but the timing and the mechanisms involved in their formation is still under debate. Proposed models involve successive accretion-collision episodes (Jordan, 1988), collision of island arcs comprising depleted material (Ashwal & Burke, 1989), basal accretion by cooling asthenospheric material (Thompson et al., 1996) and buoyant subduction and imbrication by lithospheric-scale stacks (Helmstaedt & Schulze, 1989). The latter is commonly accepted but Arndt et al. (2009) suggest that the SCLM formed from the residues of melting of subsequent plumes rising from the deep mantle.

In the early 1990s, before the discovery of diamondiferous kimberlites, differences were noted between the Slave craton and other Archean cratons, particularly with the Superior craton (Padgham & Fyson, 1992). For instance, the Slave craton, located in the northwestern part of the Canadian Shield (Figure 3.1), is relatively small (700×500 km) representing only one tenth of the Superior craton. The Slave craton shows a relatively high abundance of meta-sedimentary rocks, a higher percentage of potassium-rich plutonic rocks and a high proportion of felsic volcanic to mafic rocks (Padgham & Fyson, 1992). In addition, it shares some similarities and differences with other Archean cratons such as the Kaapvaal craton in South Africa. For instance, they both consist of crustal sections or terranes that were accreted during several tectonic episodes in Archean time. Then, the crystalline basement of the Slave craton consists of older rocks (4.0 Ga; Acasta gneiss) than the Kaapvaal craton in South Africa (3.6 Ga; gneiss of the Swaziland-Barberton areas) and the latter exhibits a simpler (less stratified) upper mantle structure inferred from seismic data (Bostock, 1998; Gao et al., 2002).

The Slave craton extends North to South from Cape Barrow to Great Slave Lake (Figure 3.1). The crust of the Slave craton consists of 4.0 to 2.58 Ga crystalline rocks and the Slave lithosphere is thought to have cratonized between 2.85 and 2.58 Ga around an older Mesoarchean (3.2-2.8) basement, the Central Slave Basement Complex (Bleeker et al., 1999; Ootes et al., 2009; Helmstaedt, 2009; Snyder et al., 2014). The Slave craton is bounded to the east and west by Paleoproterozoic belts such as the Taltson-Thelon magmatic arcs and the Wopmay Orogen (including the Great Bear Arc) respectively. The Thelon orogen formed during the dextral-oblique collision between the Slave and Rae cratons at 2.02-1.91 Ga (Hoffman et al., 2011). The Thelon magmatic arc formed on Archean basement of the adjacent Rae craton (Hoffman, 1988). The Great Slave Lake shear zone is a vertical, strike-slip structure extending over hundreds of kilometers (Hanmer et al., 1992). The shear zone defines the southern boundary of the Slave craton and also delimits the Thelon magmatic arc of the Taltson

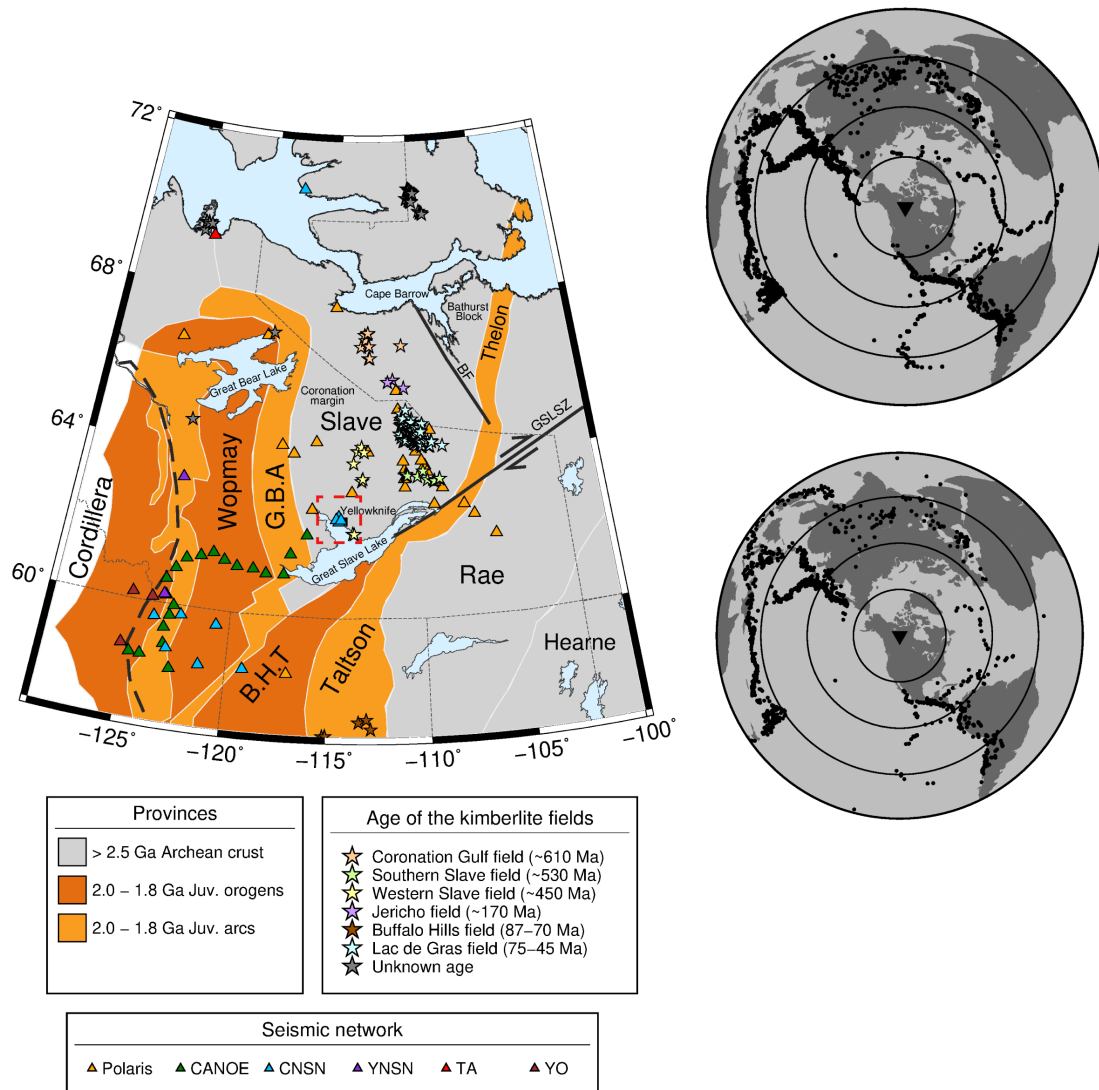


Figure 3.1.: Tectonic map of the Slave Craton and its surroundings. The red box shows the location of Yellowknife and the Yellowknife array. Also shown are the seismic stations used in this study, the age distribution of the kimberlite fields (Lockhart et al., 2004) and the Cordillera Deformation Front (black dashed line). Equidistant azimuthal projection showing the distribution of $M_W \geq 5.5$ earthquakes for P (top right - black circles) and S (bottom right - black circles) recorded between 2000 and 2017 at one or more stations of the array. Concentric circles represent epicentral distances of 30° , 60° , 90° and 120° from the center 62°N , -114°E . Abbreviations: B.H.T, Buffalo Head Terrane; G.B.A, Great Bear Arc; BF, Bathurst Fault; GSLSZ: Great Slave Lake Shear Zone

magmatic arc (Hoffman, 1988). The Great Slave Lake shear zone, the Thelon and Taltson magmatic arcs were inferred to be the same magmatic arc (Hoffman, 1988). A strong magnetic anomaly can be followed from central Alberta to the Canadian Arctic Archipelago (i.e., Victoria island) corresponding to the location of these two magmatic arcs and, therefore, supports this theory (Hoffman, 1988).

To the west, the Wopmay Orogen (Figure 3.1) is a Paleoproterozoic collisional belt that formed from the accretion of the Hottah terrane to the active western margin of the Slave craton at 1.88 Ga (Hildebrand et al., 2010; Hoffman et al., 2011). The Wopmay Orogen includes the Asiatic fold-and-thrust belt, the Coronation margin, the Great Bear magmatic arc, the Hottah terrane and the Fort Simpson terrane (Aspler et al., 2003). A geological model of the evolution of the Wopmay orogen was proposed where a westward subduction of the oceanic and continental lithosphere of the Slave plate beneath the Hottah terrane occurred at 1.9 Ga. Following an episode of slab break off at 1.884 Ga, a new eastward dipping subduction zone and arc magmatism formed resulting in the formation of the Great Bear Arc at 1.875 Ga (Hildebrand et al., 2010; Hoffman et al., 2011). An alternative model is proposed by Davis et al. (2015). Their model places the Hottah terrane in a east-dipping subduction system, south of the GSSLZ (south of its current location), adjacent to the Buffalo Head Terrane, between 2.5 and 2.2 Ga before rifting from the margin. Following this episode of rifting, the Hottah terrane drifted to the north and eventually collided with the Slave craton margin at 1.885 Ga producing the Calderian orogeny. Their model also suggests a plausible 1.9 Ga low-angle subduction between the Buffalo Head Terrane and the Taltson magmatic arc.

In 1995, Fipke et al. (1995) discovered the presence of "indicator minerals" such as Cr-diopside crystals at Point Lake, north of the Lac de Gras area. Based on this discovery, they assumed that these indicator minerals were from a kimberlite pipe in the vicinity. This location is known today as one of the major diamond mines in the Slave craton, the Ekati mine. Since this discovery, a large number of kimberlite pipes have subsequently been discovered across the Slave craton (Figure 3.1). The propensity of diamondiferous kimberlite pipes in the central Slave Craton raises many questions regarding their structural environment and source. Here we investigate the relationship between the known diamondiferous areas and their underlying mantle structure through development of a teleseismic body-wave tomography model. In particular, we determine the P- and S-wave velocity structure beneath the Slave Craton using two independent tomographic inversions of P- and S-wave relative delay times.

3.3 PREVIOUS GEOPHYSICAL STUDIES

Following the discovery of diamondiferous kimberlite pipes in the central Slave Craton in the 1990s, the composition and structure of the underlying mantle have been extensively studied (e.g., [Bostock, 1998](#); [Jones et al., 2003](#); [Snyder et al., 2004](#); [Snyder & Bruneton, 2007](#)). Seismic velocity discontinuities within the crust and upper mantle were revealed through analysis of P-to-S conversions recorded in the P-coda of teleseismic earthquakes (i.e., receiver functions). [Bostock \(1998\)](#) calculated receiver functions for five broadband seismic stations at the Yellowknife array. These data reveal that the underlying upper mantle structure is well-stratified with three distinct seismic structures denoted as H, X and L. The H structure is observed at 70-80 km depth (~ 10 km thick), exhibits a strong localized anisotropy and is interpreted as a layer of Archean crust obducted below an existing protocontinent. The X structure is hypothesized to be a sequence of layers observed at depths from 120 to 150 km. The L structure is a dipping layer located toward the center of the Slave craton at 170 to 230 km depths. The H and X structures are interpreted as markers of the formation of the Slave craton from flat subduction processes resulting in a near-horizontal mantle stratigraphy. The L structure might be related to a final stage of the craton assembly. At greater depths extending from the base of the upper mantle through the transition zone, [Schaeffer & Bostock \(2010\)](#) observe well defined and sharp 410 and 660 km discontinuities using a combination of P- and S-receiver functions. Both discontinuities demonstrate little-to-no topography, suggestive of typical transition zone conditions (i.e., no evidence for hotspots or present-day subduction).

In order to investigate the upper mantle structure underlying the Slave craton, [Bank et al. \(2000\)](#) developed a teleseismic P-wave travel-time tomography model. Their model shows the presence of a fast velocity anomaly at 100 km depth beneath the central Slave craton interpreted as the core of the craton, the Central Slave Basement Complex ([Bleeker et al., 1999](#)). The Lithosphere-Asthenosphere Boundary (LAB) is thought to lie at 250 ± 50 km in the central Slave craton ([Bank et al., 2000](#)). Furthermore, a fast P-wave anomaly is also revealed in the eastern part of the Slave craton from 100-300 km depth. Two low-velocity anomalies are located in the central and northern Slave craton and are interpreted as regions of trapped fertile mantle (i.e., a mantle material with a lherzolitic composition in comparison with the surrounding harzburgitic lithosphere).

Based on a combination of teleseismic methods such as receiver functions and tomographic inversion of relative P-wave delay times, [Snyder et al. \(2004\)](#) show the presence of 4 seismic discontinuities at depths of 38, 110, 140 and 190 km beneath the central Slave craton. These discontinuities are thought to bound 2 anisotropic layers beneath the central Slave craton.

The 38 km discontinuity is interpreted as the Moho in agreement with other studies (Griffin et al., 1999; Jones et al., 2003), whereas the 110 and 140 km discontinuities are thought to reflect the top and bottom of a depleted harzburgite layer that may contain graphite, i.e., the Central Slave Mantle Conductor, referred to as CSMC, (Jones et al., 2001, 2003). The body-wave tomography model did not reveal the presence of the CSMC.

Regions of the Earth's mantle are seismically anisotropic at different depths within the crust and the mantle (e.g., Hess, 1964). Seismic anisotropy is a useful tool that reveals ordered structures, called fabrics, and has become an important tool for studying dynamic processes in the mantle (Long & Silver, 2009). Based on the tendency of a shear wave to split upon entering an anisotropic layer into two orthogonal components (i.e., shear-wave splitting, or birefringence), Bank et al. (2000) find an average delay time of 1 second with a fast-axis direction of $N60^\circ \pm 20^\circ$ that is coherent with the present-day absolute plate motion of North-America. As a consequence, they relate the observed pattern of anisotropy within the lithosphere with the influence of the plate motion. Snyder & Bruneton (2007) show the presence of two anisotropic layers located within the mantle underlying the Slave craton. The first anisotropic layer is located at 80-120 km depths and is consistent with geological observations. This layer is interpreted to be related to a phase of deformation during the formation of the Slave craton at 2610-2580 Ma. The fast polarized axis of the deeper anisotropic layer shows a clear alignment with present-day North American plate motion and is interpreted to be due to lattice-preferred orientation of olivine in the asthenospheric mantle.

Over the last decade, the deployment of the USArray Transportable Array in the US drastically improved tomographic images over North America allowing the development of new seismic tomography models (e.g., Burdick & Lekić, 2017; Schaeffer & Lebedev, 2014; Yuan et al., 2014). Continental-scale shear-wave velocity models confirmed the presence of a thick cratonic root beneath the Canadian Shield up to 150-250 km depth (Schaeffer & Lebedev, 2014; Yuan et al., 2014). Another feature shown by those shear-wave models is the sharp velocity gradient in western Canada at lithospheric depths (i.e., 100-200 km) between the Phanerozoic Cordillera and the Archean craton.

More recently, Chen et al. (2009) show a spatial correlation between a seismic discontinuity and the conductive layer observed at depths ranging from 80-140 km beneath the Slave craton (Jones et al., 2001, 2003). This layer is interpreted as a thin conductive layer with interconnected films of graphite deposited along grain boundaries. However, the graphite film interpretation has been proven to be an unviable explanation for the the slow seismic anomaly and/or the high conductivity anomaly because the graphite film would be destroyed by its high interfacial energy at this depth (Yoshino et al., 2018). Another potential conductive

agent could be the presence of carbonatite melt within the peridotitic mantle underlying the Slave craton (Yoshino et al., 2018).

3.4 METHOD

We develop a new teleseismic body-wave tomography model of the upper mantle surrounding the Slave craton in northwestern Canada. This new model is motivated by the addition and combination of data from several new temporary broadband seismograph networks in northwestern Canada that were not available in previous studies.

3.4.1 Data

Seismic data were obtained from the Incorporated Research Institutions for Seismology (IRIS) and the Canadian Hazard Information Service (CHIS). Our data set consists of P- and S-wave arrival time delays (or delay times) from teleseismic earthquakes recorded at stations of the following permanent and temporary networks: the Canadian National Seismic Network (CNSN), the CANadian NOrthwest Experiment (CANOE), Portable Observatories for Lithospheric Analysis and Research Investigating Seismicity (POLARIS), the Yukon-Northwest Seismograph Network (YNSN), the USArray Transportable Array (TA) and the Yukon Observatory (YO). The P-wave data set consists of 20,547 delay times from 3,013 earthquakes of magnitude $M_W \geq 5.5$ in the teleseismic epicentral distance range (Δ) $30^\circ \leq \Delta \leq 120^\circ$. For the S-wave data set, we select 933 earthquakes for the direct S-wave delay times and 768 earthquakes for SKS core phase delay times, resulting in 6,140 direct S-wave delay times and 3,381 SKS delay times. The earthquake epicentral distribution is dominated by sources in the Northwestern Pacific and South America (Figure 3.1), which greatly influences seismic path coverage and resolution at depth (see later sections).

3.4.2 P- and S-wave picking

Automated picking of the first P arrival was performed on vertical component seismograms that were tapered and filtered with a non-zero phase, 2-pole Butterworth bandpass filter with corner frequencies of 0.4 Hz and 2.0 Hz. Direct S waves were picked similarly on the transverse component seismograms and were filtered between 0.02 Hz and 0.4 Hz. The SKS waves were picked on the radial component seismograms using the same bandpass filter as the direct S waves (e.g., Bastow et al., 2005, 2008).

In order to obtain accurate picks, we implement a Kurtosis-based picker (Baillard et al., 2014) in combination with the Automated and Interactive Measurement of Body wave Arrival Times (hereafter referred to as AIMBAT) developed by Lou et al. (2013). The alignment of the seismograms is done in a four-step procedure by AIMBAT. First, the seismograms are iteratively aligned using the theoretical arrival time, t_0 (predicted arrival time from IASP91 Kennett & Engdahl, 1991). The alignment is achieved by the Iterative Cross Correlation and Stack (ICCS) algorithm and produces a new pick, t_1 . Following this step, the user manually picks the phase arrival time in the array stack. We use Baillard et al. (2014)'s automatic picker to perform this step. This algorithm allows the user to automatically pick the onset of a seismic phase arrival based on a characteristic function obtained from the sliding-window kurtosis function. We apply this method to automatically pick the seismic phase arrivals in the array stack, producing a new pick, t_2 . The third step refines the alignment, in each seismogram, by using as input time t_2 and determines relative delay times using the Multi Channel Cross Correlation, MCCC (VanDecar & Crosson, 1990). This automated phase alignment can accurately recover sub-sample delay times. For both P and S data sets, we use thresholds to keep or reject delay time estimates. In particular we reject waveforms with a cross-correlation coefficient lower than 0.8 and 0.7 for P and S data sets, respectively. The network-averaged standard deviation of the P and S-delay times is 0.02 s and 0.08 s, respectively.

This method is highly sensitive to noise in the data; seismograms with a low signal-to-noise ratio result in meaningless picks. We therefore discard every seismogram with a SNR lower than 1.5 for the P-phase arrival, and 2.5 for the direct S and SKS phases. Figure 3.2 illustrates the procedures for two different earthquakes. The 27 February 2004 M_W 5.6 earthquake was recorded by 8 stations and, due to the high noise levels in the data, it is very difficult to identify the P-wave arrival and recover meaningful delay times. The 08 September 2004 M_W 6.1 earthquake was recorded by 40 stations across our study area. The P-wave arrival is clearly defined on every seismogram and the picks are all aligned at the onset of the emergent phase. In order to avoid delay times from low SNR seismograms as input to the model, we further visually check every pick and discard those without a clearly identified seismic phase arrival.

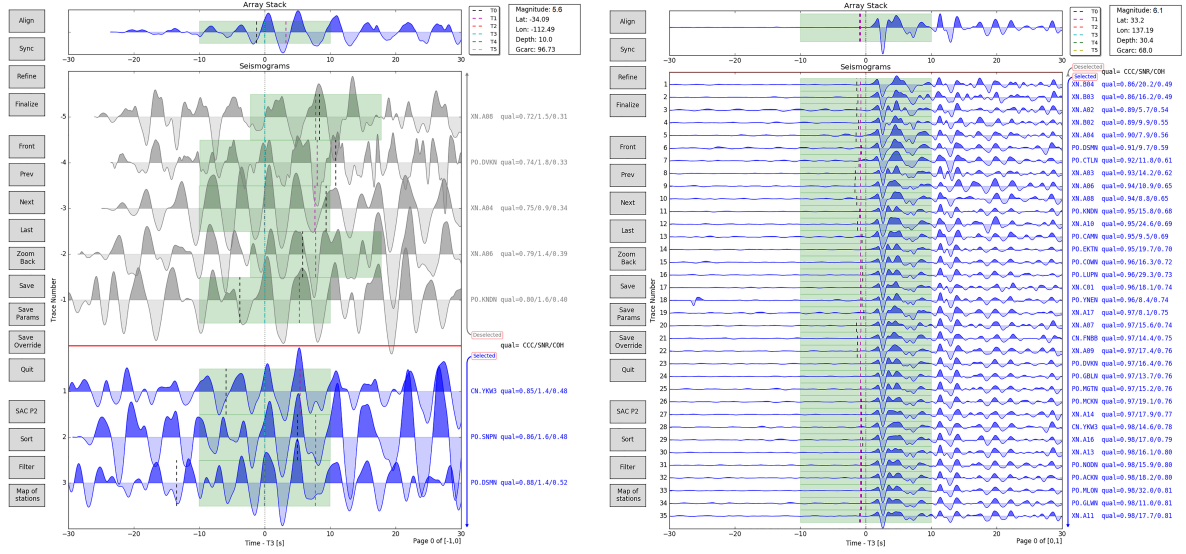


Figure 3.2.: AIMBAT's Graphical User Interface (GUI) for two different earthquakes for P-wave delay time analysis. At the top of the GUI is the array stack computed with the ICCS algorithm. The seismograms are sorted by quality (CCC: Cross-Correlation Coefficient ; SNR: Signal-to-Noise Ratio ; Coh: Coherency) and centered on the time pick produced by the MCCC, t_3 . The rejected seismograms are displayed in grey. The green shading shows the time window used for the cross-correlation. Left: The 27 February 2004 M_W 5.6 earthquake. Right: The 08 September 2004 M_W 6.1 earthquake.

3.4.3 Relative delay times

The relative delay time δt at a station is the difference between the measured (t_3) and predicted relative arrival times (t_0) and is given by:

$$\delta t = t_3 - (t_0 - \bar{t}_{net}) \quad (3.1)$$

where $\bar{t}_{net}$ is the mean expected travel time for the stations that recorded the same event (i.e., average IASP91 arrival time). Figure 3.3 shows the average relative station delay times for all P and S phases. There is a clear concordance between P and S relative station delay-time values, apart from a couple of stations located in the southwestern region of the study area. Negative delay-time values are observed in the central Slave craton and also in the northern part of the craton. The northern region of the Slave craton and southern region of the Wopmay Orogen are dominated by early arrivals. We observe a smooth transition from negative ($\delta t \sim -0.5s$) to positive ($\delta t \sim 0.5s$) delay-time values along the CANOE network from northeast to southwest, respectively. The positive delay-time values are all located in

the southwestern corner of the study area, which is similar to the results of [Mercier et al. \(2009\)](#).

In the inversion, seismic velocity anomalies are calculated relative to the background mean of the model beneath the Slave craton, as opposed to a global mean value ([Bastow et al., 2005, 2008, 2013](#)), and are represented by colored areas (red for low-velocity anomaly regions and blue for high-velocity anomaly regions).

3.4.4 *Inversion procedure*

In this study, we use the inversion procedure of [VanDecar et al. \(1995\)](#) to recover three-dimensional models of relative velocity anomalies. This inversion procedure has been used in studies over different tectonic settings in various regions of the world such as Ethiopia ([Bastow et al., 2005, 2008](#)), Western ([Mercier et al., 2009](#)) and Eastern Canada ([Boyce et al., 2016](#)).

This teleseismic travel-time inversion procedure constructs a large set of linear equations, where every equation relates the unknown velocity perturbation to the measured delay time within the model grid. Our model grid is parameterised with splines under tension constrained by a regular grid of 301,455 knots from 51° to 78° N, 90° to 138° W and 0 to 1300 km depths. The average knot spacing in the inner region of the model is 14 km between 57° and 72° N, 27 km between 100° to 128° W and 33 km between 0 and 600 km depths. During the inversion procedure, the code simultaneously inverts the relative delay times for velocity perturbations, source static terms and station static terms. The source and station static terms are free parameters in the inversion procedure. The source static terms reflect small variations in backazimuth and incidence angle due to distant heterogeneities and source mislocations ([Bastow et al., 2005](#)). The station static terms absorb the travel time anomalies associated with the region directly beneath the station where a lack of intersecting rays prevents the resolution of crustal structure ([Bastow et al., 2005](#)). The linear system of equations is under-determined (i.e., more unknowns than equations) and the solution is non-unique. In order to deal with the non-uniqueness of the problem, new regularized equations are added to the solution. The weighting of the regularization equations allows for controlling the trade-off between model smoothness and data misfit. For both P and S models, we invert the linear system of equations using 2000 conjugate gradient iterations within 10 Huber down-weighting iterations.

We plot the trade-off curves (or “L-curve”) between the Root Mean Square (RMS) residual reduction and the RMS model roughness to find the optimal final P-wave and S-wave velocity

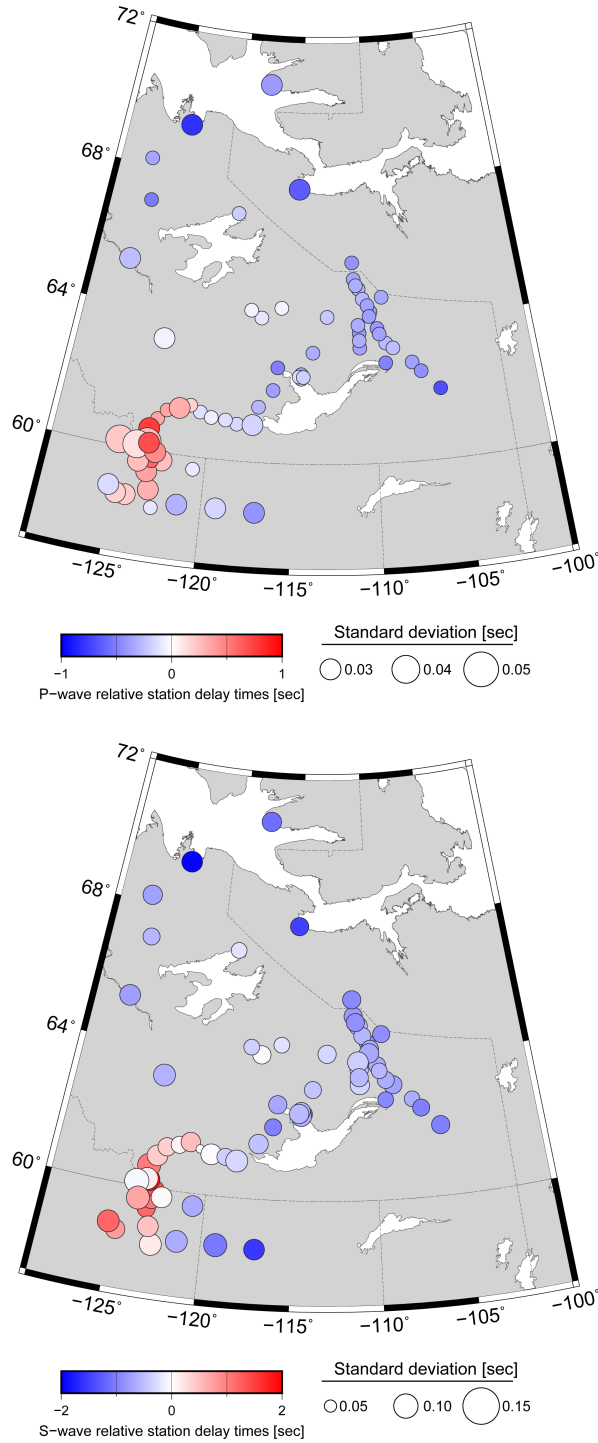


Figure 3.3.: P (top) and S (bottom) average relative station delay times. Blue and red circles correspond to early (high relative velocity) and late (low relative velocity) arrivals, respectively. The size of the circle corresponds to the standard deviation in seconds. Note the high relative velocities beneath the central Slave Craton toward the East.

models (see trade-off curves in the supplementary document, Figures A.4 and A.5). The optimal model lies in the bottom-left region of the trade-off curve. We chose flattening and smoothing parameters of $\lambda_{flat,P} = 1200$ $\lambda_{smooth,P} = 16500$ and $\lambda_{flat,S} = 1000$ and $\lambda_{smooth,S} = 16000$ for the optimal P- and S-wave models, respectively. Our final models explain 97% and 89% of the RMS of the relative arrival time residuals for P and S, respectively.

3.4.5 Checkerboard resolution test

In order to investigate the robustness of our model, we construct a synthetic checkerboard resolution test. We create layers of alternating positive and negative seismic velocity perturbations ($\delta V = \pm 5\%$) at 150, 350 and 550 km depths. Each anomaly has a radius of 75 km and is defined by a Gaussian function across their diameter. We then, invert these data using the same source-receiver geometry to investigate which parts of the model are best resolved.

The P-wave model recovers the anomalies but they are not quite distinct from each other (Figure 3.4; c-d), meaning that our model suffers from poor lateral resolution between 150 and 350 km depths. We do observe an improved recovery of anomalies at greater depths due to increased lateral extent of the crossing rays (Figure 3.4-e). In cross-section, the P-wave model has better resolution beneath the stations located in the Wopmay orogen (Figure 3.4-f); beneath the central Slave Craton, anomalies are recovered but they are smeared (Figure 3.4-h). The smearing patterns that are observed beneath the central Slave craton show dipping features that are likely due to the effect of the ray geometry of our model.

The S-wave model shows the same general pattern of anomaly recovery, though anomalies have lower amplitude than the P-wave model and they are more laterally smeared. The anomaly recovery beneath the Wopmay Orogen shows a better resolution at shallower depths (see S-wave checkerboard test in the supplementary document, Figure A.1-f). The vertical smearing is similar to that in the P-wave model (Figures 3.4 and A.1).

These plots also reveal the effect of the dominant back-azimuths in teleseismic data. In all cases, resolution is better for structural boundaries oriented in the NE-SW direction (e.g., perpendicular to the POLARIS network) as opposed to NW-SE direction (e.g., perpendicular to the CANOE network). In addition, as depth increases the area with highest resolution moves away from the locations of seismic stations toward the NW and SE, as expected for the greater ray density along these paths.

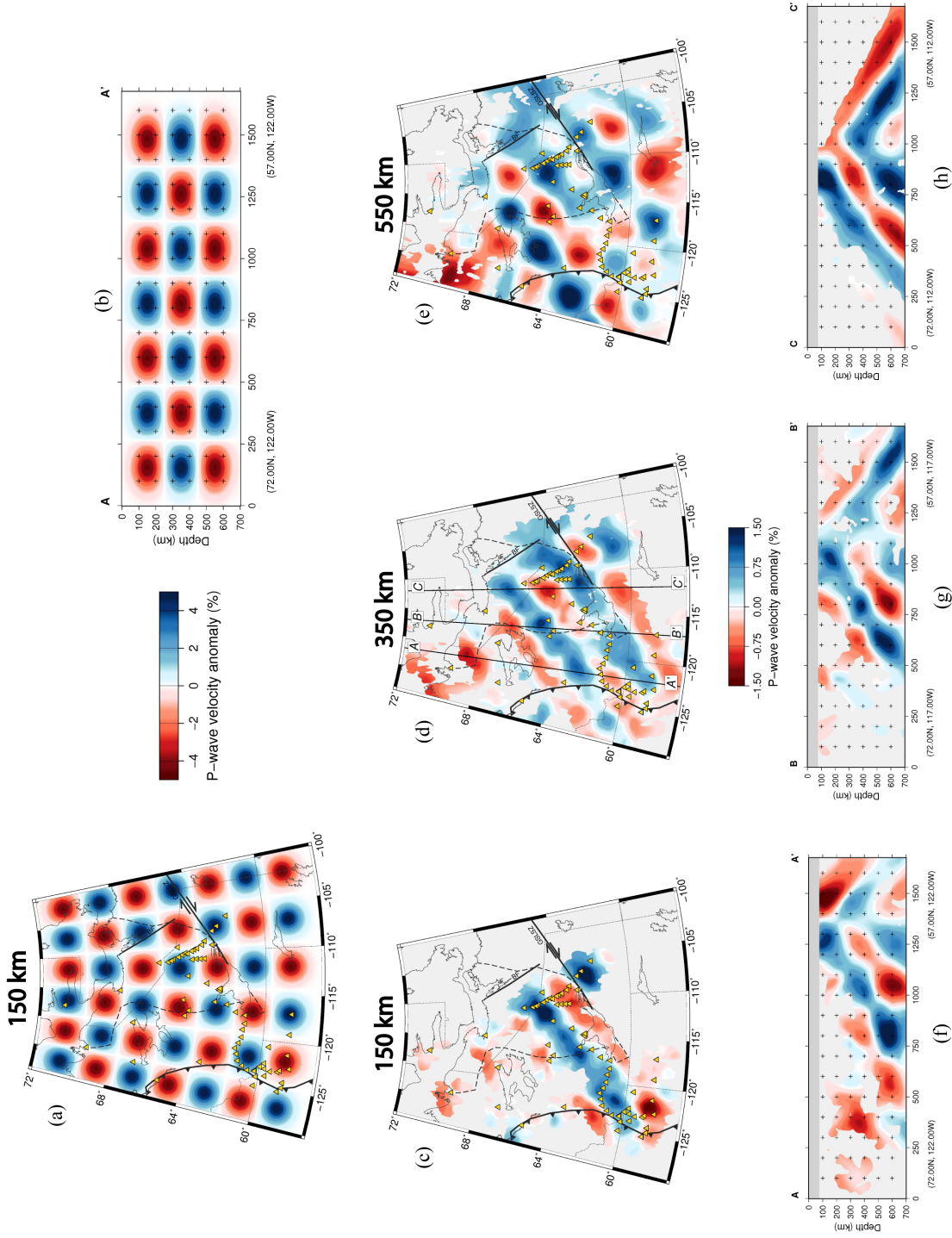


Figure 3.4.: P-wave checkerboard resolution test with spheres of radius 75 km defined by Gaussian functions across their diameter. Each sphere corresponds to an anomaly of $\delta V_p = \pm 5\%$ and they are distributed in successive layers of positive and negative anomalies at depths of 150, 350 and 550 km. (a and b) 150 km depth slice and cross section through the synthetic checkerboard test (input). (c, d and e) Output of the checkerboard resolution test : depth slices at 150, 350 and 550 km depths. (f, g and h) Cross sections through the retrieved checkerboard model (the cross-section locations are displayed on the 350 km depth slice). The Cordillera Deformation front, the Great Slave Lake Shear Zone and the Bathurst fault are shown in black and areas with less than 4 ray hits are displayed in light gray.

3.4.6 Structural tests

We perform a second synthetic test, in order to verify whether observed anomalies of both P- and S-wave models can be resolved or not. Using the same source-receiver geometry, we placed several synthetic anomalies to simulate the final preferred P- and S-wave models. Those anomalies are defined by a Gaussian function across their width, through our tomographic models. In the P-wave structural test model (Figure A.2), we placed a relatively fast region ($\delta V_P = +2\%$) along the NWT/Nunavut border from 65°N 105°W over 900 km at a depth ranging from 100 to 300 km. In addition, we placed another relatively fast region ($\delta V_P = +2\%$) following the strike of the Great Slave Lake shear zone from 62°N 105°W over 700 km at a depth ranging from 150 to 400 km. Also, we placed a series of relative slow anomalies ($\delta V_P = -2\%$) beneath (i) the central Slave craton from 63°N 111°W over 200 km at a depth ranging from 100 to 400 km depth, (ii) the Great Bear Lake and its surroundings from 64.5°N 118°W over 550 km at depth ranging from 100 to 700 km depth, and (iii) to the west of the Cordillera Deformation front from 58°N 126°W over 550 km at depth ranging from 100 to 400 km depth.

The P-wave model recovers the different anomalies (Figure A.2, 250 km depth slice). The slow anomalies underlying the Slave craton and the Great Bear lake region are well retrieved (Figure A.2, transects B-B', G-G' and H-H') with almost no smearing. Recovered amplitudes of the slow anomaly placed west of the CDF are lower than the ones retrieved in the central Slave craton (Figure A.2, transects E-E', G-G'). The fast anomalies are more poorly retrieved and clearly vertically smeared (Figure A.2, transects B-B' and H-H'). This observation is not surprising because of very sparse station coverage in the area and the geometry of the rays.

The input S-wave structural test differs only in the location of a slow anomaly that extends from the central Slave craton to the Great Bear lake region (Figure A.3). The output S-wave model shows that vertical smearing occurs where we placed relatively fast anomalies beneath the northern and southern regions at the edge of our model (Figure A.3, transects B-B' and H-H'). On the other hand, the relatively slow anomalies are well retrieved and less affected by vertical smearing effects except the slow anomaly underlying the Cordillera domain (Figure A.3, transects E-E' and G-G').

3.5 TOMOGRAPHIC MODELS

Here we show depth slices at 150, 250, 350, 450, 550 and 650 km depths and transects through our tomographic models (Figures 3.5 to 3.8). Previous surface-wave tomography

models show that the seismic velocity structure beneath the Slave craton is faster than the global average (e.g., Frederiksen et al., 2001; van der Lee & Frederiksen, 2005; Schaeffer & Lebedev, 2014). In the following sections, "fast" and "slow" P- and/or S-wave anomalies refer to "high" and "low" velocity anomalies relative to the background mean velocity of the study area (Bastow, 2012). Therefore, a slow anomaly observed in our tomographic models is still faster than the global average.

3.5.1 *P-wave model*

The slices at 150 km and 250 km depths reveal a very fast P-wave anomaly ($\delta V_P = \sim 0.8\%$) on the eastern side of the Slave craton (labelled A in Figure 3.5). They also show an alternating pattern of relatively fast and slow P-wave anomalies over a fine scale region, which is consistent with past studies (Bank et al., 2000; Snyder et al., 2004) beneath the POLARIS array (Figure 3.5). This alternating pattern of relatively fast and slow anomalies is also observed on the transects C-C' (Figure 3.6). Directly beneath the Lac de Gras area, we observe a relatively slow P-wave anomaly ($\delta V_P = \sim -0.7\%$) at depths ranging from 150 to 250 km. In cross-section, we observe two distinct slower anomalies (Figure 3.6, transect B-B') dipping to the north and they appear to extend to ~ 300 km depth (labelled B and C).

A relatively slow P-wave anomaly underlying the northwestern region of the Slave craton extends to the Coronation margin. This region remains relatively slow until 650 km depth where the anomaly changes from slow to slightly fast (Figure 3.5). To the North we observe a relatively fast P-wave anomaly and its maximum observed amplitude ($\delta V_P = \sim 0.85\%$) widens and extends southeastward with increasing depth (see 450, 550, 650 km depth slices). This relatively fast P-wave anomaly eventually underlies the northern region of the Slave at 650 km depth (Figure 3.5).

Beneath the Great Bear Lake, the P-wave model displays a relatively slow P-wave anomaly that extends from 75 km to the mantle transition zone (Figure 3.6, transect A-A'). South of the Great Bear Lake, a surprising relatively fast P-wave anomaly (labelled D) underlies the Wopmay Orogen (Figure 3.5, 150, 250 km depth slices). In cross-section, this anomaly is well-defined (Figure 3.6, transect A-A') and extends from 100 km to 300 km depths. It is directly underlain by a relatively slow P-wave anomaly at greater depths. The same relatively fast P-wave anomaly is observed on transect D-D' (Figure 3.6). On both transects (A-A' and D-D'), the transition from slow to fast is laterally sharp and nearly vertical.

South of the Great Slave Lake Shear Zone, within the Rae craton, the P-wave model reveals a relatively fast anomaly (labelled E). This anomaly changes from fast to slow at 550 to 650 km depths. In cross-section, we observe this slow anomaly extending from 500 km to 700 km depth (Figure 3.6, transect D-D'). The Thelon and Talston magmatic arcs are underlain by a relatively fast P-wave anomaly (Figure 3.5). Structural tests revealed that this region is affected by vertical smearing (Figure A.2).

The Cordillera appears to be underlain by relatively slow P-wave anomalies ranging from $\delta V_P = -0.6\%$ to -1% at 150 and 250 km depth (labelled F). In cross-section, the transect D-D' shows a series of relatively slow P-wave anomalies beneath the Cordillera domain that extends to 600 km depth but is also affected by vertical smearing (Figure A.2). The very slow anomaly at 650 km depth is the same one that extends beneath Great Bear Lake (Figure 3.5). We also observe a transition from relatively slow to fast P-wave anomalies in the southwestern part of the study area (Figure 3.5).

3.5.2 *S-wave model*

Our S-wave model displays some differences with the P-wave model. First, the general amplitude of the positive anomalies is less prominent. In contrast to the P-wave model, the S-wave model does show a relatively fast S-wave anomaly but not as fast ($\delta V_S \approx 0.3\%$; feature A), on the eastern side of the Slave craton. The alternating pattern of fast and slow anomalies is not as easy to observe (Figure 3.8, transect C-C'). Moreover, our S-wave model reveals a relatively slow S-wave anomaly in the asthenosphere beneath the Lac de Gras area ($\delta V_S \approx -1.1\%$; Figure 3.8, feature H, transects B-B', C-C' and E-E'). This anomaly is not observed in the P-wave model.

A relatively slow S-wave anomaly extends to the Coronation margin and is much slower than the observed P-wave anomaly at the same location. We also do not observe the change from slow to fast anomaly at 650 km depth. To the north, however, we do observe a relatively fast P-wave anomaly, but due to low ray hit count, it is difficult to tell if the fast S-wave anomaly is displaced southeastward with depth (Figure 3.7). Beneath Great Bear Lake, the S-wave model contains a relatively slow anomaly that extends from 75 km to the mantle transition zone (Figure 3.8, transect A-A').

The same relatively fast S-wave anomaly beneath the Wopmay Orogen is present south of the Great Bear Lake (Figure 3.7, feature D). In cross-section, this fast anomaly extends deeper than in the P-wave model to 400-500 km depth on transect A-A' (Figure 3.8) and extends to the Mantle Transition Zone at ~ 660 km depth on transect D-D'. This difference

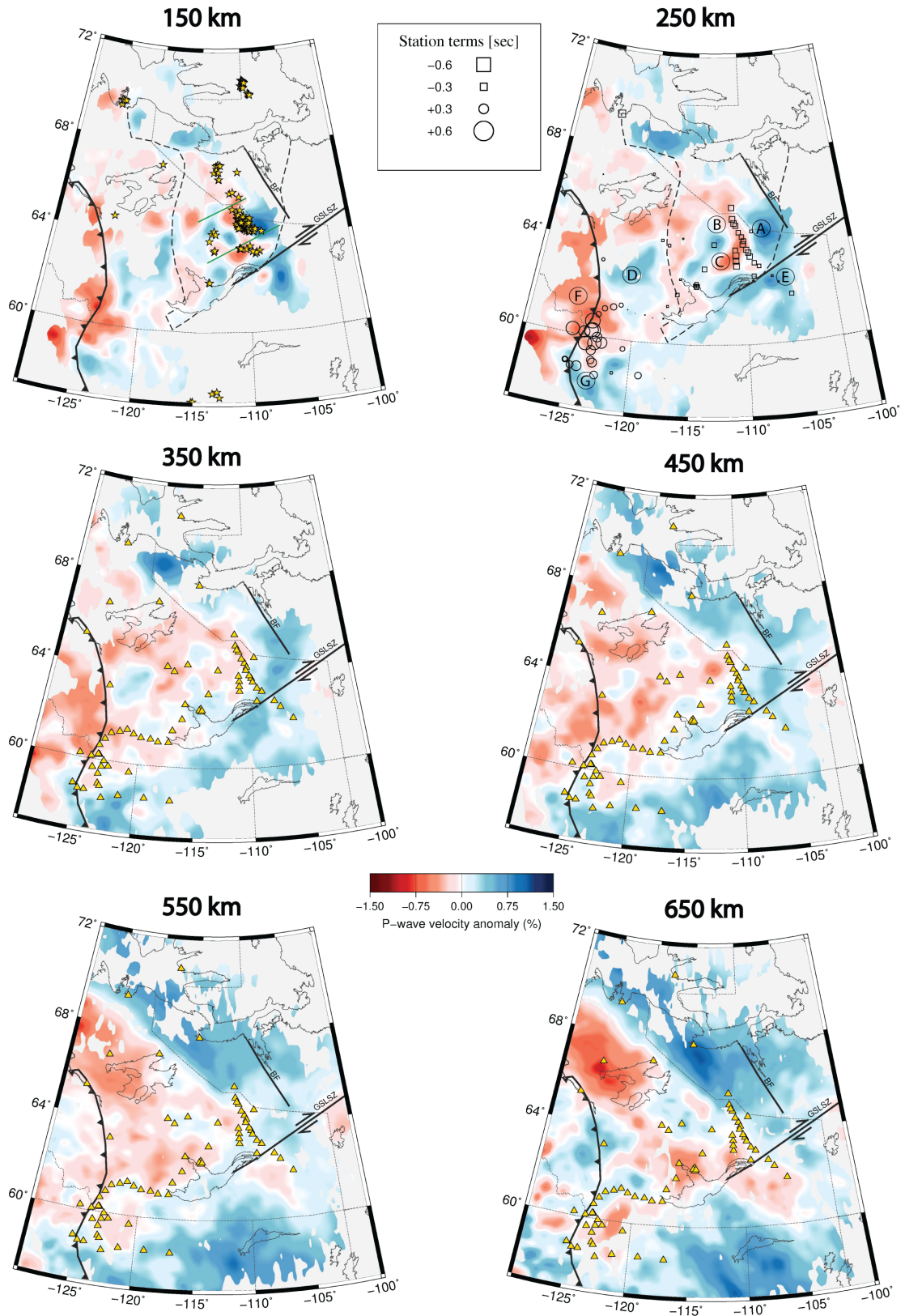


Figure 3.5.: P-wave model depth slices. Superimposed on the 250 km depth slice are the station static correction terms. Yellow triangles are station locations and the black lines indicate the Cordillera Deformation Front, the Great Slave Lake Shear Zone and the Bathurst fault. Yellow stars show the location of kimberlite pipes on the 150 km depth slice. The two green solid lines are geochemical boundaries defined by Grütter et al. (1999). Labels A-G on the 250 km depth slice refer to features specifically mentioned in the text.

3.5 TOMOGRAPHIC MODELS

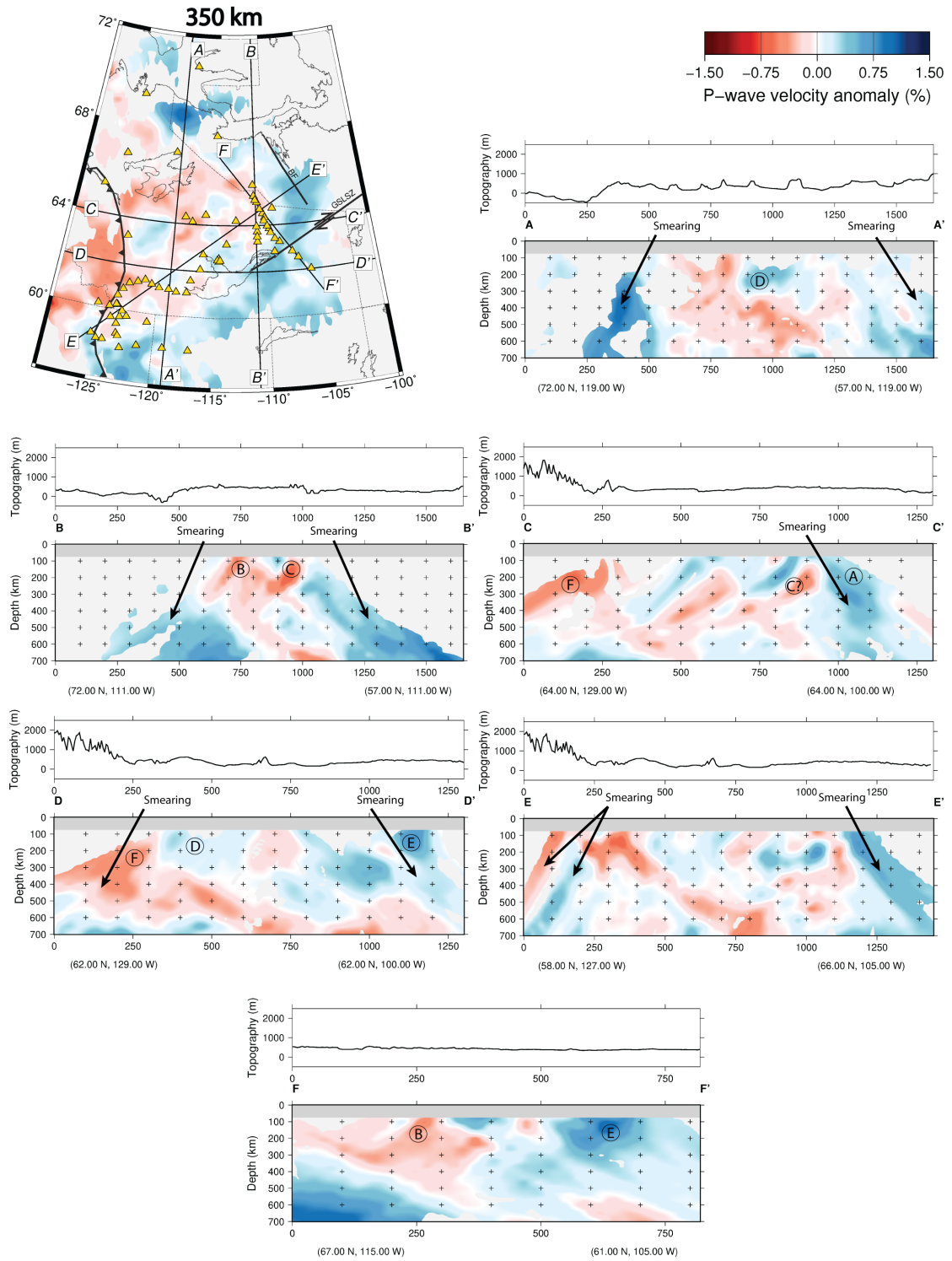


Figure 3.6.: Vertical cross sections through the P-wave model.

with the P-wave model is likely due to vertical smearing (Figure A.3). On both transects (A-A' and D-D'), this transition from slow to fast is still laterally sharp and nearly vertical.

Both P- and S-wave models reveal a fast anomaly ($\delta V_S \approx 0.3\%$; Figure 3.7), south of the Great Slave Lake Shear Zone within the Rae craton (feature E). However, at 550 and 650 km depths, this anomaly changes from fast to slow. In cross-section, we observe this slow anomaly that extends from 450 km to 700 km depth (Figure 3.8, transect D-D'). Both P- and S-wave velocity models show relatively slow seismic velocity anomalies beneath the Cordillera (Figure 3.8, feature F, transects C-C' and D-D') but are vertically smeared.

Some of the differences between the P- and S-wave velocity models can be attributed to the different number of rays used and/or to the geometry of the ray paths, since a direct P-wave does not sample Earth's interior the same way as an SKS core phase or a direct S wave.

The station static terms are plotted on the 250 km depth slice for both P- and S-wave models (Figures 3.5 and 3.7). They show that relatively fast structures (negative times) are associated with the Slave craton whereas relatively slow structures are related to juvenile structures such as the Wopmay Orogen and the Great Bear Arc.

3.6 DISCUSSION

3.6.1 Slave craton

The Slave craton appears to have a complex heterogeneous upper mantle structure compared to other Archean cratons such as the Superior craton (Frederiksen et al., 2007; Boyce et al., 2016). In our tomographic models, the eastern and northern regions of the Slave craton are dominated by faster P- and S-wave velocities, whereas the central Slave craton appears to be relatively slower.

Our P-wave and S-wave models reveal a very singular upper mantle structure beneath the central Slave craton. It displays an alternating pattern of relatively fast and slow anomalies in the eastern and central part of the Slave craton (Figures 3.5 and 3.7). The diamondiferous kimberlites in the Slave craton seem to be concentrated on steep V_P and V_S gradients along the margins of both relatively fast and slow seismic velocities (Figures 3.5 and 3.7; 150 km depth slice). For instance, the highly diamondiferous Lac de Gras field is located on a steep V_P gradient with, on the eastern side, a relatively fast P-wave anomaly ($\delta V_P \approx 0.8\%$) and on the western side a relatively slow P-wave anomaly ($\delta V_P \approx -0.7\%$; Figure 3.5, 150 km depth slice). We computed the lateral velocity gradient through our P- and S-wave models,

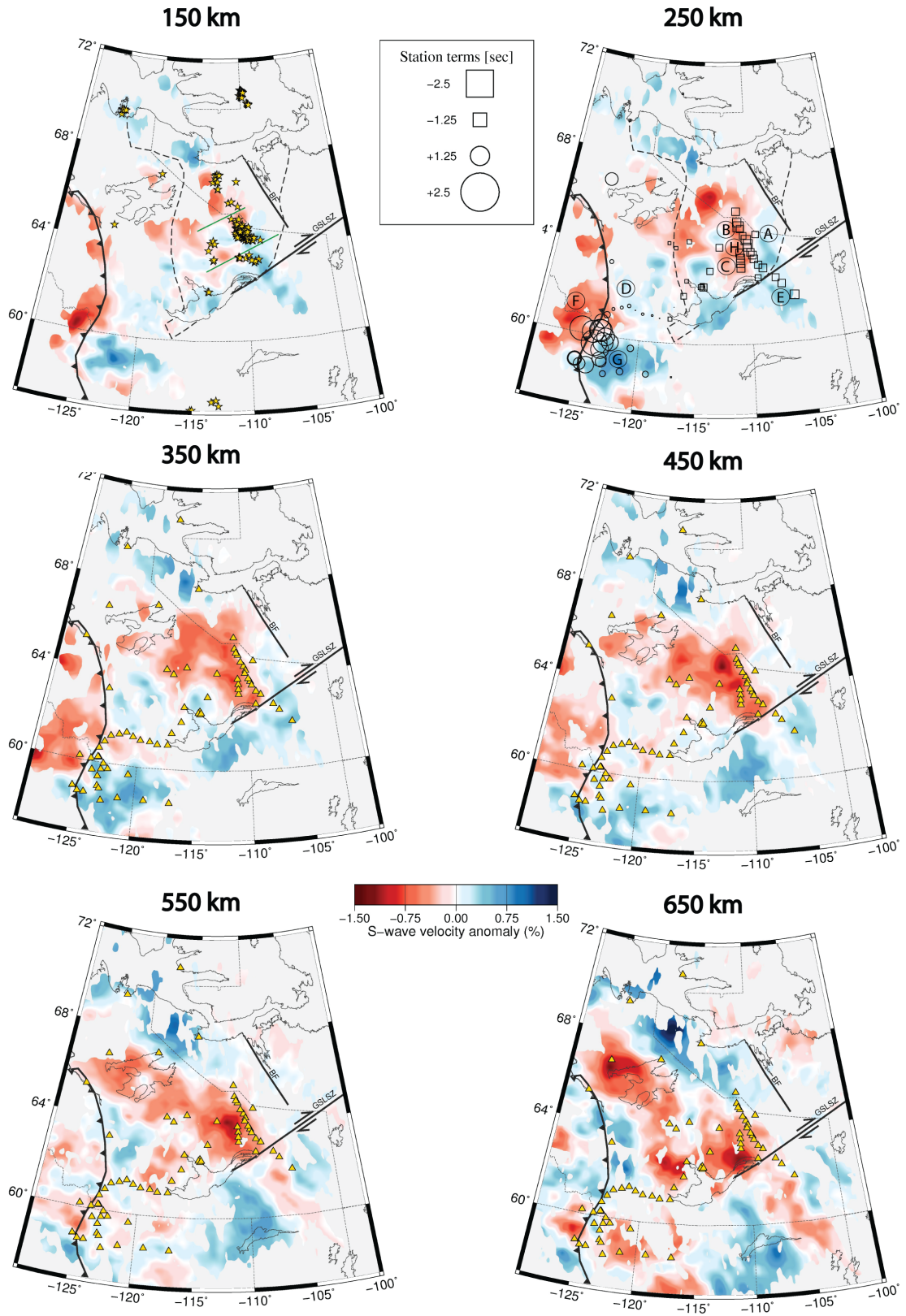


Figure 3.7.: S-wave model depth slices. Format is the same as [Figure 3.5](#).

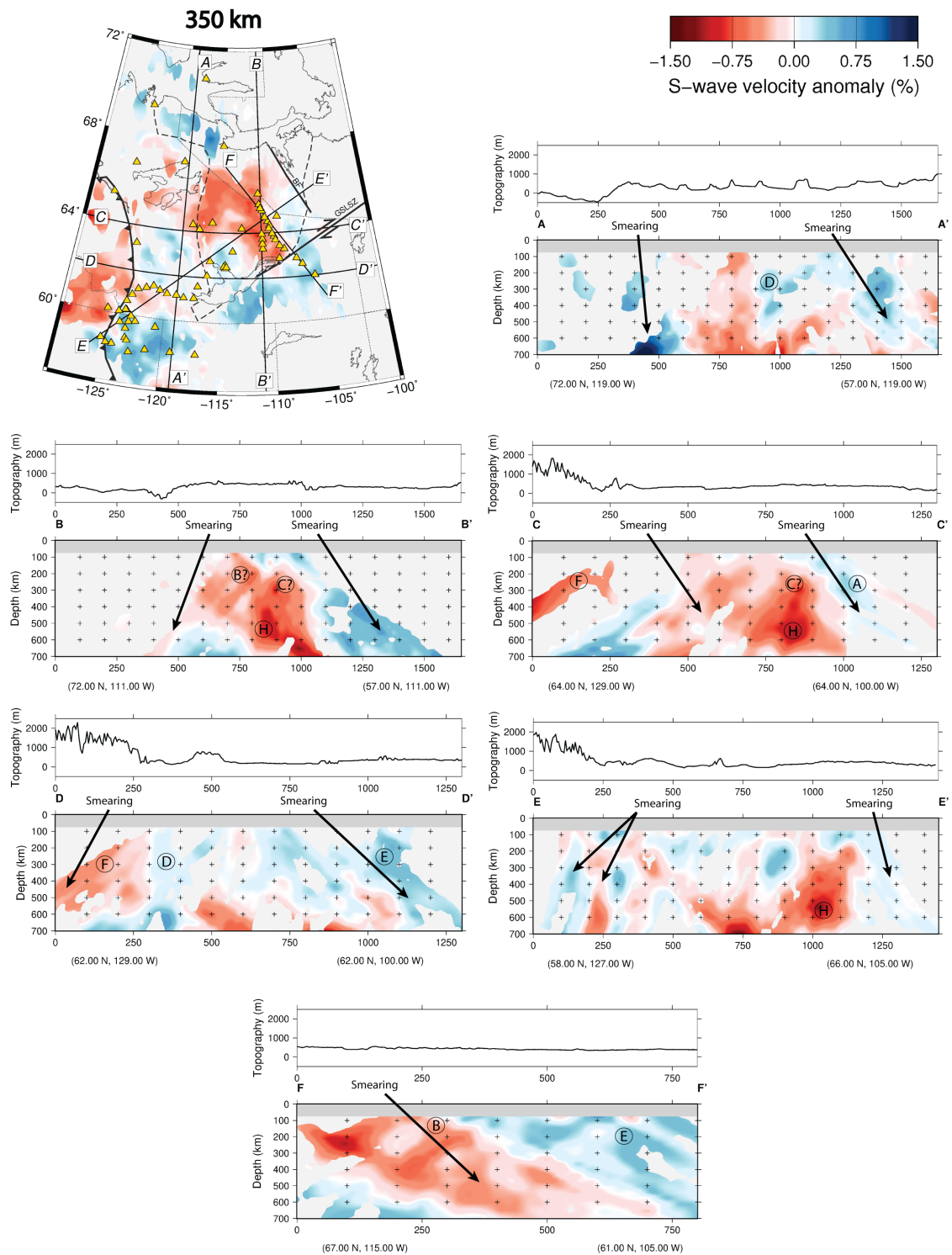


Figure 3.8.: Vertical cross sections through the S-wave model.

and observe that the Lac de Gras kimberlite field is mainly located on the margin of steep V_P gradient at 150 km depth (Figure 3.9). The Lac de Gras kimberlite field also corresponds to the latest eruptions in the Eocene (Figure 3.1). This spatial correlation between steep seismic velocity (V_P or V_S) gradients and locations of kimberlites has already been shown in other Archean cratons such as the Kaapvaal craton in southern Africa (e.g., Begg et al., 2009; Griffin et al., 2009). Nonetheless, this does not appear to be a firm rule, as the western Slave kimberlite field (Figure 3.9) lies within a region where no lateral V_P gradient is observed, at 150 km depth.

Seismic velocity anomalies can be caused by various factors: changes in chemical and/or mineralogical composition (including partial melting), anisotropy of seismic velocity and temperature variations, with the latter factor being likely dominant in the upper mantle. However, temperature may explain seismic velocity variations at global/large-scale in the continental lithosphere but small-scale seismic variations observed in cratonic lithosphere may instead be related to differences in mineralogy or bulk composition (Poupinet et al., 2003). Furthermore, as the kimberlite pipes are located in a cold and old Archean craton, the seismic velocity variations are generally interpreted in terms of composition rather than temperature. Shirey et al. (2002) showed that the variations of P-wave velocities at depth correlate with differences in the composition of diamonds within the diamond stability field beneath the Kaapvaal craton. The diamonds located above a relatively fast P-wave anomaly are dominantly peridotitic and those that sit above a relatively slow P-wave anomaly are dominantly eclogitic. The relatively slow regions are associated with either a higher Fe, Ca, clinopyroxene content or with the presence of metasomatizing fluids that hydrated the surrounding peridotitic mantle. The Lac de Gras area comprises the largest kimberlite cluster in the Slave craton. Cartigny et al. (2009) showed that, of the 85 diamonds from the Panda kimberlite field (Ekati Mine, Lac de Gras area), 73 are dominantly peridotitic, 8 are dominantly eclogitic and 4 are from the lower mantle, which is consistent with the fast P-wave anomaly located beneath the Lac de Gras area. This observation is in agreement with what is observed in southern Africa.

The central Slave craton appears to have some distinct geochemical and geophysical features compared to the rest of the craton (Snyder et al., 2014). Magnetotelluric surveys in the central Slave craton reveal the presence of a conductive layer at depths of 80 to 120 km and striking NE-SW (Jones et al., 2001, 2003). This conductive layer is also co-located with an ultra-depleted harzburgitic layer defined by Grütter et al. (1999) and a seismic discontinuity dipping to the south from 96 to 124 km depth within the central Slave craton (Chen et al., 2009). Unfortunately, this seismic discontinuity is too fine-scale to be imaged by our tomographic model but the lateral extent of this conductive layer is co-located with

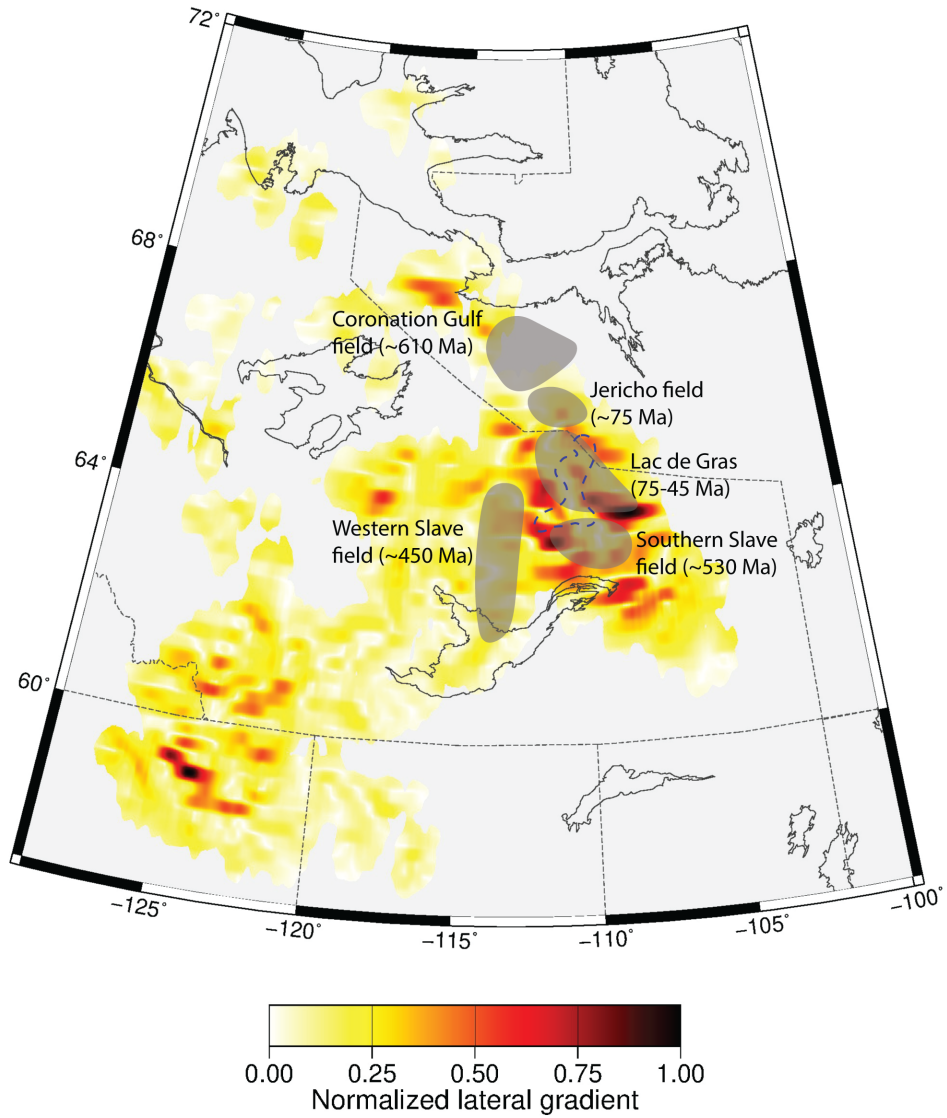


Figure 3.9.: Map showing the normalized lateral velocity gradient computed from our P-wave model at 150 km depth. Light yellow color and red-black color correspond to low and high lateral gradient respectively. Also shown the age distribution of the kimberlite fields in the Slave craton (Lockhart et al., 2004). The lateral extent of the conductive anomaly is outlined in purple (from Chen et al., 2009).

the region, in the central Slave craton, where a very low lateral V_P gradient is observed (Figure 3.9).

Furthermore, our P-wave model reveals two fine structures located in the lithosphere beneath the Lac de Gras kimberlite cluster, with relatively slow anomalies (features B and C) that extend from 75 km to 350 km depth with an apparent dip to the north (Figures 3.5 and 3.6; transect C-C'). This region of the model is well-resolved by the structural test and is therefore less affected by smearing effects (Figure A.2).

Unfortunately, our S-wave model does not image the same lithospheric structures with as much detail (labelled B? - C?). These two anomalies were already imaged in a previous tomography model (Bank et al., 2000) at a lower resolution; these authors interpret them as trapped regions of fertile mantle, which supports the model describing buoyant subduction and imbrication by lithospheric-scale stacks with trapped wedges of fertile mantle (Kusky, 1990). According to several studies, metasomatism occurred within the Slave craton (e.g., Griffin et al., 2003; Pearson et al., 2006; Stachel & Harris, 2008; Aulbach et al., 2011, 2013; Snyder et al., 2014; Eeken et al., 2018). Metasomatism is a direct consequence of fluid circulation and could explain the relatively slow P-wave anomalies but especially the relatively slow S-wave anomalies beneath the central Slave craton (e.g., Eeken et al., 2018). As a consequence, we propose that the observed slow P- and S-wave anomalies in the central Slave craton are related to metasomatised regions within the lithosphere.

Within the asthenosphere (i.e., between ~250 and 410 km depth), our S-wave model reveals a relatively slow anomaly (Figures 3.7 and 3.8; feature H, transect C-C') whose origin remains unclear. We note however that the location of this anomaly is spatially correlated with the Lac de Gras kimberlite field at the surface. It is worth noting that the last kimberlite eruptions occurred between 75 and 45 Ma within the central Slave craton (Heaman et al., 2003); it appears very unlikely that this slow anomaly sustained over millions of years within the asthenosphere beneath the cold and stable lithospheric root of the craton. Therefore, we assume that the observed slow anomaly within the asthenosphere cannot explain the actual Lac de Gras kimberlite cluster, but we suggest that the slow anomaly may be related to a potential kimberlite magma ascent in the asthenosphere. It has been shown that the underlying stratified upper mantle beneath the Slave craton is also observed in other regions (Gawler craton - Australia O'Reilly & Griffin, 2010) where ultra-deep diamond inclusions have been reported. It was suggested that the combination of both a stratified SCLM and the presence of ultra-deep diamonds may indicate that the lithospheric upper mantle beneath the Slave craton was formed by plume subcretion (O'Reilly & Griffin, 2010).

Within the Slave craton, Grütter et al. (1999) proposed a classification of the craton in three domains based on geochemical composition. The northern domain has a high propor-

tion of mantle eclogite, the central domain contains an ultra-depleted layer of harzburgite and the southeastern Slave craton has a lower proportion of mantle eclogites. Our tomographic model does not show any strong correlation with these three geochemical domains, implying that mineralogical composition is not a factor explaining the observed anomalies in the region. Previous body-wave tomographic models did not show any correlation either (Bank et al., 2000; Snyder et al., 2004). Shirey et al. (2002) showed that body wave velocities correlated with age as well as peridotite/eclogite petrogenesis of diamonds, where the Proterozoic Karoo has slower wave speeds than either the Kaapvaal or Zimbabwe cratons. We note that the Slave craton has slower S-wave velocities than the Proterozoic provinces surrounding it, in contrast to the findings by Shirey et al. (2002). We do not observe a contrast in P-wave velocities between the Slave and Proterozoic provinces.

Both P-wave and S-wave models show that the linear feature separating fast and slow velocities associated with the Great Slave Lake shear zone continues to the SW into BC at 250-450 km depths (Figures 3.5 and 3.7). This would be consistent with some previous tectonic interpretations (Hoffman, 1988) if the Archean Rae lithosphere lies to the south of this apparent boundary with Proterozoic Hottah/Ft. Simpson lithosphere lying to the north, perhaps in a former embayment of the shield margin (Cook et al., 1999). However, this interpretation would be in contradiction with the observation that the Archean lithosphere has slower seismic velocities compared to Proterozoic lithosphere. Note also that some recent studies argue that the Great Slave Lake Shear Zone is not vertical to great depths; the most recent strand (MacDonald fault) is vertical to about 15 km depth (Snyder et al., 2014).

3.6.2 Cordillera

Archean cratons are characterized by an underlying thicker and colder lithosphere than more juvenile mantle beneath younger geological formations. In Canada, the seismic velocity gradient from the Phanerozoic lithosphere beneath the Cordillera to the colder lithosphere beneath the Slave craton is strong and sharp. This sharp transition zone was imaged by regional-scale and global-scale surface wave tomography studies (e.g., Frederiksen et al., 2001; Schaeffer & Lebedev, 2014; Bao & Eaton, 2015; McLellan et al., 2018; Zaporozan et al., 2018) and a regional-scale body-wave tomography model (Mercier et al., 2009). This seismic transition was documented to follow the Cordillera Deformation Front (Figure 3.1). Our P- and S-wave tomography models show a relatively slow P- and S-wave anomaly (feature F) beneath the Cordillera juxtaposed against a relatively fast P- and S-wave anomaly (feature D) beneath the western margin of the craton (Figures 3.6 and 3.8 transect D-D').

This sharp transition seems to be nearly vertical on transect F-F' (Figures 3.6 and 3.8) and also coincides with rapid variations in topography at the surface. This sharp transition has been previously noted by a past body-wave tomography study (Mercier et al., 2009) and was suggested that this transition corresponds to the boundary between the Phanerozoic lithosphere and the Proterozoic lithosphere. We note however that the sharp transition is not as evident near the Great Bear Lake, where resolution is lower due to sparse station coverage.

3.6.3 Lithospheric thickness

The lithospheric mantle (<300 km depth) beneath cratonic regions is generally defined by fast seismic velocities, low geothermal gradient and chemically depleted composition (Artemieva & Mooney, 2001; Artemieva, 2009). However, questions remain about the nature and depth to the base of the lithosphere and the transition to the asthenosphere, commonly known as the LAB. The thickness of the lithosphere can be imaged using a variety of observables, including seismic velocities, heat flow, electrical conductivity and xenolith data (e.g., Jones et al., 2003; Kopylova & Caro, 2004; van der Lee & Frederiksen, 2005). For example, in North America, van der Lee & Frederiksen (2005) suggest that the average lithospheric thickness is ~250 km based on their global-scale surface-wave tomographic model. Similar depths are estimated beneath the Slave craton (Bank et al., 2000) and the Kaapvaal craton in southern Africa (Fouch et al., 2004). More recently, Chen et al. (2007) retrieved the shear-wave velocity from POLARIS and Yellowknife array data by inversion of Rayleigh waves. The authors show that the cratonic root beneath the Slave craton extends to 220 km depth. However, Schaeffer & Lebedev (2014) pointed out that estimates based on seismic tomographic observations have the potential to be biased due to vertical smoothing leading to an over-estimation of the lithosphere thickness.

Our P- and S-wave models do not show any evidence for the presence of a sharp LAB beneath the Slave craton, mainly due to large vertical smearing and low sensitivity of near vertically-propagating body-waves to horizontal boundaries. Based on receiver function observations across Australia, Ford et al. (2010) show that the definition of the LAB is more robust beneath the Eastern Phanerozoic region of Australia than in cratonic Australia. Furthermore, the inferred LAB is not consistently observed beneath the Slave craton (Schaeffer & Bostock, 2010; Snyder et al., 2014).

3.7 CONCLUSION

By inverting a large data set of P- and S-wave relative delay times, we develop robust teleseismic P and S body wave tomography models over the Slave craton region with the most complete coverage to date. The eastern and northern regions of the Slave craton are dominated by faster P- and S-wave velocities, whereas the central Slave craton appears relatively slower. The P-wave model reveals an alternating pattern of positive and negative anomalies over a broad region within the central Slave craton co-located with the Lac de Gras kimberlite cluster. The diamondiferous kimberlites in the central Slave craton are concentrated on steep V_P and V_S gradients along the margins of both relatively fast and slow seismic velocities. Also, the P-wave model reveals two fine structures located within the lithosphere beneath the Lac de Gras kimberlite cluster, with relatively slow anomalies that extend from 75 km to 350 km depth with an apparent dip to the north. These relatively slow P- and S-wave anomalies are associated with metasomatised regions of the lithosphere. The S-wave model shows a pervasive slow anomaly beneath the central Slave craton in the asthenosphere. This anomaly cannot be related to the kimberlite pipes at the surface due to the timing of the last kimberlite eruptions (i.e., 75-45 Ma) and absolute tectonic motion. Its origin remains unclear, but may be related to a potential kimberlite magma ascent in the asthenosphere.

Our tomographic models are based on a good earthquake epicentral distribution but are still dominated by sources in the northwestern Pacific and western South America. Unfortunately, because of the absence of major plate boundaries in the Pacific Ocean and beneath Africa, there are not enough earthquakes that sample the Earth's upper mantle in this direction and as a result, structures following this orientation are not well imaged by our models. In order to improve our future models, the deployment of seismic stations is required in the southeastern and northeastern regions of the study area, which would fill some major gaps and allow us to better constrain the upper mantle structure beneath the Slave craton.

THE UPPER MANTLE STRUCTURE OF NORTHWESTERN CANADA FROM TELESEISMIC BODY-WAVE TOMOGRAPHY

4.1 SUMMARY

The Northern Canadian Cordillera (NCC) is an actively deforming orogenic belt in northwestern Canada. Geochemical and geophysical data show that the NCC is underlain by a thin and hot lithosphere, in contrast with the adjacent cold and thick cratonic lithosphere to the east. This juxtaposition of cold/hot and thick/thin lithosphere across a narrow transition zone has important implications for regional geodynamics. The recent deployment of US-Array Transportable Array and other seismic stations across Alaska, USA and northwestern Canada, allows us to image lithosphere and upper mantle three-dimensional seismic velocity structure at significantly improved resolution. Our model reveals a broad high-velocity anomaly across northern Yukon and Northwest Territories, which is interpreted as buried cratonic lithosphere, and which we refer to as the Mackenzie craton. Another prominent high-velocity anomaly is imaged beneath northeastern British Columbia and is interpreted to indicate cratonic lithosphere beneath the Northern Rocky Mountains. These two mechanically strong lithospheric blocks, also suggested by regional magnetic data, are interpreted to buttress the ends of the Mackenzie Mountains fold and thrust belt, guiding intervening cordilleran mantle flow toward the Canadian Shield and controlling the arcuate geometry of the Mackenzie Mountains fold and thrust belt. Both P- and S-wave models also reveal the signature of a northward-dipping, subducting Wrangell slab across the southern region of the Alaska/Yukon border. Strong P- and S-wave velocity contrasts across the Tintina Fault

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suggest that it is a lithosphere-scale shear zone that extends into the upper mantle beneath the NCC and demarcates distinct regions of lithospheric mantle.

4.2 INTRODUCTION

The Northern Canadian Cordillera (NCC) of northwestern Canada is an accretionary orogen consisting of a succession of terranes accreted to the Laurentian passive margin. The NCC started to develop with the break-up of Rodinia at ~ 750 Ma (Johnston, 2008; Monger & Price, 2002; Nelson & Colpron, 2007; Nelson et al., 2013), which led to the formation of a passive intraplate continental margin on western Laurentia. During the middle Devonian, the passive margin was converted to a convergent margin with associated arc and back-arc settings, which eventually accreted and formed the Intermontane domain (Nelson & Colpron, 2007). A transpressive boundary persisted up to present time favoring the amalgamation of terranes on the margin. Based on correlations made from geological and geochemical observations and fossil faunas, the terranes forming the metamorphic hinterland of the NCC, such as the Yukon Tanana and Stikinia terranes, are thought to be the product of a rifting episode that occurred on the margin of Laurentia between late Paleozoic and early Mesozoic (see Fig. 1 ; Colpron & Nelson, 2009).

The Cordillera is divided in five domains (see Fig. 1 in, Nelson & Colpron, 2007; Nelson et al., 2013) or morphogeological belts (Monger & Price, 2002) reflecting tectonic deformation over ~ 750 Myr. From east to west, these five domains are described as follows: Ancestral North America (Laurentia) consisting of autochthonous and parautochthonous rocks of Laurentia and its miogeoclinal sedimentary cover; Peri-Laurentian (Intermontane), which consists of allochthonous peri-cratonic rocks; Insular and northern Alaska occupying more out-board positions in the Canadian Cordillera and sharing Palaeozoic and older affinities with northern Baltica, Siberia and the northern Caledonides (Colpron & Nelson, 2009); and Out-board terranes representing later Mesozoic and Cenozoic terranes along the present North American plate margin.

During the accretion of Cenozoic terranes (~ 50 Ma), prominent margin-parallel, right-lateral strike-slip faults such as the Denali and the Tintina Faults accommodated most of the oblique relative motion between the Pacific and North American plates. The Tintina Fault marks a principal geological boundary between autochthonous rocks of Laurentia and its passive margin and the allochthonous rocks of the Intermontane super-terrane (Nelson & Colpron, 2007). Approximately 400 - 430 km of cumulative displacement occurred during the Paleocene on the Tintina Fault (Gabrielse et al., 2006; Saltus & Hudson, 2007) and 370 km on the Denali Fault since the Early Cretaceous (Lowey, 1998).

A block of thick oceanic crust, the Yakutat block, detached from the western margin of North America during a rifting episode at ~ 50 Ma (Bauer et al., 2014; Pavlis et al., 2004). It migrated northward along the Queen Charlotte-Fairweather fault system and eventually

impinged obliquely on the North American plate in the Gulf of Alaska. The block was partially subducted beneath the North American plate, triggering volcanism in the Wrangell volcanic field. [Finzel et al. \(2011\)](#) suggest that the western part of the Yakutat block started colliding at 43 Ma and stopped around 23 Ma but the eastern part started colliding at 10 Ma and continues to present, driving the exceptional high-elevation relief of the St. Elias-Chugach mountain ranges. This collision zone presents one of the highest deformation rates in North America ([Marechal et al., 2015](#); [Pavlis et al., 2019a](#)).

Seismicity across the NCC is mainly focused around the Yakutat collision zone but also 800 km inland within the Mackenzie Mountains fold and thrust belt and in the Richardson Mountains further north. However, the region between the Tintina and Denali Faults exhibit a lower seismic rate ([Mazzotti et al., 2008](#); [Leonard et al., 2008](#)). GPS data and seismicity rates suggest a current shortening rate of 1 to 5 mm/yr across the Mackenzie Mountains thrust belt ([Leonard et al., 2007](#); [Mazzotti et al., 2008](#)). Further north, the Richardson Mountains exhibit a right-lateral strike-slip displacement rate of 5 ± 2.5 mm/yr with moderate seismic activity ([Mazzotti & Hyndman, 2002](#); [Hyndman et al., 2005a,b](#)). South of the Mackenzie Mountains, the Liard transfer zone (LTZ) is inherited from an asymmetric extensional rifting episode that occurred in the Neoproterozoic between 709 and 650 Ma on the western margin of Laurentia ([Lund, 2008](#)). Modern seismicity is observed well north of the LTZ but is almost absent in the south. [Mazzotti & Hyndman \(2002\)](#) explain this pattern of deformation through an orogenic float model whereby Yakutat-North America collision stress is propagated across the NCC as far inland as the Cordilleran Deformation Front (CDF). The stress transfer in this hypothesis is facilitated by a weak decollement in the lower crust produced by elevated temperatures ([Hyndman, 2017](#)). This activates thrust faults in the Mackenzie Mountains and generates associated seismicity. However, this hypothesis does not take into account any role for Denali and Tintina Faults and does not explain why the modern pattern of seismicity across the NCC is segmented.

A variety of geophysical studies indicate that the NCC is underlain by a flat ([Cook et al., 2004](#); [Cook & Erdmer, 2005](#)) and shallow Moho at 32 ± 2 km depth ([Tarayoun et al., 2017](#)), and crust with a similarly shallow Curie depth ([Gaudreau et al., 2019](#)). The region has high heat flow (≥ 100 mW/m²; [Lewis et al., 2003](#)), high temperatures at the Moho ($\sim 900^\circ\text{C}$) in contrast with $\sim 400^\circ\text{C}$ in the Canadian Shield ([Lewis et al., 2003](#); [Audet et al., 2019](#)) and < 20 km effective elastic thickness ([Audet et al., 2007](#); [Flück et al., 2003](#)). Several magnetotelluric (MT) studies have been conducted over the NCC, in particular spanning the Tintina Fault (i.e., [Ledo et al., 2002, 2004](#); [Jones et al., 2005](#)) that reveal complex resistivity variations throughout the NCC. An interesting feature displayed by the MT models is a highly resistive crustal anomaly beneath the Tintina Fault along two transects (Transects 2

and 3 in Ledo et al., 2004; Jones et al., 2005; Dehkordi et al., 2019). This anomaly was interpreted to indicate a lack of circulating fluids or precipitation of graphite in the mid to lower crust (Ledo et al., 2004). This interpretation is consistent with the lack of seismic reflectors beneath the Tintina Fault, in the crust, observed in SNORCLE (Slave Northern Cordillera Lithospheric Evolution; Cook et al., 2004) seismic profiles. Jones et al. (2005) suggest that the lack of seismic reflectors and the high resistivity anomaly imply that the Tintina Fault is "healed" and thus no longer a zone of crustal weakness. However, this is inconsistent with the seismic activity exhibited along the fault zone. Recently, Dehkordi et al. (2019) reprocessed MT data in the vicinity of the Tintina Fault and produced an image of a NE dipping high resistivity anomaly extending to 40 km depth, consistent with past MT studies.

Seismic anisotropy from teleseismic shear-wave splitting shows that azimuthal polarization directions that are parallel to the Tintina and the Denali Faults, which may be a manifestation of the long-term shearing that has occurred along these fault systems (Rasendra et al., 2014; Tarayoun et al., 2017; Witt et al., 2019). Furthermore, fast-axis orientations remain parallel to the faults at uppermost mantle depths suggesting that the Tintina and the Denali Faults shear features extend well into the NCC lithosphere (Audet et al., 2016; McLellan et al., 2018), but such studies do not yet fully incorporate seismic data from the EarthScope USArray-Transportable Array stations across Alaska and Yukon. Seismic tomography at various resolution scales generally reveals low seismic velocities within the Cordillera lithosphere transitioning to high velocities within the cratonic lithosphere across a boundary roughly coinciding with the surface expression of the Rocky Mountain Trench, which is the southern extension of the Tintina Fault in the Southern Canadian Cordillera (Frederiksen et al., 1998, 2001; van der Lee & Frederiksen, 2005; Bedle & van der Lee, 2009; Mercier et al., 2009; Schaeffer & Lebedev, 2014; Bao et al., 2014; Chen et al., 2018; McLellan et al., 2018; Zaporozan et al., 2018). Regional seismic tomography studies conducted over Alaska, USA show numerous details on the underlying lithosphere. In particular, coherent high phase velocity anomalies (Ward, 2015) and fast shear wave velocities (Jiang et al., 2018; Ward & Lin, 2018) are well defined north of the Tintina Fault at the Alaska/Yukon border.

Important questions remain regarding the nature and the origin of the lithospheric mantle underlying the NCC and past and present geodynamics of the region (e.g., Boggs et al., 2018). The recent deployment of the USArray-TA and other seismic stations across Alaska, USA and Yukon and Northwest Territories, Canada, now allow for higher resolution three-dimensional seismic velocity models of upper mantle structure. We perform a tomographic inversion of teleseismic P- and S-wave relative delay times to obtain three-dimensional seismic velocity models of upper mantle structure (from 75 to 600 km depth). This method has lower depth

resolution but excellent lateral resolution compared to surface-wave tomographic inversions. Here, however, the method is mainly sensitive to large and sharp lateral contrasts in seismic velocities, which are difficult to reconcile with a purely thermal effect and therefore also involve changes in composition. We then discuss the potential relations with surface geology and the role of upper mantle structure in the geologic history and ongoing deformation of the NCC.

4.3 METHOD

4.3.1 *Data*

Our study area is covered by 320 broadband seismic stations of the following permanent or temporary seismic networks (Figure 4.1): the Alaska Regional Network (AK), the National Tsunami Warning Center Alaska Seismic Network (AT), the Alaska Volcano Observatory (AV), the Canadian Northwest Seismic Experiment (CANOE, network code XN), the Canadian National Seismograph Network (CNSN, network code CN), the International Miscellaneous Stations (IM), the Yukon-Northwest Seismic Network (YNSN, network code NY), the Portable Observatories for Lithospheric Analysis and Research Investigating Seismicity (POLARIS, network code PO), the Earthscope-USArray Transportable Array (TA), the Collaborative Research: Relating glacier-generated seismicity to ice motion, basal processes and iceberg calving (network code XF), the St. Elias Erosion/Tectonics Project (STEEP, network code XZ), the Yukon Observatory (YO), the Dynamics of Lake-Calving Glaciers: Yakutat Glacier, Alaska (5C) and the Earthscope-USArray Flexarray Mackenzie Mountains Transect (network code 7C; Baker et al., 2019). Seismic data were obtained from the Incorporated Research Institution for Seismology (IRIS) and the Canadian Hazard Information Service (CHIS). Our P-wave data set consists of 127,138 delay times from 2,627 earthquakes with magnitude $M_W \geq 5.5$ and great circle angular distance (Δ) $30^\circ \leq \Delta \leq 90^\circ$ that occurred between 2000 and 2018 (Figure 4.1). Our S-wave data set consists of 38,924 direct S-wave delay times and 20,183 SKS core-refracted phase delay times from 1,488 teleseismic earthquakes, respectively. Our total S-wave data set therefore consists of 59,107 delay times.

4.3.2 *Travel time measurements*

Travel times are measured at all stations that commonly record an identified teleseismic earthquake. Preliminary P-wave, S-wave and SKS arrival times are predicted from a 1-D

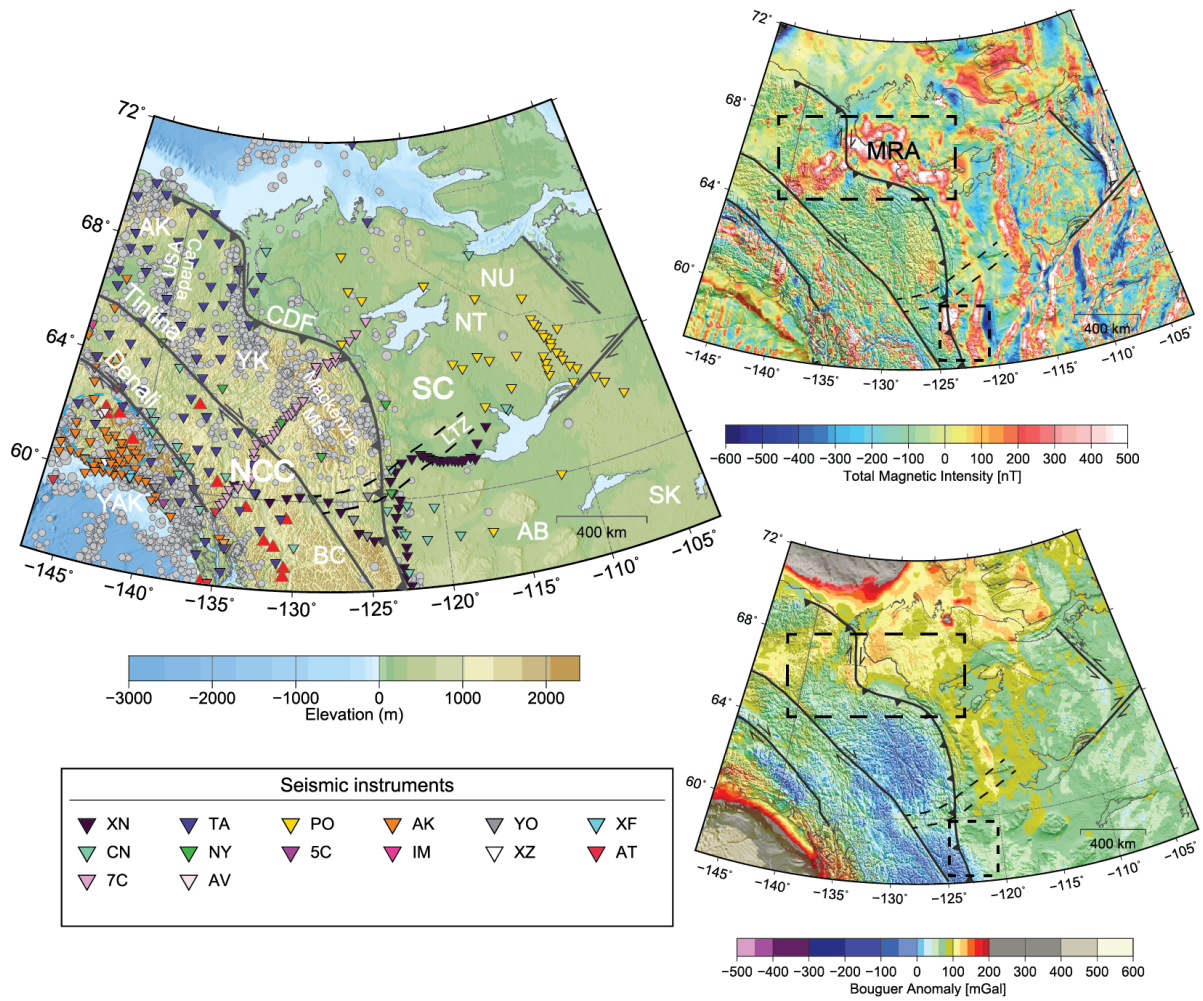


Figure 4.1.: (Left) Study area map showing the seismic station locations (inverted triangles) and each station is color-coded by seismic network. Also shown are volcano locations (red triangles), main tectonic features and seismicity ($M > 2.0$) between 2000 and 2019. The Wrangell volcanic field is contoured by a blue dashed line. (Top Right) Magnetic anomaly map obtained from NAMAG (Bankey et al., 2002). (Bottom Right) Complete spherical Bouguer gravity anomaly map ($2' \times 2'$ resolution) obtained from the WGM2012 global model (Bonvalot et al., 2012). Dash boxes show the locations of the magnetic and Bouguer anomalies discussed in the text. Abbreviations: AB, Alberta; AK, Alaska; BC, British Columbia; NT, Northwest Territories; NU, Nunavut; SK, Saskatchewan; YK, Yukon; CDF, Cordilleran Deformation Front; LTZ, Liard Transfer Zone; MRA, Mackenzie River magnetic Anomaly; NCC, Northern Canadian Cordillera; SC, Slave Craton; YAK, Yakutat

reference velocity model, IASP91 (Kennett & Engdahl, 1991). We filter and taper P-wave vertical component seismograms with a non-zero phase, 2-pole Butterworth bandpass filter with corner frequencies of 0.4 Hz and 2.0 Hz. Direct S-wave transverse component and SKS-wave radial component seismograms are filtered similarly with corner frequencies of 0.02 and 0.4 Hz.

Manual picking of the first peak or trough of seismic wave energy (i.e., P-, S- or SKS-wave) is performed on each seismogram using the Automated and Interactive Measurement of Body wave Arrival Times software (hereafter referred to as AIMBAT) developed by Lou et al. (2013). AIMBAT pre-aligns the seismograms before using the Multi Channel Cross-Correlation (MCCC). Then each recording station pair i, j for a given earthquake, is cross-correlated and a relative delay time is determined (VanDecar & Crosson, 1990). Least-squares optimization is applied on the set of cross-correlation derived delay times, where the mean is forced to be zero. The final delay times are calculated by subtracting the demeaned predicted relative arrival times from IASP91. The final delay time is defined as

$$\delta t = t_i^{est} - (t_{0i} - \bar{t}_{net}) \quad (4.1)$$

where t_i^{est} is the least-squared optimized delay time for each station i (i.e., demeaned relative travel times among the subset of stations recording the same earthquake), and $t_{0i} - \bar{t}_{net}$ is the demeaned predicted arrival time for the i^{th} station based on IASP91.

The MCCC method provides a way to assess the uncertainty of the relative delay times. The standard deviation of the relative delay times associated with the i^{th} trace is

$$\sigma_i = \sqrt{\frac{1}{n-2} \sum_{j \neq i} \delta t_{ij}^2}. \quad (4.2)$$

The network-averaged standard deviations are 0.03 s and 0.09 s for the P- and S-wave delay time data sets, respectively (Figure 4.2). Similar delay-time errors are found in other studies (Boyce et al., 2016; Bollmann et al., 2019). Figure 4.2 shows the average relative station delay times for both P- and S-wave data sets. Both P- and S-wave delay-time maps show a clear concordance. Stations located in the Slave craton and its surroundings appear to be dominated by negative relative delay times whereas the Cordillera is clearly dominated by positive delay times with some noticeable variations: (i) the Yukon-Tanana Upland, which is part of the Ancestral North America super-terrane (as defined by Nelson & Colpron, 2007), shows negative delay times, (ii) northern Yukon is marked by P-wave delay times around 0 but negative S-wave delay times, (iii) and the very southwestern corner of our study area is

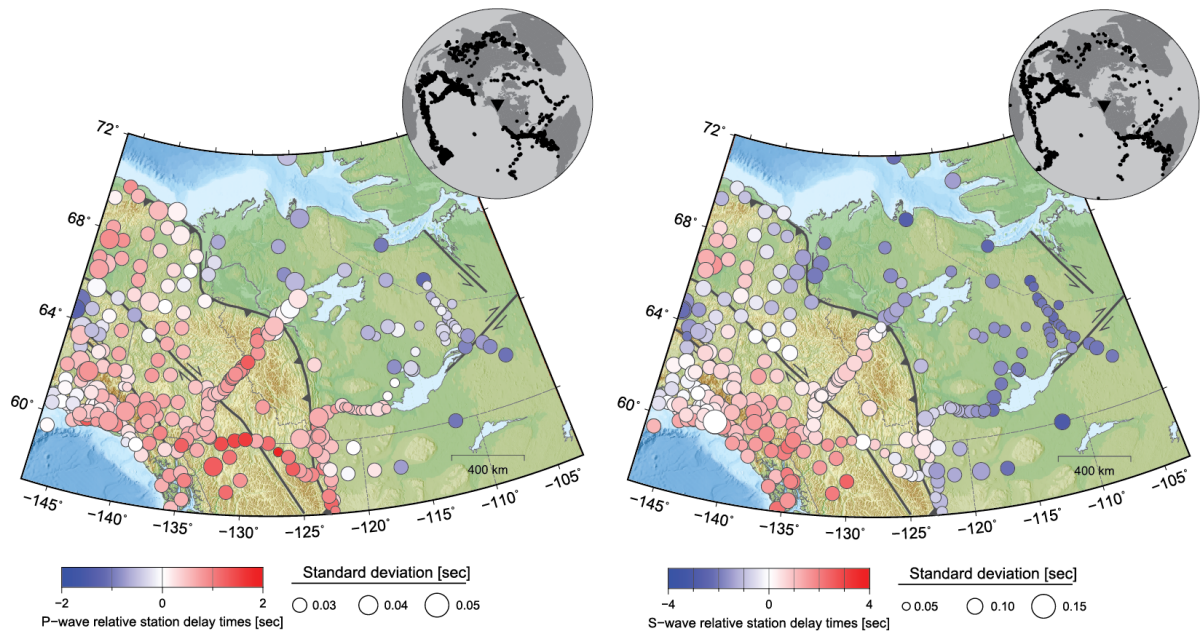


Figure 4.2.: P (left) and S (right) average relative station delay times. Blue and red circles denote early (high relative velocity) and late (low relative velocity) arrivals, respectively. The size of the circle corresponds to the average standard deviation in seconds calculated from Equation. 2. Also shown is the equidistant azimuthal projection displaying the distribution of $M_W \geq 5.5$ earthquakes for P and S recorded between 2000 and 2018 at one or more stations of the array.

marked by negligible delay times. The eastern branch of the CANOE network (Figure 4.1) shows positive P-wave delay times and negative S-wave delay times which is consistent with the range of V_P/V_S values (1.75-2.0) obtained from $H - \kappa$ technique for the same stations (see Fig. 7, Tarayoun et al., 2017). Stations on the northeastern end of the Mackenzie Mountains transect (Figure 4.1) show negative S-wave delay times and mark a transition with the observed positive delay times in the Cordillera but this observation does not appear as clearly for the P-wave data set.

4.3.3 Inversion procedure

We use VanDecar's inversion procedure to resolve the three-dimensional isotropic velocity structure underlying northwestern Canada (VanDecar et al., 1995). This tomographic software has been applied to a variety of tectonic settings such as subduction zones (e.g., Bostock & VanDecar, 1995), cratonic regions (e.g., Frederiksen et al., 2007; Mercier et al., 2009; Villemaire et al., 2012; Bastow et al., 2013; Boyce et al., 2016; Esteve et al., 2019; Bollmann et al., 2019), and rift settings (e.g., Bastow et al., 2008; White-Gaynor et al., 2019). Relative delay times are inverted simultaneously for slowness perturbations, station static terms and source terms. The formulation of the inverse problem is given as

$$P_{ij}^{k-1} \Delta s_j^{(k)} + h_r^{(k)} + e_q^{(k)} = \Delta t_i^{(k-1)}, \quad (4.3)$$

where $P_{ij}^{k-1} = [\frac{\delta t_i}{\delta s_j}]_{s=s^{(k-1)}}$, $\Delta s_j^{(k)}$ is the perturbation to slowness $s^{(k-1)}$ at node j , $h_r^{(k)}$ is the time correction for station r at iteration k , $e_q^{(k)}$ is the time correction for event q at iteration k , and $\Delta t_i^{(k-1)}$ is the travel-time residual for i^{th} ray relative to iteration $(k-1)$ of the model.

P- and S-wave models use identical parameterization, with splines under tension constrained at a series of regular knots. A spline is a smooth curve that is pinned to pass through the data points called knots. The grid extends from 51° N to 77° N in latitude, from -156° W to -94° W in longitude and from the surface to 1300 km depth. This grid has 27, 76 and 103 knots in the depth, latitude and longitude direction resulting in a total of 211,356 knots. The inner grid is sampled every 0.25° of latitude, 0.5° of longitude and 33 km in depth.

The inverse problem posed by Equation 3 has a nontrivial model null space in this study (211,356 slowness perturbation unknowns, 127,138 relative P-wave delay times and 59,107 relative S-wave delay times), and a unique, but biased, solution is thus obtained by im-

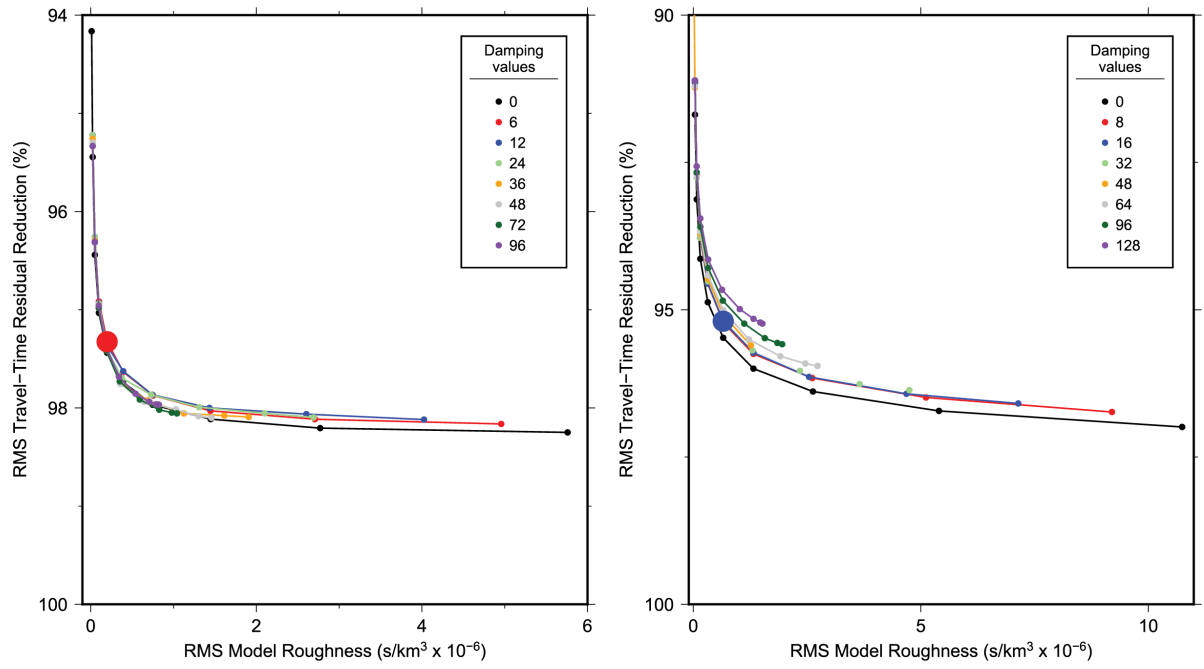


Figure 4.3.: Trade-off curves between percentage relative travel-time residual reduction and model roughness for the P- (left) and S-wave (right) models. The trade-off curves are constructed from 72 inversions with different weights applied to the damping, flattening and smoothing parameters. Each color denotes a different damping value. The preferred P- and S-wave models are shown with a larger symbol.

plementing model regularization. In addition to minimizing the two-norm fit to the data, the VanDecar method incorporates three types of regularization into the objective function: damping (i.e., model 2-norm minimization), flattening (i.e., first-derivative smoothing) and smoothing (i.e., second-derivative smoothing). To find a preferred model, we examine the trade-off curves between regularization and data fit. We first choose the weights for the flattening and smoothing parameters without damping, and then determine the damping weight associated with the same percentage of residual reduction obtained solely using the flattening and smoothing parameters. Subsequently, we run a series of inversions that explain varying percentages of the initial percentage root mean square (RMS) residuals. The result is the exploration of an optimal trade-off region from 72 inversions (Figure 4.3). In this study, we use regularization parameters that put the preferred model slightly above the highest curvature of the trade-off area and thus conservatively favors a slightly more regularized model and somewhat worse fit to the data than the formal corner of the trade-off curve when plotted on linear axes. For both models, we use the same number of conjugate gradient iterations (2000) and the same number of Huber down-weighting iterations (10) to obtain a more robust solution. Huber's method (Huber, 1964) gradually reduces the weight, or "down-weights", observations with large outliers (i.e., modifies the error function to minimize the influence of very large residuals). Our final models explain 97.33 % and 95.19 % of the RMS relative arrival time residuals for P and S, respectively.

4.4 INVERSION RESOLUTION

The resolution of a tomographic inversion problem depends on the model parameterization, as well as on raypath geometry and density. These factors ultimately determine the degree of rank deficiency and ill-posedness in the unregularized system (e.g., Aster et al., 2018). Incorporating regularization produces a unique model but imposes solution bias. In teleseismic body wave studies, raypaths have very steep incidence angles within the uppermost mantle, which results in better lateral resolution than vertical resolution. As a consequence of limited vertical resolution, models are affected by inherent vertical smearing (e.g., Bastow et al., 2008; Youssof et al., 2015; Boyce et al., 2016; Liddell et al., 2018).

To assess model resolution, we utilize a checkerboard test model that consists of an alternating pattern of positive and negative anomalies ($\delta V = \pm 5\%$) at 200, 400 and 600 km depth (Figures B.2 and B.3). Each anomaly has a radius of 50 km and is defined by a Gaussian function across their diameter. We add Gaussian noise with standard deviations of 0.03 and 0.1 s to the P-wave and S-wave data sets, respectively and then invert these

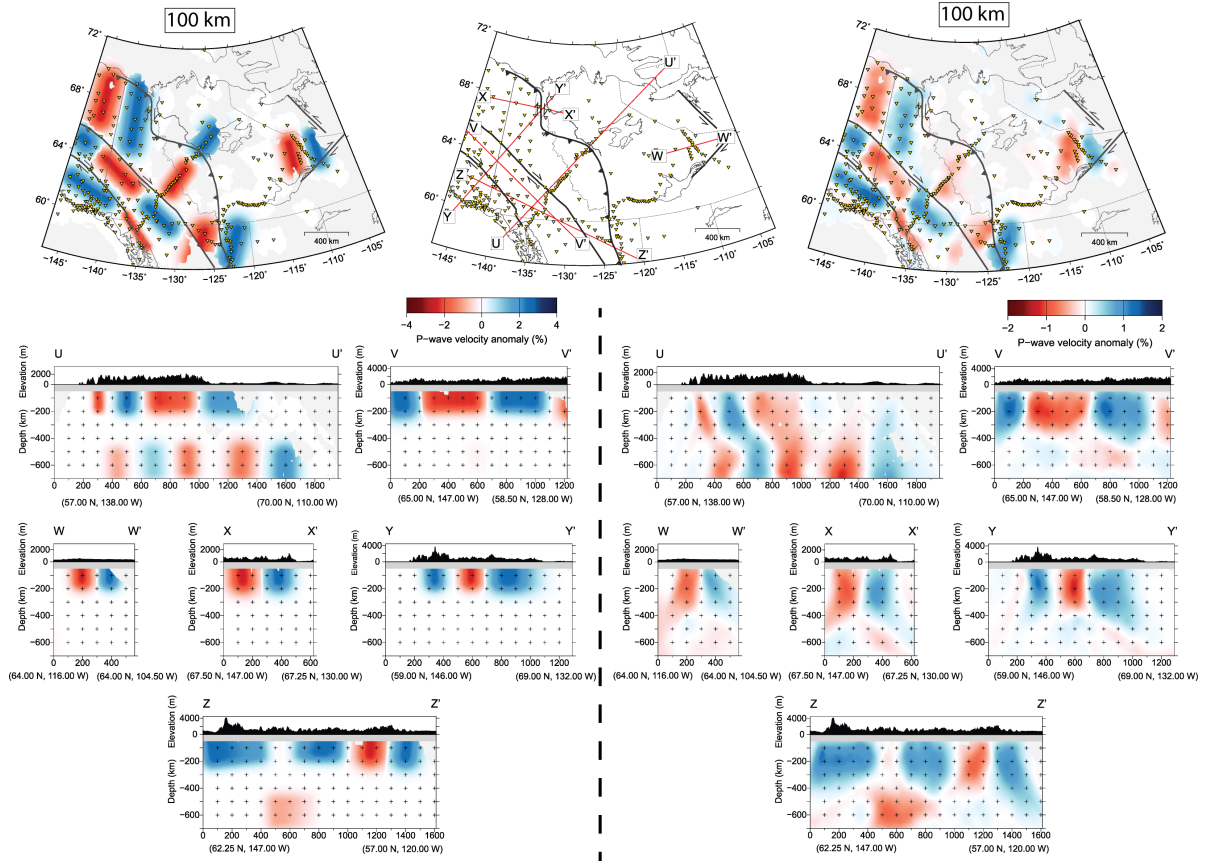


Figure 4.4.: Cross sections and depth slices through the P-wave structural resolution test. Structural resolution test showing input anomalies (left) and output anomalies (right) through our P-wave model. Synthetic anomalies are defined by Gaussian functions across their widths. Also shown: station locations (yellow inverted triangles) and tectonic structures from Figure 4.1. Areas with less than 4 ray hits are displayed in light gray and the first 50 km are masked.

data using the same experiment raypath geometry and model parameterization to assess the recovery of the test model and infer resolution features.

Recovered models show a clear distinction in resolution between the western region (the Cordillera) and the eastern region (the Canadian Shield). The station coverage in the west is clearly improved by the deployment of the USArray TA stations (Figure 4.1), with a relatively dense (~ 85 km) and regular station spacing in eastern Alaska and western Yukon (Figures B.2 and B.3). The recovery of the anomalies is excellent in the southwestern region of both models. The amplitude recovery of P- and S-wave models is about 50 to 60 %, which is typical for a regularized least square tomographic model. The amplitudes are most accurate in the region of dense ray path coverage and the recovery becomes better at intermediate depths due to the increase in crossing ray paths. East of the Cordilleran Deformation Front,

the station coverage is much sparser. The position of anomalies is well retrieved beneath the POLARIS network (Figure 4.1) and the CANOE network. Stripes of anomalies oriented northeast-southwest show resolution artifacts due to poor station coverage and reflect the back-azimuth distribution of the teleseismic rays. The recovered S-wave model generally displays better recovery of the anomalies and their amplitudes (Figure B.3), in particular along the northern part of the Alaska/Yukon border and the relatively dense Mackenzie Mountains transect. The observed better recovery for the S-wave model could be due to the addition of SKS core phases in our S-wave data set.

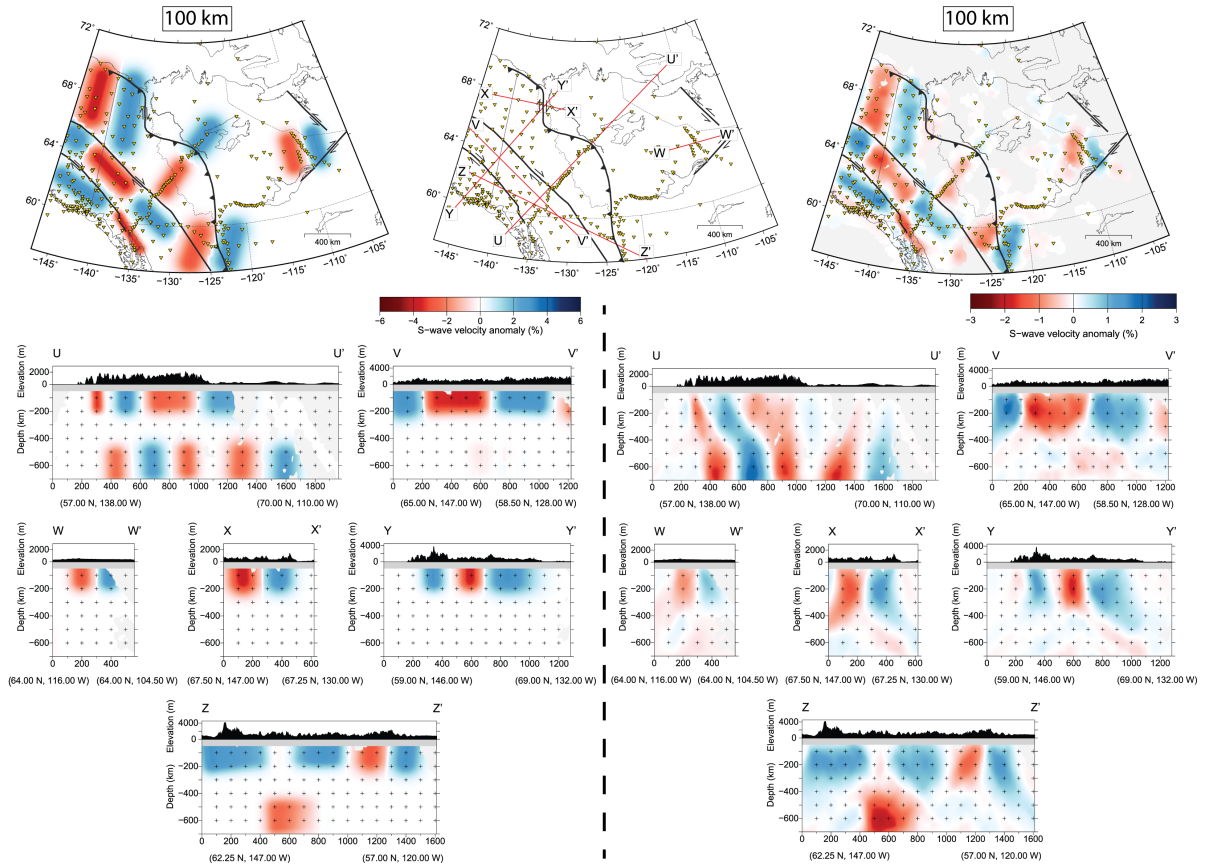


Figure 4.5.: Cross sections and depth slices through the S-wave structural resolution test. Format is the same as Figure 4.4.

In addition to the checkerboard test, we assess the resolution of longer-wavelength features and the effects of limited resolution in our real model by performing a synthetic structural test (Figures 4.4 and 4.5). This test incorporates structures that mimic particular features in the recovered model (see following section) and applies the identical procedure as employed in the checkerboard tests. The elongated synthetic anomalies are defined by Gaussian velocity anomalies across their width. For the P-wave structural test, we model fast ($\delta V_P = 3\%$) and

slow ($\delta V_P = -3\%$) anomalies from the surface to 200 km depth across the study area. We also reproduce anomalies between 450 km and 650 km depth with different amplitudes. From southwest to northeast, we model a slow region beneath the Denali Fault ($\delta V_P = 1.5\%$), a fast region beneath the Tintina Fault ($\delta V_P = 1\%$), a slow region beneath the Mackenzie Mountains ($\delta V_P = -1\%$), a slow region beneath the Great Bear Lake ($\delta V_P = -1.5\%$), and a fast region beneath the northern region of the Slave craton ($\delta V_P = 2\%$). Outside these test anomalies, the background P-wave and the S-wave velocity anomaly is set to zero. We compute the S-wave structural test in a similar way except that the amplitudes are slightly different (Figure 4.5). The fast regions between the surface and 200 km depth were assigned a $\delta V_S = 3\%$ and the slow regions $\delta V_S = -3\%$ or $\delta V_S = -4\%$. At greater depth, the anomaly velocities are $\delta V_S = \pm 2.5\%$.

Both P- and S-wave structural test output models indicate excellent lateral resolution between the surface and 200 km depth and at greater depths between 450 km and 650 km depth (Figures 4.4 and 4.5). The only region where the input anomalies are not retrieved with the same resolution is along the eastern part of the Mackenzie Mountains transect (between the Tintina Fault and the Cordilleran Deformation Front). At this location we juxtaposed slow and high-velocity anomalies to investigate the resolution of the transition between the Cordillera and the western craton boundary. Failure to recover this contrast in velocity anomalies may indicate that a longer recording period is required to get a high-resolution model of the cordillera-craton transition zone and/or that regularization effects from surrounding sparsely sampled model parameters are reducing model recovery along the transect. Individual cross-sections show that our P- and S-wave models suffer from expected vertical smearing effects due to the near-vertical raypaths (Figures 4.4 and 4.5). Within the Cordillera, the shallower set of anomalies are smeared down to 300 km depth (vertical smearing of ~ 100 km) and retrieved anomalies in the mantle transition zone are smeared up to the shallower anomalies and seem to connect with them. Slave craton anomalies are smeared down to ~ 400 km depth (vertical smearing of ~ 200 km). The indicated resolution scales indicate that care must be taken when interpreting the depth extent of anomalies.

4.5 UPPER MANTLE STRUCTURE

We present depth slices through our tomographic models at 100, 150, 200 and 300 km depth (Figures 4.6 and 4.7). Using a relative delay time approach leads to a key drawback, which is the loss of information on the background mean velocity (see Equation. 1). Our models have a zero mean corresponding to the average velocity of northwestern Canada. As

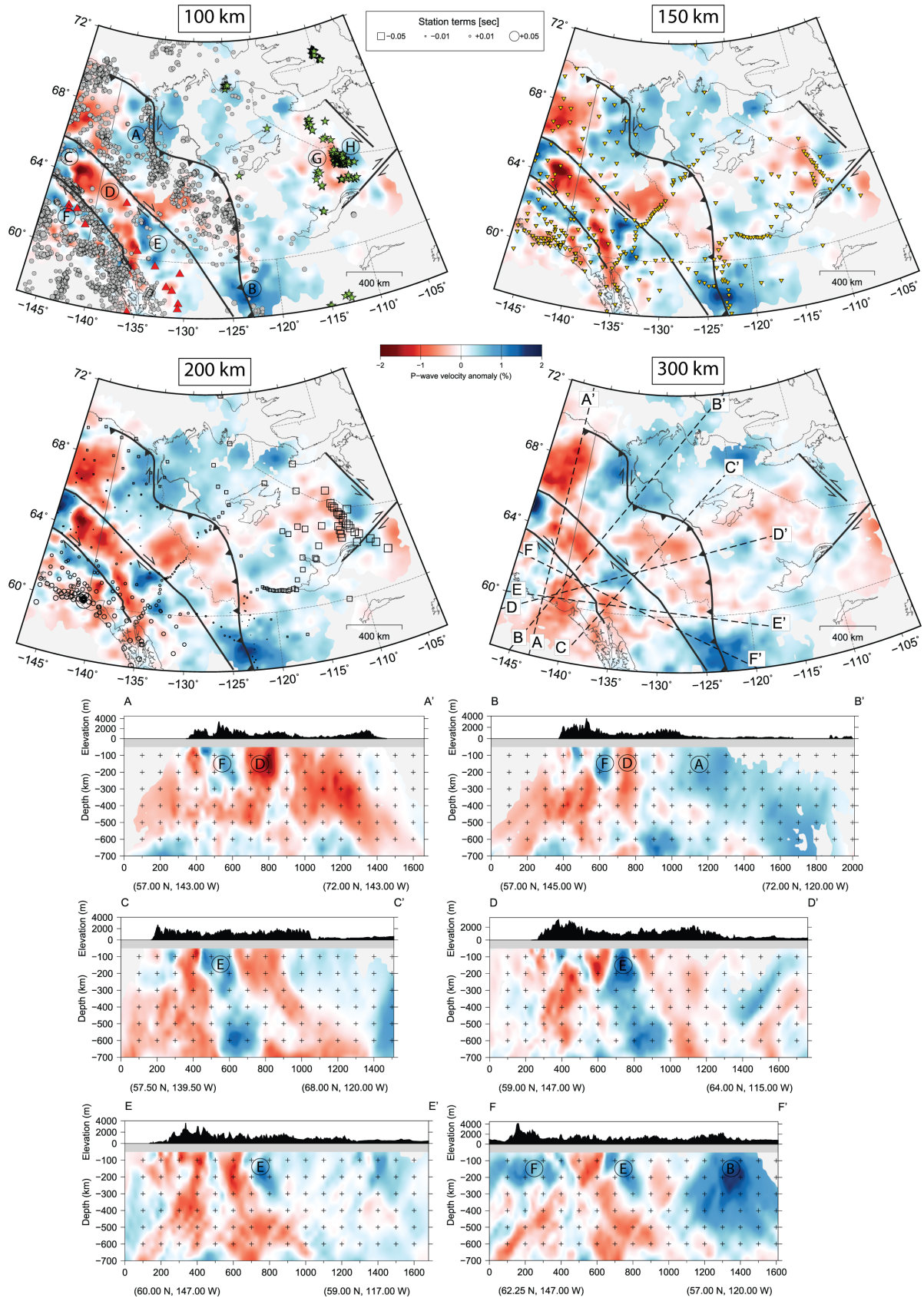


Figure 4.6.: Depth slices and cross sections through the P-wave model. Superimposed on the 100 km depth slice the regional seismicity, the volcano locations and the distribution of kimberlite fields in the western Canadian Shield. Also shown: stations static terms after removal on the 200 km depth slice and the transect locations on the 300 km depth slice. Tectonic structures from Figure 4.1. Areas with less than 4 ray hits are displayed in light gray.

a consequence, the recovered amplitudes cannot be compared with the global mean value (Bastow, 2012). For example, at 100 km, the SL2013sv model shows a fast velocity anomaly ($\delta V_S = 6\%$) at 58°N , -123°W (Schaeffer & Lebedev, 2013) whereas, at the same location, our S-wave model shows an average S-wave velocity anomaly of $\delta V_S = 2\%$. We removed the station static terms from our models to avoid contamination of mantle structure by the crustal velocity anomalies (Figures 4.6 and 4.7).

4.5.1 *P-wave model*

The P-wave velocity structure underlying northwestern Canada appears to be highly heterogeneous (Figure 4.6). Overall, cratonic lithosphere east of the CDF is characterized by fast P-wave anomalies whereas the cordilleran lithosphere west of the CDF displays slow P-wave anomalies. A fast P-wave anomaly underlies the region across all of northern Yukon and the northwestern part of the Northwest Territories (labeled A). This anomaly extends down to an apparent depth of 400 km and is laterally well-resolved but vertically smeared, as revealed by our structural test. We also observe a slow P-wave anomaly in the Brooks Range of eastern Alaska adjacent to the fast P-wave anomaly in northern Yukon (labeled A). The P-wave velocity contrast is well-resolved and sharp in this area. To the south, a relatively slow P-wave anomaly extending down to 400 km depth is located beneath the Mackenzie Mountains but the resolution tests show that the synthetic anomalies were not retrieved very well in this particular region. Further to the south, the P-wave model reveals a pervasive, vertically smeared high-velocity anomaly ($\delta V_P \geq 1.5\%$) beneath northeastern British Columbia extending down to an apparent depth of 500 km (labeled B). Between the Tintina Fault and the Denali Fault, a series of well-resolved high to low to high-velocity anomalies from northwest to southeast are revealed by our model and extend down to 300 km depth (labeled C, D and E). These anomalies are laterally well-resolved and are slightly smeared (~ 100 km). At greater depths (400 km and below), a well-resolved fast P-wave anomaly ($\delta V_P = 2\%$) sits in the mantle transition zone. South of the Denali Fault, a 150 km wide and 600 km long fast P-wave anomaly extends to 300 km depth (labeled F). This anomaly is located beneath the Wrangell Volcanic field in Alaska, USA. Relative slow and fast P-wave anomalies exist beneath the POLARIS network in the Slave craton at 100 km depth (labeled G and H).

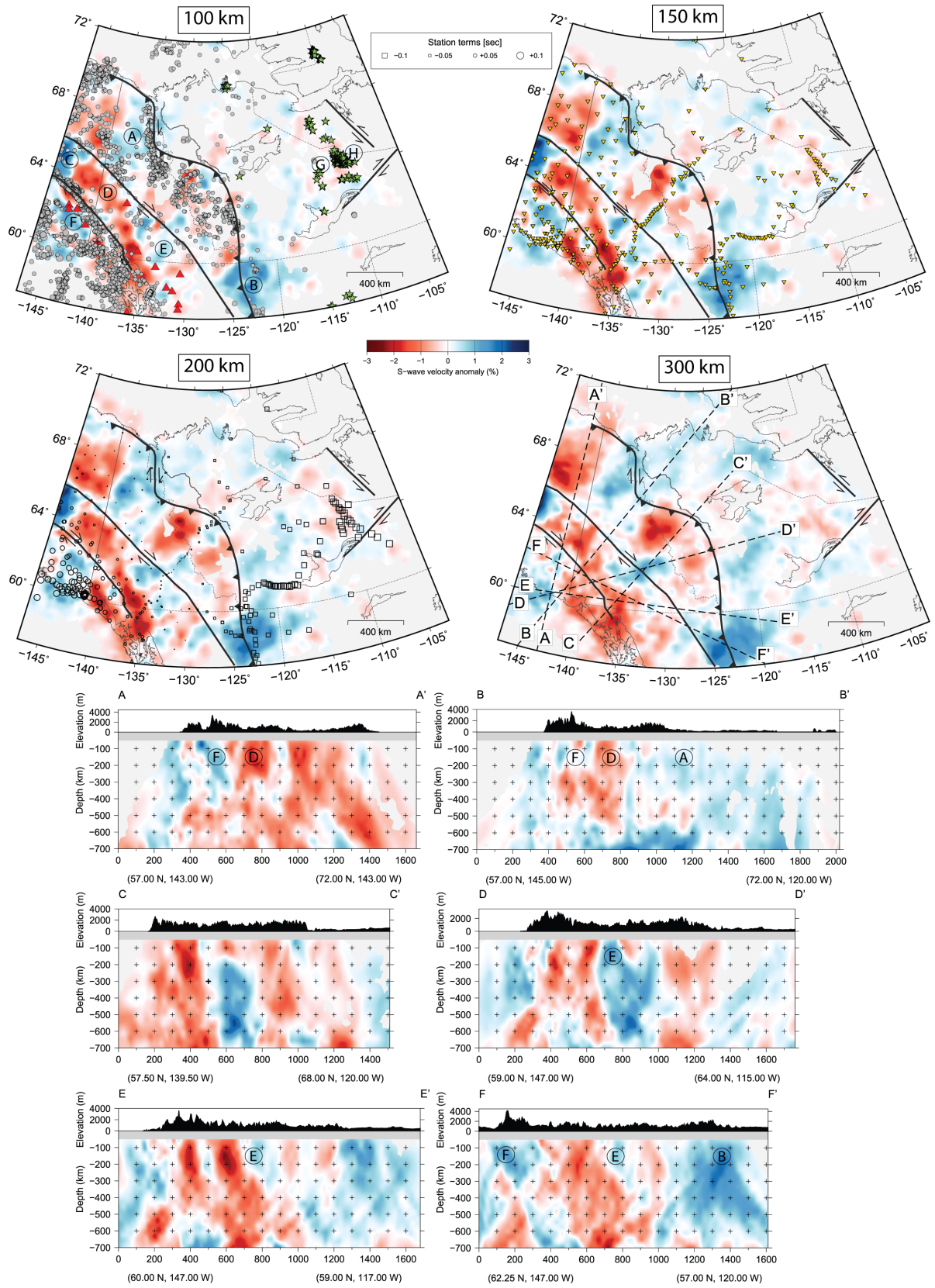


Figure 4.7.: Depth slices and cross sections through the S-wave model. Format is the same as Figure. 6.

4.5.2 *S-wave model*

The S-wave model reveals similar features with some differences (Figure 4.7). Across northern Yukon and Northwest Territories, a similar fast S-wave anomaly extends down to an apparent depth of 400 km (labeled A). The Brooks Range is underlain by a slow S-wave anomaly and a similar S-wave velocity contrast is observed between the fast and slow S-wave anomalies beneath Yukon and the Brooks Range, respectively. To the south, a pervasive slow anomaly underlies the Mackenzie Mountains. Its maximum amplitude is located beneath 100 and 300 km depth ($\delta V_S = -2\%$ at 64.5°N , -132°W). Further to the south, northeastern British Columbia is underlain by a similar fast S-wave anomaly as shown in the P-wave model (labeled B). The S-wave model also reveals laterally well-resolved structures beneath the Tintina Fault and the Denali Fault up to 200 km depth (labeled C, D and E). Then at greater depths (400 km and below), a prominent fast S-wave anomaly is located beneath the Tintina Fault beneath the central segment of the Mackenzie Mountains transect network. South of the Denali Fault, a fast S-wave anomaly extends down to ~ 250 km depth beneath the Wrangell Volcanic field. To the east and beneath the POLARIS network in the Slave craton, a slow anomaly ($\delta V_P \leq -1\%$) extends down to the mantle transition zone. However, as revealed by our resolution tests, this region of the model is affected by vertical smearing due to the lack of crossing rays. Furthermore, the high-velocity anomaly underlying the eastern Slave craton does not appear as prominent.

4.6 DISCUSSION

Our tomographic models reveal five prominent features: (i) a broad high-velocity anomaly beneath northern Yukon that extends down to 400 km apparent depth (labeled A), (ii) a high-velocity anomaly located beneath northeastern British Columbia (labeled B), (iii) a series of fast and slow anomalies between the Tintina Fault and the Denali Fault extending down to 300 km depth (labeled C, D and E), (iv) a high-velocity anomaly beneath the Wrangell Volcanic field (labeled F) and (v) juxtaposed slow and high-velocity anomalies beneath the Slave craton (labeled G and H). We use magnetic (Bankey et al., 2002) and gravity (Bonvalot et al., 2012) data to help interpret these features. We plot the magnetic anomaly map and the complete spherical Bouguer gravity anomaly map adapted to our study area, which both give an indication of the distribution of the subsurface magnetic and density anomalies, respectively (Figure 4.1). Positive magnetic anomalies can reflect the presence of buried crystalline basements beneath a sedimentary cover. Similarly, negative magnetic

anomalies may be the signature of thick sedimentary basins. Low Bouguer anomalies are generally interpreted to be related to lower density materials and vice versa. Low density materials are commonly associated with granitic intrusions or sedimentary rocks and high-density materials with mafic intrusions.

4.6.1 *Geophysical evidence for the Mackenzie craton*

Both P- and S-wave models show a sharp seismic velocity contrast between the eastern part of the Brooks Range and a high-velocity body in northern Yukon. This well-resolved high-velocity anomaly extends west of the CDF in northern Yukon (Figures 4.6 and 4.7, labeled A). This region is part of Laurentia and has a thick sedimentary cover (see Fig. 1; Nelson & Colpron, 2007). Based on geological mapping (Ross, 1991), it was postulated that beneath the sedimentary cover there exists a rigid crystalline basement (2.1 to 1.8 Ga). As pointed out by Schaeffer & Lebedev (2014), the surficial expression of the CDF changes across the Richardson Mountains and is shifted further to the west. Schaeffer & Lebedev (2014)'s continental-scale surface-wave tomography model shows high seismic velocities (characteristic of a cratonic lithosphere) beneath this area and therefore strongly suggest the presence of an unexposed craton in northern Yukon. Both our P- and S-wave models show a relatively high velocity anomaly distributed over northern Yukon and Northwest Territories at 100 km and deeper. Seismic velocities are mainly sensitive to temperature and, as a result, the relatively high velocity beneath northern Yukon is interpreted as a cold thermal anomaly within the lithosphere. Precambrian cratonic lithosphere is typically underlain by thick and high-velocity lithospheric roots, cold and depleted in Fe-rich (i.e., basaltic) materials (Jordan, 1978).

Regional magnetic studies conducted in northwestern Canada (Saltus & Hudson, 2007; Pilkington & Saltus, 2009) reveal pronounced high-amplitude and long-wavelength magnetic anomalies across Yukon and Northwest Territories north of the Mackenzie Mountains fold and thrust belt (Figure 4.1). This Mackenzie River magnetic anomaly (MRA, Pilkington & Saltus, 2009) is oriented approximately NE-SW and is 650 km long and roughly 200 km wide, with maximum amplitudes between 400 and 500 nT. The high velocity anomaly observed in our tomographic models and the MRA are both located beneath northern Yukon and Northwest Territories. The MRA is interpreted as mafic lower crust, implying a basaltic depletion of the underlying upper mantle, which would become mechanically resistant against edge erosion of its western boundary through upper mantle convection (Saltus & Hudson, 2007).

The Bouguer gravity map shows a clear distinction between the NCC and the Canadian Shield; the Cordillera is dominantly underlain by low Bouguer anomalies whereas the Canadian Shield shows high Bouguer anomalies. However, northern Yukon and Northwest Territories lie within a region where a long-wavelength high Bouguer anomaly is observed (80 to 140 mGal). Another moderate high Bouguer anomaly is observed at 65°N , -136°W (40 to 80 mGal) and contrasts with the long-wavelength low Bouguer anomaly underlying the Mackenzie Mountains. We suggest that the long-wavelength high Bouguer anomaly is indicative of a subsurface dominated by mafic rocks and that the moderate high anomaly reflects the signature of a Proterozoic sedimentary basin (see Fig. 3 Nelson & Colpron, 2007).

Results from SKS splitting analyses reveal a large-scale northeast-southwest alignment of the fast axis directions beneath the eastern branch of the CANOE network in the Canadian Shield (Snyder & Bruneton, 2007; Courtier et al., 2010; Audet et al., 2016). Venereau et al. (2019) calculated the SKS splitting parameters for EarthScope USArray-TA stations in Alaska and Yukon. The fast axis directions for the TA stations located in Yukon are subparallel to the large-scale alignment of the fast axis directions in the Canadian Shield. This implies the two structures have recorded a similarly aligned fabric and potentially share a lithospheric mantle structure of similar origin.

All geophysical evidence presented in this study strongly suggest that the lithosphere beneath northern Yukon is mechanically strong. Therefore, we interpret the geophysical anomalies beneath northwestern Yukon and Northwest Territories as being indicative of cratonic lithosphere. This supports the existence of an unexposed Precambrian craton, which we refer to as the Mackenzie craton.

4.6.2 Cratonic buttressing of the NCC

Both P- and S-wave models show well-resolved high-velocity anomalies extending west of the CDF in the Cordillera domain at two locations; in northeastern British Columbia and across northern Yukon and Northwest Territories (Figures 4.6 and 4.7, labeled A and B). Another regional teleseismic body wave tomography study imaged the same high-velocity anomaly in northern British Columbia beneath the eastern branch of the CANOE network at 100 km depth (see Figs. 8 and 9, Mercier et al., 2009), which was interpreted as cratonic lithosphere. This is also shown in Audet et al. (2019), where the thick fast lithosphere of the Canadian Shield extends beneath the northern Rocky Mountains.

The magnetic anomaly map shows another long-wavelength magnetic high, co-located with the fast P- and S-wave anomalies (Figure 4.1). This magnetic high is located beneath the northeastern region of British Columbia between 57°N and 60.5°N and has the same maximum amplitude as the MRA. The Bouguer anomaly map also shows a westward extension of a moderate Bouguer anomaly (40-60 mGal) west of the CDF at the Yukon/British Columbia border. Both magnetic anomalies coincide with regions where the surficial expression of the CDF is marked by geometrical changes along strike and are interpreted as a mafic lower crust that formed in an extensional rift setting (Pilkington & Saltus, 2009). The magnetic high and the Bouguer anomaly in northeastern British Columbia coincide with the Liard transfer zone inherited from an asymmetric extensional rifting episode that separates the lower and upper plates of the asymmetric rifting system (Lund, 2008). North of the LTZ, the lower plate margin, which consists of the broader Mackenzie Mountains and the Selwyn sedimentary basin, is underlain by slow P- and S-wave velocities. South of the LTZ, the upper plate margin formed on a detachment fault and also corresponds to a narrower sedimentary cover in British Columbia with underlying fast P- and S-wave velocities.

Seismic anisotropy results from teleseismic SKS splitting analysis suggest that the lithospheric mantle of the former Laurentian margin underlies the NCC south of the LTZ (Audet et al., 2016). Fast P- and S-wave anomalies at 100 km depth extend up to 63° N and –135° W in southern Yukon. A sharp slow-to-fast transition occurs at this location and could mark the western limit of the North American lithospheric mantle. The inferred extent of the North American lithospheric mantle is slightly further north compared to the results of Audet et al. (2016) because the deployment of the Mackenzie Mountains transect network (Baker et al., 2019) brings improved resolution to this region.

In summary, south of LTZ, geophysical data suggest that a mechanically strong and resistant lithospheric mantle underlies a mafic crust. At the northern end of the CDF, an unexposed Mackenzie craton is interpreted beneath northern Yukon and the adjacent Northwest Territories. A lithospheric geodynamic model constrained by geology, GPS and seismic data shows that the upper mantle flow direction beneath the Mackenzie Mountains is north-eastward (i.e., orthogonal to the CDF; Finzel et al., 2014). These results strongly suggest that the two cratonic lithospheres act like buttresses on both ends of the Mackenzie Mountain fold and thrust belt and are coherent with regional magnetic studies (Saltus & Hudson, 2007; Pilkington & Saltus, 2009). This would suggest that the two pieces of strong cratonic lithospheres were pinned and channeled mantle flow beneath the Mackenzie Mountains, which gave the modern arcuate shape to the CDF (Figure 4.8). Thinner lithosphere beneath the Mackenzie Mountains also implies hotter geotherms there and greater buoyancy, maintaining the high elevations (in contrast to northern Yukon and Liard Basin).

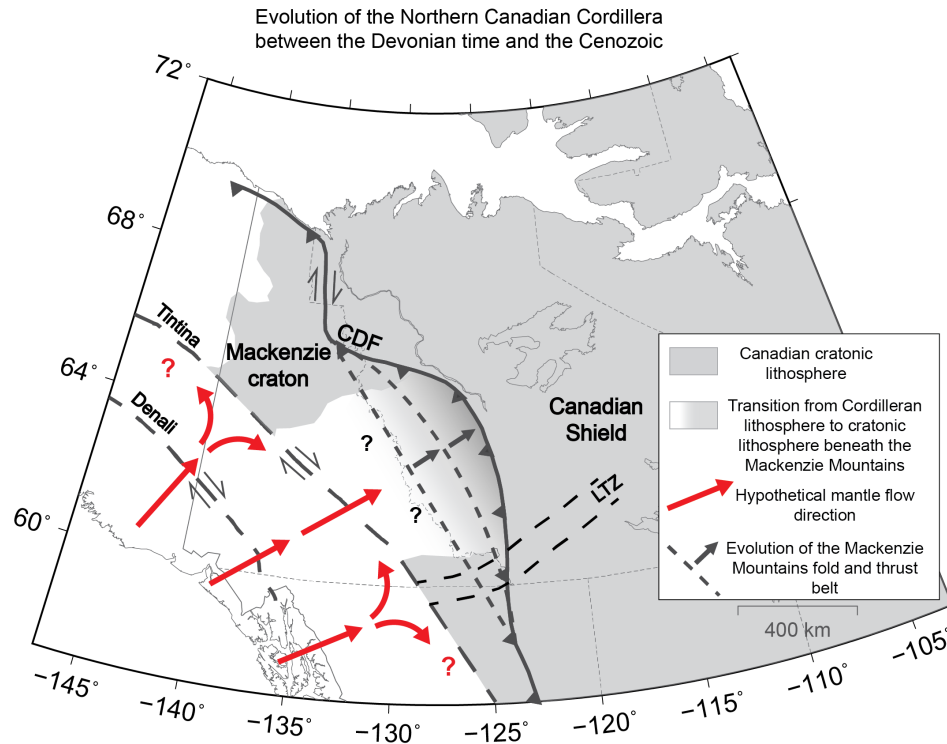


Figure 4.8.: Schematic model depicting cratonic buttressing and hypothesized mantle flow and Mackenzie Mountains regional lateral displacement from Devonian time to the Cenozoic. Tectonic structures from Figure 4.1. The Denali and Tintina faults (dashed lines) became active at the end of the Cretaceous period and during the Paleocene, respectively.

4.6.3 The Tintina Fault

The tectonic evolution of the NCC involves the displacement of accreted terranes along major right-lateral strike-slip fault zones such as the Denali and Tintina Faults. These major tectonic structures accommodated most of the oblique relative motion between the Pacific and North American plates during the Eocene. The Tintina Fault is interpreted to have accommodated 420 to 490 km of horizontal displacement during the Paleocene, based on geological and geophysical evidence (Gabrielse et al., 2006; Zelt et al., 2006; Snyder et al., 2005; Nelson & Colpron, 2007; Hayward, 2015). The crustal structure of the Tintina Fault has been imaged by the SNORCLE seismic reflection (Cook et al., 1999, 2004; Cook & Erdmer, 2005) and magnetotelluric profiles (Ledo et al., 2002, 2004). Line 3 of SNORCLE reveals that the Tintina Fault is characterized by a lack of seismic reflectors to 30-35 km depth, especially in the lower crust (Snyder et al., 2005). This would suggest the Tintina

Fault cuts the entire crust and is coherent with magnetotelluric results (Ledo et al., 2004; Dehkordi et al., 2019).

Our P- and S-wave models reveal a series of anomalies across the region between the Tintina and the Denali faults (Figures 4.6 and 4.7; labeled C, D and E). These laterally well-resolved anomalies are coherent and extend down to 300 km depth; however, these are affected by inherent vertical smearing (Figures 4.4 and 4.5). Laterally, there is a strong P- and S-wave velocity contrast across the Tintina Fault (Figures 4.6 and 4.7; transects B-B' and C-C'). This could reflect differing lithospheric mantle across the region implying that the Tintina Fault is a deeper feature extending down to the upper mantle and is therefore coherent with seismic anisotropy results (Audet et al., 2016). Geochemical evidence from xenoliths bolsters this contention by displaying distinct variations in isotopic compositions of erupted alkaline basalts on either side of the Tintina Fault that are interpreted to reflect compositionally different mantle sources (Abraham et al., 2001).

4.6.4 *The Wrangell subduction zone*

NW-SE elongated high velocity P- and S-wave anomalies extending down to 300 km depth are located near the USA/Canada border (62° N and -144° W) beneath the Wrangell volcanic field (Figures 4.6 and 4.7, labeled F), reflecting the signature of the Wrangell Slab (Gou et al., 2019; Bauer et al., 2014; Wang & Tape, 2014; Martin-Short et al., 2016; Martin-short et al., 2018). This anomaly is laterally well-resolved but has vertical smearing of ~ 100 km. The northward-dipping slab geometry is imaged to be nearly flat over 200 km and then dips nearly vertically to 300 km depth beneath the Wrangell volcanic field. This imaged slab geometry contradicts some other tomographic studies where either the imaged slab signature does not appear clearly (Martin-Short et al., 2016; Martin-short et al., 2018; Wang & Tape, 2014) or has a lesser flat extent (Gou et al., 2019). However, another recent seismic tomography study imaged a high-velocity anomaly that was similarly interpreted as a slab segment sinking vertically beneath the Wrangell volcanic field (Jiang et al., 2018). As pointed out by Wang & Tape (2014), the Yakutat microplate currently moves northwestward which means that the Wrangell slab is subducting obliquely. The Wrangell slab thus may be connected to the Yakutat slab to the west and could represent the early stage of a slab tear (Pavlis et al., 2019a). Improved local resolution is likely necessary to investigate this hypothesis.

4.6.5 Kimberlite fields in the Slave craton and surroundings

The mineral compositions of diamond host rocks such as kimberlites have been extensively studied (e.g., [Stachel & Luth, 2015](#)). Diamonds form within sub-continental lithospheric mantle where necessary pressure and temperature conditions are achieved (the diamond stability field) or at greater depths within the asthenosphere ([Motsamai et al., 2018](#)). The locations of kimberlite fields in the Slave craton and its surroundings are superimposed on 100 km depth slices as green stars in Figures 4.6 and 4.7. Kimberlite magmatism in the region occurred in different episodes over a long time scale. For example, the oldest known kimberlites in the northern Slave craton have ages of 613 Ma and the most recent, Lac de Gras area (central Slave craton), kimberlites have ages between 75 and 45 Ma ([Nowicki et al., 2004](#); [Lockhart et al., 2004](#)). Kimberlite fields in the Slave craton form a north-south trending corridor.

It has generally been observed that kimberlite fields overlie the margins of steep V_P and V_S gradients within the cratonic lithosphere (the Australian craton, [O'Neill et al. \(2005\)](#); the Kaapvaal craton, South Africa, [Shirey et al. \(2002\)](#); the Slave craton, Canada, [Esteve et al. \(2019\)](#)). Although the overall recovery of input anomalies within the Canadian Shield is poor when compared to the Cordillera domain due to sparse station density, our P-wave model resolves juxtaposed relatively low- and high-velocity anomalies to 300 km depth (Figure 4.6, labeled G and H) in central and eastern Slave craton at 100 km depth in the region of densest kimberlite exposures. However, our S-wave model does not show a faster region beneath eastern Slave craton (Figure 4.7, labeled H). Nonetheless, the recovery appears to be better beneath the POLARIS stations (Figures B.2 and B.3). The structural test reveals that the two low- and high-velocity anomalies modeled in the Slave craton that mimic the features observed in the real model are laterally well-resolved but are characteristically affected by smearing effects (vertical smearing of ~ 200 km). Therefore, the transition from low to high velocity in the central and eastern Slave craton is a resolved feature in our models but we cannot confidently constrain its depth extent. This result is consistent with another regional teleseismic study conducted over this region ([Esteve et al., 2019](#)).

4.7 CONCLUSION

This study images the most data-rich and up-to-date P- and S-wave velocity structure beneath northwestern Canada using P- and S-wave relative delay times. The best resolved region of the model corresponds to the location of the USArray-TA network deployed over

Alaska, USA and Yukon and Northwest Territories, Canada. The main observations and interpretations from our models are as follows:

- Distribution of regional magnetic, Bouguer gravity, and seismic velocity anomalies support the existence of an unexposed “Mackenzie craton” in northern Yukon and adjacent Northwest Territories.
- Geophysical evidence for two mechanically strong lithospheric blocks, interpreted as cratonic lithosphere, acting like buttresses on both ends of the Mackenzie Mountain fold and thrust belt, which suggest the hypothesis that the two pieces of strong cratonic lithosphere channel mantle flow and perhaps gravitational collapse and lower crustal flow, controlling the modern arcuate configuration of the Mackenzie Mountains ([Figure 4.8](#)). Greater constraints on crustal and lithosphere-scale temperatures and strength from additional data, such as ambient noise surface wave inversion and regional scale body waves, will be necessary to test this hypothesis.
- Strong P- and S-wave velocity contrasts across the Tintina Fault indicate that differing lithospheric mantle is juxtaposed on either side. This suggests that the Tintina Fault has displaced upper mantle material along a sharp boundary, and demarcates a lithospheric-scale shear zone.
- Both P- and S-wave models resolve the Wrangell slab. The slab geometry is observed to subduct steeply beneath the Wrangel volcanic field at an oblique angle relative to the present northwestern motion of the Yakutat block.
- The distribution of kimberlite fields in the Slave craton correlates with the margin of fast and slow V_P mantle anomalies.

SEISMIC EVIDENCE FOR CRATON CHISELING AND DISPLACEMENT OF LITHOSPHERIC MANTLE BY THE TINTINA FAULT IN THE NORTHERN CANADIAN CORDILLERA

5.1 SUMMARY

The Northern Canadian Cordillera (NCC) of northwestern Canada is segmented by several margin-parallel, right-lateral strike-slip faults that accumulated several hundred kilometers of displacement between the Late Cretaceous and the Eocene. The depth extent of these faults, notably the Tintina Fault (TF), has important implications for the tectonic assemblage and evolution of NCC lithospheric mantle, but geophysical models and geochemical data remain inconclusive. Using a recent three-dimensional P-wave seismic velocity model, we resolve a series of sharp (≤ 10 km) P-wave velocity contrasts ($\sim 4\%$) at uppermost mantle depths beneath the surface trace of the TF. Seismic anisotropy data that represent upper mantle fabrics reveal similar changes in the orientation and magnitude of anisotropy in the vicinity of the TF. These data suggest that the TF is a lithospheric-scale shear zone. After restoration of 430 km of right-lateral displacement along the TF, fast P-wave anomalies align with the outline of the North American craton margin. We propose that the fast anomaly structure currently located in eastern Alaska represents a fragment of the Mackenzie craton that has been chiseled and displaced to the northwest by the TF between the Late Cretaceous and the Eocene. A second cratonic fragment currently located in the southern NCC may be associated with the Cassiar terrane at upper mantle depth. These observations provide the first evidence that large lithospheric-scale shear zones cut through refractory mantle and produce major lateral displacement of cratonic mantle material within Cordilleras worldwide.

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5.2 INTRODUCTION

The amalgamation and evolution of continental lithosphere is traditionally understood through the lens of the Wilson cycle (Wilson, 1966), which thermally rejuvenates the margins of old continental blocks through repeated episodes of opening and closing of ocean basins. In particular, most rifted cratonic blocks have been sutured by collision, either with other cratonic blocks or with younger continental lithosphere (Currie & van Wijk, 2016, and references therein). Understanding the evolution of cratonic margins can give important clues for plate reconstructions in the Proterozoic (e.g., Merdith et al., 2017), but there are significant challenges related to their episodic reactivation. Furthermore, the recognition of margin-parallel strike-slip shear zones that mobilize material along the margin may further complicate geological interpretation. In particular, the depth penetration of these faults may have important consequences for tectonic reconstructions and the role of the upper mantle in controlling their evolution. Consequently, the reconstruction of cratonic rifted margins in the Wilson cycle remains poorly constrained due to the difficulty in preserving evidence of their former state following continental collision.

One notable exception is the western margin of the Canadian Shield (CS) in North America, which has been spared from continental collision since the Mesoproterozoic (Whitmeyer & Karlstrom, 2007). It is now juxtaposed with the Northern Canadian Cordillera (NCC) in northwestern Canada, which is an active transpressional accretionary orogen. The NCC initially developed in the Neoproterozoic with the formation of the Laurentian passive margin, following break-up of the supercontinent Rodinia. This passive margin persisted up to Devonian time and subsequently transitioned to an east-west convergent plate boundary (Nelson & Colpron, 2007). Multiple Cenozoic margin-parallel, right-lateral strike-slip fault zones cut through both accreted terranes and passive-margin sequences of the NCC. In particular, the Tintina Fault (TF) has produced 430 km of horizontal displacement between the Late Cretaceous and the Eocene (Gabrielse et al., 2006) (Figure 5.1).

Despite decades of research, there exists no clear consensus on the origin of the lithospheric mantle in the NCC and its relationship to accreted terranes and North American basement in the overlying crust (e.g., Pavlis et al., 2019b). In particular, it remains unclear if the TF displaced entire lithospheric blocks along a transcurrent boundary or only a thin sheet of exotic upper crustal material (e.g., Snyder et al., 2002; Cook et al., 2004; Hayward, 2015, 2019). Resolving this conundrum would have profound implications on the origin of lithospheric mantle (e.g., Abraham et al., 2001) and tectonic evolution models (e.g., Cook et al., 2004; Hayward, 2015; Hildebrand, 2009; Johnston, 2008). A recent teleseismic tomography model (Estève et al., 2020a) suggests the juxtaposition of differing lithospheric

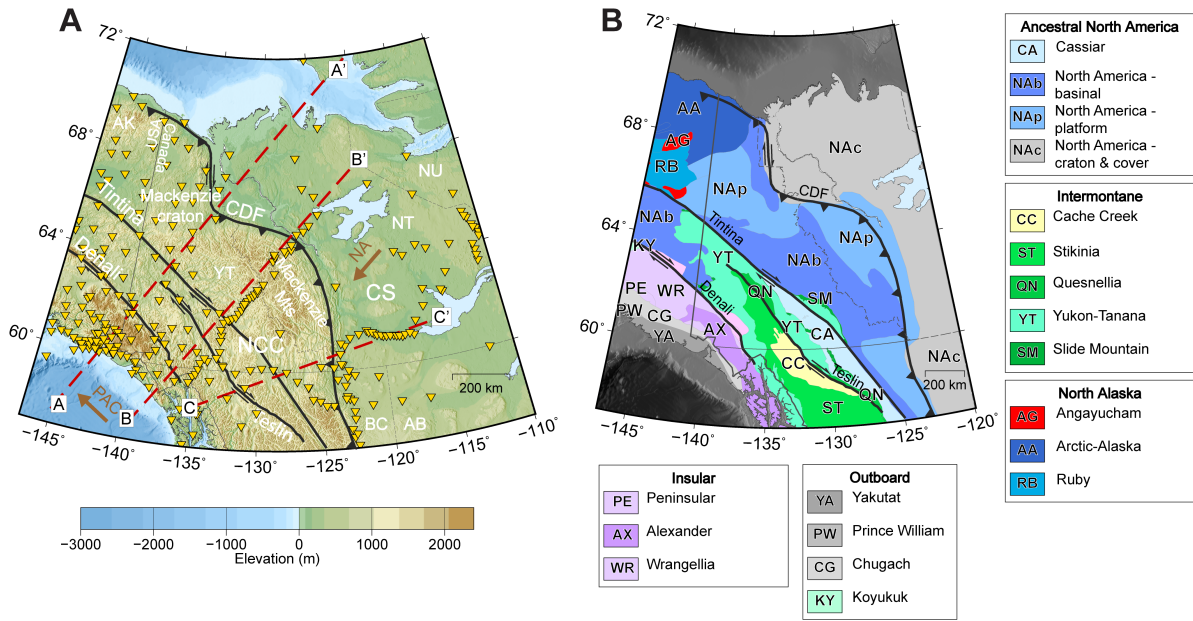


Figure 5.1.: **A**: Topographic map of northwestern Canada showing the seismic stations (inverted triangles) used to develop the teleseismic P-wave model. Also shown are the main tectonic structures and the cross-section locations (red dashed line) displayed in Figure 5.3. Brown arrows show the absolute plate motion directions of Pacific (PAC) and North American (NA) plates, from DeMets et al. (2010). **B**: Surface terrane maps of the NCC, modified from Nelson & Colpron (2007). Abbreviations: AB, Alberta; AK, Alaska; BC, British Columbia; NT, Northwest Territories; NU, Nunavut; YT, Yukon; CS, Canadian Shield; CDF, Cordilleran Deformation Front; NCC, Northern Canadian Cordillera.

mantle across the TF, which was loosely interpreted as evidence for penetration of the TF into the uppermost mantle.

In this work, we investigate this question in more detail by combining the P-wave tomography model with existing and new seismic anisotropy measurements in the region. We further restore 430 km of right-lateral displacement of upper mantle seismic velocities across the TF. Taken together, these results provide additional compelling evidence that large lithospheric-scale faults can cut entirely through cratonic margin lithosphere, with important consequences for our understanding of cratonic lithosphere evolution and preservation through time.

5.3 TELESEISMIC CONSTRAINTS

We use the most complete three-dimensional, teleseismic P-wave tomography model to date (Estève et al., 2020a) to focus on the uppermost mantle structure surrounding the

TF. This model represents relative, isotropic seismic velocity anomalies with respect to the IASP91 global model (Kennett & Engdahl, 1991) and has high (10s of km) lateral resolution but generally poor vertical resolution. The P-wave model reveals a laterally heterogeneous seismic velocity structure beneath the NCC, which remains coherent after averaging the model between 66 and 200 km (Figure 5.2-A). The model supports the presence of a newly identified Mackenzie craton in northwestern Canada (Schaeffer & Lebedev, 2014), which acts as a stiff buttress that helps control the arcuate shape of the Cordilleran Deformation Front (CDF) (Figures 5.1 and 5.2-A; Estève et al., 2020a). From northwest to southeast, a series of fast-slow-fast P-wave anomalies are observed between the TF and the Denali Fault. In cross-section (Figure 5.3), these P-wave anomalies are defined by sharp velocity gradients (≤ 10 km) beneath the TF and extend to upper mantle depths.

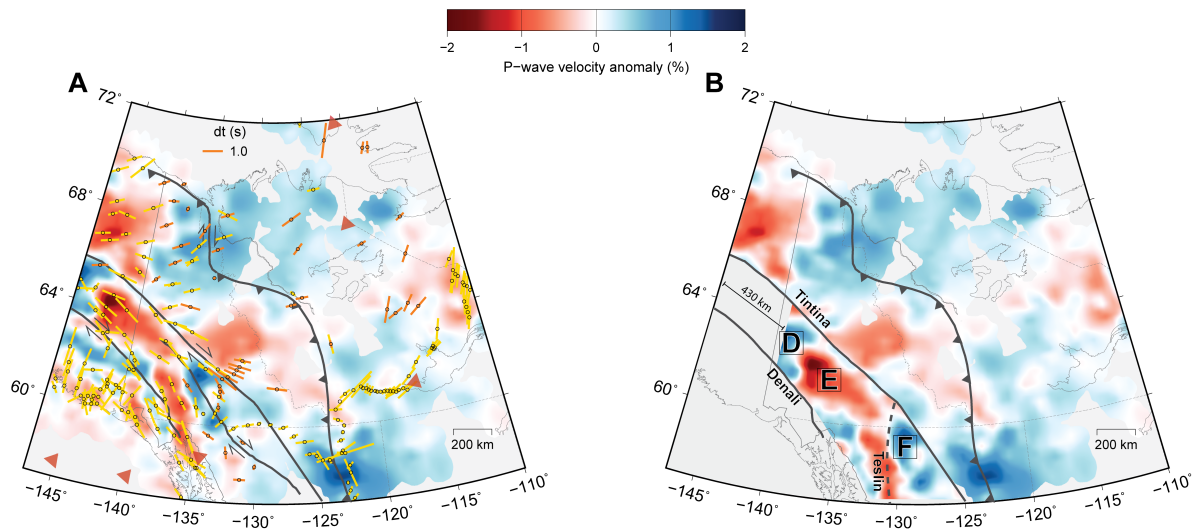


Figure 5.2.: **A**: P-wave anomaly model averaged between 66 and 200 km depth. Also shown are the average SKS splitting parameters for each station. Colored bars represent the fast axis direction of propagation scaled by the delay time. Gold-colored bars are compiled from Audet et al. (2016); Courtier et al. (2010); Rasendra et al. (2014); Snyder & Bruneton (2007); Venereau et al. (2019); orange bars are from this study. **B**: Restoration of the 430 km of right-lateral displacement on the Tintina Fault. The dashed-line shows the approximate location of the Teslin Fault obtained from the pre-Late Cretaceous restoration of Nelson & Colpron (2007). Labels D, E and F show the location of features discussed in the text.

Constraints on upper mantle dynamics can be obtained from estimates of seismic anisotropy, which is sensitive to the coherent alignment of deformation-induced rock fabric. In the upper mantle, seismic anisotropy is principally attributed to the lattice preferred orientation of olivine during shear (e.g., Nicolas & Christensen, 1987), and may reflect either present-day

mantle convection, and/or “fossilized” deformation processes in the uppermost mantle. We determine seismic anisotropy from shear-wave splitting analysis of core-refracted shear waves (e.g., SKS and SKKS), which is defined by two parameters, the fast direction of wave propagation (or fast axis ϕ) and the delay time (δt) between the fast and slow components. These parameters reflect integrated properties along the nearly vertical ray path on the receiver side, and therefore have poor vertical resolution.

We compile 184 station-averaged shear-wave splitting estimates from five studies carried out across northwestern Canada (Audet et al., 2016; Courtier et al., 2010; Rasendra et al., 2014; Snyder & Bruneton, 2007; Venereau et al., 2019) and augment these observations with new SKS splitting estimates for 45 seismic stations in the region (Figure 5.3-A) (Supplemental Materials). Previous studies note a large-scale coherent northeast-southwest fast axis alignment in the CS, interpreted as fossilized fabric within the North American lithospheric mantle and/or sub-lithospheric mantle flow subparallel to the present-day absolute motion of North America (e.g., Courtier et al., 2010). Fast axis directions in northern Yukon and Northwest Territories are also subparallel to the fabrics observed in the CS providing further evidence of the Mackenzie craton. Another notable feature is the progressive clockwise rotation of fast axis directions near the TF associated with increasing delay times approaching the fault along most of its length (Figures 5.2 and 5.3). Along transect B-B', splitting estimates in the North American craton show moderate delay times ($\delta t \sim 0.5$ s) with an east-northeast orientation (Figure 5.3-A). When approaching the TF, the fast axes rotate clockwise to align with its strike ($\phi \sim -60^\circ$); delay times increase to ≥ 1.5 s (Figure 5.3-B). Fast axis directions aligning with CS fabrics are observed in the southernmost NCC, which have been interpreted as North American lithosphere extending west of the TF (Audet et al., 2016), consistent with the juxtaposition of fast anomaly structure across the TF.

5.4 TINTINA FAULT: LITHOSPHERIC-SCALE SHEAR ZONE

The seismic velocity model indicates a strong P-wave velocity contrast at upper mantle depths over a short distance (≤ 10 km) across the TF. It seems difficult to relate such contrast to a purely thermal anomaly. Since the TF was last active in the Eocene (Nelson & Colpron, 2007), a late Eocene abrupt temperature contrast across the fault would be appreciably diffuse by now. For example, the characteristic length scale of a thermal discontinuity under one-dimensional diffusion since 37 Ma is approximately 22 km (Supplemental Materials). Instead, we propose that this likely reflects compositionally unique blocks of lithospheric mantle that are now separated by the TF. A compositional origin is consistent with estimated

5.4 TINTINA FAULT: LITHOSPHERIC-SCALE SHEAR ZONE

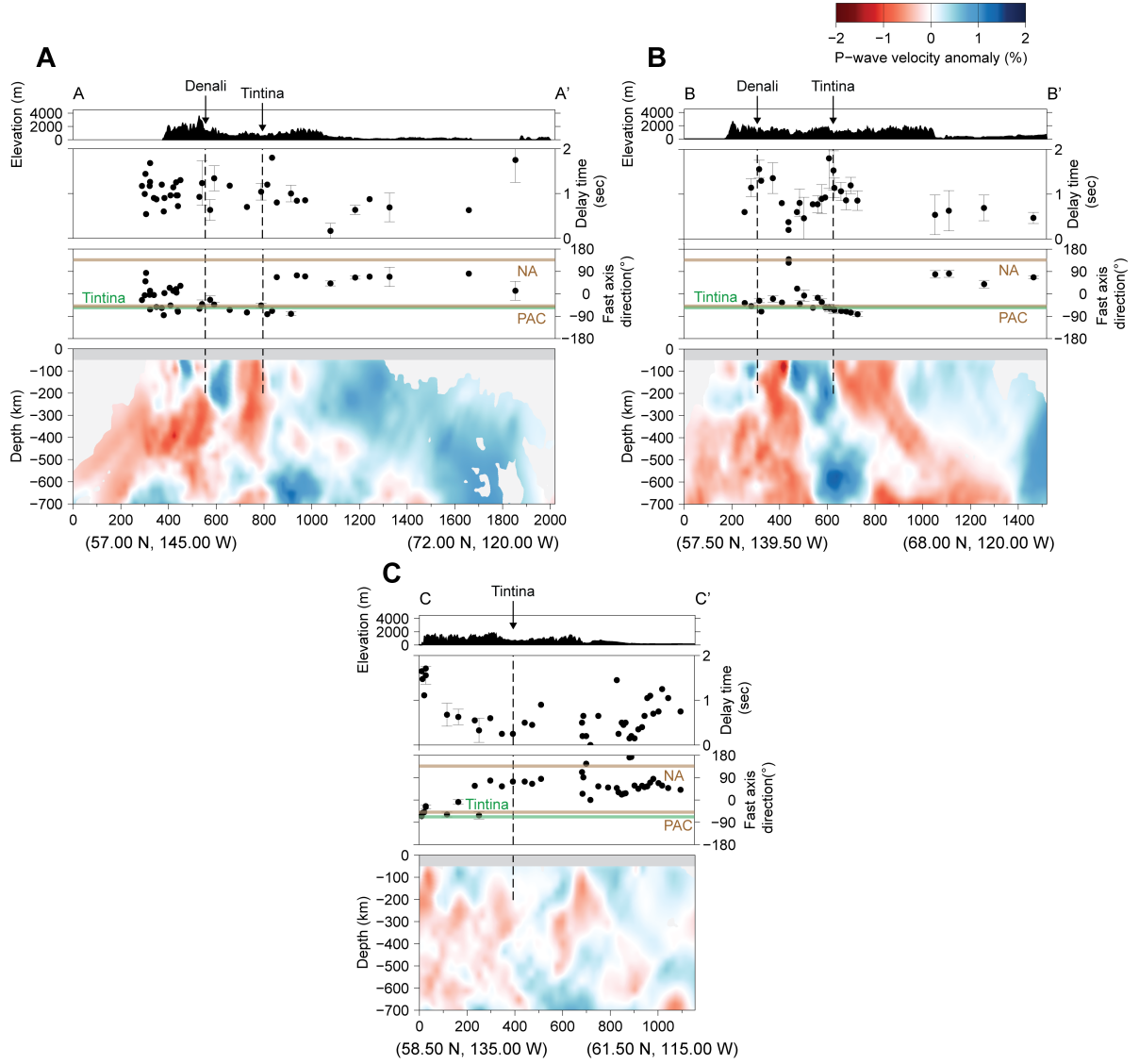


Figure 5.3.: Topographic profile (top), shear wave splitting parameters (middle two panels), and P-wave model of [Estève et al. \(2020a\)](#) (bottom) along the A-A', B-B' and C-C' cross-sections (see [Figure 5.1](#) for cross-section locations). The vertical dashed lines show the location of the Tintina and Denali faults. The approximate strike of the Tintina Fault and the current absolute motion of the North American (NA) and Pacific (PA) plates from [DeMets et al. \(2010\)](#) are shown as green and brown horizontal lines, respectively.

seismic velocities from cratonic and NCC xenoliths that show a $\sim 2\%$ difference (Supplemental Materials). Other studies have observed such patterns of large velocity contrasts across major faults in orogenic belts. For examples, seismic tomography of the Altyn Tagh Fault in Tibet reveals a series of slow and fast P-wave anomalies extending to uppermost mantle depths. The slow anomaly is interpreted to reflect shearing in the lithospheric mantle (Wittlinger et al., 1998).

Furthermore, shear wave splitting estimates show that in crossing the TF, fast axes rotate to align with the fault zone. This rotation is accompanied by increasing delay times, which is consistent with greater alignment of shear-induced fabrics (Figure 5.3-A and B) and implies that deformation is recorded through a large portion of the uppermost mantle column. Similar patterns of upper mantle seismic anisotropy are observed along other mantle-penetrating strike-slip faults (e.g., Vauchez et al., 2012), such as the Great Glen Fault (Bastow et al., 2007), the Altyn Tagh Fault (Li et al., 2011), and the Alpine Fault (Duclos et al., 2005). We note that, in the southernmost NCC, the observed decrease in δt across the TF has been explained as interference between TF and CS fabrics, consistent with lithospheric-scale shearing (Figure 5.3-C).

Based on these constraints, we propose that the TF marks a boundary between lithospheric mantle regions with distinct compositions and that shearing to uppermost mantle depths has occurred along the TF.

5.5 CRATON CHISELING BY THE TINTINA FAULT

Given the known displacement of the terranes along the TF between the Late Cretaceous and the Eocene, we investigate the continuity of the upper mantle structures across the TF by restoring 430 km of horizontal motion (Gabrielse et al., 2006; Hayward, 2015) from the vertically averaged P-wave anomaly model (Figure 5.2). After restoration (Figure 5.2-B), we observe close spatial alignment and continuity of three major P-wave anomaly structures across the TF: 1) a positive anomaly to the northwest, coincident with the southeastern limit of the Mackenzie craton (labeled D); 2) a negative anomaly structure in the central part of the NCC (labeled E) and 3) a second positive anomaly to the southeast which displays anisotropy consistent with CS fabrics across the TF (labeled F).

We cannot determine whether or not the restored seismic structures are compositionally similar. Nonetheless, we speculate that the restored contour of positive P-wave anomalies represents the outline of the North American lithosphere margin at pre-Late Cretaceous times (e.g., Nelson & Colpron, 2007) (Figure 5.4-A). We propose a conceptual tectonic evolution

model of the upper mantle in northwestern Canada (Figure 5.4-B) whereby fragments of North American cratonic lithosphere were chiseled and displaced to the northwest by the TF from the Late Cretaceous to the Eocene. The northwest fragment, currently located in eastern Alaska, has been previously interpreted by regional teleseismic tomography studies as an aseismic region of the subducted Pacific slab (Jiang et al., 2018; Martin-Short et al., 2016). Instead, we propose that it represents a piece of the Mackenzie craton whose northwest translation may have limited the inboard extent of flat-slab subduction in Alaska. Our model further suggests that the southeast fragment can be associated with the Cassiar terrane extending to upper mantle depths, and would represent another piece of cratonic lithosphere that has been displaced by the TF (Calvert et al., 2017). This model proposes a re-examination of lithospheric reconstructions at the former rifted margin of Laurentia.

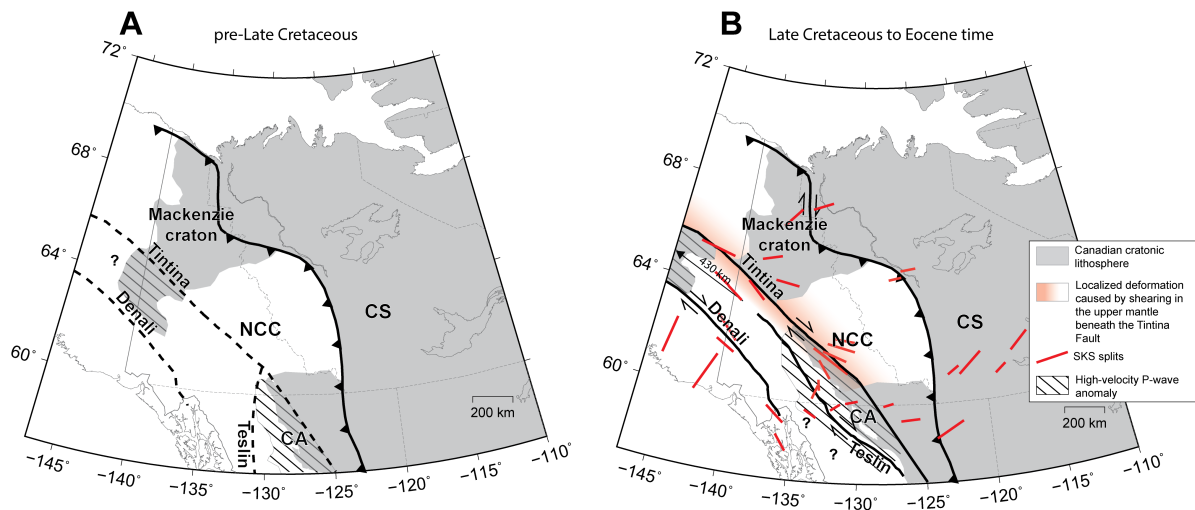


Figure 5.4.: **A:** Mid-Jurassic to Late Cretaceous NCC showing the extent of cratonic mantle before chiseling by the Tintina Fault (e.g., Nelson & Colpron, 2007). **B:** Late Cretaceous to Cenozoic NCC showing the 430 km of right-lateral displacement along the Tintina Fault. Red bars show inferred upper mantle fabrics from seismic anisotropy data at selected locations.

5.6 CONCLUSION

In this paper, we use teleseismic data collected in the NCC to constrain upper mantle velocity structure and dynamics in northwestern Canada and investigate the evolution of the cratonic margin. We conclude that the TF is a major lithospheric-scale shear zone that extends down to uppermost mantle depths. We further propose that the TF has chiseled off small pieces of the Laurentian craton between the Late Cretaceous and the Eocene. This would imply that

large lithospheric-scale shear zones are able to cut through small pieces of refractory cratonic mantle and transport them over several hundred kilometers. This could be the first evidence for this type of process, since the other referenced faults do not have cratonic material in the uppermost mantle that can serve as clear marker for displacement and prove that the offset extends to those depths.

Part III

CRUSTAL AND UPPERMOST MANTLE STRUCTURE OF NORTHWESTERN CANADA

AZIMUTHALLY ANISOTROPIC RAYLEIGH-WAVE GROUP VELOCITY MAPS ACROSS NORTHWESTERN CANADA FROM BAYESIAN INVERSION

6.1 SUMMARY

The recent deployment of temporary seismic stations across northwestern Canada allows us to investigate the seismic velocity structure and dynamics of the crust and the uppermost mantle. We measure fundamental-mode Rayleigh wave group velocity dispersion from regional earthquakes. We invert the dispersion data set using trans-dimensional Bayesian tomography in order to estimate azimuthally anisotropic group velocity maps at periods ranging from 10 to 60 s. Our results show that short period group velocity anomalies correlate well with geological features in northwestern Canada. We infer from our group velocity maps that the Mackenzie Mountains and the gulf of Alaska are underlain by a thicker crust. At long periods, the group velocity structure of the NCC is highly heterogeneous. In particular, a broad arcuate low group velocity structure across the NCC is interpreted to be a hot thermal anomaly in the lower crust and uppermost mantle potentially reflecting up-welling of dry and anhydrous materials from the underlying asthenosphere. Moreover, northern Yukon is interpreted to be underlain by a cold lower crust and uppermost mantle supporting the westward extent of North American craton in the NCC. Lastly, the Cordillera-Craton lithospheric boundary extends west of the Cordillera Deformation Front in the northwestern Mackenzie Mountains.

6.2 INTRODUCTION

Northwestern Canada has recorded more than 2.5 Gy of tectonic evolution with the formation of the cratonic core of North America and the development of the Phanerozoic Canadian Cordillera, which started with the break-up of supercontinent Rodinia around ~ 750 Ma

(Nelson & Colpron, 2007). Conversion of the passive margin occurred during the middle Devonian, leading to the development of subduction and arc magmatism. From Late Jurassic - Early Cretaceous onward, the developing orogen underwent right-lateral transpression, which produced 800 km of horizontal displacement on the Teslin, Tintina and Denali faults (e.g., Gabrielse et al., 2006; Hayward, 2015). At roughly 50 Ma, the Yakutat terrane, a small composite oceanic-continental terrane, detached from North America during a rifting episode (Plafker et al., 1994). At 30 Ma, the Yakutat terrane started to migrate northward along the Queen Charlotte-Fairweather fault system and collided with North America in the gulf of Alaska generating rapid uplift, intense seismicity, crustal thickening and the highest mountain ranges in North America. In the Early Miocene (~ 20 Ma), a mid-oceanic ridge subducted beneath North America, leading to the development of a slab window beneath the NCC (Thorkelson et al., 2011).

The present-day Canadian Cordillera is characterized by a thin and hot lithosphere (e.g., Audet et al., 2019; Hyndman, 2017) with high surface heat flow (Lewis et al., 2003), shallow Curie depth (Gaudreau et al., 2019) and low effective elastic thickness (Audet et al., 2007; Flück et al., 2003). Mantle xenolith and seismic imaging studies suggest minimum Moho temperatures near 800°C and lithospheric thickness lower than 65 km in its northern extent (Audet et al., 2019; Francis et al., 2010). Seismic tomography studies spanning various portions of the Canadian Cordillera show an abrupt change in seismic velocities at uppermost mantle depths between the Cordillera and the adjacent cratonic lithosphere (e.g. Mercier et al., 2009; Bao et al., 2014; Schaeffer & Lebedev, 2014; Zaporozan et al., 2018; Chen et al., 2019). Except for the highest elevation areas in the St-Elias-Chugach Mountains and the Mackenzie Mountains (Audet et al., 2020), the Canadian Cordillera does not have deep crustal roots, as might be expected in a continental collision (Tarayoun et al., 2017); however, it has been demonstrated that thermal expansion sustained by vigorous small-scale convection in the hot asthenospheric mantle can dynamically and gravitationally support its high elevations (Hyndman & Currie, 2011).

The deployment of recent temporary seismic networks across northwestern Canada, such as the Yukon-Northwest Seismograph Network, the Earthscope-USArray Transportable Array, and the Mackenzie Mountains Earthscope Project (Baker et al., 2019) allows us to develop high resolution seismic tomography models. Here, we use Rayleigh-wave group velocity maps at periods ranging from 10 to 60 s, obtained from recordings of regional earthquakes, in order to investigate the crust and uppermost mantle seismic velocity structure throughout northwestern Canada.

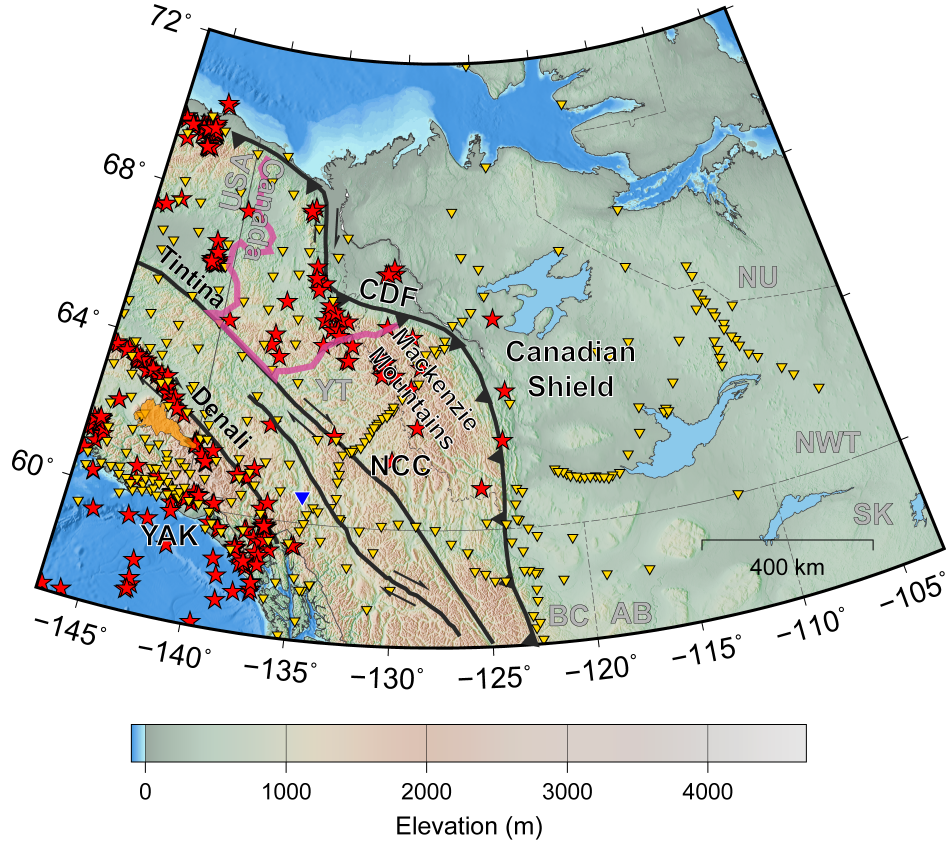


Figure 6.1.: Topographic map of northwestern Canada. Seismic stations (inverted yellow triangles) and regional earthquakes (red stars) used in this study. Inverted blue triangle shows station WHY. Black lines mark major tectonic structures in northwestern Canada. Orange fill shows the location of the Wrangell volcanic field. Purple line shows the westward extent of the Mackenzie craton from [Estève et al. \(2020a\)](#).

6.3 DATA AND METHOD

6.3.1 Data

Seismic data were downloaded from the Incorporated Research Institution for Seismology (IRIS) and the Canadian Hazard Information Service (CHIS) for 310 temporary and permanent seismic stations across northwestern Canada. The deployment time periods vary across the study area. For example, the CANOE seismic stations (the Canadian Northwest Seismic Experiment) were deployed for less than 2 years while other stations recorded decades of seismic data (e.g., station WHY located in Whitehorse, Yukon; [Figure 6.1](#)). Our data set consists of 469 regional earthquakes with magnitude $M_W \geq 4.5$ over a time period from July 1993 to September 2019 ([Figure 6.1](#)). We extract 30-minute vertical component seis-

mograms for all available seismic stations. Earthquakes that occurred less than 200 km away from a given station were not considered in the data set because surface-waves are difficult to observe at such close spacing at the periods considered in this work. We then remove the instrument response from each seismogram, detrend and apply a zero-phase bandpass filter with corner frequencies of 0.008 and 0.5 Hz to suppress microseismic noise and long-period instabilities.

6.3.2 *Dispersion measurements*

Surface wave group velocities from regional earthquakes are useful in imaging crustal and uppermost mantle structure since regional earthquakes excite surface-wave periods that are sensitive to crustal/uppermost mantle depths (Figure 6.2). In addition, unlike phase velocities, group velocity measurements do not require information on the source-time function of the earthquake. However, the uncertainties associated with group velocities are usually larger than those of phase velocities. In particular, earthquakes that occurred between 1993 and 2010 have poorly constrained epicentral locations due to a historically sparse seismograph distribution in northwestern Canada. Despite larger uncertainties, the analysis of surface wave group velocities remains a valuable tool since it allows for a wider range of earthquake magnitudes, and therefore, provides valuable information on the subsurface velocity structure in seismic tomography inversions.

We measure fundamental-mode Rayleigh wave group velocities from regional earthquakes using the multiple filter analysis implemented by Herrmann (2013), which allows the user to manually pick the velocity-period envelope range for each seismogram (Figure 6.3). The longest-operating stations yield more than 140 group velocity dispersion curves. Overall, group velocity dispersion measurements are made between 10 and 100 s, but we invert only for periods ranging between 10 and 60 s where path coverage is sufficient to resolve azimuthal anisotropy.

Before the inversions we perform an outlier analysis by, first, removing dispersion measurements that have a propagation path length shorter than 3 wavelengths. Subsequently, we examine our dispersion measurements within each source bin (with a radius of 45 kilometers) recorded at each station individually and remove outlier measurements using a median absolute deviation threshold of 2. Our final data set consists of 5703 group velocity dispersion curves.

Figure 6.4 shows the spatial distribution of our group velocity dispersion measurements across northwestern Canada at 9 periods ranging between 10 and 60 s. Overall, the path

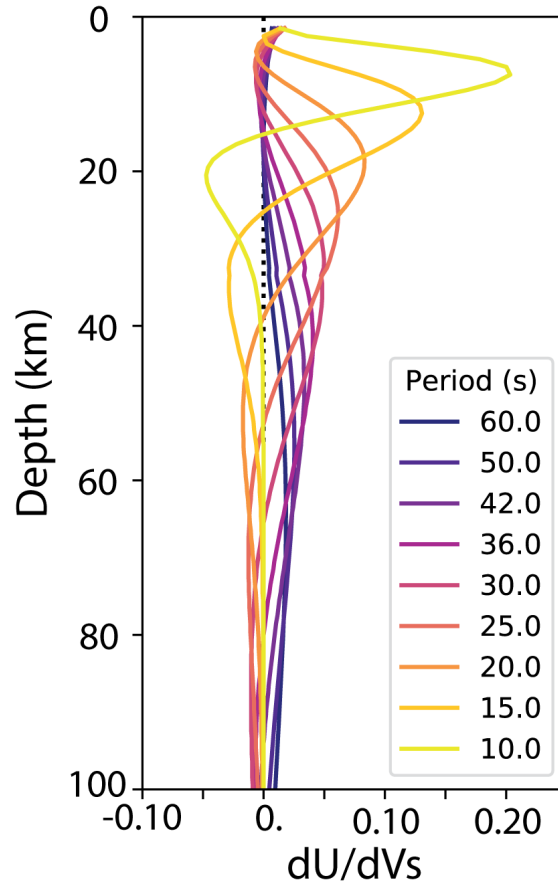


Figure 6.2.: Rayleigh wave group velocity sensitivity kernels at 10, 15, 20, 25, 30, 36, 42, 50 and 60s, calculated for a simple model with a 33 km thick crust and upper mantle velocities from the global velocity model AK135 (Kennett et al., 1995).

coverage is excellent through the NCC and is sparser in the Canadian Shield due to the lack of permanent seismic stations. At all periods, there is a clear trend in our data set showing group velocities increasing eastward, toward the Canadian Shield. Source-station paths going through the NCC show a wide range of velocities, reflecting a heterogeneous subsurface seismic velocity structure. In particular, at longer periods, source-station paths traveling across the Richardson and the Mackenzie Mountains shows moderate-to-high and low-to-moderate group velocities, respectively. This data set is the input to the tomographic inversion.

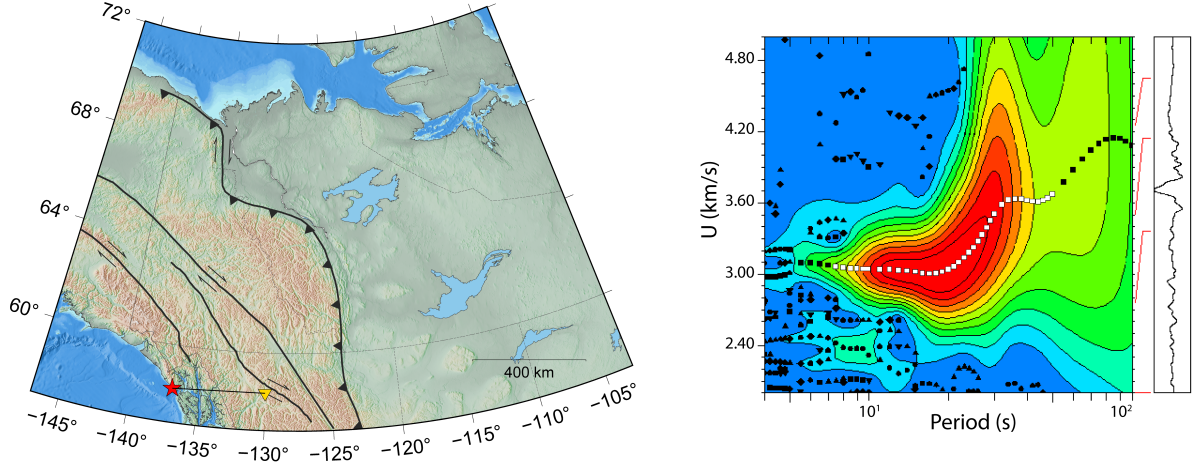


Figure 6.3.: (Left) Topographic map shows the source (red star) and the seismic station (yellow inverted triangle) location for a particular source-station path. (Right) Example of an individual surface-wave group-velocity dispersion measurement using the multiple filter analysis (Herrmann, 2013). Colors represent the energy of the surface wave envelope as a function of group velocity and period. White squares show the manually-selected period band for this event at this station. The diagram on the right shows the seismogram.

6.3.3 Inversion procedure

Inversion of surface wave group velocity dispersion measurements is performed individually at 9 periods ranging between 10 and 60 s within a Bayesian framework in order to estimate group velocity maps. Here we use the trans-dimensional tomography method described by Gosselin et al. (2020), which solves simultaneously for isotropic group velocity and azimuthal anisotropy. The velocity of a surface wave propagating through an anisotropic medium at a given period (T) and azimuth (ψ) is given by :

$$C(T, \psi) = C_0(T) + A1(T) \cos(2\psi) + B1(T) \sin(2\psi) + A2(T) \cos(4\psi) + B2(T) \sin(4\psi) \quad (6.1)$$

where $C_0(T)$ is the isotropic group velocity at a given period T , 2ψ is the variation of the group velocity with π periodicity, and 4ψ is the variation with $\pi/2$ periodicity. In our implementation, the variation of group velocity with $\pi/2$ periodicity is neglected to simplify computations (Gosselin et al., 2020).

We use the reversible-jump Markov-chain Monte Carlo method (rjMCMC) to generate a large ensemble (100 million) of samples which approximate the posterior probability density

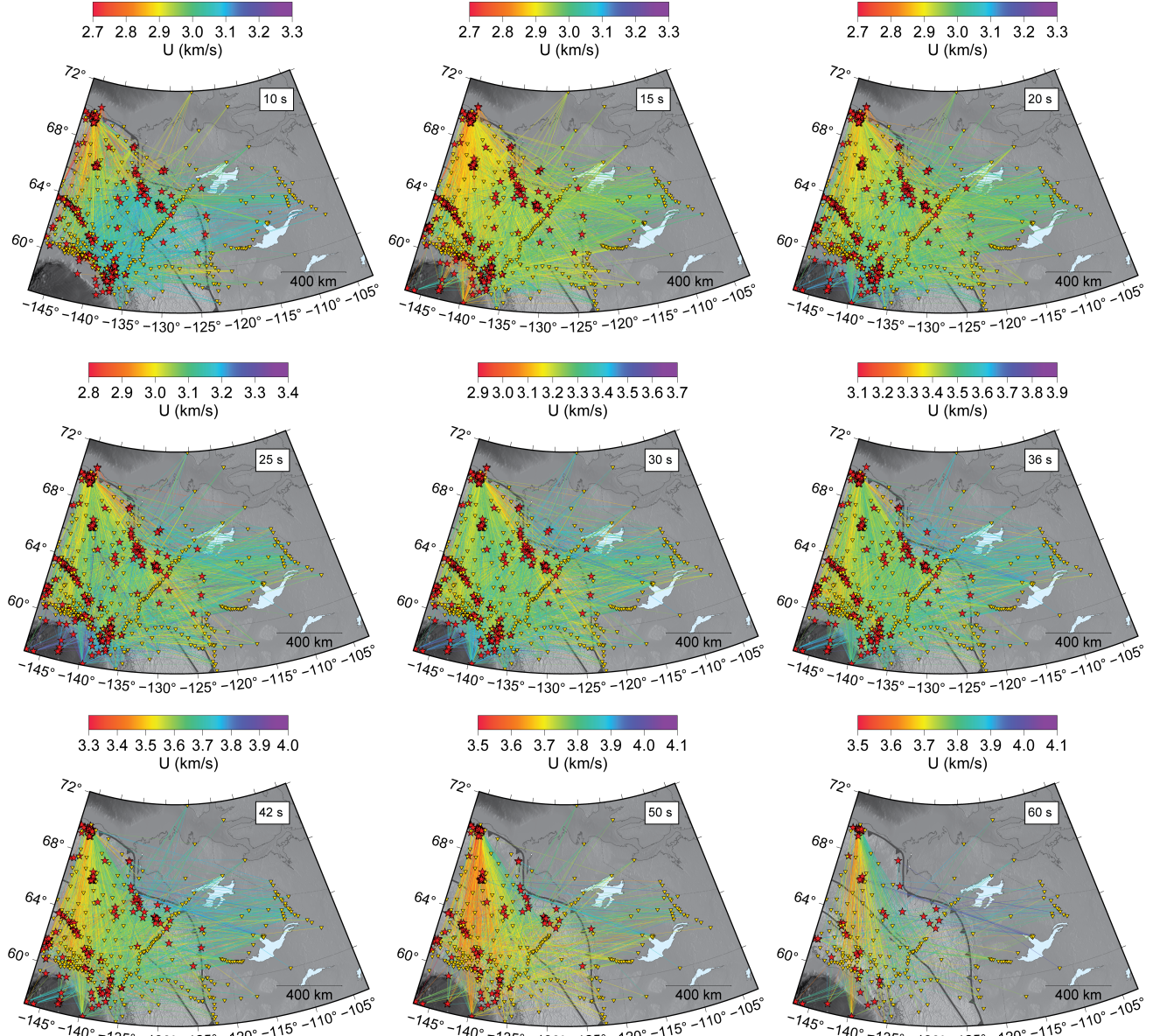


Figure 6.4.: Source-station group velocity dispersion maps plotted at periods of 10, 15, 20, 25, 30, 36, 42, 50 and 60 s. Note that the color scale differs between maps. Earthquakes and seismic stations are shown as red stars and inverted yellow triangles. Black lines show the main tectonic structures such as the Denali Fault, the Tintina Fault and the Cordillera Deformation Front.

(PPD) of the model parameters. Initial models are randomly chosen and 20 chains generate 5 million samples each with 2.5 million removed as burn-in samples. We calculate the mean and standard deviation of the ensemble of models at every location on the map as a representative model with associated uncertainty. The group-velocity maps are parameter-

ized by an unknown number of nodes. The number and configuration of nodes is solved as part of the inversion. The group velocity between nodes is linearly interpolated using Delaunay triangulation (Hawkins et al., 2019). This parameterization scheme is mapped over a regular integration grid with a 30 kilometer spacing during forward calculations (i.e., surface wave travel-time predictions). The uncertainty σ associated with our surface wave data set is defined as a linear relationship between a source-to-station distance Δ_i and two hyper-parameters a and b for a given path i (Bodin et al., 2012).

$$\sigma_i = a\Delta_i + b \quad (6.2)$$

6.4 GROUP VELOCITY MAPS OVER NORTHWESTERN CANADA

6.4.1 *Isotropic group velocity structure*

We present our mean isotropic absolute group velocity maps for periods ranging between 10 and 60 s across northwestern Canada (Figure 6.5). We also show the isotropic group velocity uncertainty across northwestern Canada obtained from the standard deviations from the ensemble of rjMCMC samples from the inversion Figure 6.6. In the following sections, “high” and “low” group velocity anomalies refer to fast or slow anomalies relative to the mean group velocity of the inversion data at each period. Overall, at all periods, the NCC appears to show greater variability in group velocity in comparison with the Canadian Shield.

At short periods (10, 15, 20 s; Figure 6.5), we observe a low group velocity anomaly beneath the Richardson Mountains and the northeastern region of Alaska. The central region of the NCC is underlain by moderate-to-high velocity anomalies. In particular, a high velocity anomaly is located in the southern part of the Mackenzie Mountains. Two low velocity anomalies encompass the Mackenzie Mountain fold and thrust belt on both ends. The isotropic group velocity uncertainty is low across northwestern Canada except the area in close proximity to the Cordillera Deformation Front (CDF), which exhibits higher uncertainties with respect to the rest of the NCC (Figure 6.6). To the east, a broad high velocity anomaly is located across the Canadian Shield.

At intermediate periods (25, 30, 36 s; Figure 6.5), a low group velocity anomaly is located beneath the Mackenzie Mountains. We observe a high velocity anomaly directly to the east of the Richardson Mountains in Northwest Territories. At 36 s, a band of high velocity extends from Northwest Territories to northeastern Alaska. A broad low velocity anomaly extends from the gulf of Alaska to the Denali Fault. Few to no structures are observed in the region across the Tintina and the Denali faults. We note that the isotropic group velocity

uncertainty is high (≥ 0.2 km/s) east of the CDF and in the central part of the Mackenzie Mountains (Figure 6.6).

At longer periods (42, 50, 60 s; Figure 6.5), a high group velocity anomaly extends across northern Yukon and Northwest Territories. At 60 s, it extends further south in Yukon and stops at the Tintina Fault. The Mackenzie Mountains are still underlain by a low velocity anomaly, although, its location is shifted to the west closer to the Tintina Fault. At 42 s, the slow velocity anomaly south of the Denali Fault is still present. At 60 s, we observe a succession of high and low anomalies in the region across the Tintina and Denali faults.

6.4.2 Azimuthal Anisotropy

Figure 6.7 shows the mean magnitude and directions of 2ψ azimuthal anisotropy estimated from the inversion at periods ranging from 10 to 50 s. Figure 6.8 shows the uncertainties in the fast-axis direction of azimuthal anisotropy. We focus our discussion of anisotropy results on regions of the model having at least one crossing ray and showing a low isotropic group velocity uncertainty for consistency. Moreover, the inversion cannot resolve azimuthal anisotropy at 60 s due to the sparse path coverage.

At short periods (< 20 s), the fast axis directions are oriented northwest-southeast, subparallel to the the Tintina and Denali faults. Fast axis directions for the Mackenzie Mountains are northwest-southeast and we observe a counter-clockwise rotation of fast axis directions in northern Yukon. East of the CDF, west of the Great Slave lake, the fast axis directions are north-south and rotate to be oriented northwest-southeast to the west of the Great Bear lake. The fast-axis directions mark the changes in direction of the CDF from south to north, as observed by McLellan et al. (2018). At longer periods, the northeast-southwest fast axis direction in northern Yukon persists and extends further south, across the Tintina Fault. Unfortunately, the inversion cannot resolve anisotropy in the Canadian Shield at long periods due to a sparser path coverage.

We note that some regions of the model show high uncertainties in fast-axis direction (Figure 6.8). In particular, at 15 s, a region in the central part of the NCC with dense path coverage shows no anisotropy (Figure 6.7) and high uncertainties (Figure 6.8). Figure 6.9 shows the azimuthal path coverage between 0° and 360° at two locations in the study region. First, at the Yukon/British Columbia border where the uncertainty is the lowest at 15 s (Figure 6.8). The path coverage is excellent and this region of the model is sampled uniformly in all directions (Figure 6.9). Second, in the central part of the NCC where we observe a high uncertainties in fast-axis direction at 15 s (Figure 6.8). Most of the surface

6.4 GROUP VELOCITY MAPS OVER NORTHWESTERN CANADA

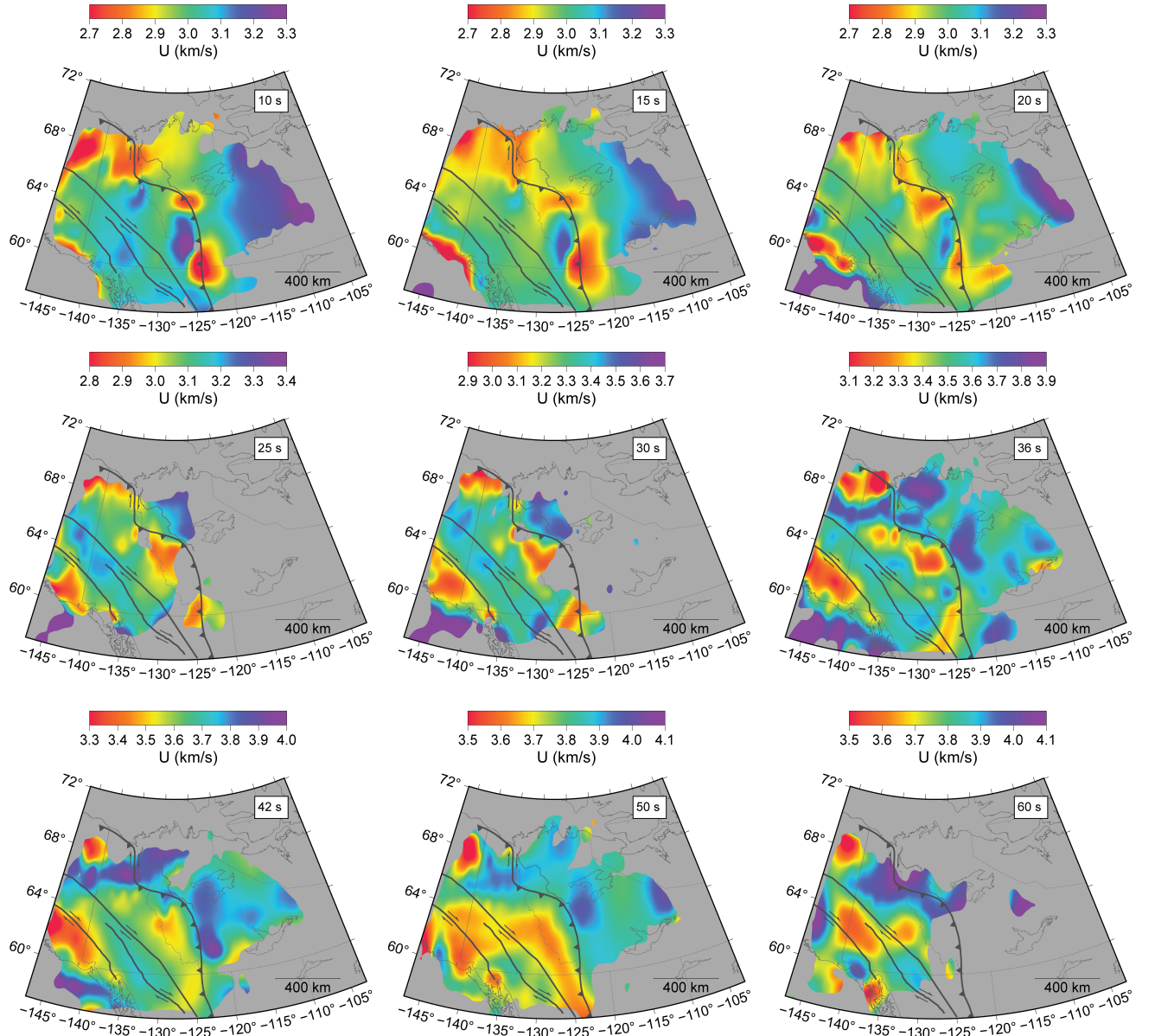


Figure 6.5.: Mean isotropic group velocity maps at periods ranging from 10 to 60 s estimated from the ensemble of rjMcMC samples. Note that the color scale differs and therefore mean velocity between maps. Areas with a high uncertainty and low hit count are masked in gray.

wave paths come from the northwestern direction ($\sim 280 - 290^\circ$; [Figure 6.9](#)). As there is a bias in the azimuthal path coverage, the inversion cannot confidently resolve anisotropy for this location, or any locations with poor azimuthal coverage.

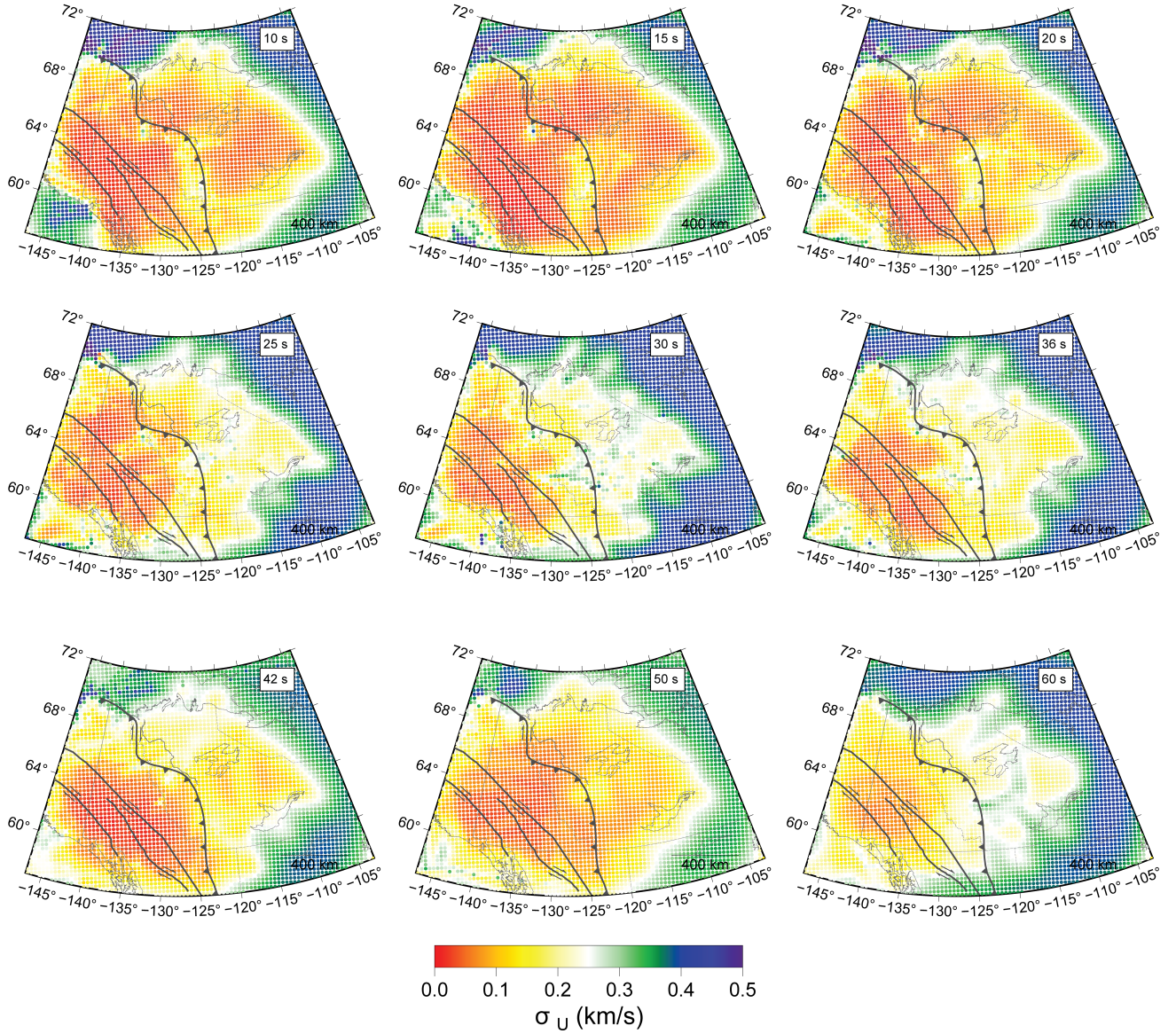


Figure 6.6.: Isotropic group velocity uncertainty estimated from the ensemble of rjMCMC samples at periods ranging from 10 to 60 s at each point of the underlying integration grid.

6.5 DISCUSSION

6.5.1 Isotropic crust and upper mantle structure

Group velocity anomalies reflect compositional changes in the crust and correlate well with geological features such as volcanic fields, regions consisting mainly of plutonic rocks and sedimentary basins at short period. At longer periods, group velocities are sensitive to crustal

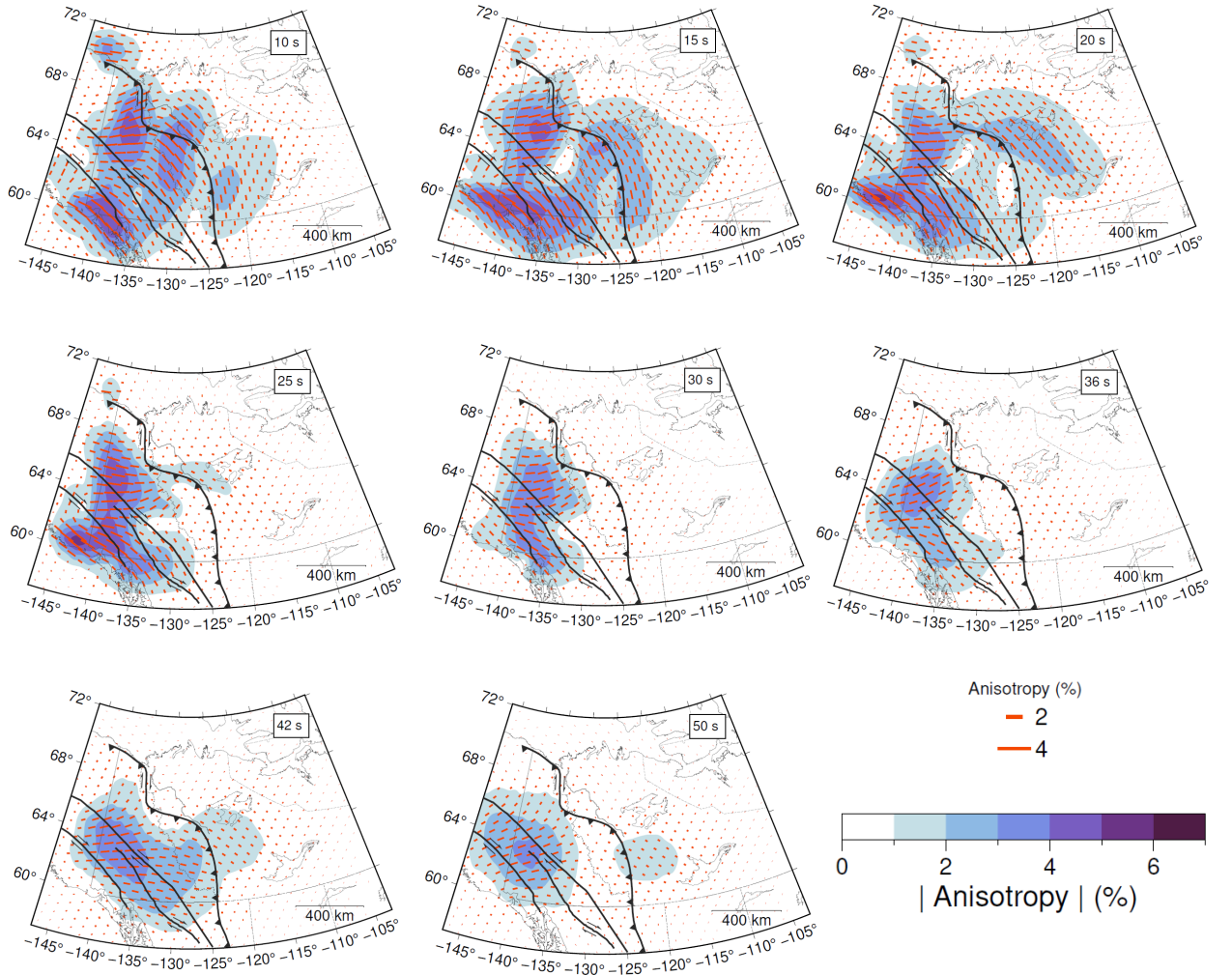


Figure 6.7.: Azimuthal anisotropy in group velocity over northwestern Canada. The mean magnitude and directions of 2ψ azimuthal anisotropy estimated from the ensemble of rjMcMC samples for inversions at periods ranging from 10 to 50 s are shown.

thickness variations and thermal anomalies located in the lower crust and/or uppermost mantle. Typically, regions underlain by low group velocity anomalies are interpreted to have a thicker crust and vice versa (Shapiro, 2005; Moschetti et al., 2007).

At short periods, our group velocity maps show high group velocity anomalies throughout the Canadian Shield, which likely reflects the igneous and metamorphic composition of the

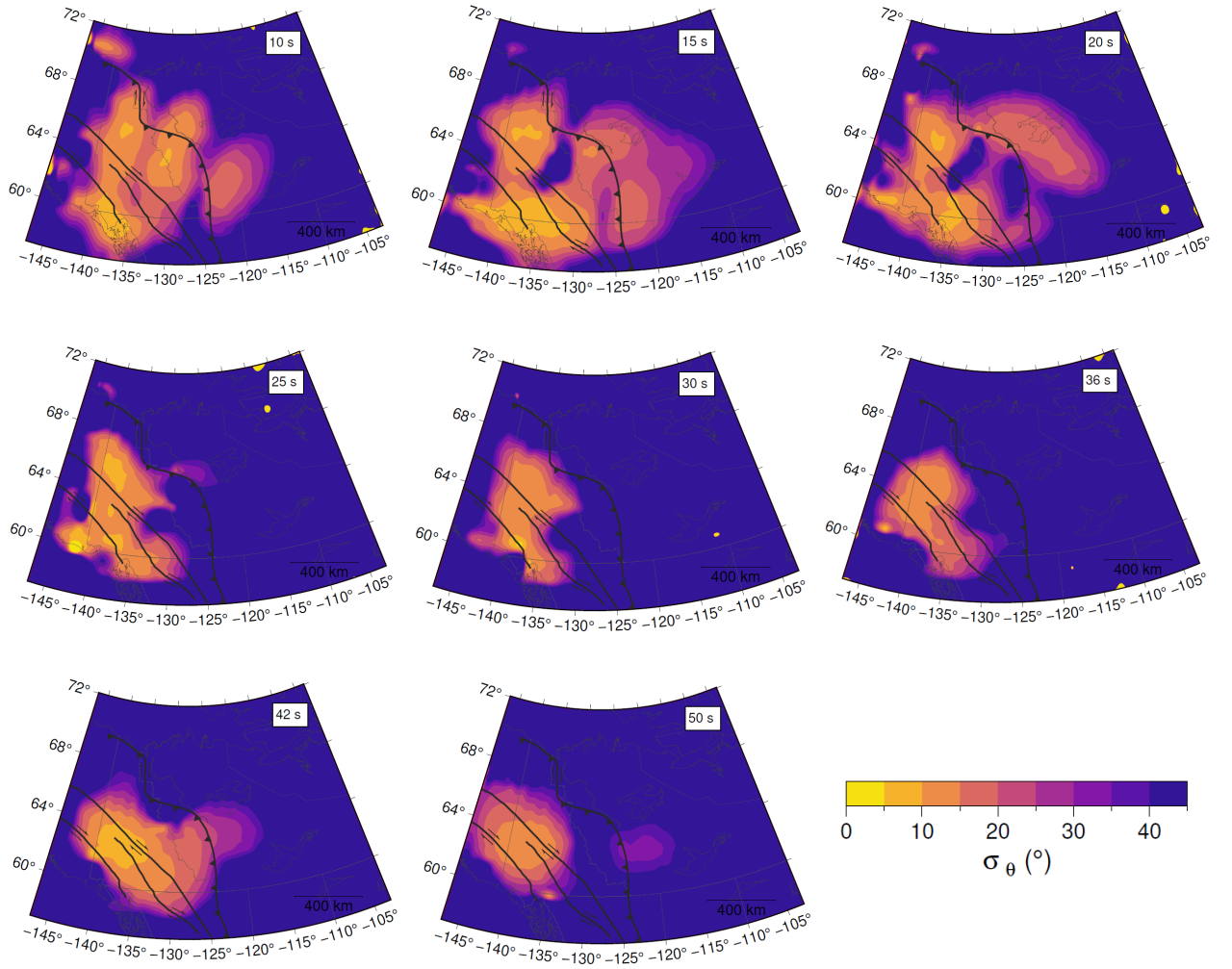


Figure 6.8.: Uncertainty of fast-axis direction of azimuthal anisotropy over northwestern Canada estimated from the ensemble of rjMCMC samples for inversions at periods ranging from 10 to 50 s are shown.

rocks within the craton. Sedimentary basins appear as low group velocity anomalies at short periods due to the low shear wave velocities of sediments (Moschetti et al., 2007). Low velocity anomalies correlate with the locations of thick sedimentary cover at 10 and 15 s in our group velocity maps such as the low velocity anomaly located at the Yukon/British Columbia border (Mooney & Kaban, 2010). Another interesting feature observed in our

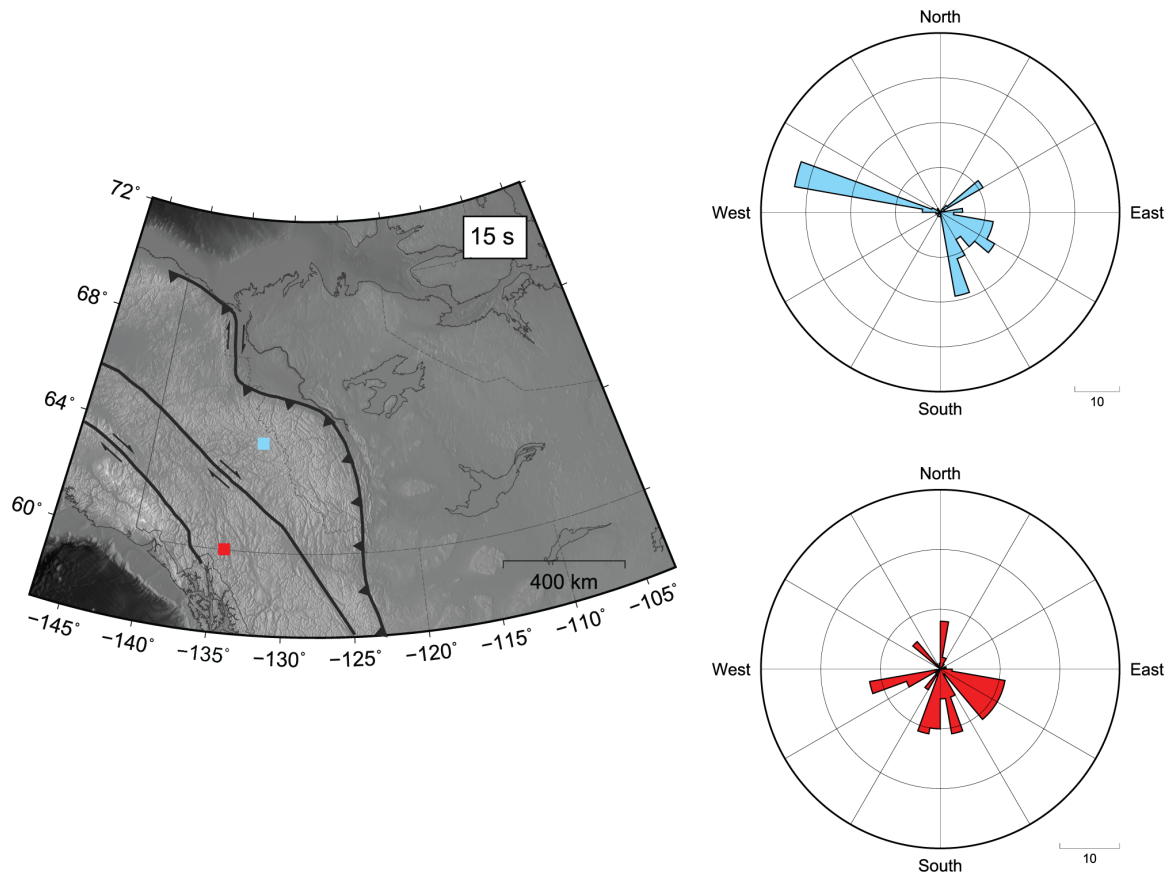


Figure 6.9.: (Left) Topographic map of northwestern Canada showing the locations of two grid points highlighted in blue and red. (Right) Rose diagrams showing the azimuthal path coverage for the two grid points located on the topographic map.

group velocity maps is the velocity anomaly located beneath the Richardson Mountains. At short periods, the region is underlain by a low velocity anomaly likely caused by the thick sedimentary cover (Mooney & Kaban, 2010). At longer period, the same feature appears as a high velocity anomaly in our model, which could indicate a thin crust in the area.

Our group velocity maps at periods of 30, 36 and 42 s reveal a low group velocity anomaly beneath the southeastern region of Alaska. As the low velocity anomaly is observed at intermediate periods, which are sensitive to crustal thickness variations, a plausible explanation for this low velocity anomaly is that it represents an area of thicker crust. This explanation is consistent with the current tectonics of the gulf of Alaska, as it is characterized by the collision between the Yakutat block and North America resulting in rapid uplift, intense seismicity and crustal thickening (e.g., Marechal et al., 2015). Moreover, this low group velocity anomaly also encompasses the Wrangell volcanic field (Figure 6.1), a recent volcanic arc

triggered by the subduction of the Yakutat block (Brueseke et al., 2019). Another potential explanation could be that this low velocity anomaly results from warmer lower crust and uppermost mantle. However, the Wrangell volcanic field does not extend as far east as the low velocity anomaly (Figure 6.1).

We also observe a pervasive low group velocity anomaly beneath the Mackenzie Mountains for periods between 20 and 60 s, although we note that the size and position of this low velocity anomaly vary at longer periods. The Mackenzie Mountains have recorded shortening and compressional deformation from Late Cretaceous to Early Cenozoic (Mazzotti et al., 2008; Hayward, 2015; Enkelmann et al., 2019). Furthermore, based on a number of geophysical studies, it has been suggested that the Mackenzie Mountains are underlain by basement rocks of Laurentian affinity (e.g., Snyder et al., 2002; Cook et al., 2004; Cook & Erdmer, 2005; Hayward, 2015). Therefore, we suggest that the low group velocity anomaly is indicative of a thicker crust beneath the Mackenzie Mountains, in agreement with crustal thickness estimates obtained using receiver functions across this region (Audet et al., 2020). The crustal thickening could be related to the thrusting of terranes at the Mackenzie Mountain fold and thrust belt and/or the stacking of crusts of Cordilleran and Laurentian affinity, or a combination of both.

At 50 s, which is predominantly sensitive to deep crustal and uppermost mantle structure, an arcuate low group velocity anomaly is observed beneath the NCC, suggesting that it is underlain by a hot lower crust and uppermost mantle. A positive thermal anomaly located in the deep crust and uppermost mantle beneath the NCC could be consistent with the up-welling of hot and dry asthenospheric mantle materials in the NCC through the opening of the slab window due to the subduction of a mid-oceanic ridge at 20 Ma (Thorkelson et al., 2011).

A high group velocity anomaly underlies northern Yukon and extends to the Tintina Fault at 60 s. A similar fast velocity anomaly is observed in teleseismic tomographic studies over Alaska and northwestern Canada (Jiang et al., 2018; Estève et al., 2020a). Estève et al. (2020a) suggest that this high velocity anomaly reflects the westward extent of the North American craton in the NCC and that it acts as a stiff buttress against the intervening mantle flow. This high velocity anomaly probably has a thermal origin, resulting from colder lower crust and uppermost mantle, consistent with the hypothesis of the Mackenzie craton in northern Yukon.

Lastly, high group velocity anomalies extend west of the CDF in the northwestern region of the Mackenzie Mountains, although we acknowledge the sparse path coverage at 60 s. The westward extent of fast velocity anomalies could show that the Canadian cratonic lithosphere extends westward in the Mackenzie Mountains and connects with the fast velocity anomaly

interpreted as the Mackenzie craton, as it was suggested by [McLellan et al. \(2018\)](#). This implies that the Cordillera-Craton lithospheric boundary is located westward of the CDF, as has been observed in the Southern Canadian Cordillera ([Chen et al., 2019](#)).

6.5.2 *Seismic Anisotropy*

Seismic anisotropy in the crust is more challenging to interpret compared to the upper mantle. This is attributed to the heterogeneous crustal composition compared to the homogeneous upper mantle. However, upper crustal seismic anisotropy is interpreted to be caused by several mechanisms such as vertically-aligned fluid-filled cracks parallel to the direction of maximum horizontal compressive stress, microcracks, and preferentially oriented pore space ([Crampin, 1994](#)). At greater crustal depths, seismic anisotropy is attributed to metamorphic rocks containing aligned anisotropic minerals such as micas, amphiboles or orthopyroxenes (i.e., lattice preferred orientation) ([Moschetti et al., 2010](#)). At upper mantle depths, seismic anisotropy results from the lattice preferred orientation of olivine crystals, providing constraints on the upper mantle flow ([Tommasi et al., 2000](#)).

At short periods, the fast axis directions are oriented northwest-southeast and are aligned with the main tectonic structures throughout the NCC with one noticeable exception in northern Yukon where the fast axis directions rotate to be northeast-southwest. A similar pattern of anisotropy is observed in a Rayleigh wave phase velocity tomography study over the NCC ([McLellan et al., 2018](#)). [Rasendra et al. \(2014\)](#) suggest that this coherent northwest-southeast alignment of fast axis directions reflects the signature of a fabric associated with the orientation of accreted Mesozoic terranes onto the margin. Moreover, fast axis directions are subparallel to the surface expression of the Denali and Tintina faults, which reflect long-term shearing along those faults ([Rasendra et al., 2014](#); [Tarayoun et al., 2017](#); [McLellan et al., 2018](#)).

At longer periods, the northwest-southeast alignment of fast axis directions is still present in the southern region of the NCC and is overprinted by a northeast-southwest alignment in the northwestern region of the NCC. This observed northeast-southwest alignment of fast axis directions across the northwestern region of the NCC may be consistent with the modeled basal traction inferred to be generated by mantle flow beneath Alaska and northwestern Canada ([Finzel et al., 2015](#)). The orogenic float model ([Oldow et al., 1990](#)) has been invoked in the NCC to explain thrust faulting in the Mackenzie Mountains and stress transfer to the CDF ([Mazzotti & Hyndman, 2002](#)). However, our results may support the hypothesis that

stresses can be transmitted via large-scale mantle flow throughout the northwestern region of the NCC rather than a mechanically-weak lower crust.

6.6 CONCLUSION

In this chapter, we successfully invert Rayleigh wave group velocity dispersion measurements from regional earthquakes within a Bayesian framework in order to estimate azimuthally anisotropic group velocity maps throughout northwestern Canada. Short period group velocity anomalies correlate well with geological features. The present-day NCC was previously thought to be characterized by a thin crust (~ 33 km; Hyndman, 2017), however our results suggest that there are crustal thickness variations within the NCC. In particular, low group velocity anomalies beneath the Mackenzie Mountains and the gulf of Alaska are interpreted to reflect a thicker crust. At 50 s, a broad arcuate low group velocity anomaly beneath the NCC is interpreted to be a hot thermal anomaly in the lower crust and uppermost mantle. Such thermal anomaly could reflect uprising hot and dry asthenospheric materials caused by the opening of a slab window in the Cenozoic. A long period high group velocity anomaly underlies northern Yukon, which we interpret as a cold thermal anomaly located in the lower crust and uppermost mantle. This supports the hypothesis of the Mackenzie craton, which marks the westward extent of the North American craton in the NCC. Our results also suggest that the Cordillera-Craton boundary extends west of the CDF in the northwestern Mackenzie Mountains. Fast axis directions of azimuthal anisotropy are subparallel to the main structures across the NCC at short periods. At long periods, the northeast-southwest alignment of fast axis directions may be consistent with large-scale mantle flow. The following step in this analysis will be to retrieve the pseudo 3-dimensional shear wave velocity structure from our group velocity maps in order to better constrain the location of these velocity anomalies within the crust and the uppermost mantle.

Part IV

CONCLUSION AND REFERENCES

CONCLUSIONS

The lithosphere of northwestern Canada recorded a complex tectonic evolution over 2.5 Gy from the formation of ancient cores of continental lithosphere such as the Slave craton to the Phanerozoic Cordilleran Orogeny (Hoffman, 1988). These major tectonic events shaped and modeled northwestern Canada. The deployment of recent seismic networks such as the Earthscope USArray-Transportable Array, the Mackenzie Mountain Earthscope Project, and the Yukon Northwest Seismograph Network enables the development of seismic tomography models at an unprecedented high resolution in northwestern Canada. In this thesis, we investigate the crustal and upper mantle seismic velocity structure and the upper mantle dynamics of northwestern Canada. The main observations and interpretations from this thesis are as follows.

In Chapter 3, we investigate the upper mantle seismic velocity structure beneath the diamondiferous Slave craton from the inversion of teleseismic P- and S-wave travel times. The P-wave model reveals that the central Slave craton is underlain by an alternating pattern of small-scale high and low velocity P-wave anomalies in the lithospheric mantle. In particular, two low-velocity P-wave anomalies encompass the latest kimberlite eruptions in the Lac de Gras kimberlite field, which are interpreted to reflect metasomatisation in the lithospheric mantle. Furthermore, the Lac de Gras kimberlite field is located where steep V_P and V_S gradients occur. The S-wave model displays a pervasive low-velocity anomaly in the asthenosphere, which may reflect a potential kimberlite magma ascent in the asthenosphere.

In Chapter 4, we extend our tomography models of the Slave craton westward to include all of northwestern Canada. Both P- and S-wave models reveal a laterally heterogeneous seismic velocity structure beneath the NCC. The tomography models support the presence of a newly identified Mackenzie craton in northwestern Canada (Schaeffer & Lebedev, 2014), which acts as a stiff buttress that helps control the arcuate shape of the Cordilleran front during orogeny. Moreover, strong P and S wave velocity contrasts across the Tintina Fault

suggest that it is a lithosphere-scale shear zone that extends into the upper mantle beneath the NCC and demarcates distinct regions of lithospheric mantle.

[Chapter 5](#) provides seismic evidence for the trans-lithospheric nature of the Tintina Fault in the NCC. Using our 3-dimensional P-wave model, we resolve a series of sharp P-wave velocity contrasts at uppermost mantle depths beneath the surface trace of the Tintina Fault. Estimates of seismic anisotropy reveal similar changes in the orientation and magnitude of anisotropy in the vicinity of the Tintina Fault. Together, these results suggest that the Tintina Fault is a lithospheric-scale shear zone. Moreover, after restoration of 430 km of right-lateral displacement on the fault, two high-velocity P-wave anomalies align with the outline of the North American craton margin. We propose that the high-velocity anomaly structure currently located in eastern Alaska represents fragment of the Mackenzie craton that has been chiseled and displaced to the northwest by the TF between the Late Cretaceous and the Eocene. A second cratonic fragment currently located in the southern NCC may be associated with the Cassiar terrane at upper mantle depth.

[Chapter 6](#) presents a new crustal and uppermost mantle seismic velocity model obtained from the inversion of surface-wave group-velocity dispersion measurements within a Bayesian framework. The group velocity maps for periods ranging between 10 and 60 s reveal a heterogeneous crustal velocity structure across the NCC. In particular, the Mackenzie Mountains and the gulf of Alaska are underlain by a low group velocity anomaly, which we interpret to reflect a thicker crust in comparison with the rest of the NCC. The crustal overthickening in the Mackenzie Mountains could be attributed to the thrusting of terranes at the Mackenzie Mountain fold and thrust belt and/or to the stacking of crusts of Cordilleran and Laurentian affinity or a combination of both, whereas in the gulf of Alaska we attribute it to the underthrusting of the Yakutat block beneath the North American margin. At long periods (≥ 50 s), a broad arcuate low group velocity anomaly extends across the NCC, which we interpret as a hot thermal anomaly located in the lower crust and uppermost mantle. This may reflect up-welling of dry and anhydrous materials from the underlying asthenosphere and is consistent with the slab window hypothesis. Northern Yukon is characterized by a high group velocity anomaly at 60 s, which is interpreted as the westward extent of the North American craton. The Cordillera-Craton lithospheric boundary extends west of the CDF in the northwestern Mackenzie Mountains.

Our seismic tomography models reveal a complex seismic velocity structure from the crust to the upper mantle. In particular, our teleseismic body-wave tomography models and isotropic Rayleigh-wave group-velocity map at 60 s show a high seismic velocity anomaly underlying northern Yukon ([Figure 7.1](#)). Interestingly, these are co-located with high-amplitude long-wavelength magnetic (MRA - Mackenzie River Anomaly, [Pilkington & Saltus, 2009](#)) and

Bouguer anomalies (Figure 4.1), low heat flow (≤ 45 mW/m²) and deep Curie point depth estimates (Witter et al., 2018). Lastly, SKS splitting measurements across northern Yukon and Northwest Territories are subparallel to the general trend of fast axes observed in the Canadian Shield. This observation implies that the two regions have preserved similar fabrics. These geophysical evidence are interpreted to reflect the presence of the Mackenzie craton, the westward extent of the North American craton in the NCC.

The seismic signature of the cratonic fragment of the Mackenzie craton currently located in eastern Alaska does not appear as clearly in our surface-wave tomography model as in our teleseismic body-wave tomography models (Figure 7.1). At the longest periods considered in Chapter 6 (42, 50 and 60 s), which are predominantly sensitive to uppermost mantle structure, group velocity anomalies are small in magnitude but large in spatial extent, reflecting the effect of long wavelength surface wave data. Therefore, the seismic signature of the cratonic fragment may be too small to be seen at those periods. Note that regional surface-wave tomography models over northwestern Canada and Alaska do not show it either (Wang & Tape, 2014; Ward, 2015; McLellan et al., 2018; Feng & Ritzwoller, 2019).

Figure 7.2 shows all available SKS splitting measurements (red bars) across northwestern Canada and the Rayleigh-wave group-velocity 2ψ azimuthal anisotropy at 50 s (blue bars). The general pattern of seismic anisotropy across northwestern Canada obtained from SKS splitting measurements differs from our azimuthally-anisotropic surface-wave tomography model. Seismic anisotropy derived from surface-wave tomography and SKS splitting measurements does not always agree with each other (Bao et al., 2016; Foster et al., 2020; Montagner et al., 2000; Wüstefeld et al., 2009; Becker et al., 2012) and this may be due to the different resolution of body and surface waves. Teleseismic core-refracted shear waves (i.e., SKS, SKKS) provide seismic anisotropy measurements with a good lateral, but poor vertical resolution due to the ray geometry (i.e., steep incident angle beneath the station) as opposed to surface-wave tomography, which provides a relatively good vertical resolution, but a poor lateral resolution (Montagner et al., 2000). Surface-waves will tend to average out anisotropy, whereas SKS splitting measurements will resolve local lateral variations. Then, teleseismic shear wave splitting usually assumes only one anisotropic layer beneath the seismic station. However, we observe with our azimuthally-anisotropic surface-wave model that this is a poor assumption and that the direction of seismic anisotropy can rotate as a function of depth. Lastly, seismic anisotropy derived from SKS splitting measurements and surface-wave tomography cannot be directly compared, as the two methods provide information on seismic anisotropy along vertical and horizontal propagation directions (Wüstefeld et al., 2009). Nonetheless, SKS splitting measurements and azimuthally-anisotropic surface-wave model can be related (Montagner et al., 2000; Wüstefeld et al., 2009; Becker et al.,

2012) through the comparison between the two types of data and by calculating predicted splitting parameters from the anisotropic tomography model. Coincident predicted SKS fast directions with observed SKS fast directions are generally interpreted in terms of vertically coherent lithospheric deformation while any discrepancy between the two would indicate depth-variant crust-mantle deformation.

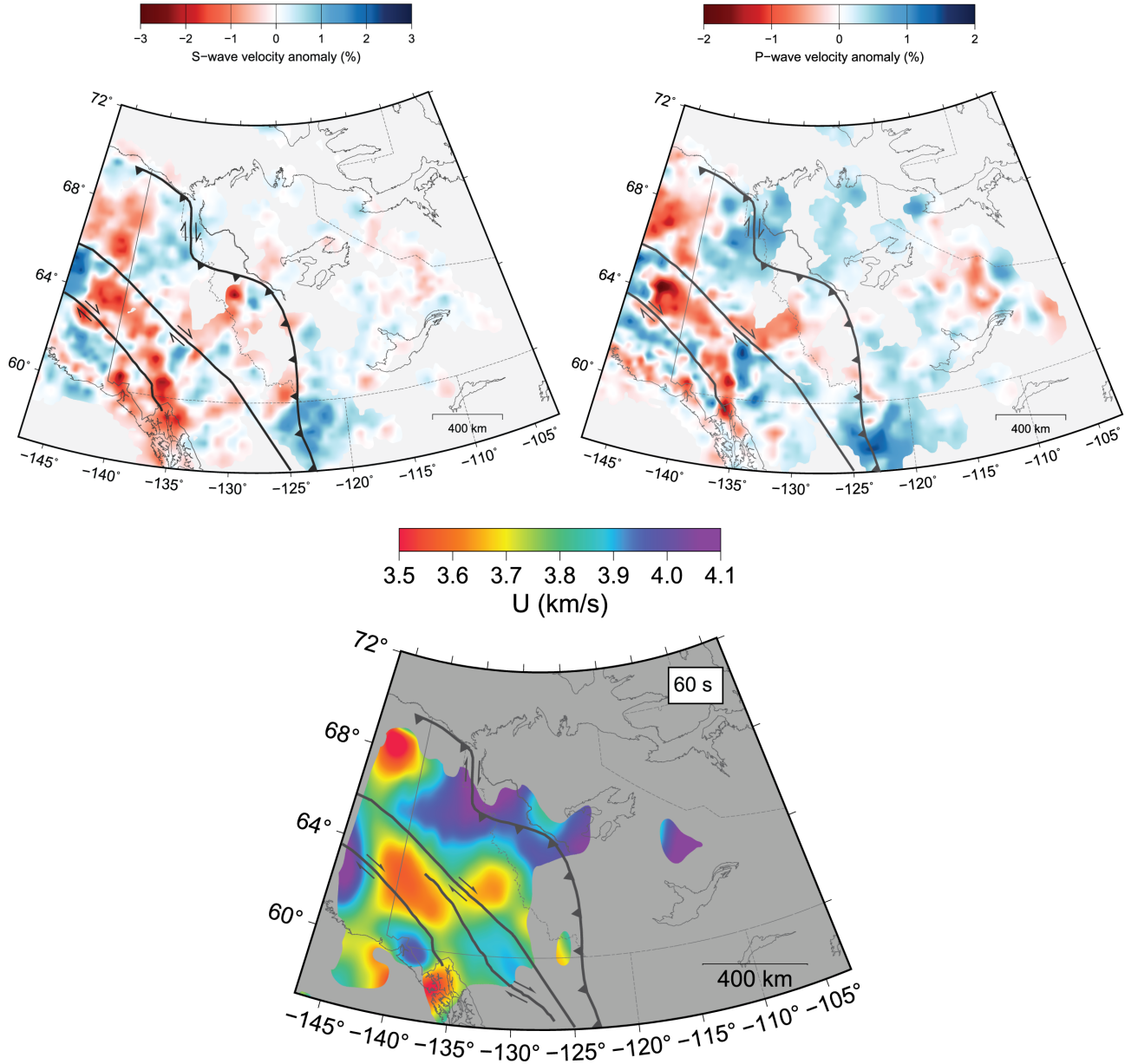


Figure 7.1.: Teleseismic P- (top left) and S-wave (top right) models at 100 km depth across north-western Canada from (Estève et al., 2020a). Mean isotropic group velocity map at 60 s (bottom).

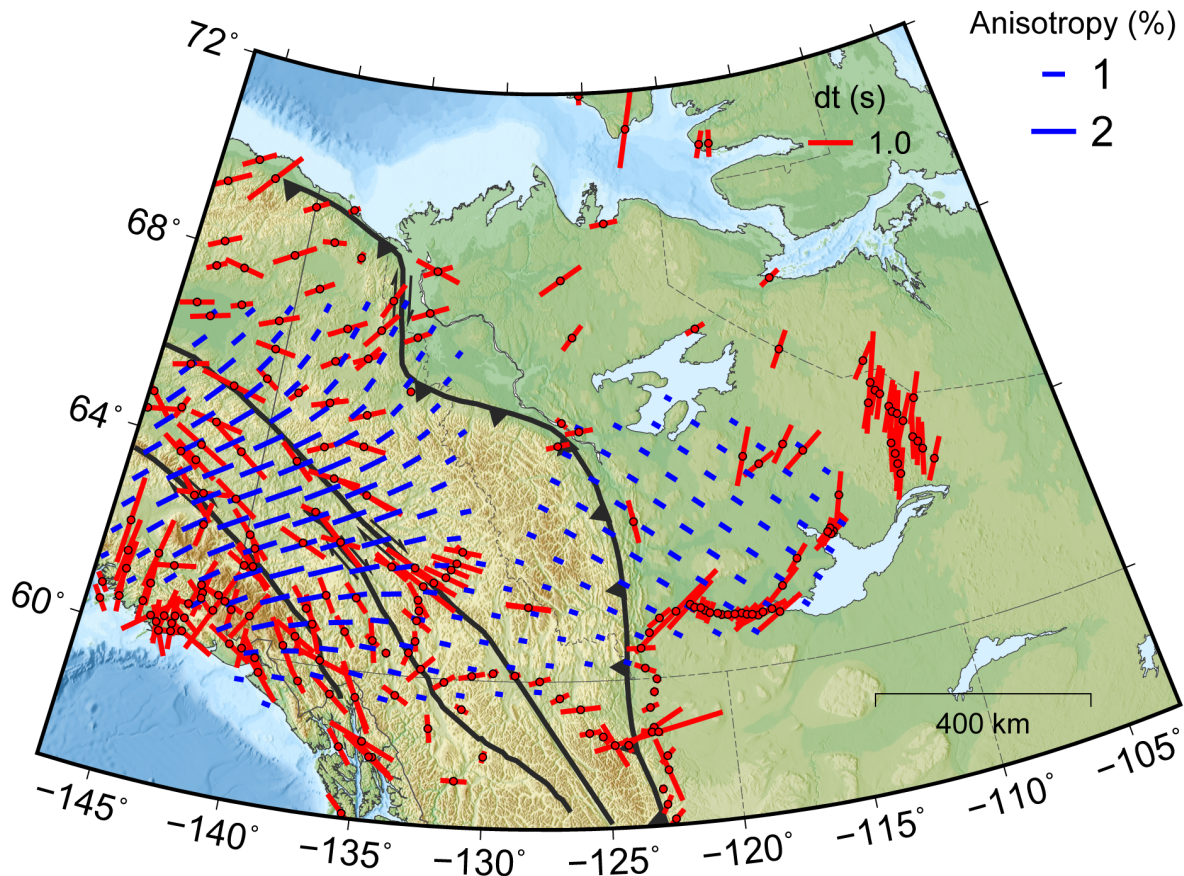


Figure 7.2.: Rayleigh-wave group-velocity 2ψ azimuthal anisotropy at 50 s (blue bars) across northwestern Canada. Red bars show compiled SKS splitting measurements from Audet et al. (2016); Courtier et al. (2010); Rasendra et al. (2014); Snyder & Bruneton (2007); Venereau et al. (2019); Estève et al. (2020b).

The passive-source seismic studies that we conducted in this thesis shed light upon the crustal and upper mantle seismic velocity structure of northwestern Canada and, in particular, across the NCC due to a denser station coverage. The results obtained provide new insight into the fundamental questions formulated in Chapter 1, however, other fundamental questions remain unanswered such as:

- What is the depth range of strength concentration within the NCC?
- What is the viscosity structure of the lithosphere in the NCC ?
- Is there some sort of linkage between the seismicity occurring in the Richardson and the Mackenzie Mountains ?

These questions are tied to the necessity of collecting new data given the sparse temporal and spatial sampling of the region. For example, the deployment of new seismic stations

across northwestern Canada in target areas would help in addressing some of these questions. The current distribution of seismic stations is uneven across northwestern Canada. The rapid increase in seismic stations operating in the NCC led to the formulation of new hypotheses on the tectonic evolution of the region. However, the station coverage in regions such as Northwest Territories remains sparse leading to unconstrained areas in geophysical models and therefore limits the understanding of the tectonic evolution of the region. In particular, the work done in the Slave craton highlighted the limitations of the tomographic method in regions where the coverage of seismic stations is sparse.

Moreover, future work should investigate the subsurface seismic structure of the NCC through the development of new seismic tomography methods to enhance the quality of the resulting subsurface images. For instance, by performing a joint inversion of receiver functions and surface wave tomography model to obtain a refined 3-dimensional shear wave velocity model (e.g., [Ward & Lin, 2018](#)) and by using the Moho refracted phase, P_n , to estimate temperature and viscosity in the uppermost mantle (e.g., [Schutt et al., 2018](#)) across the NCC. Recently, [Boyce et al. \(2019\)](#) successfully inverted absolute P-wave arrival times from teleseismic earthquakes to obtain the absolute 3-dimensional upper mantle seismic velocity structure across eastern Canada. Applying such methods to northwestern Canada would be complementary to absolute shear wave velocity models (e.g., [Schaeffer & Lebedev, 2014](#)) and would enable further understanding of the cratonic lithosphere evolution and preservation through time.

Lastly, future efforts should be made to better constrain the spatial and depth distribution of the seismicity across the NCC. Estimates of lithospheric thickness, temperature and composition would be helpful for future geodynamic modelling and magnetotelluric studies would be complementary to seismic studies in the region.

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Part V

APPENDICES

SUPPORTIVE INFORMATION FOR "UPPER MANTLE STRUCTURE UNDERLYING THE DIAMONDIFEROUS SLAVE CRATON FROM TELESEISMIC BODY WAVE TOMOGRAPHY"

A.1 LIST OF FIGURES

- [Figure A.1](#) shows the S-wave checkerboard test. This resolution test permits to assess the robustness of the model.
- [Figure A.2](#) shows the P-wave structural test. We performed this resolution test in order to assess the effect of vertical and/or horizontal smearing occurring through our P-wave model.
- [Figure A.3](#) shows the S-wave structural test. We performed this resolution test in order to assess the effect of vertical and/or horizontal smearing occurring through our S-wave model.
- [Figure A.4](#) shows the P-wave model trade-off curve between relative travel-time residual reduction (%) and model roughness ($\text{s/km}^3 \times 10^{-7}$).
- [Figure A.5](#) shows the S-wave model trade-off curve between relative travel-time residual reduction (%) and model roughness ($\text{s/km}^3 \times 10^{-7}$).
- [Figure A.6](#) shows the normalized lateral velocity gradient computed from the P-wave model at 100, 150, 200, 250, 300 and 350 km depth. Light yellow color and red-black color correspond to small and high lateral gradient respectively.
- [Figure A.7](#) shows the normalized lateral velocity gradient computed from the S-wave model at 100, 150, 200, 250, 300 and 350 km depth. Light yellow color and red-black color correspond to small and high lateral gradient respectively.

A.2 LIST OF TABLES

Table A.1 shows the weights applied to the flattening and smoothing parameters associated with the model roughness and the travel-time residual reduction for the P-wave model. The smoothest models do not fit data (*i.e.*, model I) whereas models with the minimum residual reduction are rough (*i.e.*, model A).

Table A.2 shows the weights applied to the flattening and smoothing parameters associated with the model roughness and the travel-time residual reduction for the S-wave model. The smoothest models do not fit data (*i.e.*, model I) whereas models with the minimum residual reduction are rough (*i.e.*, model A).

A.3 DATASET INFORMATION

We also provide a "tecto-manuscript-data.tar.gz" file containing P- and S-wave tomographic models, a README file and a guide to VanDecar's code.

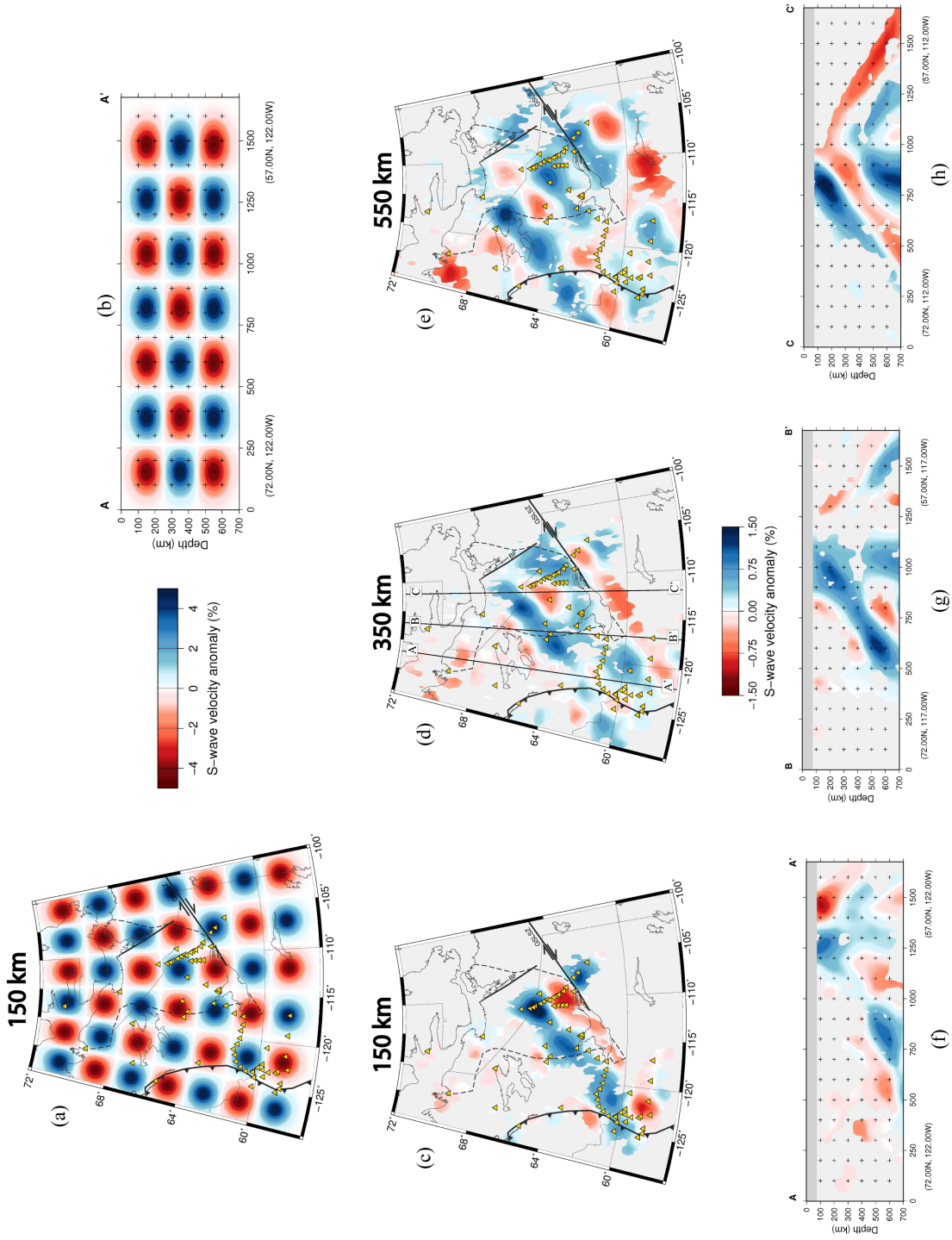


Figure A.1.: S-wave checkerboard resolution test with spheres of radius 75 km defined by Gaussian functions across their diameter. Each sphere corresponds to an anomaly of $\delta V_S = \pm 5\%$ and they are distributed in successive layers of positive and negative anomalies at depths of 150, 350 and 550 km. (a and b) 150 km depth slice and cross section through the synthetic checkerboard test (input). (c, d and e) Output of the checkerboard resolution test : depth slices at 150, 350 and 550 km depths. (f,g and h) Cross sections through the retrieved checkerboard model (the cross-section locations are displayed on the 350 km depth slice). The Cordillera Deformation front, the Great Slave Lake Shear Zone and the Bathurst fault are shown in black and areas with less than 4 ray hits are displayed in light gray.

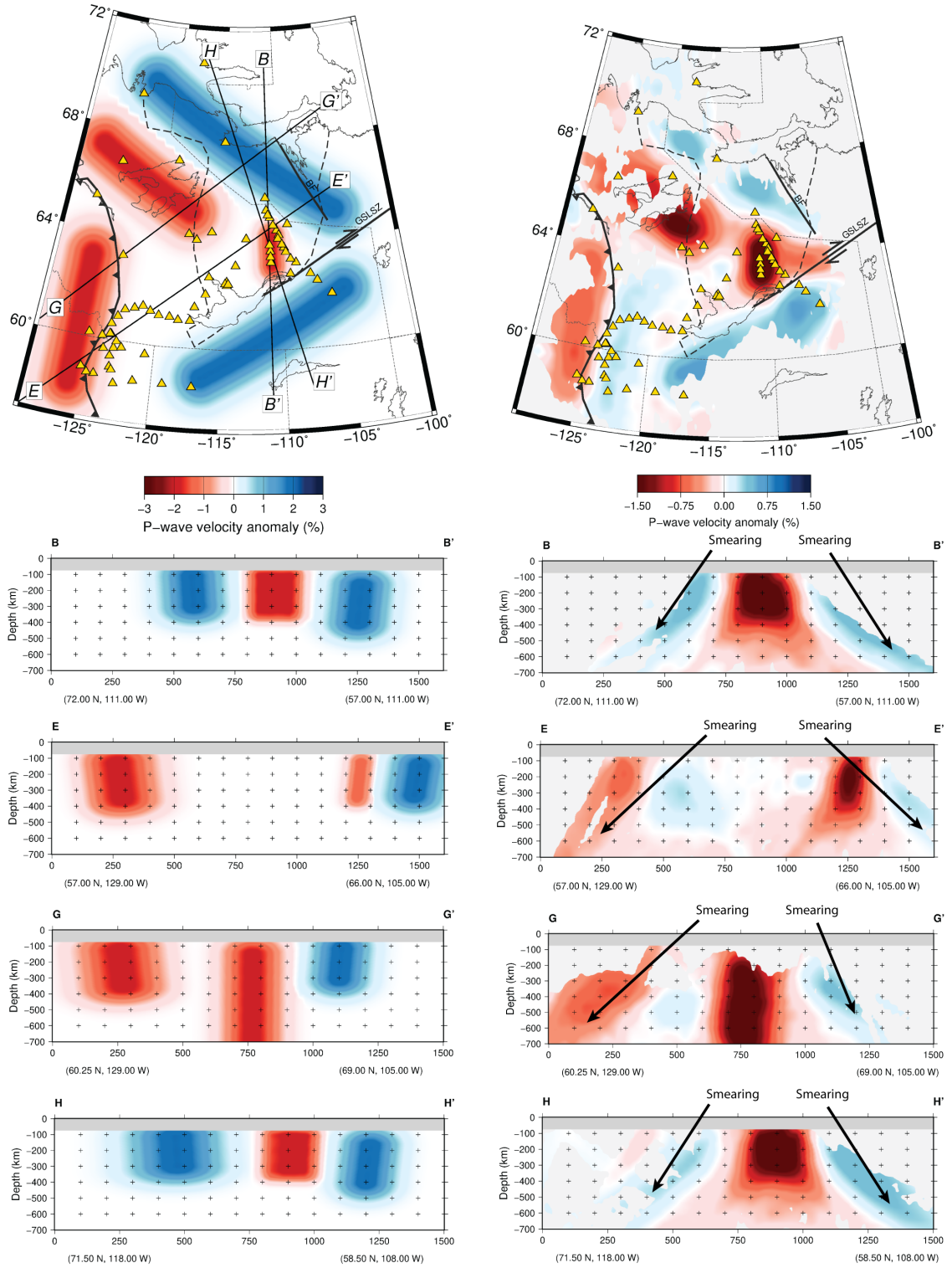


Figure A.2.: Cross-sections and depth slices through the P-wave structural resolution test. Structural resolution tests showing input anomalies (left) and output anomalies (right) through our P-wave model. Synthetic anomalies are defined by Gaussian functions across their widths. Also shown, station locations (yellow triangles), the Cordillera Deformation front, the Great Slave Lake Shear Zone and the Bathurst fault in black. Areas with less than 4 ray hits are displayed in light gray and the first 75 km are masked.

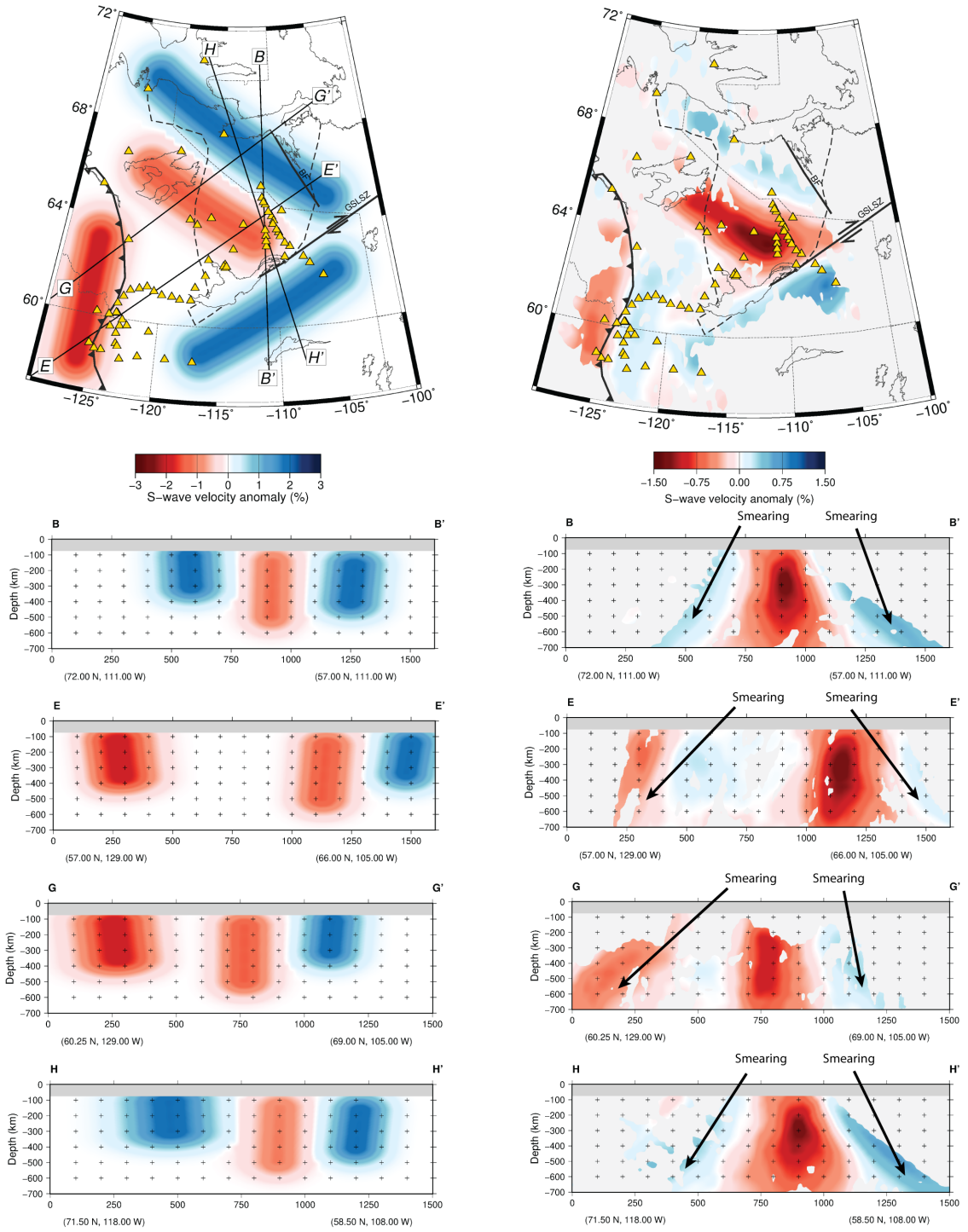


Figure A.3.: Cross-sections and depth slices through the S-wave structural resolution test. Same format as [Figure A.2](#).

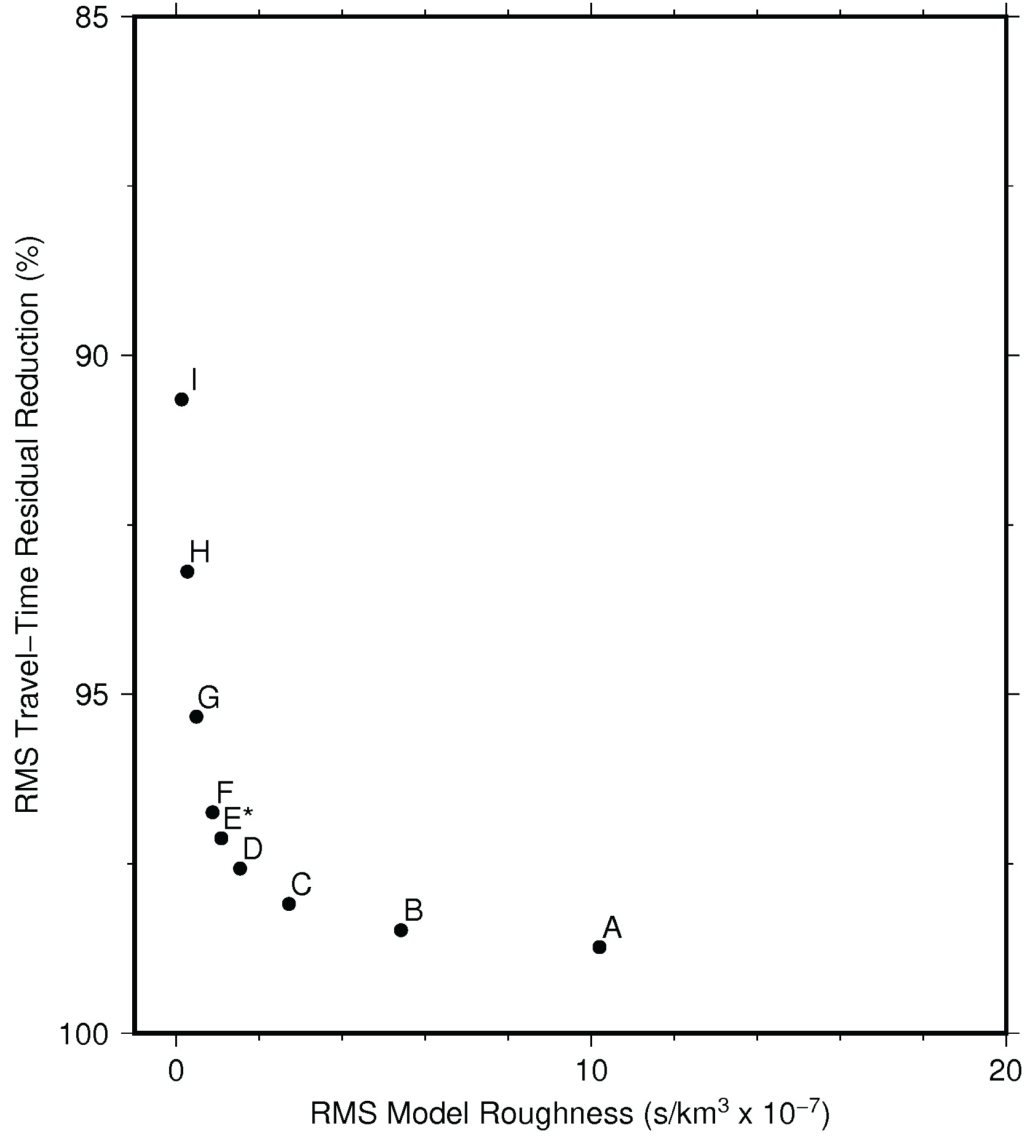


Figure A.4.: Trade-off curve between relative travel-time residual reduction (%) and model roughness ($\text{s/km}^3 \times 10^{-7}$) for the P-wave model. The letters A-I correspond to inversions with different flattening (λ_{flat}) and smoothing parameters (λ_{smooth}). The letter E* represents our preferred model.

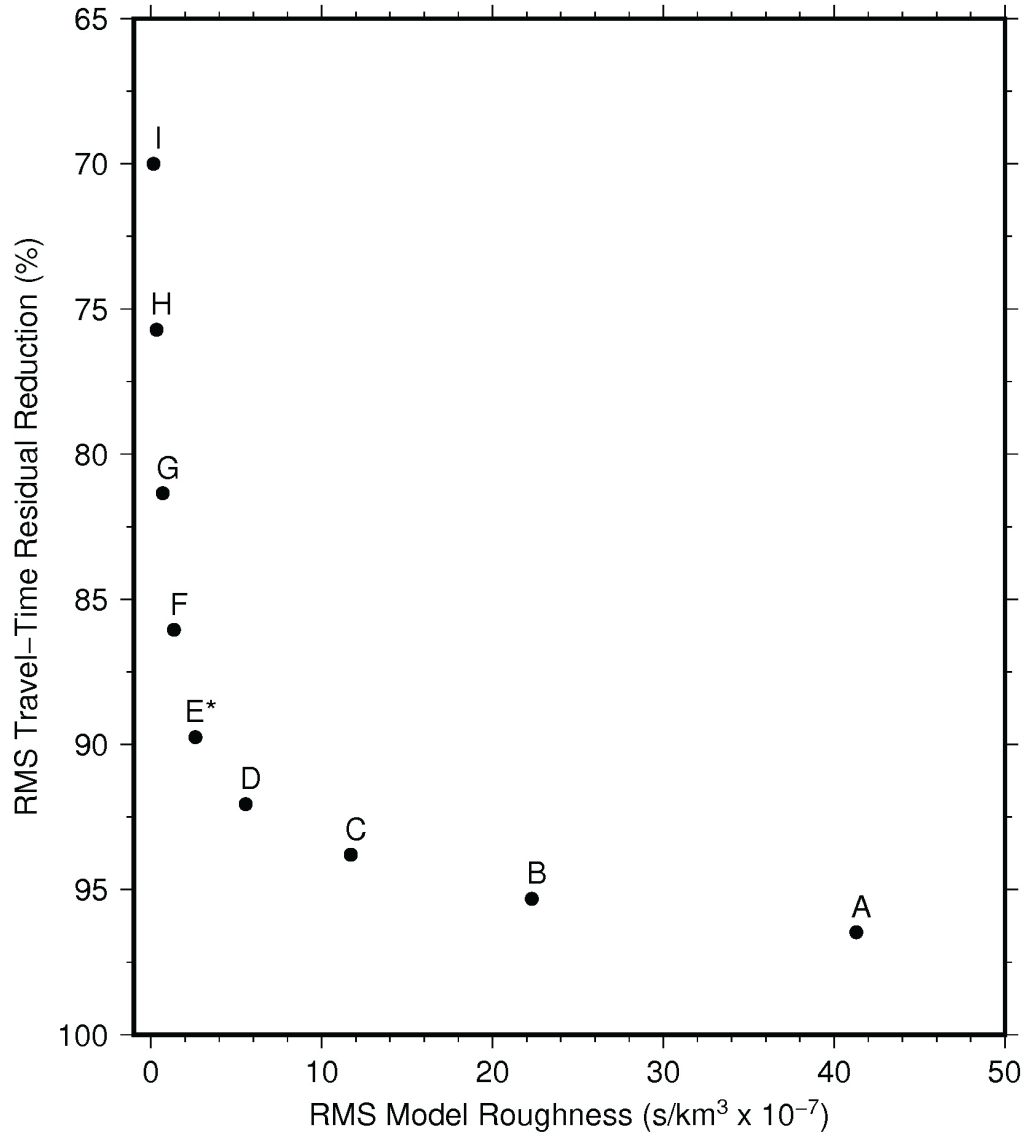


Figure A.5.: Trade-off curve between relative travel-time residual reduction (%) and model roughness ($\text{s/km}^3 \times 10^{-7}$) for the S-wave model. The letters A-I correspond to inversions with different flattening (λ_{flat}) and smoothing parameters (λ_{smooth}). The letter E* represents our preferred model.

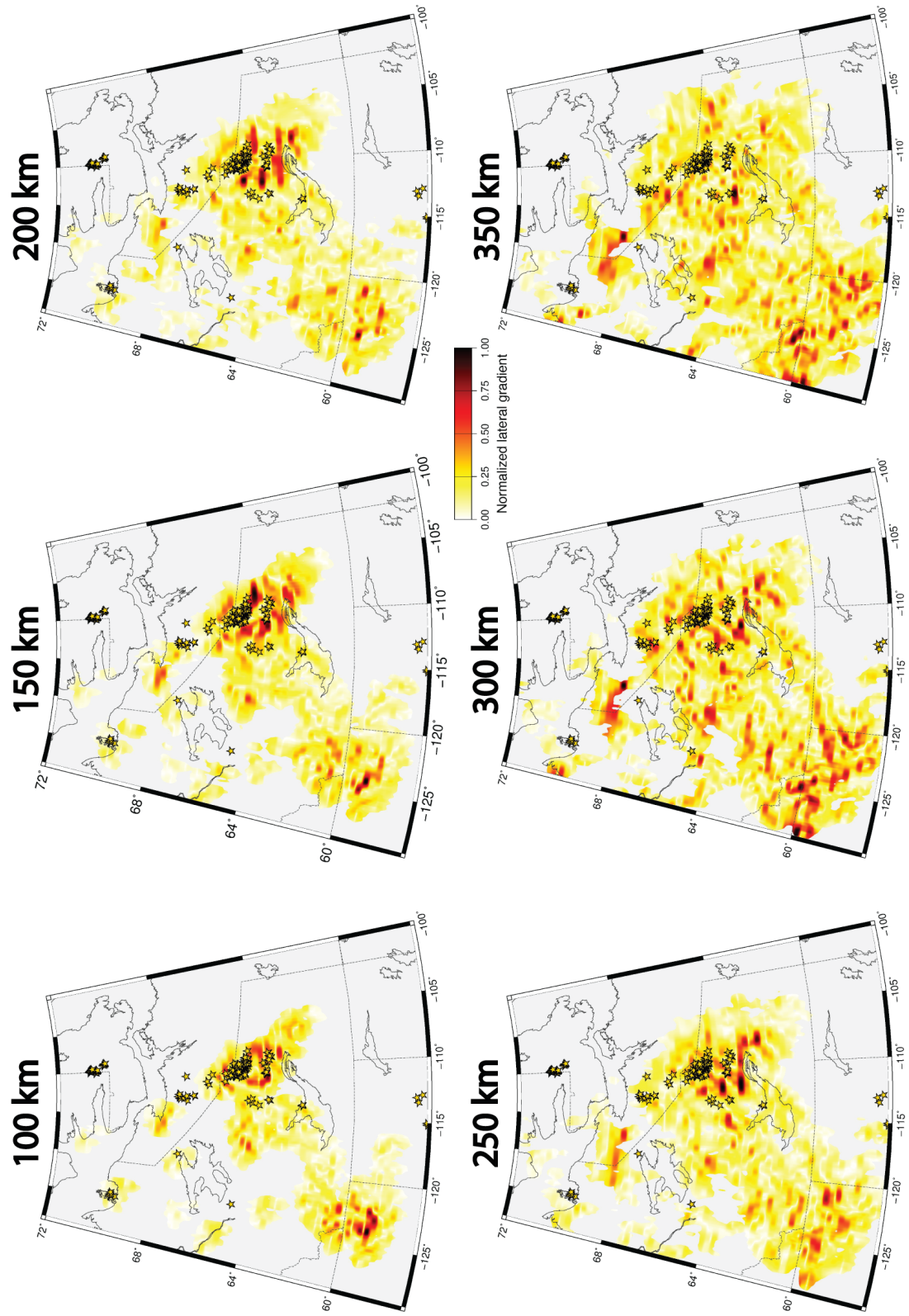


Figure A.6.: Maps showing the normalized lateral velocity gradient computed from the P-wave model at 100, 150, 200, 250, 300 and 350 km depth. Light yellow color and red-black color correspond to small and high lateral gradient respectively. Yellow stars show the location of kimberlite pipes.

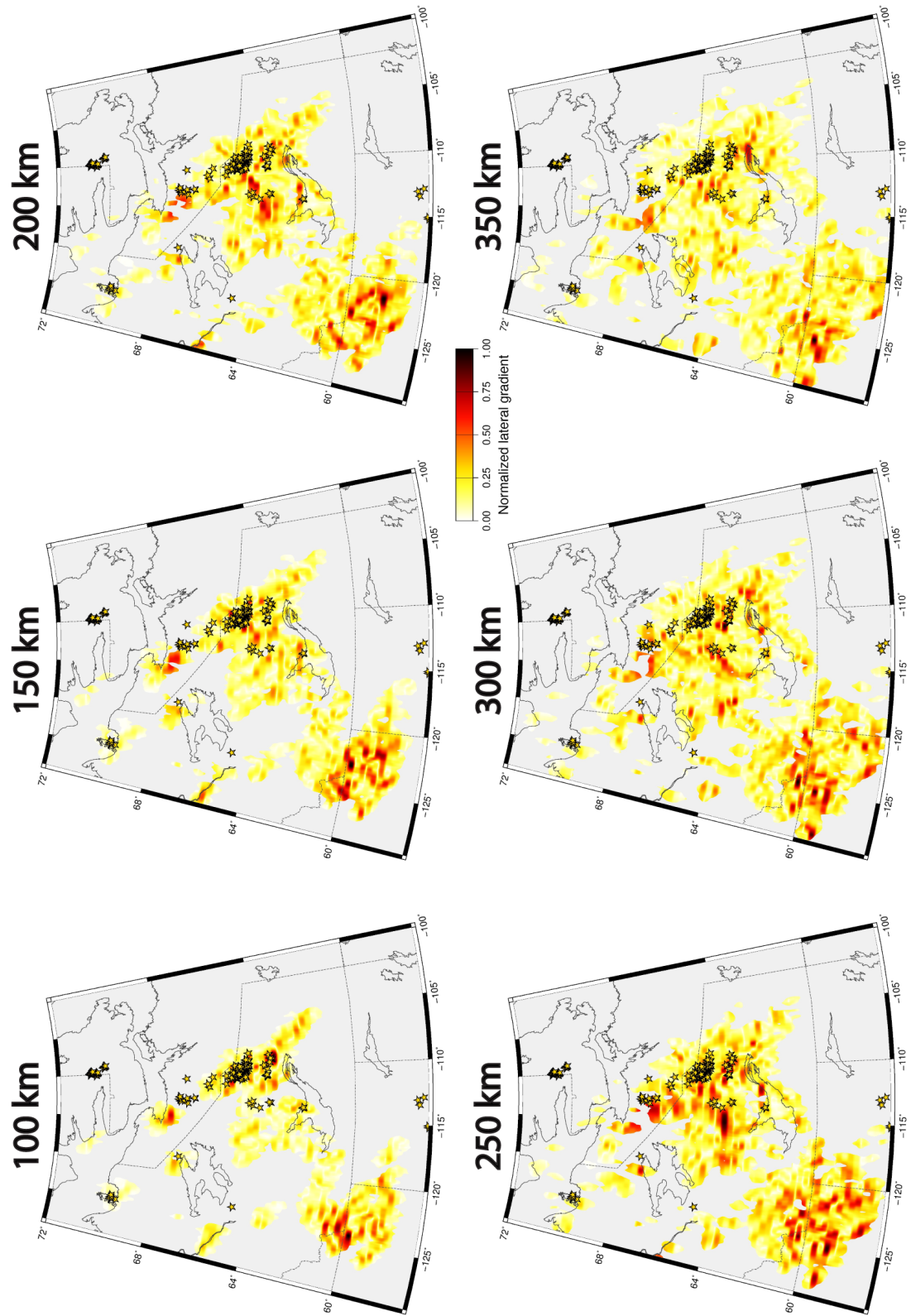


Figure A.7.: Maps showing the normalized lateral velocity gradient computed from the S-wave model. Same format as [Figure A.6.](#)

Table A.1.: Trade-off curve parameters for the P-wave model

flattening	smoothing	RMS model Roughness	RMS reduction (%)	Model
$(s/km^3 * 10^{-7})$				
100	1125	10.20	98.73	A
200	2250	5.42	98.48	B
400	5500	2.72	98.09	C
800	11000	1.54	97.57	D
1200	16500	1.09	97.12	E*
1600	22000	0.87	96.74	F
3200	44000	0.49	95.33	G
6400	88000	0.27	93.19	H
12800	176000	0.13	90.65	I

Table A.2.: Trade-off curve parameters for the S-wave model

flattening	smoothing	RMS model Roughness	RMS reduction (%)	Model
$(s/km^3 * 10^{-7})$				
62.5	1000	41.30	96.47	A
125	2000	22.30	95.32	B
250	4000	11.70	93.80	C
500	8000	5.55	92.05	D
1000	16000	2.61	89.74	E*
2000	32000	1.35	86.05	F
4000	64000	0.69	81.34	G
8000	128000	0.33	75.71	H
16000	256000	0.15	70	I

SUPPORTIVE INFORMATION FOR "UPPER MANTLE STRUCTURE OF NORTHWESTERN CANADA FROM TELESEISMIC BODY WAVE TOMOGRAPHY"

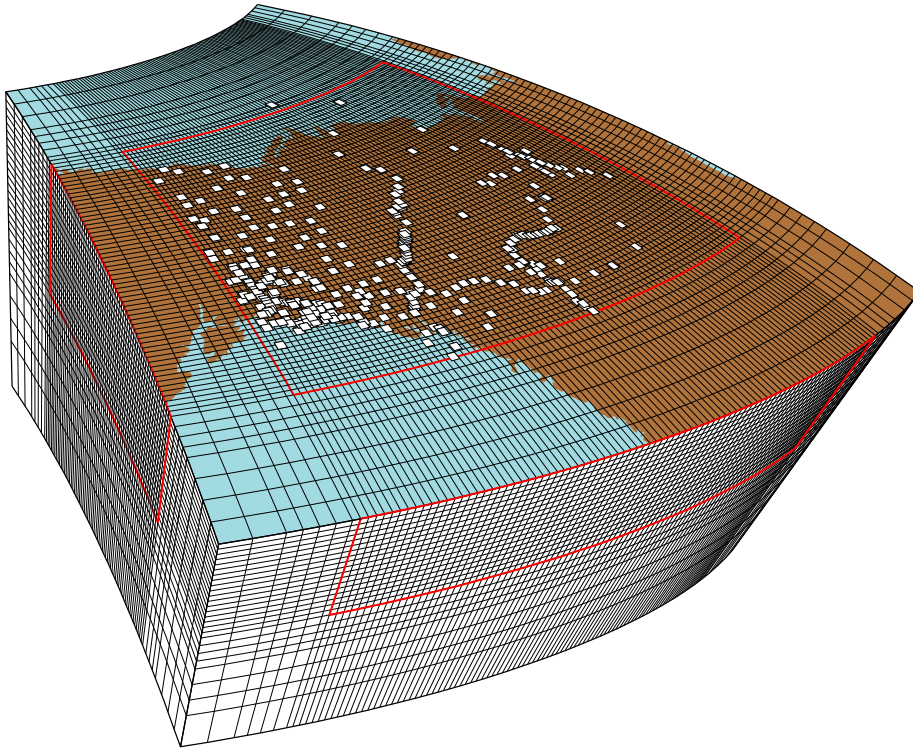
Contents of this file

1. Figures B1 to B7.

Introduction

This supporting information provides details on the three dimensional representation of the parameterization grid used to represent the P- and S-wave velocity structure beneath north-western Canada ([Figure B.1](#)). It includes P and S-wave resolution tests, (Figures [B.2](#) and [B.3](#), respectively) and the ray density maps for both P- and S-wave models (Figures [B.4](#) and [B.5](#)). We also include depth slices at 75 km, 400 km, 500 km and 600 km for completeness for both P- and S-wave models (Figures [B.6](#) and [B.7](#)). Our P- and S-wave delay time data sets and teleseismic P- and S-wave tomography models are available in the attached .zip folder.

Knot Locations for Taut Spline Parameterization



Number of knots in radius = 27; latitude = 76; longitude = 103
Total number of knots = 211356

Figure B.1.: Three dimensional grid used to compute the P- and S-wave velocity structure beneath northwestern Canada. The red box represents the study area and is defined from $57^{\circ}N$ to $72^{\circ}N$, $-147^{\circ}W$ to $-103^{\circ}W$ and 0 to 1300 km depth. The intersecting black lines and the white squares indicate node positions and seismic station locations.

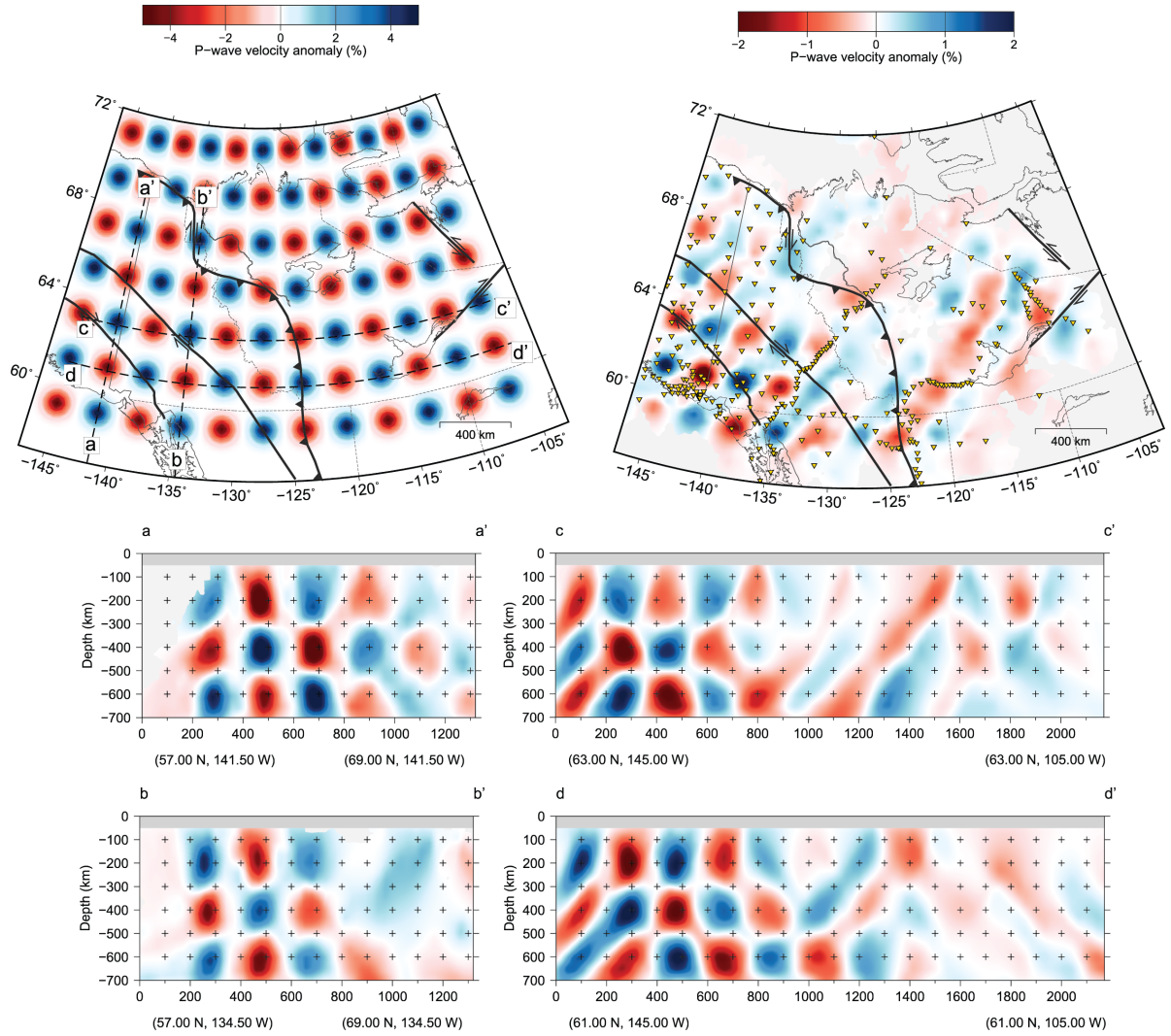


Figure B.2.: P-wave checkerboard resolution test. (Top left) Map showing the distribution of synthetic anomalies at 200 km depth (input). Also shown the cross section locations and same tectonic structures as [Figure 4.1](#). (Top right) Output of the checkerboard test at 200 km depth. (Bottom) Cross sections through the retrieved checkerboard model. Areas with less than 4 ray hits are displayed in light gray and the first 50 km are masked.

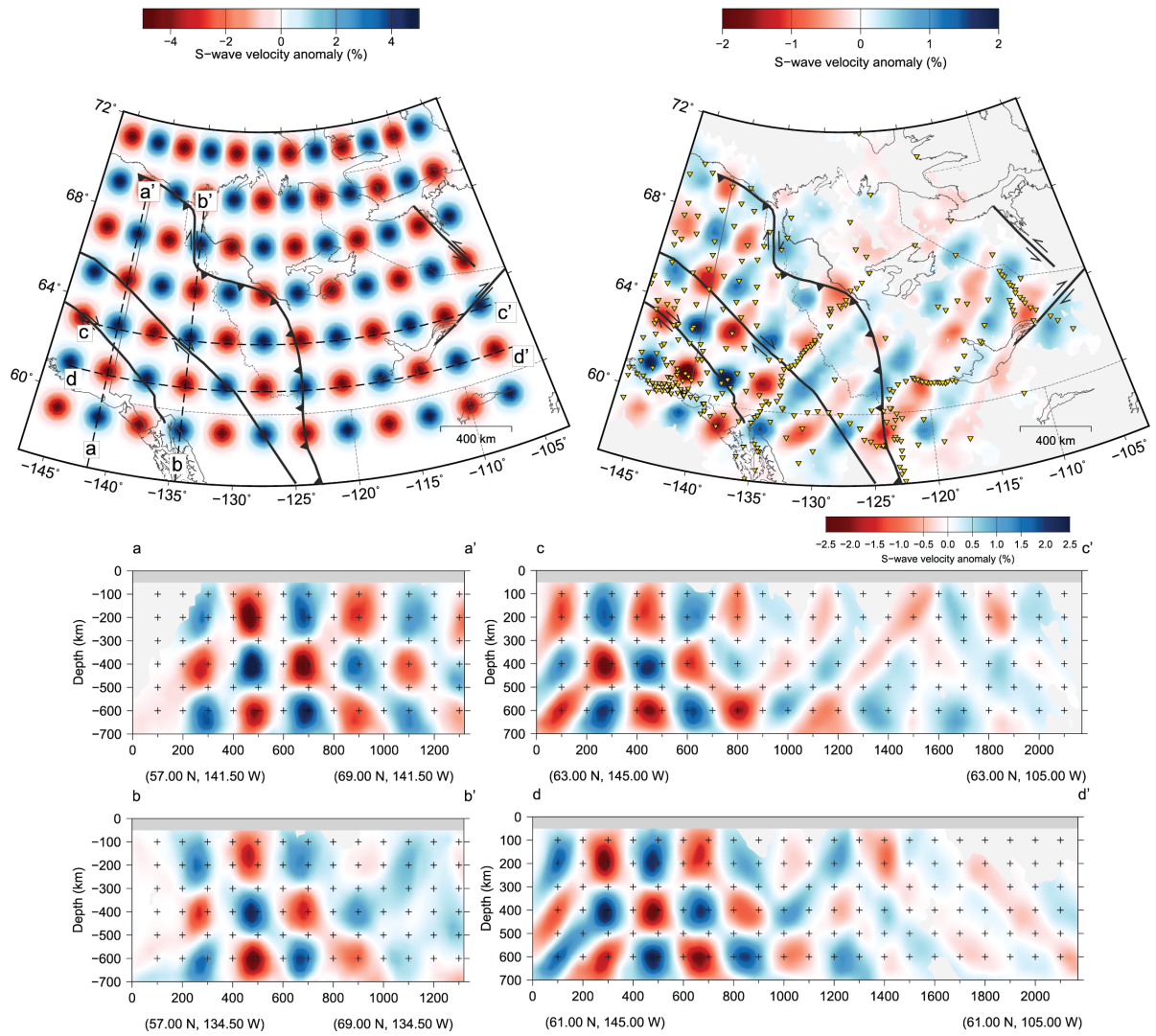


Figure B.3.: S-wave checkerboard resolution test. Format is the same as [Figure B.2](#).

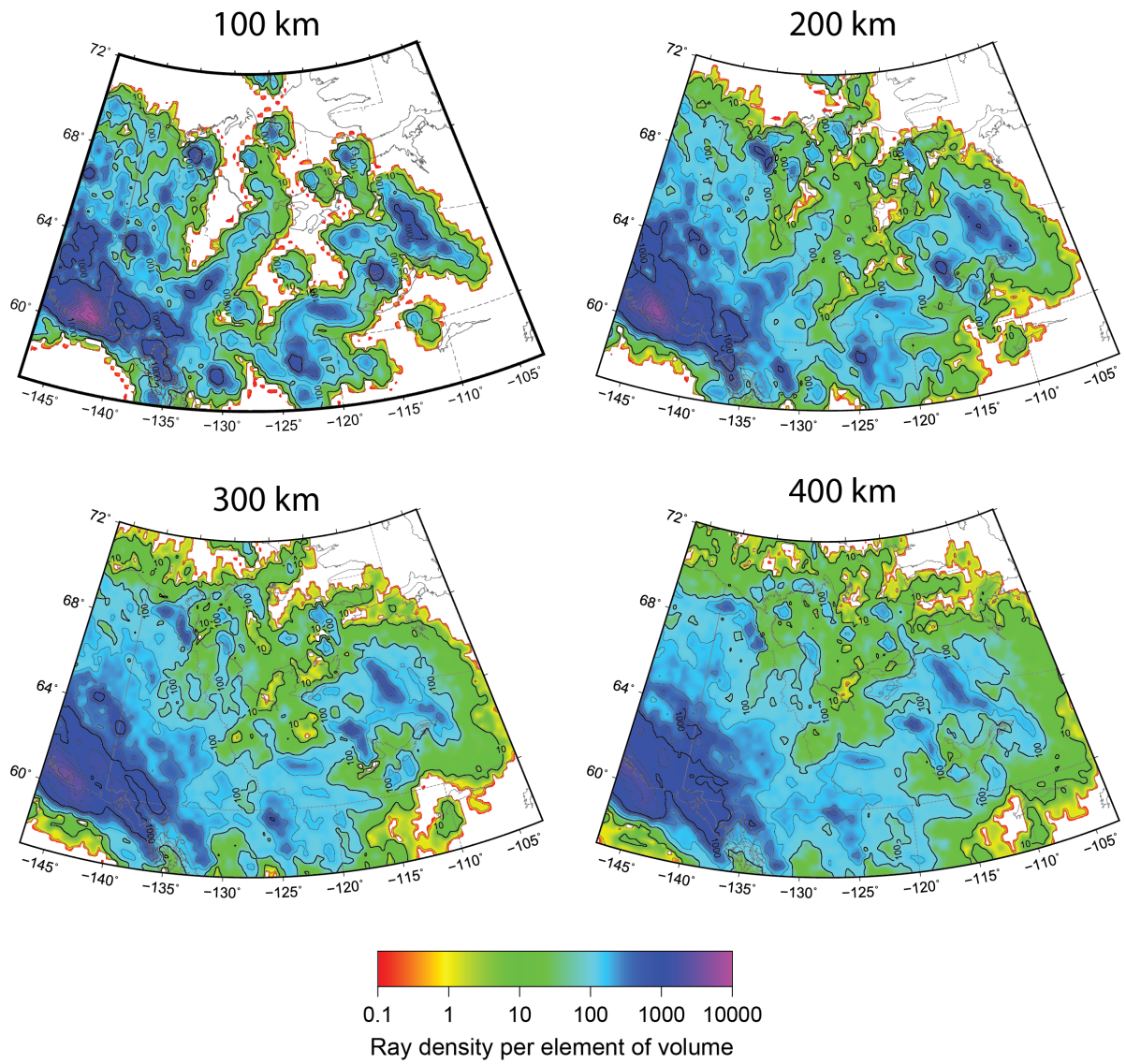


Figure B.4.: Ray density maps through our P-wave model at four different depths 100, 200, 300 and 400 km depth. Blue/purple color represents a high ray density whereas red/yellow color shows a low ray density.

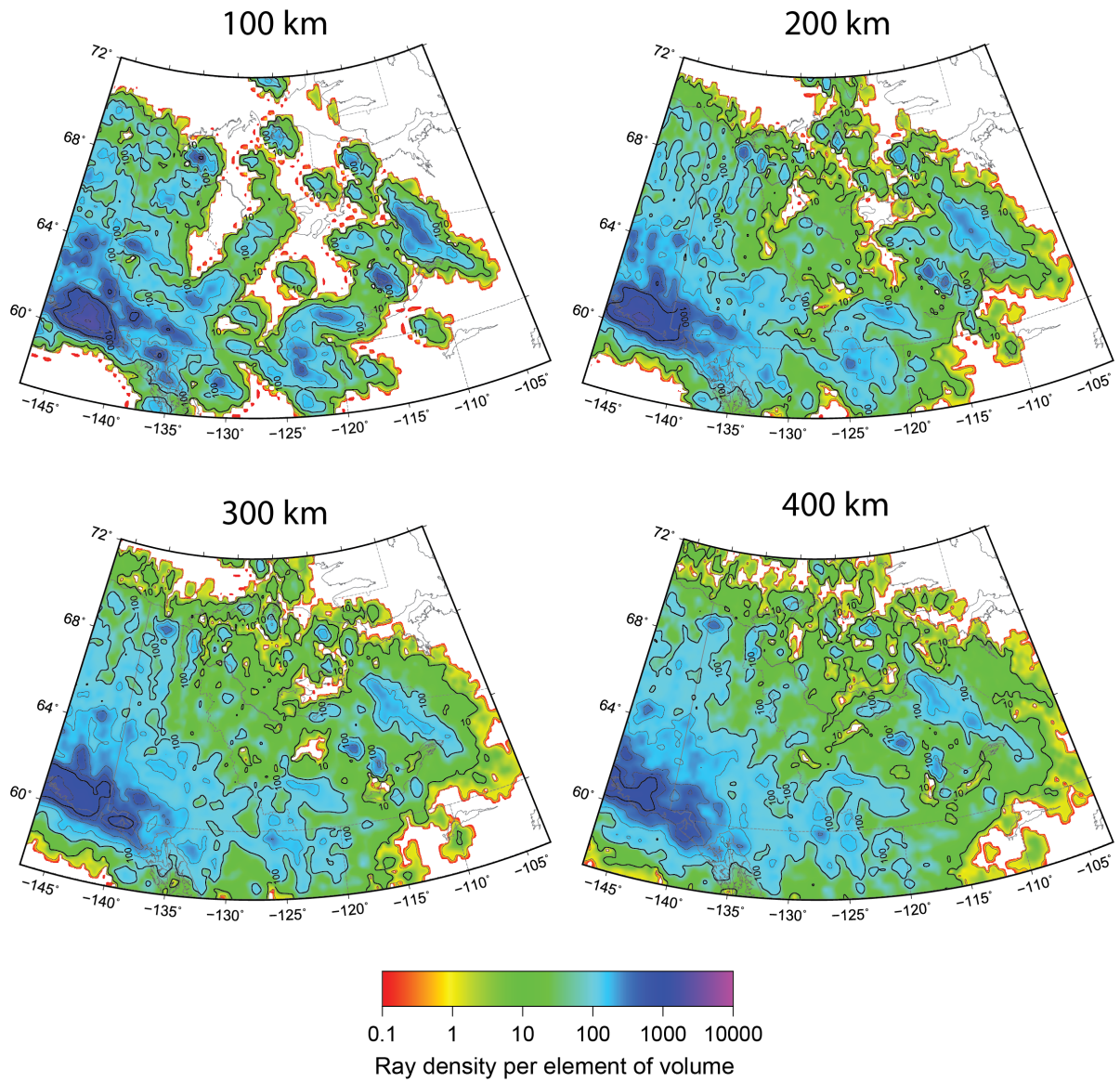


Figure B.5.: Ray density maps through our S-wave model at four different depths 100, 200, 300 and 400 km depth. Blue/purple color represents a high ray density whereas red/yellow color shows a low ray density.

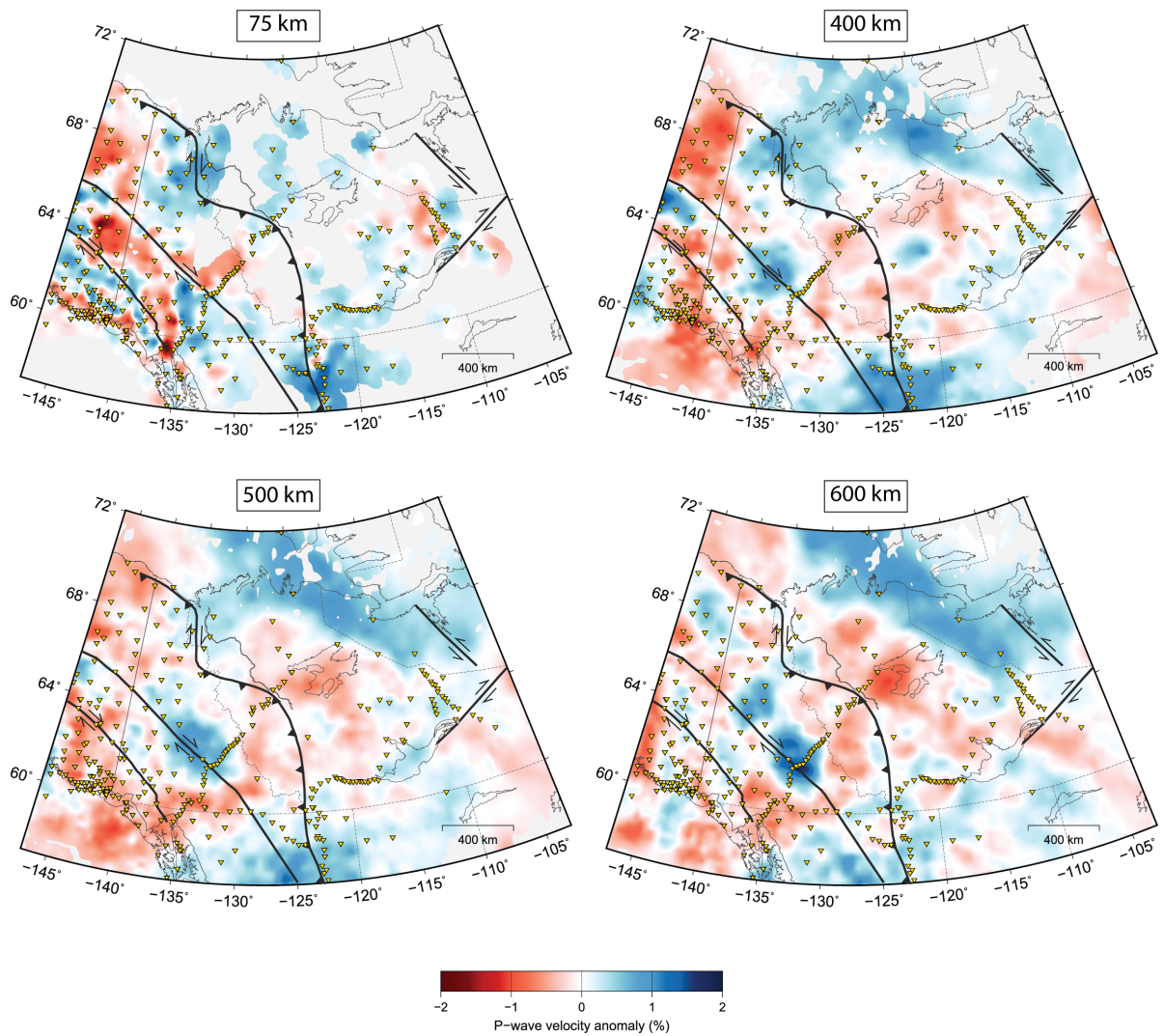


Figure B.6.: Depth slices and cross sections through the P-wave model. Tectonic structures from [Figure 4.1](#). Areas with less than 4 ray hits are displayed in light gray.

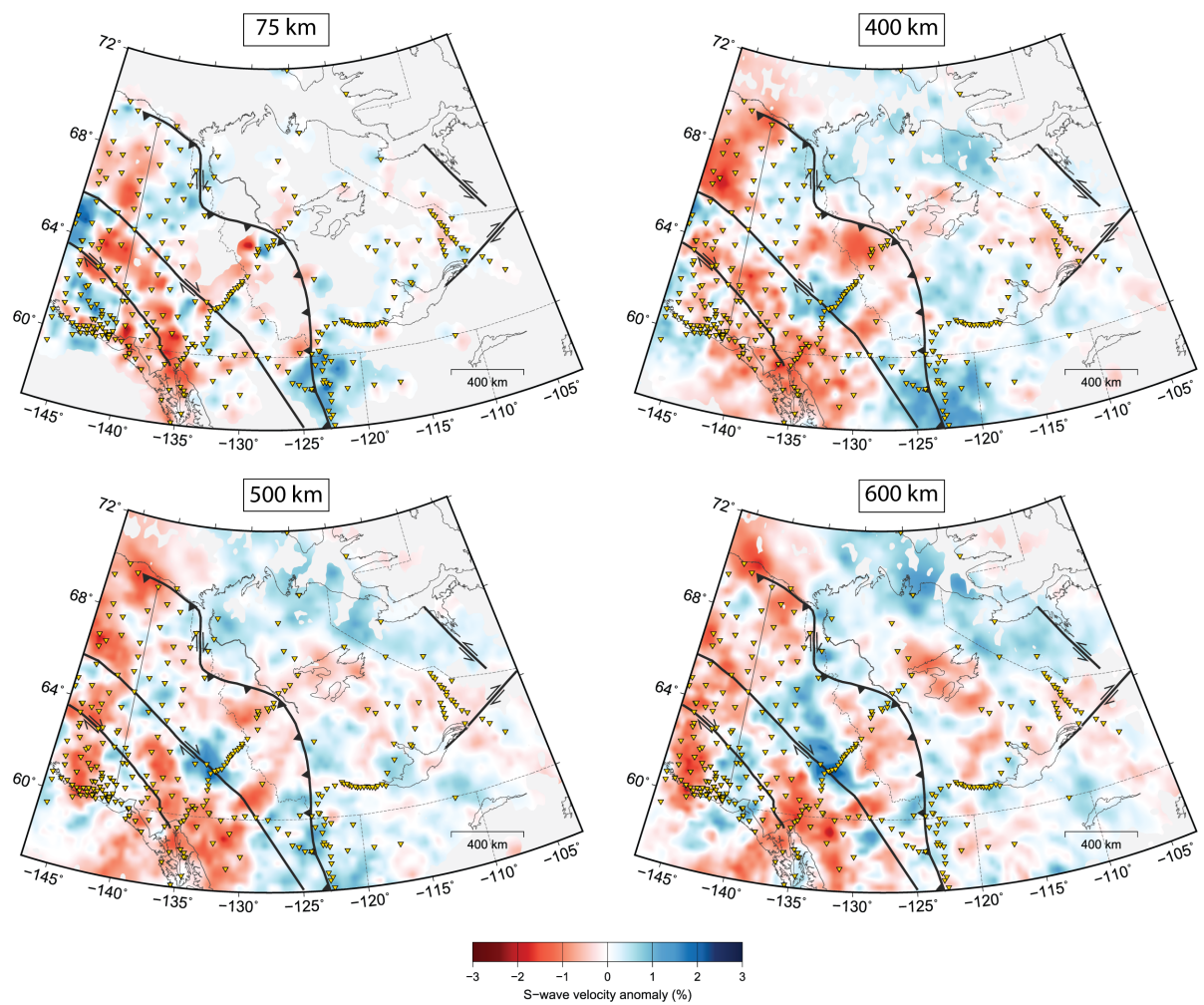


Figure B.7.: Depth slices and cross sections through the S-wave model. Same format as [Figure B.6](#).

SUPPLEMENTAL MATERIALS FOR SEISMIC EVIDENCE FOR CRATON CHISELING AND DISPLACEMENT OF LITHOSPHERIC MANTLE BY THE TINTINA FAULT IN THE NORTHERN CANADIAN CORDILLERA

C.1 TELESEISMIC SHEAR WAVE SPLITTING ANALYSIS

Shear waves travelling through Earth have the property to split into two perpendicularly polarized waves because anisotropic media are birefringent. One shear wave will travel along the fast axis direction of propagation (ϕ) whereas the other perpendicular shear wave will travel along the slow axis direction of propagation. As the two shear waves propagate through an anisotropic medium, a delay time will accumulate (δt). At the Earth's surface, it is therefore possible to measure the two splitting parameters (δt and ϕ). The magnitude of the delay time depends on the thickness of the anisotropic medium and the fast axis direction is related to the orientation of the anisotropic medium. Shear wave splitting measurements provide information on the upper mantle deformation. However, this technique has limitations. For teleseismic shear waves, the calculated splitting estimates is an integrated value along the ray path on the receiver side from the Core-Mantle Boundary (CMB) up to the surface. Therefore, teleseismic shear waves record both lithospheric and asthenospheric deformations.

In this study, we calculated splitting estimates for 45 stations located across northwestern Canada (Table C.1). We used teleseismic earthquakes of magnitude $M_W \geq 6.0$ in the angular epicentral distance $85^\circ \leq \Delta \leq 120^\circ$ that occurred between January 2014 and August 2019. We only selected waveforms with a Signal-to-Noise Ratio (SNR) greater than 7.5 dB. Waveforms were then rotated into the three-dimensional ray system (LQT). We calculated SKS splitting parameters for each station using SplitPy (Audet & Schaeffer, 2019), a Python version of SplitLab (Wüstefeld et al., 2008). SplitPy removes the effect of seismic anisotropy using two different techniques: the rotation-correlation technique (RC; Bowman & Ando, 1987) and the minimum energy technique (SC; Silver & Chan, 1991). The two techniques

calculate a grid-search for the splitting parameters (δt and ϕ), which best removes the effect of splitting. The rotation-correlation technique uses the maximization of the cross-correlation coefficient between the waveforms on the radial Q and transverse T components. The minimum energy technique searches for the minimum energy of displacement on the transverse component.

In order to determine if the resulting splits are either null or non-null measurements, SplitPy compares the delay times ($\rho = \delta t_{RC} / \delta t_{SC}$) and the azimuth of fast axis estimates ($\delta\phi = \phi_{SC} - \phi_{RC}$) from both techniques. A teleseismic null measurement requires that (1) the SNR of the tangential component of the waveform is $< 3dB$ or (2) $\delta\phi \approx 45^\circ$ and ρ is small. A teleseismic non-null measurement requires that (1) the $\rho > 0.7$ and (2) $\delta\phi < 22.5^\circ$. Null and non-null measurements are then qualitatively evaluated as “Good”, “Fair” and “Poor”. For non-null measurements, we have $0.8 < \rho < 1.1$ and $\delta\phi < 8^\circ$ for “good” measurements; $0.7 < \rho < 1.2$ and $\delta\phi < 15^\circ$ for “fair” measurements; and “poor” otherwise. For nulls, we have $0 < \rho < 0.2$ and $37^\circ < \delta\phi < 53^\circ$ for “good” nulls; $0 < \rho < 0.3$ and $32^\circ < \delta\phi < 58^\circ$ for “fair” nulls; and “poor” otherwise.

Figure C.1 shows an example of a “good” measurement for an earthquake that occurred 24 September 2016 ($M_W = 6.9$). The center panels show results for the rotation-correlation technique (RC). The lower panels show results for the minimum energy technique (SC). We separately calculated a vector average of all “fair” and “good” non-null measurements for both techniques, and obtain final estimates by vector averaging the results of both techniques into a single estimate of δt and ϕ for each station. Figure C.2 show all the resulting “fair” and “good” non-null measurements for the RC and the SC techniques, respectively.

Null measurements provide additional information on the upper mantle structure beneath the seismic station. A null measurement occurs if the shear wave propagates in an isotropic medium and/or the shear wave polarization is aligned with the fast or slow axis direction in the anisotropic medium (Wüstefeld & Bokermann, 2007). Figure C.3 shows the back-azimuth of all “good” null measurements (black bars) along with the vector-averaged estimates (red bars). The back-azimuth distribution of all “good” null measurements are either aligned or perpendicular to the fast axis direction, which gives us confidence in our results. Figure C.4 shows all multievent average results of teleseismic shear-wave splitting from this study and from the compilation of shear-wave splitting estimates (Audet et al., 2016; Courtier et al., 2010; Rasendra et al., 2014; Snyder & Bruneton, 2007; Venereau et al., 2019). Figure C.5 shows the stations for which the splitting parameters are projected along cross sections A-A', B-B' and C-C' (Figure 5.3).

C.2 TELESEISMIC P-WAVE TOMOGRAPHY

In this study, we use the teleseismic P-wave tomography model of [Estève et al. \(2020a\)](#). The P-wave data set consists of 127,138 delay times from 2,627 earthquakes with magnitude $M_W \geq 5.5$ spanning the angular epicentral distance range $30^\circ \leq \Delta \leq 90^\circ$ that occurred between 2000 and 2018. The three-dimensional isotropic velocity structure underlying north-western Canada is calculated using VanDecar's inversion procedure [VanDecar \(1991\)](#). The relative delay times are inverted simultaneously for slowness perturbations, station static terms and source terms.

The P-wave model is parameterized with splines under tension constrained at a series of regular nodes. The grid extends from $51^\circ N$ to $77^\circ N$ in latitude, from $-156^\circ W$ to $-94^\circ W$ in longitude and from the surface to 1300 km depth. The grid has 27, 76 and 103 knots in the radius, latitude and longitude direction resulting in 211,356 knots. The inner region of the grid is sampled every 0.25° of latitude, 0.5° of longitude and every 33 km.

This inverse problem is strongly under-determined and is solved through regularization. Here, three types of regularization are applied: smoothing, flattening and damping. In order to find the preferred model, a trade-off area is generated from 72 inversions (see [Estève et al., 2020a](#), for more details). We use 2000 conjugate gradient iterations and 10 down-weighting iterations in order to get a stable solution. The final P-wave model explains 97.33 % of the RMS of the relative arrival time residuals.

Resolution of such tomographic inversion problem depends strongly on the ray geometry, and the back-azimuth distribution of teleseismic earthquakes. For a qualitative assessment of resolution, synthetic checkerboard and structural tests are performed. We generate a synthetic model consisting of an alternating checkerboard pattern with slow and fast P-wave anomalies that are $\pm 5\%$ at 200, 400 and 600 km depth ([Figure C.6-A](#)). These synthetic anomalies have a radius of 50 km are defined by a Gaussian function across their diameter. Gaussian noise with a standard deviation of 0.03 s was also added to the P-wave data set in order to mimic the noise that occurs in the real travel-time measurements.

The recovery of the synthetic anomalies is excellent across the Tintina and Denali faults and in the southwestern region of the P-wave model ([Figure C.6-B](#)). East of the Cordillera Deformation Front (CDF), stripes of anomalies oriented northeast-southwest show the effect of a poor station coverage and reflect the back-azimuth distribution in the teleseismic data.

[Figure C.7](#) shows the input and output of the P-wave structural test. This structural test is performed to assess the robustness of longer-wavelength features and the effect of smearing (inherent to the technique) in the real model. The output synthetic model shows excellent

lateral resolution between the surface and 200 km depth and at greater depths between 450 km and 650 km depth.

C.3 CALCULATION OF THERMAL LENGTH SCALE

To estimate the characteristic thermal diffusion length, we use the formula $l = \sqrt{\kappa\tau}$, where l is the diffusion length, κ is the thermal diffusivity of peridotite, and τ is the time over which the diffusion occurs, assumed to be 37 M.yrs. Peridotite is highly ($\sim 20\%$) anisotropic in thermal diffusivity. If we assume the measured shear wave splitting fast axis corresponds to the most thermally diffusive axis (Gibert et al., 2003), then diffusivity should be minimized along the direction perpendicular to the Tintina Fault. For a 1000°C temperature (diffusivity decreases with increasing temperature), at a 1 GPa pressure (diffusivity increases slightly with pressure) a reasonable diffusivity for spinel lherzolite is $5 \times 10^{-7} \text{ m}^2 \text{ s}^{-1}$. This leads to a conservative estimate for characteristic length of 22 km. If the maximum thermal diffusivity axis is not aligned parallel to the Tintina Fault, or if diffusivity anisotropy is minimal, then diffusivity and the characteristic length will be larger (Gibert et al., 2003).

C.4 COMPARISON OF NCC WITH CRATONIC XENOLITHS

To investigate the feasibility of seismically observable compositional variations in the lithosphere, we consider two sets of xenoliths. The first is a collection of 345 xenoliths associated with the Canadian Cordillera (Francis et al., 2010) which we take to approximate the composition of the lower velocity non-cratonic mantle in the region. For a representation of the composition of cratonic mantle, we use 45 harzburgite xenoliths from Wiedemann Fjord, Greenland (Canil, 2004). These xenoliths have been noted to have similar olivine Mg# to the Parry Peninsula xenoliths from the Mackenzie craton (Liu et al., 2018).

We use *Perple_X* (Connolly, 1990) to calculate the P-wave velocity for each sample using the *stx11ver.dat* database (Stixrude & Lithgow-Bertelloni, 2011), at 10000 kbar and 800°C , roughly at the conditions of the uppermost mantle lithosphere. We chose to use the *stx11ver.dat* database because it is widely used in the seismological literature (e.g., Ma et al., 2020) and does a fairly good job of estimating velocity for a given composition. However, it approximates the composition of the mantle as only six oxides: SiO_2 , MgO , FeO , CaO , Al_2O_3 , and Na_2O . An alternative approach is to use the *hp62.dat* database (Holland & Powell, 2011). This considers a much fuller range of composition but is not widely used in seismology because it produces less accurate mantle velocity estimates.

While neither database is ideal, our tests show in both cases there is about a 2% increase in velocity between the Cordilleran and cratonic xenoliths (Figure C.8). This suggests the hypothesis that the Tintina fault bounds compositional variations is feasible.

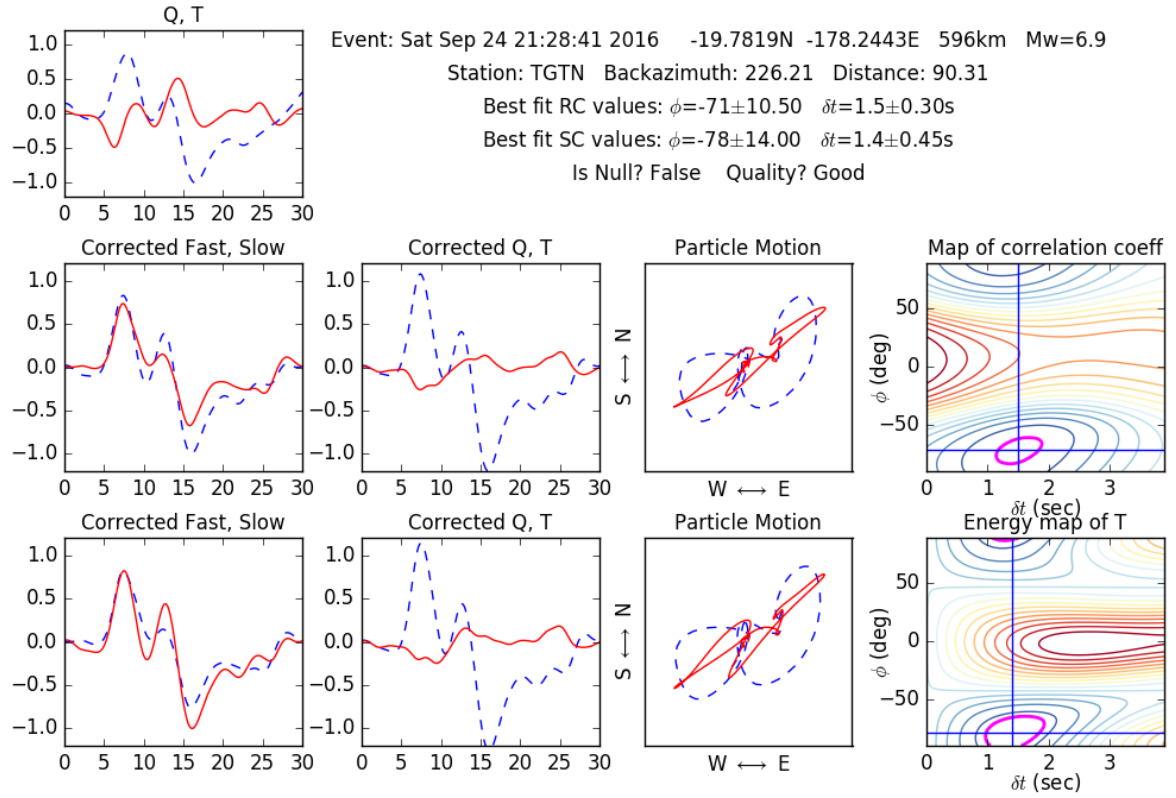


Figure C.1.: Example of a single measurement at station TGTN. The top left panel shows the longitudinal (Q, blue) and tangential (T, red) seismograms where the time window is manually picked around the predicted SKS phase arrival. The text gives event details as well as splitting parameters resulting from the two techniques. The center and bottom panels display results for the RC and SC techniques, respectively. For each technique, from left to right, the first panel displays the fast (blue) and slow (red) components of the seismogram. The second and third panels display the radial (Q, blue) and tangential (T, red) components and the particle motion before (blue) and after (red) removing the effect of splitting, respectively. The fourth panel displays either the misfit map of correlation coefficient (RC technique) or the energy map of the tangential component (SC technique). Warm and cold colors represent high and low misfit values, respectively. The magenta contours show the 95% confidence interval.

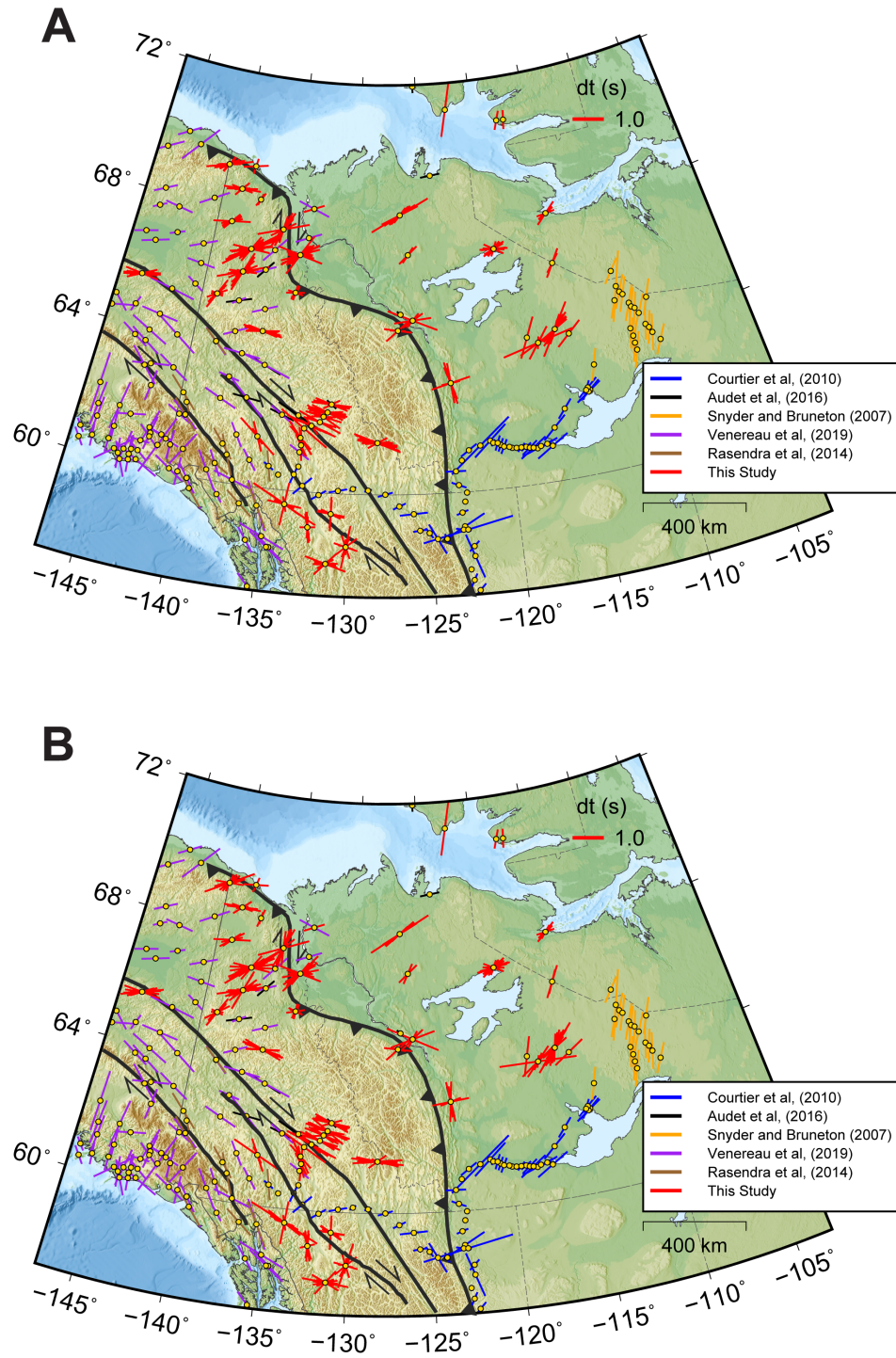


Figure C.2.: Topographic map of northwestern Canada showing all “fair” and “good” non-null SKS splitting measurements (in red) for each station for the RC (A) and the SC (B) techniques. Also shown, the compiled estimates of shear wave splitting from Audet et al. (2016); Courtier et al. (2010); Rasendra et al. (2014); Snyder & Bruneton (2007); Venereau et al. (2019).

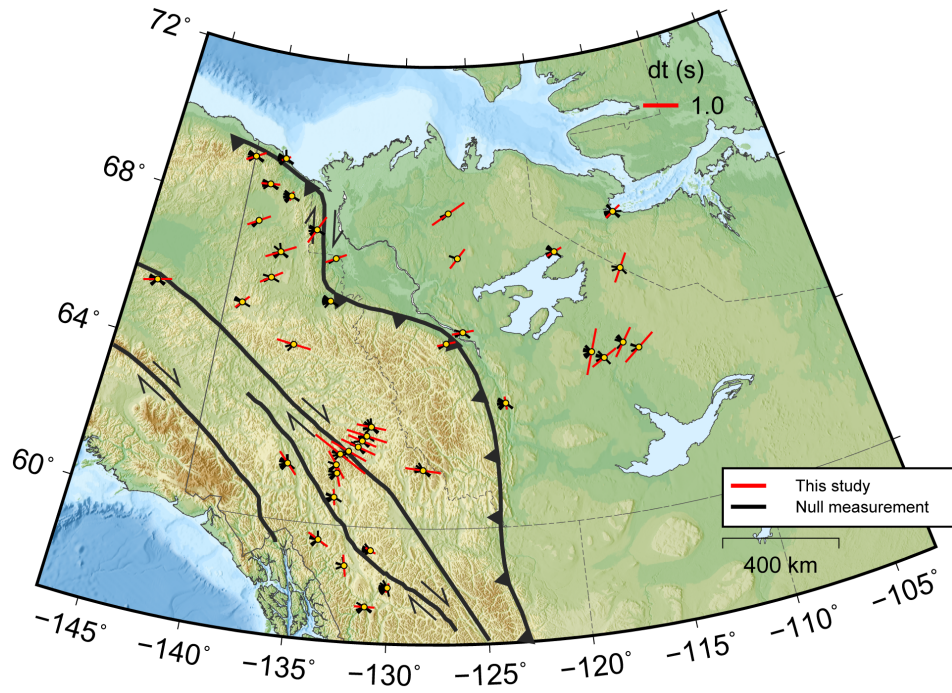


Figure C.3.: Topographic map of northwestern Canada showing the average splitting parameters (red) and all "fair" and "good" null measurements for each station (black).

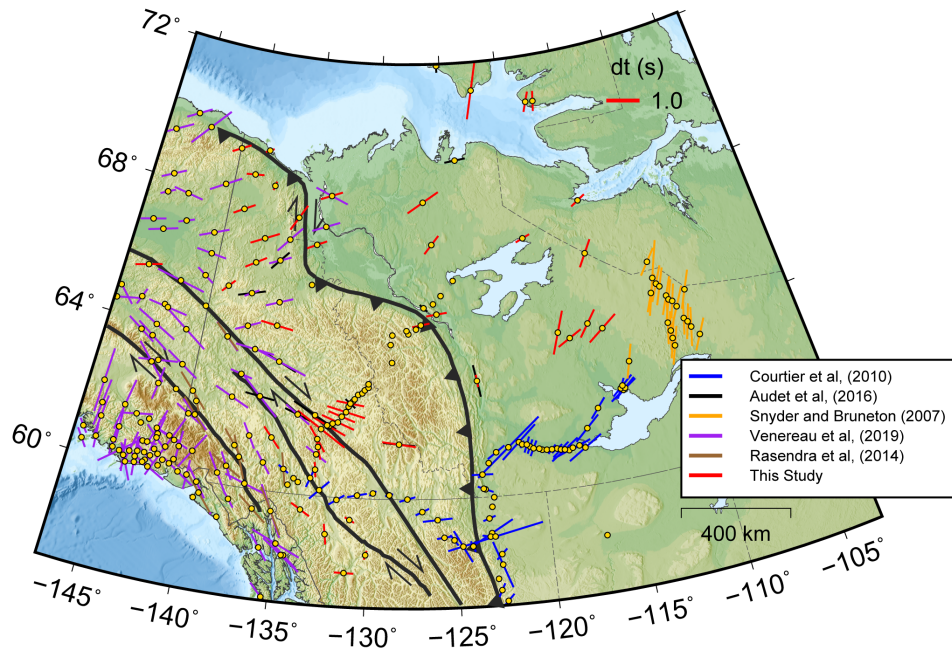


Figure C.4.: Topographic map of northwestern Canada showing the average splitting parameters (red) and all "fair" and "good" null measurements for each station (black).

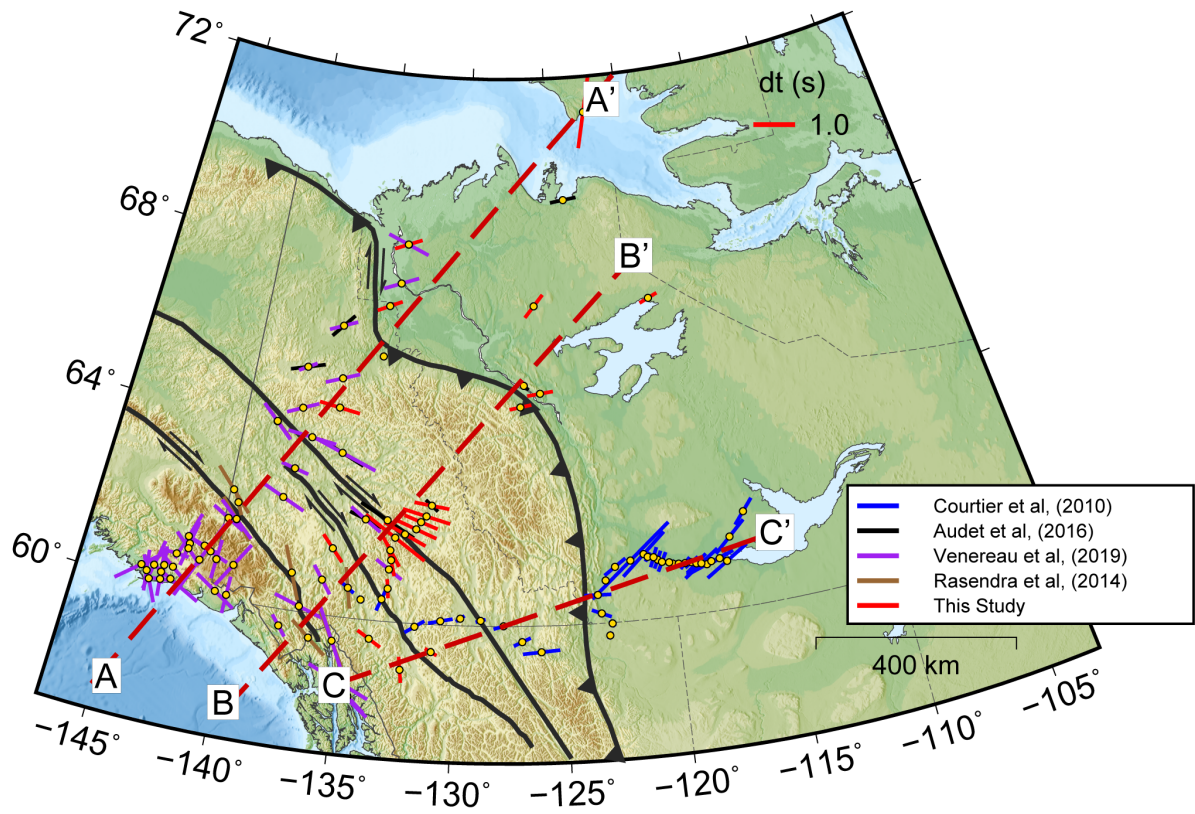


Figure C.5.: Topographic map of northwestern Canada showing the stations for which the splitting parameters are projected along cross sections A-A', B-B' and C-C' (Figure 5.3).

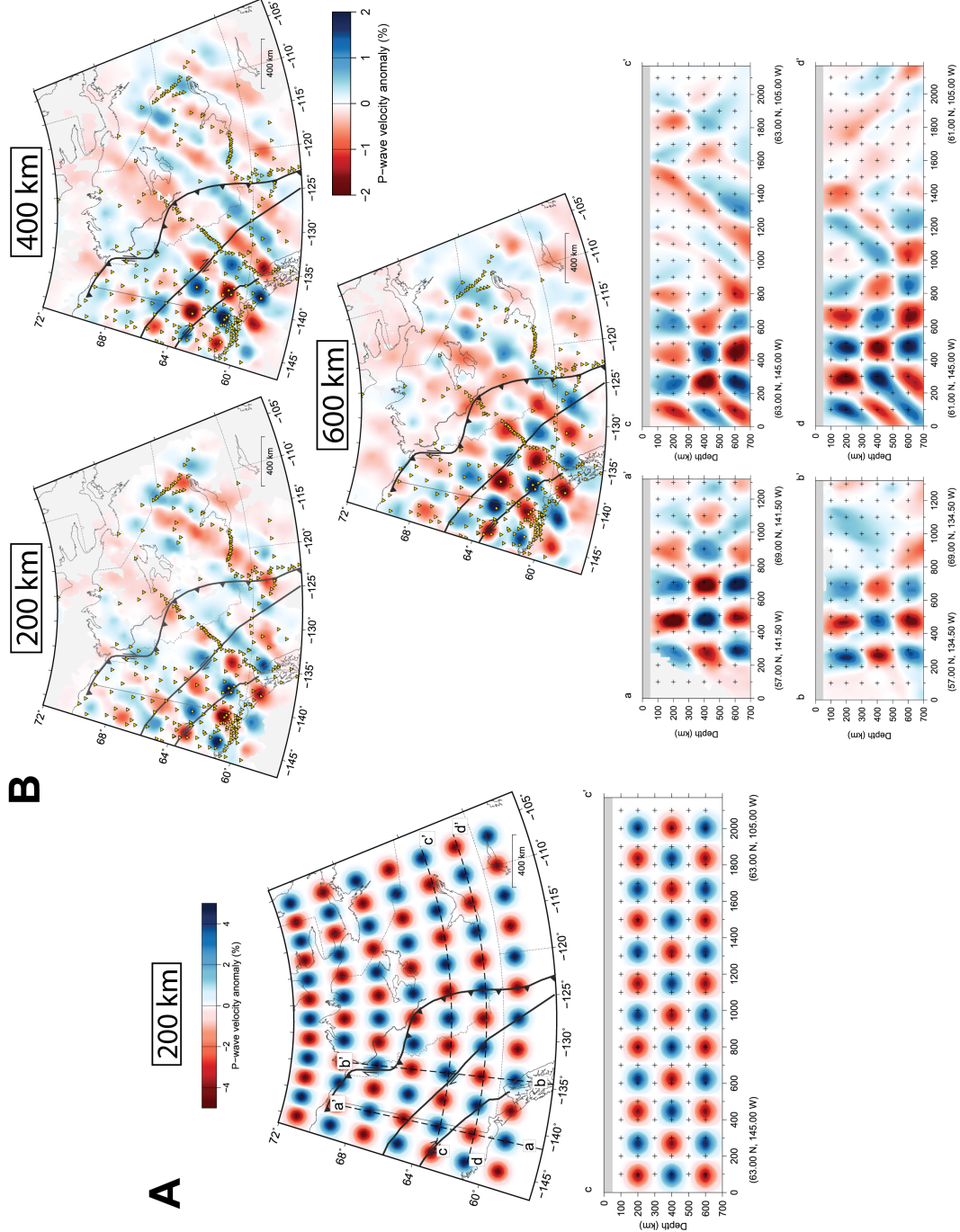


Figure C.6.: P-wave checkerboard test. A: Input map and cross section c-c' showing the distribution of synthetic anomalies through the input checkerboard model. Cross section locations are displayed on the 200 km depth slice. B: Output maps and cross sections through the retrieved checkerboard model at 200, 400 and 600 km depth. Areas with less than 4 ray hits are displayed in light gray and the first 50 km are masked. Note the excellent recovery in the western region of the P-wave model.

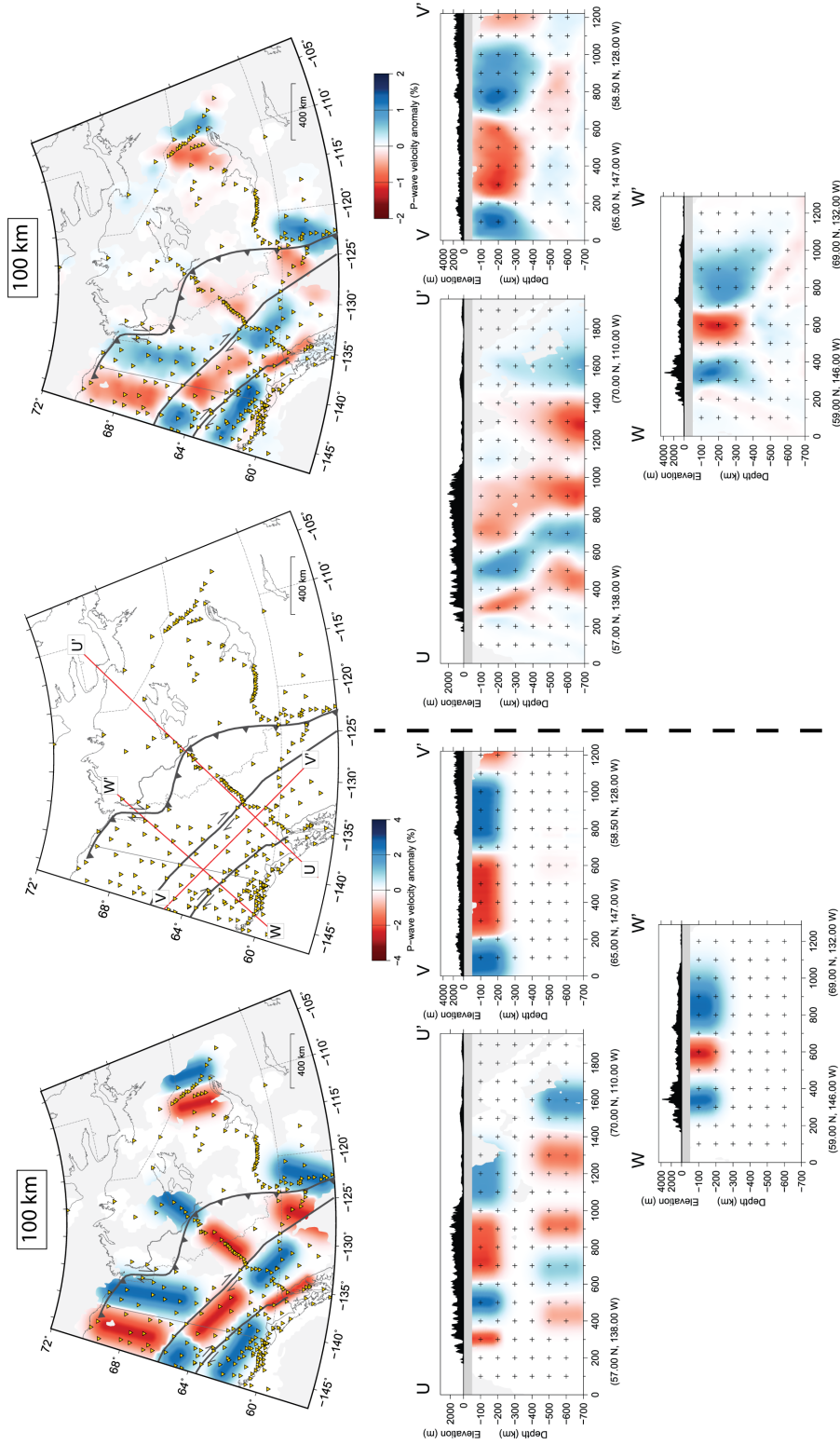


Figure C.7.: P-wave structural resolution test. Cross sections and depth slices through the P-wave structural resolution test. Structural resolution test showing input anomalies (left) and output structural P-wave model (right) through the output structural P-wave model. Synthetic anomalies are defined by Gaussian functions across their widths. Areas with less than 4 ray hits are displayed in light gray and the first 50 km are masked. Note the excellent lateral resolution in the western region of the P-wave model.

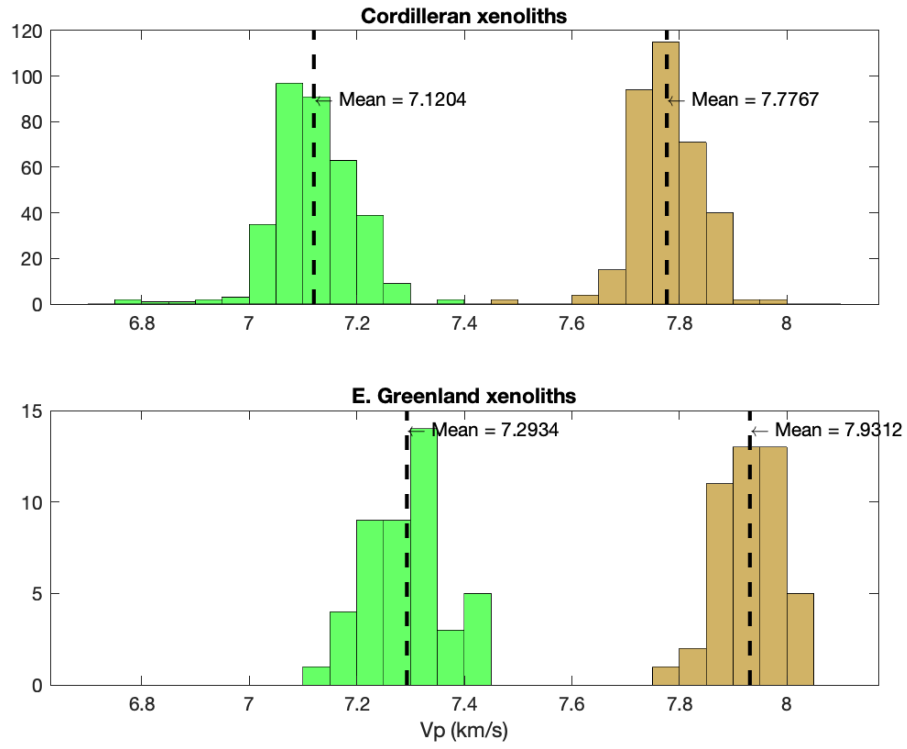


Figure C.8.: Histograms of estimated P-wave velocities for two different xenolith populations, using two different thermodynamic databases. Top subplot shows velocities for Cordilleran xenoliths (Francis et al., 2010), and bottom shows east Greenland samples, representative of cratonic compositions (Canil, 2004). The shift between the same color histograms between the top and the bottom plots provides an estimate of the magnitude of potential compositionally caused velocity changes.

Table C.1.: Station average splitting parameters

Stations					Average Station Parameter			
Network	Name	#	Lat (°)	Lon (°)	$\phi(^{\circ})$	σ_{ϕ}	Δt (s)	$\sigma_{\delta t}$
TA	D27M	17	69.243	-140.965	61.6412	6.945	0.6018	0.1448
TA	D28M	2	69.3286	-138.737	61.9368	19.8632	0.2761	0.3471
TA	E28M	11	68.6043	-139.535	84.3616	8.6143	0.5361	0.1583
TA	E29M	3	68.3889	-137.897	14.7276	27.9602	0.2765	0.5898
TA	F28M	6	67.6136	-139.872	60.5468	11.2174	0.7055	0.2602
TA	F30M	9	67.6106	-135.786	29.7252	6.5961	0.8989	0.1496

Stations					Average Station Parameter			
Network	Name	#	Lat (°)	Lon (°)	$\phi(^{\circ})$	σ_{ϕ}	Δt (s)	$\sigma_{\delta t}$
TA	G29M	27	66.9116	-138.022	64.6524	5.2003	0.9239	0.1271
TA	G31M	22	66.9227	-134.271	65.4438	6.5537	0.6352	0.1033
TA	H29M	15	66.2191	-138.369	58.9283	11.8714	0.7058	0.2467
TA	H31M	8	65.8052	-134.343	41.665	10.039	0.1652	0.1707
TA	I28M	7	65.4483	-139.936	43.3278	9.6396	0.5062	0.1439
TA	J30M	13	64.5753	-136.33	-80.7745	7.7921	0.9998	0.1841
TA	N31M	4	61.4862	-135.78	-40.5946	14.0452	0.8047	0.31
TA	P32M	7	59.5898	-133.715	-58.4571	9.8456	0.6767	0.2571
TA	Q32M	9	58.9601	-132.269	-8.062	8.8326	0.6284	0.1734
TA	R33M	8	59.3946	-130.967	-61.8987	14.3514	0.3275	0.2698
TA	S34M	10	57.9128	-131.131	-88.8816	8.9724	0.5867	0.2262
NY	MMPY	7	62.6189	-131.263	-81.9853	9.7238	0.8565	0.2184
NY	NLHD	1	71.2509	-122.333	12.299	37.9	1.750	0.500
NY	PYRD	3	64.8212	-126.907	78.5366	14.8631	0.538	0.4427
NY	TGTN	21	61.5267	-128.273	-82.2022	5.1074	1.021	0.1111
NY	TULN	4	65.1147	-125.855	81.4116	14.7038	0.63	0.4488
NY	ULUN	1	70.7532	-117.185	7.207	54.462	0.605	1.163
NY	WGLY	8	63.2281	-123.458	-6.8802	8.7427	0.3745	0.2629
PO	CLVN	2	67.039	-126.078	39.0334	16.0122	0.6915	0.2858
PO	CTLN	26	64.4287	-116.019	36.0453	4.6633	0.9406	0.1112
PO	DHRN	20	67.0331	-119.509	66.2771	6.2512	0.4697	0.1254
PO	GALN	5	64.1167	-117.314	60.6964	10.3699	0.8566	0.2024
PO	HPLN	2	66.3485	-115.323	30.3082	14.2489	0.8963	0.2816
PO	ILKN	1	64.2241	-115.129	53	15.5	1.15	0.45
PO	ROMN	1	64.3157	-118.018	19.5	28	1.35	0.75
PO	SMPN	8	68.2107	-126.654	56.2018	7.5246	1.1124	0.1823
CN	DLBC	7	58.4372	-130.027	15.8806	9.8692	0.2746	0.249

Stations					Average Station Parameter			
Network	Name	#	Lat (°)	Lon (°)	$\phi(^{\circ})$	σ_{ϕ}	Δt (s)	$\sigma_{\delta t}$
CN	HMNT	1	70.7631	-117.806	18.655	10.400	0.62	0.19
CN	INK	1	68.307	-133.53	68.31	38.97	0.69	0.32
CN	KUKN	8	67.8226	-115.09	57.5024	8.3744	0.5257	0.1497
7C	MM07	2	60.7272	-133.071	-7.0275	22.2913	0.4617	0.469
7C	MM09	4	61.3568	-133.021	-15.8346	9.5509	0.7748	0.1882
7C	MM10	3	61.5744	-133.091	-33.5526	13.7643	0.8928	0.3213
7C	MM11	1	61.8624	-132.915	-56.5	15.75	1.8	0.675
7C	MM12	6	61.9525	-132.461	-64.561	9.2019	1.1393	0.22
7C	MM17	5	62.0789	-131.948	-68.9756	7.6344	1.0641	0.1999
7C	MM18	5	62.2424	-131.734	-70.8528	9.0801	0.8617	0.2106
7C	MM19	9	62.3754	-131.495	-75.7245	7.3741	1.1952	0.1825
AK	PPD	14	65.5174	-145.525	74.0115	7.5492	0.8189	0.1759