

Two-Station Rayleigh Wave Tomography of Canada

by

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ABSTRACT.

Seismic tomography is a common method used to study the lithosphere and can reveal evidence of past and present tectonic and geodynamic processes. In largely aseismic areas such as central and eastern Canada, single-station measurements of surface-wave dispersion will provide limited and uneven resolution; two-station measurements allow the use of a larger selection of earthquakes (i.e., extending to teleseismic events) while still preserving a regional-scale measurement. This work examines seismic velocity and anisotropy variations of the Canadian lithosphere using two-station-based surface-wave tomography, to obtain new insights into the formation history of the continent.

A cross-correlation method was used to obtain new two-station dispersion curves (a total of 117) which were added to a dataset of previously published measurements. These were chosen to fill in gaps in ray coverage and to maximize azimuthal distribution, particularly across the Prairies and the Cordillera. Tomographic inversion was used to calculate anisotropic phase velocity maps for periods of 15-400 s, following outlier exclusion and resolution testing.

The phase velocity maps show that the low velocity Cordilleran lithosphere contains two southern zones (west and east) with different anisotropy, which are interpreted to be due to the influence of the subducting Juan de Fuca plate in the west. A sharp Cordillera-craton boundary with the Canadian Shield is also noted, which shifts eastward near the US border. A very low velocity anomaly is observed beneath southeastern Saskatchewan which is likely linked to a mantle suture in this area. The highest velocities are seen beneath the craton, with particularly high velocities in Nunavut (northern Canada) and the western Superior Province. Strong E-W anisotropy is observed beneath the western Superior resulting from frozen lithospheric fabric. A clear distinction between the Precambrian Shield and the lower velocity Appalachian domains is also noted. This Appalachian low velocity zone is thought to be associated with the present-day Northern Appalachian Anomaly, rather than the Great Meteor Hotspot track.

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1. INTRODUCTION.

1.1 Overview.

Seismic tomography is an imaging technique that uses seismic waves to reconstruct two-and three-dimensional images of the Earth's interior. Seismic tomography uses differences in the velocity of seismic waves as they travel in order to produce a 3D model. Seismic tomography is a useful tool, providing information on seismic velocity, which relates directly to geological properties, but is fundamentally a geophysical variable. There are several types of seismic tomography, the main ones being surface wave tomography and body-wave tomography. Seismic waves generated by earthquakes are most relevant for lithospheric-scale work, but seismic tomography can also be done using ambient noise or controlled sources, which provide good crustal sensitivity (Bensen et al., 2007).

Understanding Canadian tectonic history is synonymous with understanding the Canadian lithosphere, since plate tectonics is a lithospheric process. Imaging the lithosphere on a large scale is best done with seismic tomography. This methodology has been used in localized areas in Canada (e.g., Foster et al., 2020), and on a large scale in global models (e.g., Schaeffer and Lebedev, 2014), but there are not many models of Canada as a whole. The models that do exist are largely based on one-station surface wave methods (e.g., Schaeffer and Lebedev, 2014), which limit resolution in aseismic regions, as well as areas with low seismic station density, such as east of the Cordillera (e.g. the Prairie provinces). The two-station approach addresses this issue, and in this thesis, surface waves are analysed and the first two-station model of the entire Canadian lithosphere is presented.

Surface wave tomography is the preferred method for this work as teleseismic body wave tomography, while having good lateral resolution, lacks depth resolution. It is as a result of this teleseismic geometry that the rays are near-vertical, and thus depth resolution is worse than lateral resolution (Stewart, 1991). In contrast, surface waves have good depth resolution, though the structures are laterally averaged (Stewart, 1991). Surface waves are also preferable for this work in particular, as in areas with little to no instrumentation, some information is still recovered provided there are crossing paths in these regions. This is beneficial for imaging Saskatchewan and Manitoba which have a low station density. Surface wave tomography is commonly employed to measure the variation in velocity with depth, because of an inherent feature of surface waves – dispersion – which is the phenomenon whereby waves of different frequencies travel at different velocities. Surface wave velocity varies with period, with shorter periods being sensitive to shallow depths, and longer periods sensitive to greater depths (Stein and Wysession, 2003). Additionally, surface wave tomography is an accurate measure of lithospheric velocity as it provides a measure of absolute velocity rather than relative velocity, as is generally the case with teleseismic body wave tomography.

1.2 Two-station Tomography Method.

Surface wave tomography methods involve two key steps – surface wave dispersion determination and tomographic inversion. Single-station and two-station surface wave analysis methods are the primary methods that are used to conduct these type of analyses, however there are other methods that have been developed, such as array methods, eikonal tomography, Helmholtz tomography and ambient noise tomography methods. The two-station method, first introduced by Sato (1955), is the widely preferred technique for largely aseismic areas as well

as regions with sparse station coverage. It is advantageous as the derived dispersion curve reflects only the structure between the stations involved, regardless of the earthquake-station distance (Stewart, 1991).

For the two-station method to be employed in a study area, it requires two stations to be located within this area, as well as events lying on a great-circle path that includes these two stations. When the surface wave passes the first station, it is recorded, before continuing along this great-circle path to the second station, where it is also recorded. A dispersion curve is obtained by comparing the two recorded seismograms. The two-station method cancels out the earthquake source function phase effects (e.g., Darbyshire et al., 2007) and restricts structural sensitivity to the path segment between the two stations. The background theory of Rayleigh wave dispersion analysis has been fully described by numerous authors (e.g., Saunders et al., 1998 and Darbyshire et al., 2004).

In this study, existing two-station paths from the Canadian National Seismic Network (CNSN), Portable Observatories for Lithospheric Analysis and Research Investigating Seismicity (POLARIS) and USArray Transportable Array (TA) seismic archives are compiled, and these measurements are then used to create dispersion curves. New dispersion measurements are added to this existing dataset to improve overall resolution in largely aseismic areas. These dispersion curves are then inverted to produce phase velocity maps. In order to obtain a better understanding of the Canadian subsurface, the resulting model is compared to previously published models of Canada and its various sub-regions.

The two-station Rayleigh wave tomography approach that is used for this project models isotropic phase velocity variations and azimuthal anisotropy. Dispersion measurements are

inverted for isotropic, 2ψ and 4ψ anisotropic heterogeneity. Rayleigh wave phase velocity is much more sensitive to V_s (S wave velocity) than V_p (P wave velocity). In order to better interpret the phase velocity maps at various periods, Rayleigh wave sensitivity to V_s structure with depth for each period, previously determined for different subregions across Canada using the AK135 1-D model (Kennett et al., 1995), is used. The depth of peak sensitivity of surface waves increases with wave period, allowing progressively deeper structure to be sampled. However, longer periods also sample a wider range of depths, which is a factor that needs to be taken into account during interpretation. The tectonic subdivisions of Canada described in *chapter 2* are also taken into account during the interpretation stage.

1.3 Objectives.

This work seeks to unify existing regional models and put them into a larger context. The objective of this thesis is to examine the lithospheric thickness and velocity variations of the Canadian lithosphere using two-station Rayleigh wave tomography. Essentially, this work will map out Rayleigh wave dispersion across Canada with as much resolution as the available data permit, and interpret the resulting maps in light of Canadian tectonic subdivisions. This is accomplished by combining existing two-station paths with new, chosen paths to fill in the gaps in coverage. In order to properly image the Prairies where there is poor station coverage, seismic stations located in the US, just south of the Canadian border will be used to add new paths. New paths were also added using Canadian station pairs, in which one was located in the west and the other in the central-eastern region of Canada, with path intersections lying in the Prairie provinces. This thesis aims to reveal anomalies occurring across Canada in a broader context as compared to isolated anomalies noted in regional-scale studies of specific Canadian tectonic subdivisions.

2. GEOLOGICAL BACKGROUND.

The large-scale tectonic history of North America has been reconstructed from a combination of geological, geophysical and geochronological data, which has led to a division of the continent into tectonic provinces; the most recent and complete such division is described by Whitmeyer and Karlstrom (2007). The major tectonic subdivisions of Canada (Figure 2.1) are described, mapping the accretionary history of North America. It has been theorized that the oldest cratonic lithosphere was perhaps formed under a pre-plate tectonics regime and, in this region, later modified by processes such as plate collisions of Archean continents (e.g. Rae-Hearne with Superior), orogenesis, rifting, accretion of juvenile domains and volcanic arcs, hotspot passage and magmatism (Whitmeyer and Karlstrom, 2007).

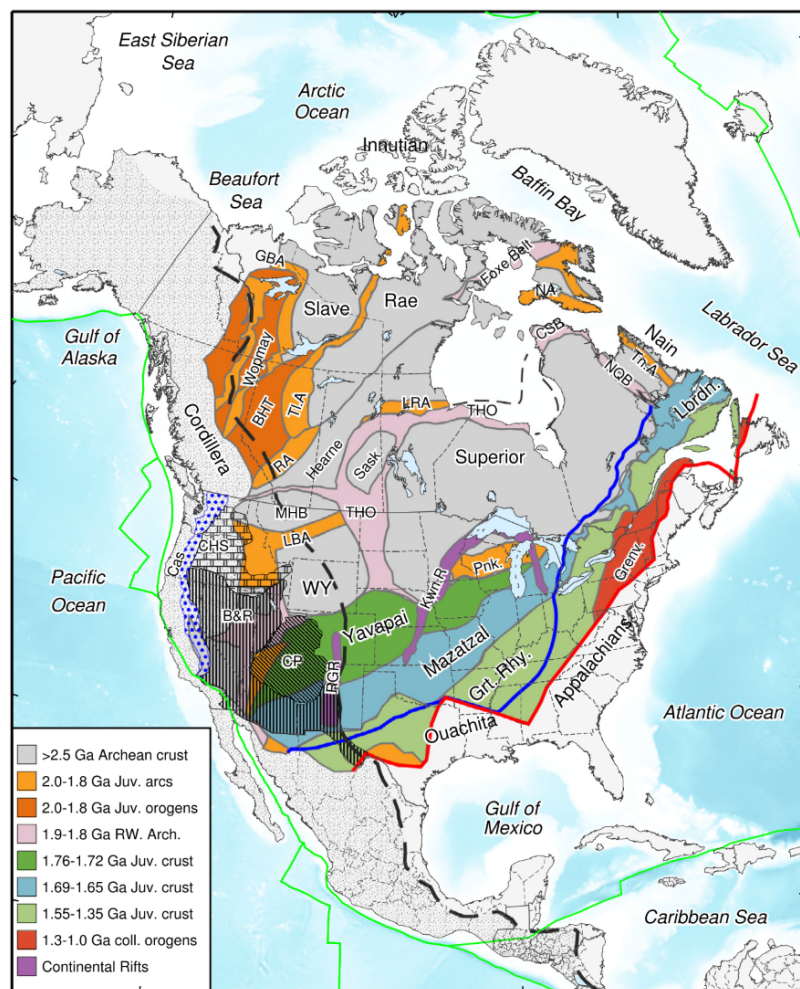


Figure 2.1. Map of North America depicting tectonic provinces (taken from Schaeffer and Lebedev, 2014). The red line indicates rift boundaries along eastern Laurentia. Abbreviations for structures interpreted in this study: THO = Trans-Hudson Orogen, MHB = Medicine Hat Bock, BHT = Buffalo Head Terrane, WY = Wyoming Province.

2.1 Central and Eastern Canada.

2.1.1 Superior Craton

The Superior Craton is the largest Archean craton preserved on Earth (Darbyshire et al., 2007), and forms the central core of the study region (Figure 2.2). It was assembled in the time period between approximately 3.0 to 2.6 Ga. Percival et al. (2012) describe the province as an accretionary orogen due to the identification of several prominent continental and oceanic terranes.

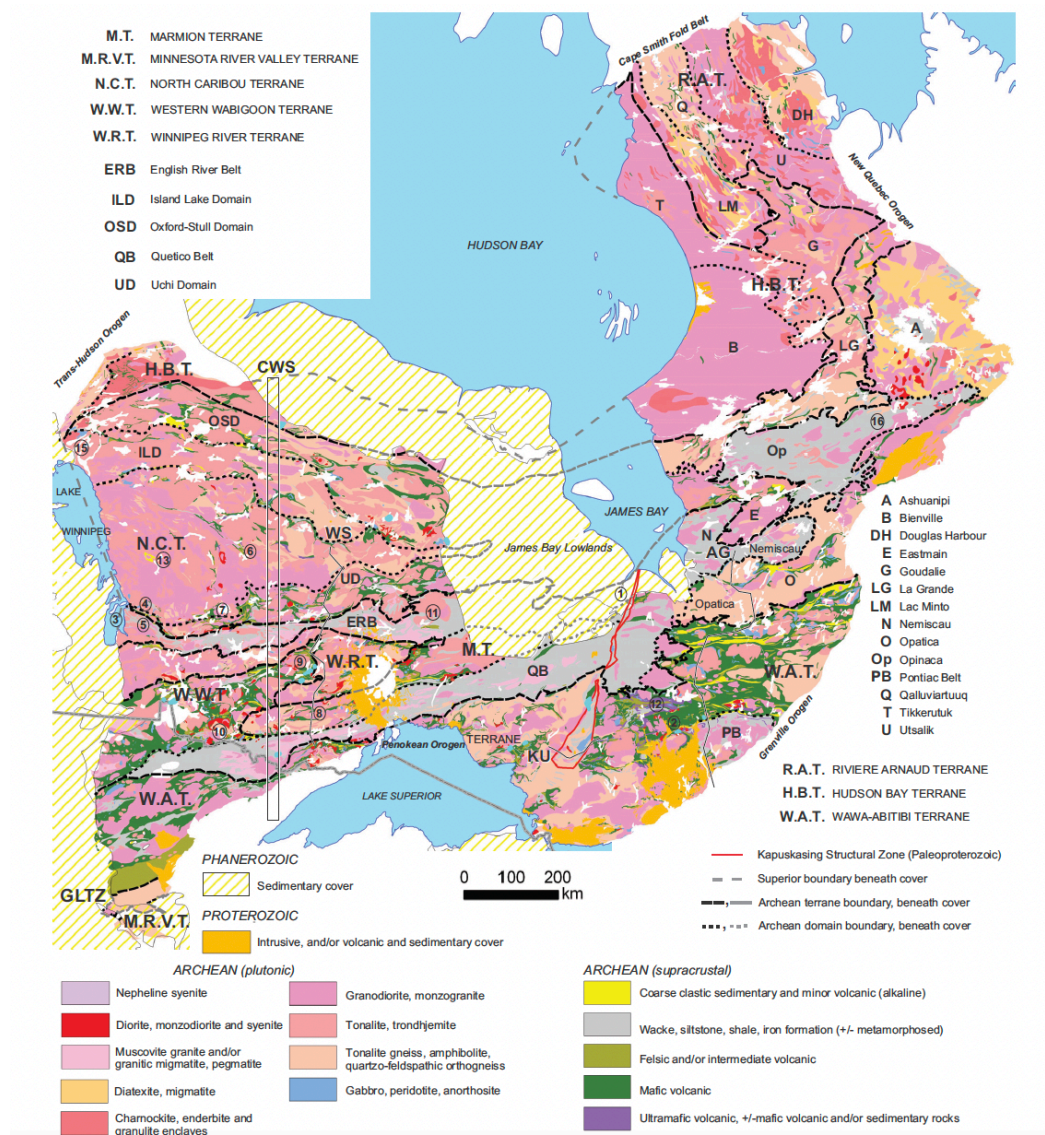


Figure 2.2. Tectonics and bedrock geology of the Superior Province (Percival et al., 2012).

Percival et al. (2006) describe five accretionary events of the western Superior. The core of the western Superior is comprised of an amalgamation of two continental blocks (the Northern Superior and the North Caribou superterrane), which according to Thurston et al. (1991) formed the nucleus for the later accreted terranes. The Northern Superior superterrane (NSS) features the oldest fragments of continental crust at an estimated age of 3.7 Ga. South of the NSS lies the 3.0 Ga North Caribou superterrane. Subsequent craton growth continued through a succession of accretions to the south. Accretions started with the Winnipeg River terrane, which features continental remnants as old as 3.4 Ga, followed by the Marmion terrane (~3.0 Ga) further south, and lastly the continental Minnesota River Valley terrane, which features very old crust at approximately 3.5 Ga (Percival et al., 2006). Terrane boundaries are broadly oriented in an E-W direction in the western and central Superior. Regions of mostly oceanic crust surround the older continental fragments across the province. Seismic studies carried out in the western and central Superior craton region feature structures which have been interpreted as fossil subduction zones (Musacchio et al., 2004).

The Abitibi subprovince, which is a collection of volcanic and plutonic material and encompasses the Earth's largest preserved greenstone belt, occupies the south-central Superior. In the northeastern Superior, there is a drastic change in the structural trends, denoted by N-S trending terranes. The Superior craton surface geology is quite complex, including tonalite-trondhejemite-granodiorite (TTG) suites, granites, high-grade metamorphic rocks and volcanic-plutonic suites (likely of oceanic origin) (Percival et al., 2012). There are two different interpretations for the assemblage of the Superior craton – the first being accretion via a similar process to modern-style plate tectonics (e.g., Percival et al., 2006), and the second being a non-tectonic regime (e.g., Bédard and Harris, 2014).

2.1.2 Grenville Province

The southeastern margin of Laurentia exhibited the behaviour of an active Andean-style margin between approximately 1.89 and 1.2 Ga, with subduction occurring below Laurentia, and with polarity eventually switching at some later point. According to Ludden and Hynes (2000) the Grenville Province originated subsequent to cumulative terrane accretions onto the southeast of the Superior Province, via several Proterozoic orogenic episodes known as the Penokean (~1.9-1.6 Ga), Labradorian (~1680 – 1660 Ma), Pinwarian (~1500-1450 Ma) and Elzevirian (~1250-1190 Ma).

The Grenville Orogeny (Figure 2.3) was a major continental collision in the formation of Rodinia, occurring in three distinct pulses – a Shawinigan, an Ottawan and a Rigolet – though there remains uncertainty regarding whether or not the Shawinigan episode was actually linked to continental collision (Ludden and Hynes, 2000). Due to this uncertainty, it is currently widely accepted that the collision which allowed for the formation of the Grenville Orogen occurred in two pulses (Ottawan and Rigolet). This collision was initiated at the southeast of Laurentia at approximately 1.11 Ga, leading to crustal shortening and later emplacement of AMCG complexes (anorthosite, mangerite, charnockite, granite and related granitoids). This was followed by a second episode of crustal shortening and AMCG complex emplacement a short time later (Corrigan and Hanmer, 1997).

Meanwhile, continued collision resulted in the formation of an extensive plateau (similar to the Tibetan plateau). The Grenville Front, northwest of the Grenville Province, also emerged during this orogeny period; it signifies the northwestern deformation limit. Crustal profile models (e.g., Ludden and Hynes, 2000) suggest that the Grenville Orogen overlies Archean

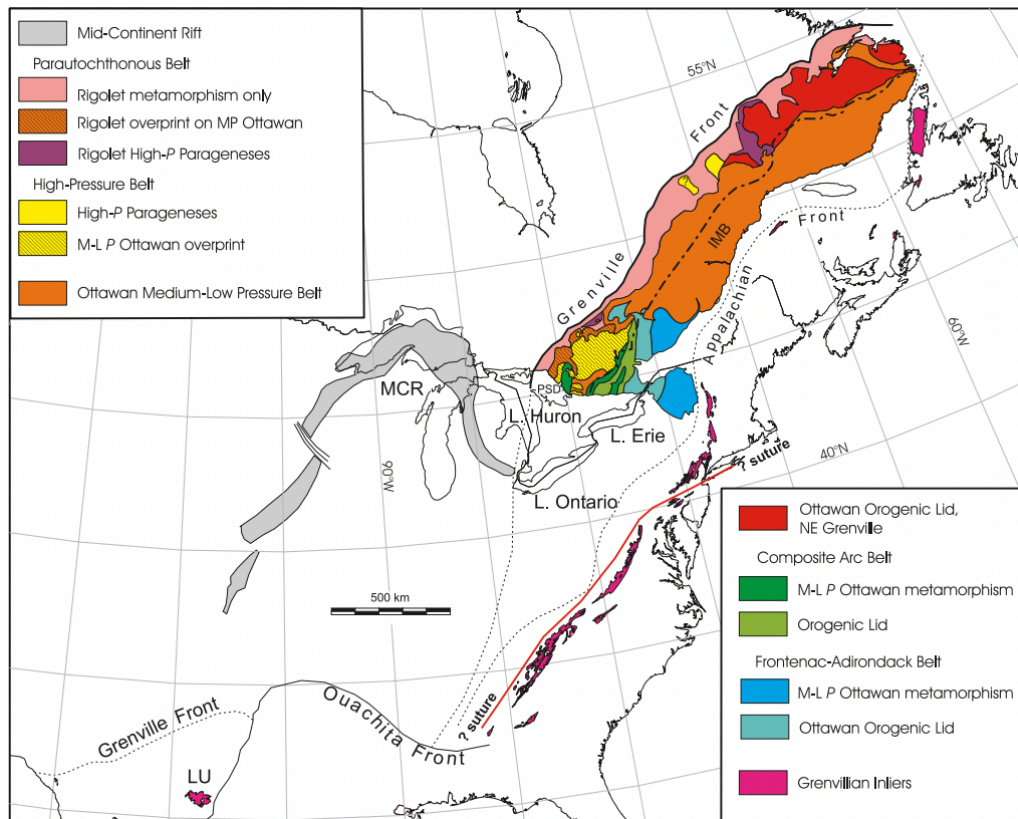


Figure 2.3. Map highlighting Grenville tectonics (Hynes and Rivers, 2010).

basement up to approximately 250 km southeast of the Grenville Front, though the structure of the margin is complicated.

2.1.3 Appalachian Orogen

The Appalachian orogenic belt, located southeast of the Grenville Province, documents the closure of the Iapetus and Rheic Oceans, which occurred between 480 and 260 Ma (van Staal et al., 2012). The Appalachian terranes were accreted to eastern Laurentia during the Palaeozoic. According to Whitmeyer and Karlstrom (2007), following an episode of failed rifting, the Appalachian mountains were initially located on the eastern side of the rift (in the vicinity of Gondwana). These mountains were later re-accreted to Laurentia during what is now termed the “Appalachian orogenic cycle”, occurring during the Palaeozoic. Following this accretion, no further notable phases of either tectonism or magmatism impacted the area until

the effect of the Great Meteor Hotspot at approximately 200-110 Ma (Heaman and Kjarsgaard, 2000). The Appalachians and the Caledonian Orogen were part of the same Palaeozoic orogenic belt during the assembly of Pangea, but were split up due to the opening of the Atlantic Ocean.

2.1.4 Trans-Hudson Orogen and Hudson Bay

The Trans-Hudson Orogen (THO) is a Palaeoproterozoic orogenic belt that extends over 4000 km from central USA into central and northern Canada, and is approximately 1.8 Ga old. According to Corrigan et al. (2009), this orogeny resulted from the closure of the Pacific-scale Manikewan Ocean (likely > 5000 km wide), concluding with the terminal collision of the Archean-aged Superior craton with the Western Churchill (Hearne and Rae) craton (Darbyshire et al., 2017, and references therein).

Preceding terminal collision, accretion of oceanic arcs and continental fragments took place along the craton margins. Because of the shape, structure and lateral extent of the THO, it has been compared to the Himalayan-Tibetan-Karakoram orogen (HKTO) (St-Onge et al., 2006). The interior of the THO is largely characterized by an assemblage of intraoceanic juvenile Palaeoproterozoic arc terranes of estimated age 1.9-1.8 Ga. Darbyshire et al. (2017) divides the Canadian region of the THO into two distinct segments – the “southern” segment (comprising Manitoba-Saskatchewan) and the “northern” segment (intersecting northern Quebec and southern Baffin Island) – whose characteristics are summarized.

At 1.92 Ga, the oceanic lithosphere of the southern segment was still undergoing subduction, leading to the accretion of continental sections, and later oceanic terrane accretion to the SE boundary of the Hearne (Hammer et al., 2010). During the terrane assembly, the Wathaman

Batholith, an Andean-scale batholith, was formed along the Hearne boundary. A collision between the Sask craton, a small Archean craton that was discovered via Lithoprobe (Lucas et al., 1993), and the accreted terranes along the southeastern Hearne, occurred at approximately 1.83-1.84 Ga, following which, terminal collision took place between the lower plate Superior craton and the upper plate Hearne during 1.83 to 1.80 Ga. According to Hammer et al. (2010) craton stabilisation (in this segment) was fully achieved by 1.70-1.65 Ga. The northern segment featured the accretion of the Meta Incognita continental section to the Rae craton at approximately 1.88-1.856 Ga (Corrigan et al., 2009). Further accretions followed, with extensive regional arc magmatism carrying on until ~1.82 Ga. The Superior and Rae (with its accreted terranes) collided between 1.82 and 1.795 Ga (Corrigan et al., 2009).

There are four large Palaeozoic intracratonic basins in North America, one of which is Hudson Bay. It overlays a significant segment of the THO with sedimentary sequences to a depth of approximately 2.5 km, and contains water to about 100 m depth. Hudson Bay is deemed to be the largest basin (of the four Palaeozoic intracratonic basins) by surface area but has the least sediment thickness. The thick, inflexible cratonic lithosphere underneath is thought to be responsible for the slight subsidence (Eaton and Darbyshire, 2010). At the centre of Hudson Bay, a cluster of NNW-trending normal faults as well as the uplift in the area, separates the basin into two relatively equal sub-basins.

2.1.5 Mid-Continental Rift

The Mid-Continental Rift (MCR) is a 1.1 Ga aged U-shaped structure that is approximately 3000 km long, which crosscuts the Archean, Palaeoproterozoic and Mesoproterozoic tectonic provinces to the south of the western Superior craton (Van Schmus and Hinze, 1985). It is

considered to be a failed rift system, speculated to be in response to stresses associated with the Grenville Orogeny (Hinze et al., 1992). A less widely accepted alternative interpretation was put forward by Stein et al. (2014) which suggests that the MCR was formed as a part of the rifting of Amazonia and Laurentia, which then became inactive following stable seafloor spreading.

The MCR is centred around a graben with the arms consisting of shallow, inwardly dipping strata. The stratigraphy of the MCR predominantly consists of mafic volcanic and plutonic rocks, belonging to the Keweenawan Supergroup, which have exposed outcrops in the Lake Superior region, and have been thoroughly studied. The volcanic units are dominated by basalt flows that are tholeiitic in nature. The Keweenawan Supergroup consists of mainly igneous rocks, but also sedimentary clastic rocks. The MCR structural geometry is made up of several axial horsts that are all aligned, which predominantly consist of mafic extrusive rocks – basalts, olivine tholeiite and rhyolites (Hutchinson et al., 1990) – all of which are overlain by late rift clastics. The basins are confined by normal faulting, while late stage reverse faulting affects the graben.

The MCR system is considered to be somewhat of an anomaly. Typically, rift systems do not contain any significant amounts of flood basalts, a category of large igneous provinces (LIPs), and these flood basalts are not usually associated with rifting and tensile/extensional events. The MCR however, contains a significant amount of igneous rocks that are predominantly flood basalt, emplaced via a hot mantle plume in two separate major episodes (Ojakangas et al., 2001). Interestingly, the MCR possesses a comparable volume of magma to typical flood basalts, but the flood basalts of the MCR are much thicker in comparison to other flood basalts due to its depositional region (Stein et al., 2018). The thickness of the MCR flood basalt reaches

a maximum of approximately 20 km, most of which is overlain by sedimentary bedding, up to 10 km in certain places (Ojakangas et al., 2001). The MCR then, although comprised of “rift-geometry”, has the composition and volume of an LIP.

2.1.6 New Quebec and Torngat Orogens

The New Quebec and Torngat orogens consist of continental sequences which record the transition from rift environments to foredeep environments (Wardle et al., 1990). Both orogens occur as a suture located between the Rae Province and the adjoined Superior and Nain Provinces. The New Quebec Orogen is characterised by a broad fold and thrust belt, though the Torngat Orogen, in contrast, is predominantly considered to be an extensive sinistral shear system.

2.2 Western Canada.

Western Canada features a Precambrian cratonic terrane with multiple younger terranes accreted onto it. The younger terranes range in age from the Early Permian to late Mesozoic-Cenozoic, while the assemblage of the craton took place largely in the Archean and Early Proterozoic. The cratonic terrane consists of three Archean-aged crustal blocks older than 2.5 Ga - the Slave, Rae and Hearne. The late Archean-aged Slave province is a granite-greenstone terrane and was a foreland both to the east and west, for the Thelon orogen and the Wopmay orogen, respectively (Hoffman, 1988). The onset of the Slave-Rae collisional zone occurred between 1.97 and 1.92 Ga, resulting in the formation of the Thelon due to the dextral-oblique colliding of the Slave and Rae provinces, and the indentation of the Rae province caused by the Slave foreland province (Hoffman, 1988). A noteworthy rifting event resulting in the

formation of rift boundaries at the western margin of Laurentia (the core of the North American craton) occurred during the Middle Proterozoic.

2.2.1 Rae and Hearne Cratons

The Rae and Hearne provinces are Archean blocks which share several common characteristics, and are located between the Superior and Slave provinces. Both the Rae and Hearne contain very old material, approximately 3.3-3.5 Ga, in addition to late Archean greenstone belts (Hoffman, 1988). The Churchill Province is comprised of the Rae and Hearne cratons, which are separated by the Palaeoproterozoic Snowbird Tectonic Zone (Hoffman, 1988).

The Rae domain is categorized by amphibolite to granulite grade metaplutonic gneisses as well as narrow supracrustal belts which sustained deformation during the Arrowsmith orogeny which onset during the early Palaeoproterozoic, during the period of 2.5-2.3 Ga (Berman et al., 2005). Overlying sedimentary (2.19-1.75 Ga) and volcanic material (1.83-1.78 Ga) was emplaced via plutonic intrusions and underwent reworking as a result of several accretionary episodes (Berman et al., 2005, 2007).

The Hearne domain consists of largely volcanic sedimentary material (2.35-1.75 Ga) and intruded plutons (1.83 and 1.78 Ga) which is underlain by low grade metamorphic supracrustal and plutonic rocks (Spratt et al., 2014). The Hearne province has sustained major Proterozoic reworking resulting from collisions and their associated impacts (Hammer et al., 2010).

2.2.2 Slave Province

The Slave Province is a relatively small cratonic segment, located in northwestern Canada, which features a Hadean to Mesoarchean basement complex and contains the Acasta gneisses which are regarded as the oldest rocks on the planet, with an estimated age of 4.027 Ga (Stern and Bleeker, 1998). The tectonic assembly of the province is estimated to have been completed by 2.60 Ga.

Unlike the majority of other Archean cratons, there are high quality exposures of the Slave craton which has allowed for superior bedrock mapping. According to Padgham and Fyson (1992) the Slave craton further deviates from other Archean cratons as it exhibits various unique facets – high volume of sedimentary material compared to volcanic material, high volumes of felsic rocks relative to mafic rocks, high volumes of sialic basement, as well as high volumes of potassium-rich granite. Additionally, the central Slave features the major diamond-producing Lac de Gras kimberlite pipes which were discovered in 1991 (Fipke et al., 1995).

The western region of the province features the Acasta gneisses, the oldest material, whereas the youngest material seems to be concentrated in the east of the province (Davis and Hegner, 1992), which has resulted in several possible interpretations regarding assembly and evolution. In the northwest, continent-building of the province was interrupted on three separate occasions resulting in continental breakup. The first incident occurred after Archean cratonization, circa 1.92-1.90 Ga during the development of the Coronation margin of the western Slave Province. The second event happened around 1.84-1.76 Ga, after the development of the Wopmay Orogen and resulted in the creation of the Fort Simpson-Wernicke basin (a Mesoproterozoic

westward-thickening basin). The third event occurred at approximately 0.75-0.55 Ga, which triggered the initial rifting and the formation of the miogeocline of the Cordillera.

2.2.3 Wopmay Orogen

The Wopmay orogen assembly occurred between 1.92 and 1.86 Ga, and is located to the west of the Slave-Rae collisional zone, adding juvenile lithosphere to the active western margin of the Slave Province. The orogen is considered to be a lithospheric plate accretionary orogen featuring arc magmatism associated with subduction and further accretion of exotic terranes (Cook and Erdmer, 2005).

There were two major episodes of arc accretion during the development of this orogen. Subduction (dipping west) was interrupted between 1.90 and 1.88 Ga due to the accretion of the Hottah terrane which resulted in the evolution of the Coronation margin of the Slave Province. East-dipping subduction was initiated and the formation of the Great Bear magmatic arc occurred between 1.875 and 1.843 Ga (Clowes et al., 2005). The Nahanni-Fort Simpson terrane was later accreted, terminating circa 1.745 Ga. The development of the Fort Simpson Basin came about due to postorogenic extension and subsidence (Cook and Erdmer, 2005).

2.2.4 Taltson and Buffalo Head

The Taltson magmatic zone occurs in the Canadian Shield and is a northerly trending belt of Archean to Palaeoproterozoic basement extending from Northern Alberta to southwestern Northwest Territories. It has been thought to be part of the result of the collision between both the Slave craton and Buffalo Head terrane with the Rae craton, which occurred between 2.0 and 1.9 Ga (Card et al., 2014). The Taltson magmatic zone is considered to be a southern

extension of the Thelon Magmatic Zone which is an orogenic belt located between the Slave and Churchill Provinces.

2.2.5 Medicine Hat Block

The Medicine Hat Block (MBH) is located between the Archean Wyoming Craton and the Archean Hearne Province (southwestern Laurentia). It is a structurally complex block of Precambrian crystalline crust, which is overlain by Phanerozoic cover. Geochemical data from crustal xenoliths provide evidence of the MHB being a distinct feature (Gifford et al., 2020). They also obtain U-Pb and Lu-Hf data from zircons that indicate protolith crystallization ages of 2.50-3.28 Ga, as well as magmatism at 1.76-1.81 Ga. This would suggest that the MBH was crucial to the assembly of western Laurentia, as an extension of the Montana Metasedimentary Terrane (MMT) of the Wyoming Province. The MBH and MMT were connected initially in the Mesoarchean, but broke apart in the early Palaeoproterozoic with the formation of a relatively small ocean basin; following its closure, the MBH and MMT rejoined during accretion to Laurentia via the Great Falls Orogeny (Gifford et al., 2020).

2.2.6 The Cordillera

The current framework of the Cordillera of western Canada represents more than 1.6 billion years of evolution. The complexity of the Cordilleran structure is largely due to the nature of the different terranes which were accreted, as well as the time at which these were accreted, rather than the metamorphism and volcanism which would have occurred relatively recently (Gabrielse and Yorath, 1991).

The Cordilleran terrane is comprised of several accreted terranes which can be categorized into the following groups: (1) oceanic terranes; (2) Meta pericratonic hinterland; (3) Cassiar

platform and shelf; and (4) North American platform and shelf (Gabrielse and Yorath, 1991) (Figure 2.4). Extensive carbonate platforms were formed along the west-dipping shelf which features uneven indentations resulting from multiple basins and embayments and the associated cratonic clastics (Gabrielse and Yorath, 1991).

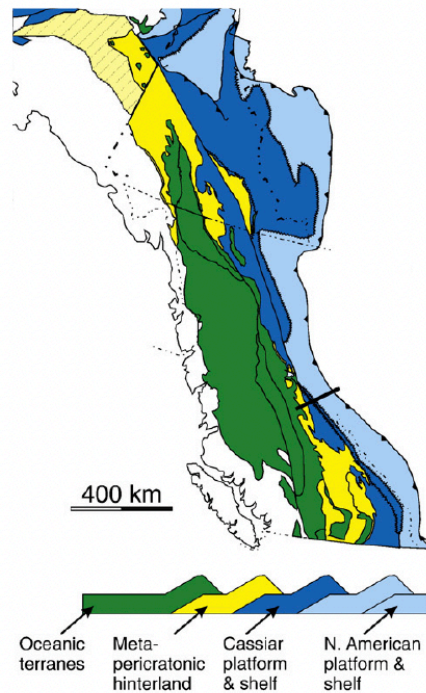


Figure 2.4. Geological map of western Canada's Cordillera (Gabrielse and Yorath, 1991).

The tectonic history of the Cordillera is as a result of the extension essential to the break-up of the Rodinia supercontinent which resulted in rifting within Laurentia (Hoffman, 1988), creating a passive margin. Phases of volcanism and graben formation during the Early Palaeozoic provide strong indications of rifting along the outer miogeocline (Gabrielse and Yorath, 1991). The Middle Devonian marked the generation of offshore magmatic arcs from the North American continental plate at the convergent margin. Ongoing subduction induced the eventual convergence of the continental plate margin with the offshore oceanic arcs, precipitating the formation of the Cordillera.

The miogeocline became separated from its basement around the time of the Middle Jurassic, underwent thickening after periods of folding, and was finally emplaced along the edge of the craton. This separation occurred as a result of the uplift, and to a lesser extent metamorphism and volcanism, that occurred within the Omineca Belt, following superterrane collisions (Gabrielse and Yorath, 1991). Within the Cenozoic, vast regions of juvenile volcanic material present in the southern Intermontane and Coast Belts provide evidence for a changing tectonic mechanism. There is extensive faulting, specifically right-lateral strike-slip faults, occurring across various fault systems within the Cordillera and representing over 1500 km of northwesterly movement (Gabrielse and Yorath, 1991).

2.3 Hotspot Tectonism.

Episodes of hotspot activity have affected the Canadian lithosphere throughout history. The Superior craton features the approximately 2.5 Ga Matachewan mafic dyke swarm (Figure 2.5), with dykes trending in a NNW to NW direction, covering a large area of the central Superior (Ernst and Bleeker, 2010). The Mackenzie dyke swarm (Figure 2.6) also appears across the Canadian Shield (part of the larger Mackenzie LIP), with dykes extending in a southeasterly direction from Arctic Canada to the Great Lakes. Its source is thought to be a mantle plume called the Mackenzie hotspot. The Great Meteor Hotspot has been thought to be responsible for the NW-SE trending magmatic track across eastern Ontario, southwestern Quebec and the northeastern US (Figure 2.7). Numerous kimberlite eruptions occurring around 190-150 Ma within the Superior craton, as well as igneous intrusions around 130-100 Ma, appear to indicate the presence of the suggested hotspot track (Heaman and Kjarsgaard, 2000).

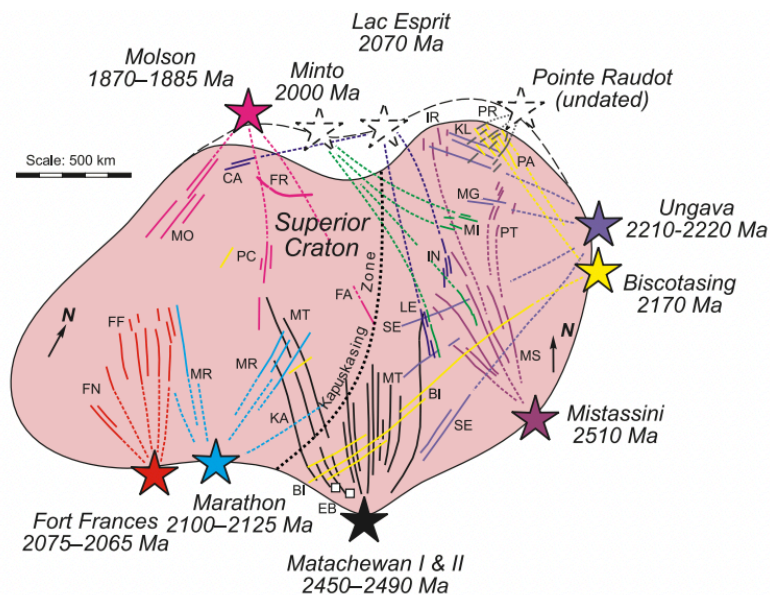


Figure 2.5 Schematic of the radiating dyke swarms associated with the breakup of the Superior craton from the ancestral landmass, Superia (taken from Ernst and Bleeker, 2010). The Superior craton features the Matachewan dyke swarm (MT).

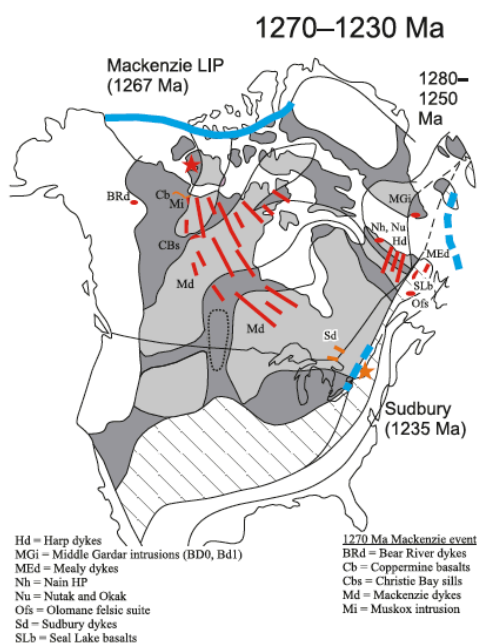


Figure 2.6. Map showing the location of the Mackenzie dyke swarm (taken from Ernst and Bleeker, 2010).

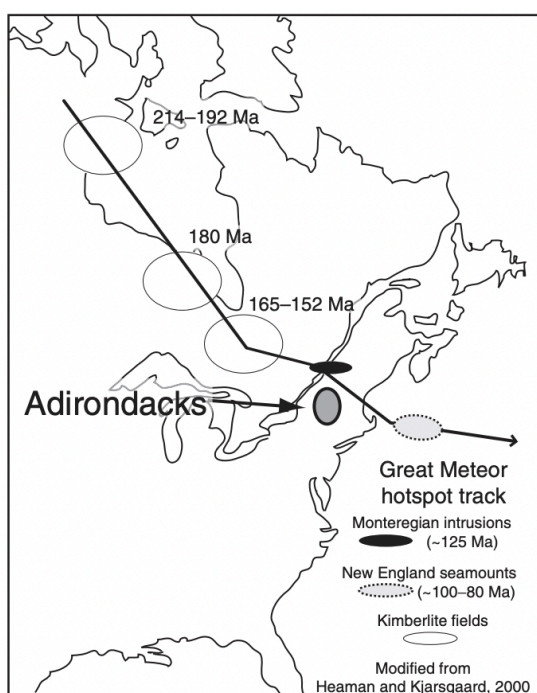


Figure 2.7. Map of the Great Meteor hotspot track modified from the findings of Heaman and Kjarsgaard, 2000 (taken from Taylor and Fitzgerald, 2011).

3. PREVIOUS GEOPHYSICAL STUDIES.

This work aims to put regional observations into a larger context. This review is geared largely toward North American models, and the corresponding observations these models reveal about the lithosphere. Observations from tomographic models produced using both surface wave and body wave measurements are examined, but only surface wave methodologies are discussed.

3.1 Surface Wave Tomography.

Various surface wave tomography methods have been used to image the Earth, the more common being one-station methods, two-station methods and array methods. One-station methods use the recording of an earthquake at a single station in order to determine the structure between the source and the receiver (Stein and Wysession, 2003). For distant sources, this usually encompasses large areas outside the study area. However, recording with the use of two instruments on a common great-circle path to the source allows for comparison of the signals at the two instruments, which will isolate the segment of the path between the two instruments. This is known as the two-station method, and results in better constraining of the local structure using distant earthquakes, as compared to one-station methods (e.g., Zaporozan et al., 2018). Array methods record seismic waves from earthquakes through the use of numerous seismometers in a well-defined, closely spaced network. This configuration allows for high-quality and homogeneous datasets to be produced, which can then be used to study the Earth's subsurface in great detail, on a local scale.

3.1.1 One-station Methods.

The partitioned waveform inversion, introduced by Nolet (1990), aimed to reduce the computational expenditure for large-scale seismic modelling problems by redefining the model parameters as path integrals over seismic velocity rather than have the model parameters be the seismic velocities. The partitioned waveform inversion method is a commonly used method in surface wave analysis and has been modified through the years to become a more efficient technique. The method uses waveform fitting to invert surface-wave recordings for an averaged 1-D structure (a non-linear inversion). Following this step, these average structures from many waveforms may be combined to obtain a 3-D structure via a linear inversion. The Nolet approach effectively removes the non-linearity in the first stage of the inversion, which contrasts with the approach commonly used for two-station methods.

A study by van der Lee and Nolet (1997) used the partitioned waveform inversion method described above. They examined the upper mantle S-velocity structure of North America to a depth of 660 km, through inversion of fundamental and higher mode Rayleigh waveforms. The resulting model (Figure 3.1) showed a high-velocity root below the North American craton, which does not exceed a depth of 250 km, a distinctive area of high velocity in the expected location of the subducted fragments of the Farallon plate (at the transition zone, at an approximate depth range of 300-400 km), and a low velocity area below the western United States, extending to a depth of 300 km.

Frederiksen et al. (2001) used the same partitioned waveform inversion method (Nolet, 1990; van der Lee and Nolet, 1997) to examine the S-wave velocity structure of the upper mantle of Canada. An inversion was performed to obtain a 3-D velocity model, found to be well-resolved to a depth of 400 km. The model (Figure 3.2) depicts two anomalies - high-velocity structure

below the Canadian shield and low velocities below the Cordillera. High velocities continue to depths of about 250 km below the North American craton.

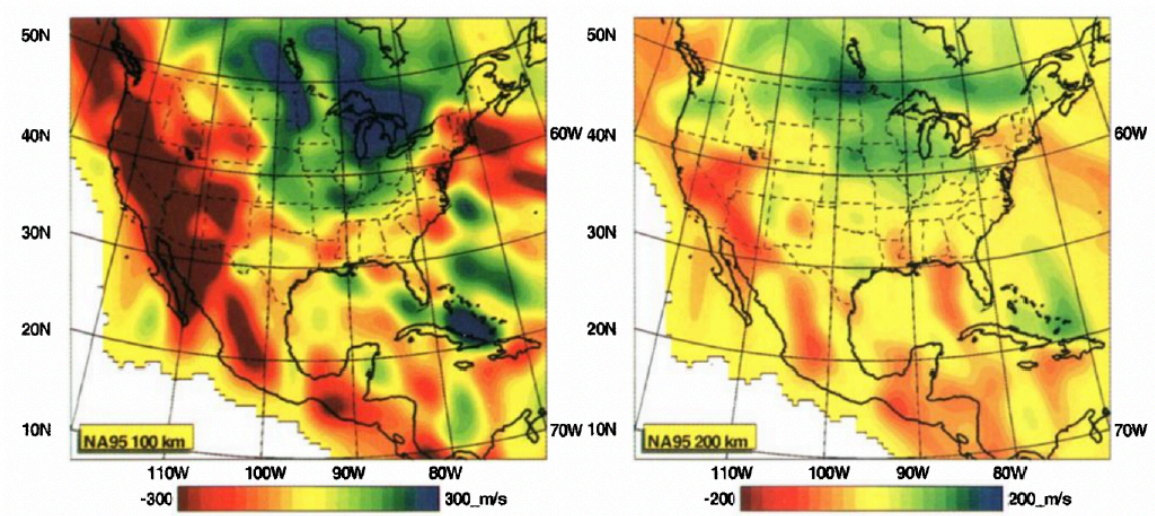


Figure 3.1. S-velocity model at depths of 100 km and 200 km (taken from Nolet and van der Lee, 1997).

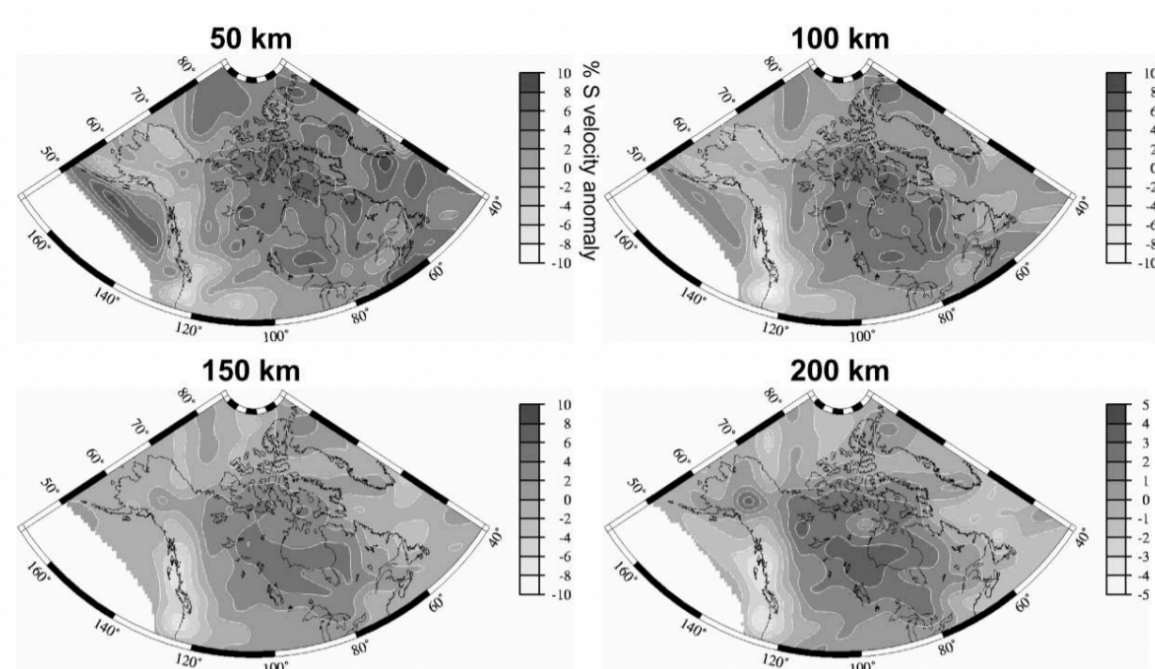


Figure 3.2. Frederiksen et al. (2001) S-wave velocity model at depths between 50 and 200 km.

Van der Lee and Frederiksen (2005) examined the 3D structure of the North American upper mantle using surface wave tomography via the partitioned waveform inversion method (Nolet, 1990). Their resulting 3D S-velocity model was consistent with previous findings (e.g., van

der Lee and Nolet, 1997; Frederiksen et al., 2001). The model found that the Pacific upper mantle was seismically slower than the Atlantic upper mantle. Additionally, beneath Precambrian North America, high S velocities suggested that the upper mantle may be up to 150% more rigid than that of Meso- and Cenozoic North America. High velocity anomalies were also recovered in the deep upper mantle, which were indicative of trailing fragments of the subducted Farallon and Kula plates. The model also found that the thickness of the high velocity region beneath the craton is approximately 250 ± 70 km.

Bedle and van der Lee (2009) examined the S velocity variations below North America, using the partitioned waveform inversion method (Nolet, 1990). The resulting model (Figure 3.3) found that the lithosphere beneath Archean regions of the craton extends to a depth of approximately 200 km, while the Proterozoic parts exhibit an average depth of approximately 175 km. The Archean region is also ~ 80 m/s faster than the Proterozoic. The western Cordillera exhibits a very sharp boundary with the craton, transitioning from slow to fast velocity over a distance of approximately 150 km. Within the transition zone, high velocity anomalies are apparent, although the anomaly is wider than has been shown in previous models (e.g., van der Lee and Frederiksen, 2005), possibly due to the additional data included for this work. This high velocity anomaly is thought to be associated with the subducted Farallon slab.

Schaeffer and Lebedev (2013) presented a global shear wave speed model of the upper mantle and transition zone, using the automated multimode inversion method of Lebedev et al. (2005). This method improves on earlier partitioned waveform methods by automating the initial waveform-fitting step. Although a global model was recovered, only results and interpretations of the North American region will be discussed. Active deformation beneath the young lithosphere of western North America is suggested by a prominent low-velocity anomaly. The

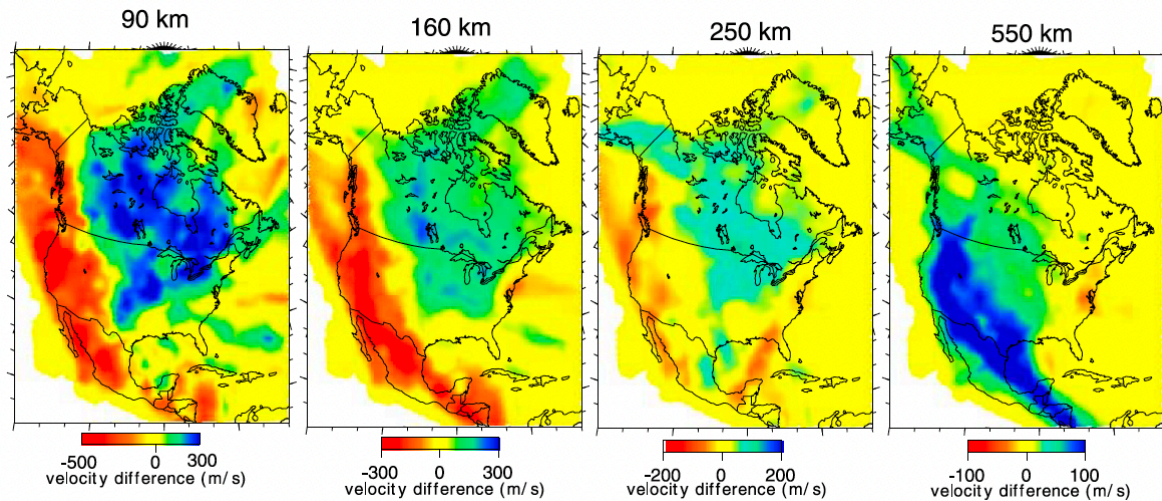


Figure 3.3. Horizontal depth slices at 90, 160, 250 and 550 km through model NA07 (Bedle & van der Lee, 2009).

model here is considered well-resolved, as it takes advantage of the combined distribution of earthquakes and stations across the planet, resulting in good raypath coverage. The model depicts the change of low to high velocities from west to east. Additionally, high velocity anomalies below North America are indicative of a complicated subduction history. The Juan de Fuca plate is imaged subducting below the Cascades at depths of 200 - 410 km. Fragments of the Farallon plate are imaged near the base of the transition zone and seem to extend eastward below the continental US towards the Great Lakes.

Schaeffer and Lebedev (2014) conducted a study examining the North American continent, seeking to further constrain their previous work (Schaeffer and Lebedev, 2013) in the North American context. This work used the automated multimode surface waveform tomography methods (Lebedev et al., 2005, Schaeffer and Lebedev, 2013). The resulting model (Figure 3.4) shows sharp velocity gradients in western Canada, which suggests that the boundary of the craton at depth maps onto the Rocky Mountain Front at the surface. Additionally, high seismic velocities unique to cratonic lithosphere, below the Great Bear Arc and Beaufort Sea, suggest the presence of the unexposed “Mackenzie craton” as proposed by researchers in

mineral exploration (Santana, 2007). This Schaeffer and Lebedev (2014) model is currently the most widely cited recent model for North America.

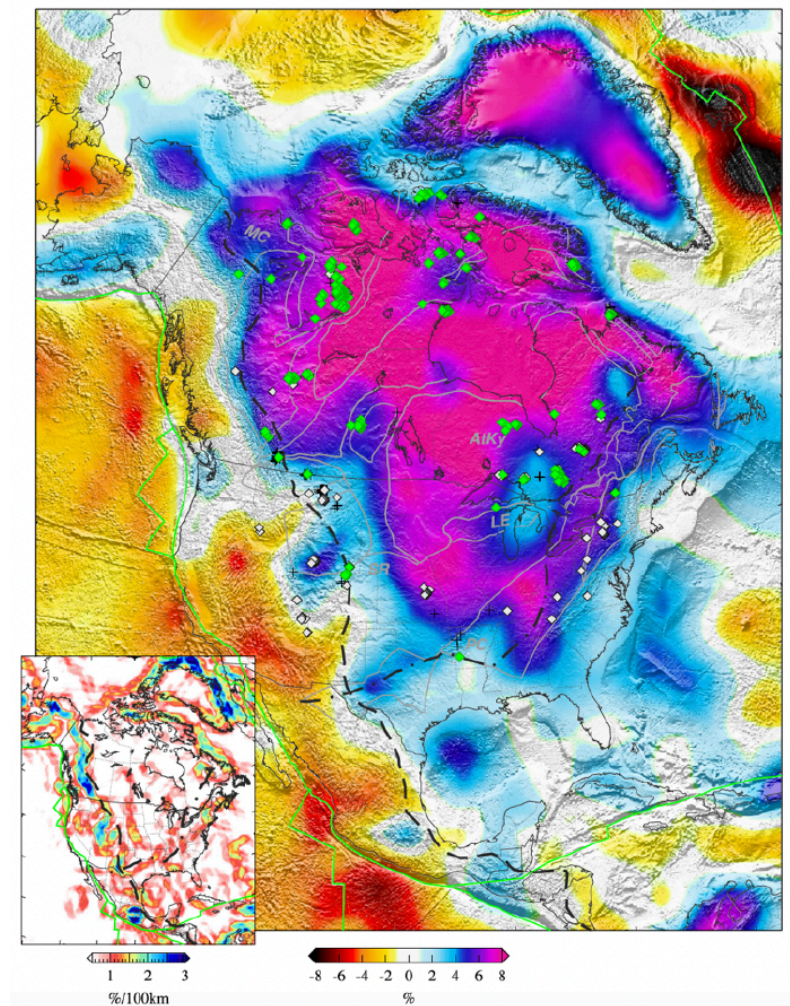


Figure 3.4. Schaeffer and Lebedev (2014) model of seismic wavespeed variation of North America at 150 km depth. The inset map shows significant lateral velocity gradients, plotted in percentage per 100 km. Diamond symbols: locations of diamondiferous (green) and non-diamondiferous (white) kimberlites and lamproites (World Kimberlite Database, Faure, 2010).

3.1.2 Two-station Methods.

This sub-section details studies which employ two-station methods at the regional-scale, in contrast to the continental and global studies described in *subsection 3.1.1*. These studies focus on the focus solely on the study area of this work, i.e. Canada and its tectonic sub-divisions.

Darbyshire (2005) used two-station fundamental mode Rayleigh wave tomography methods to examine the upper mantle structure of Arctic Canada. This two-station method contrasts with the partitioned waveform inversion method (Nolet, 1990), as with the two-station method, the dispersion maps are formed first. The phase velocity dispersion curves were obtained using a modified version of the transfer-function method of Gomberg et al. (1988) as it allows analysis of numerous seismogram pairs simultaneously. This modified method as detailed by Darbyshire et al. (2004) involves determining the phase velocity by representing the problem as a linear filter estimation, where the seismogram for a particular event at the most distant station is regarded as the convolution of the seismogram at the near station with an Earth filter, representing the dispersion curve. The dispersion curves obtained were plotted against the CANSND dispersion curve of Brune and Dorman (1963) for reference. Data from two-station paths at various Rayleigh wave periods were combined into phase velocity maps using tomographic inversion, specifically a stochastic inversion method (Franklin, 1970). The inversion performs calculations relative to a reference velocity; the average phase velocity at each period was used in this study. The resulting model (Figure 3.5) found that the shear wave velocity of the Canadian Shield was 4-5% higher than that indicated by global reference models. Additionally, continental lithosphere beneath the southern Arctic Islands, apart from Baffin, was significantly thinner than the ≥ 200 km previously reported. The Slave craton shows lower velocities than the rest of the Shield sampled in this study, which is resolved as globally fast, with a lithospheric thickness of > 200 km. It is important to note that anisotropy was neglected in the calculation and it was suggested by Debayle and Kennett (2000) that the overall pattern of lateral mantle heterogeneity across an area is not greatly affected by the inclusion or exclusion of anisotropy in calculations.

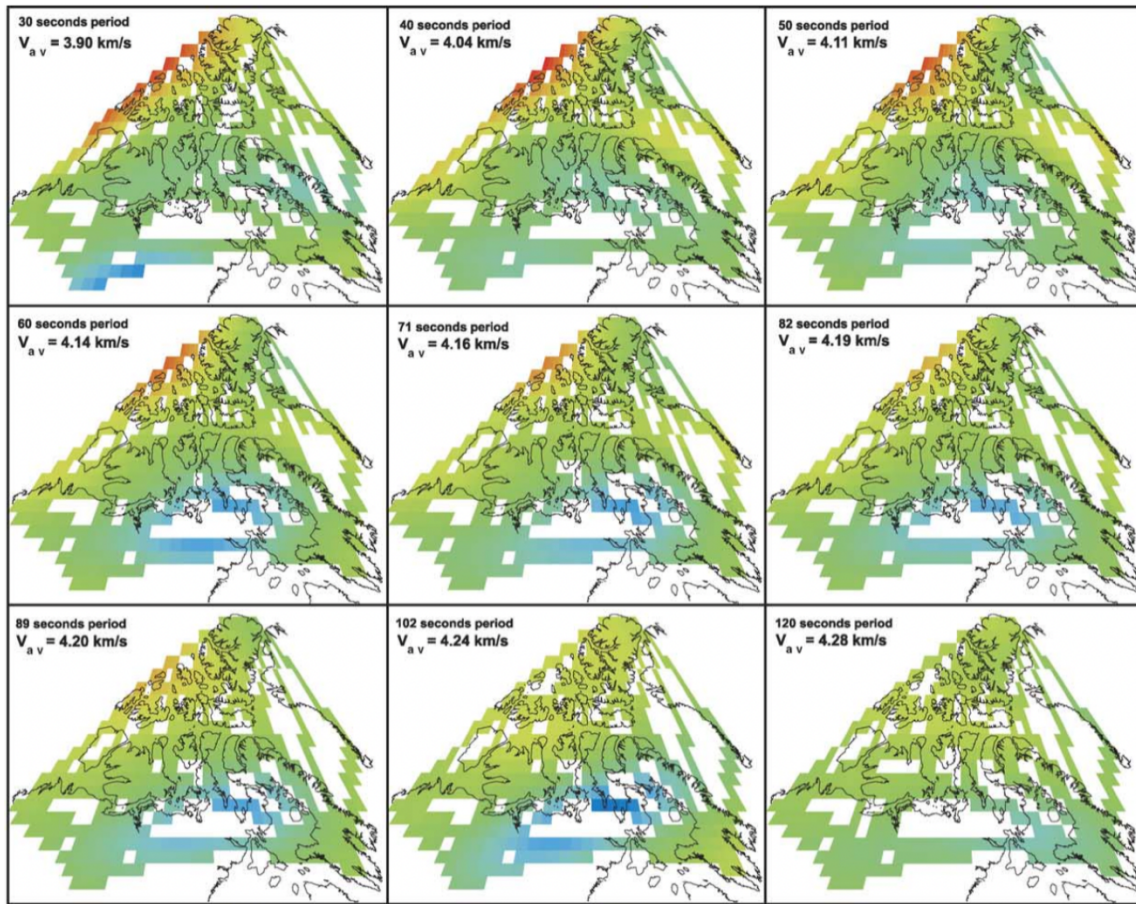


Figure 3.5. Phase velocity maps of Arctic Canada at periods ranging between 60 and 120s (modified from Darbyshire, 2005).

Darbyshire et al. (2007) used a two-station method to provide new insights into the Superior Province lithosphere. They used the same transfer-function method as Darbyshire (2005). This study found that lithospheric thicknesses vary between 140 and 200 km in the central and western Superior Province, with the thickest lithosphere, approximately 220-240 km, appearing in northeastern Ontario. Lithospheric thicknesses range from approximately 100-200 km across eastern Ontario. The huge variation may be attributed to the heterogeneous mantle structure and the presence of anisotropy across the region, particularly the Southern province, the Proterozoic Grenville Province and the Abitibi Greenstone Belt.

Darbyshire and Lebedev (2009) examined the Rayleigh wave phase velocity and azimuthal anisotropy of the upper mantle of the Superior and Grenville Provinces, using a two-station tomographic method, which solves for both isotropic and anisotropic terms (Deschamps et al.,

2008) with the use of an LSQR algorithm (Paige and Saunders, 1982). The method performs a least-squares inversion of dispersion data. The resulting model found that the isotropic phase velocity of the western Superior was largely faster than the regional average. The recovered anisotropy in this region, at periods less than 120 seconds, was interpreted to be due to the frozen lithospheric fabric associated with tectonic boundaries, while at longer periods, the anisotropy was attributed to sublithospheric flow. Interestingly, the anisotropy of the western Superior appears to be in roughly the same orientation for both the lithosphere and asthenosphere, in contrast to the rotation inferred in the eastern part of the study region. Across the majority of the central Superior Province, the model depicted phase velocities slower than the regional average, and pronounced rotation of fast direction of anisotropy at longer periods, suggesting a multi-layered structure of anisotropy in the upper mantle.

Darbyshire and Eaton (2010) examined the lithospheric root beneath Hudson Bay using Rayleigh wave dispersion. The method used to calculate the dispersion curves for this analysis is the cross-correlation method (Meier et al., 2004) – utilizing a two-station dispersion method to obtain phase velocity curves of fundamental Rayleigh wave modes. These dispersion curves were inverted for 1D depth profiles, representing the average structure along the interstation path. The modelling of the dispersion data used a two stage process patterned after that of Shapiro and Ritzwoller (2002) and Yang et al. (2008). A subset of the dispersion data was used, and individual curves underwent this linearised inversion procedure. The parametrization of the model was patterned after Yang et al. (2008). This study found high-velocity subcontinental lithosphere below which velocities decrease uniformly, tending to the global reference value. Lithospheric thicknesses in the Hudson Bay area ranged between 185 and 260 km. The Hudson Bay results suggested a layered lithosphere similar to that of the Slave craton, and a lithospheric root that is close to being buoyant at depth (Darbyshire and Eaton, 2010).

A 2013 study conducted by Darbyshire et al. examines the Hudson Bay lithosphere using a two-station method and a tomographic inversion. The method is the same as that described by Darbyshire and Lebedev (2009), to obtain phase velocity maps. From the phase velocity maps, at each grid point, a new dispersion curve is extracted, which therefore represents the 1D structure beneath that grid point. Each dispersion curve is then inverted to obtain the isotropic velocity following the Monte Carlo search protocol of Yang et al. (2008), and the anisotropic depth profile (Montagner and Nataf, 1986) for each point. The isotropic shear wave velocities for the 1D profiles were then interpolated to make the 3D final model. The anisotropic depth inversion model samples the structure of the upper mantle to depths of approximately 300 km. The study found lithospheric thicknesses ranging from 180 km in the northeast (below the rift zone in the vicinity of the Hudson Strait) to about 260 km below central-southern Hudson Bay. Within the upper mantle, beneath the intracratonic basin of central Hudson Bay, a relatively low velocity zone was observed surrounded by regions of higher velocity below the Archean aged areas (Figure 3.6). The highest velocities resolved were associated with cratonic cores, while lower seismic velocities were exhibited by the Trans-Hudson Orogen Proterozoic mantle.

Work conducted by Petrescu et al. (2017) explores the eastern Superior craton using a Rayleigh wave two-station method. The dispersion measurements were made using the cross-correlation method of Meier et al. (2004) and the same tomography method as Darbyshire and Lebedev (2009), described above. This work (Figure 3.7) found that the lithosphere of the Superior Craton is difficult to interpret anisotropically - the upper region exhibits fast seismic velocities and anisotropic fabrics that match Archean tectonic boundaries, while the lower lithosphere is more homogeneous than the upper, though anisotropically weak, which was interpreted to be due to regionally slow mantle convection at the time of formation.

A study by Zaporozan et al. (2018) investigates the Cordillera-craton boundary of western Canada using two-station surface wave analysis, making use of the cross-correlation method of Meier et al. (2004) to measure phase velocity dispersion curves. Both tomographic and depth

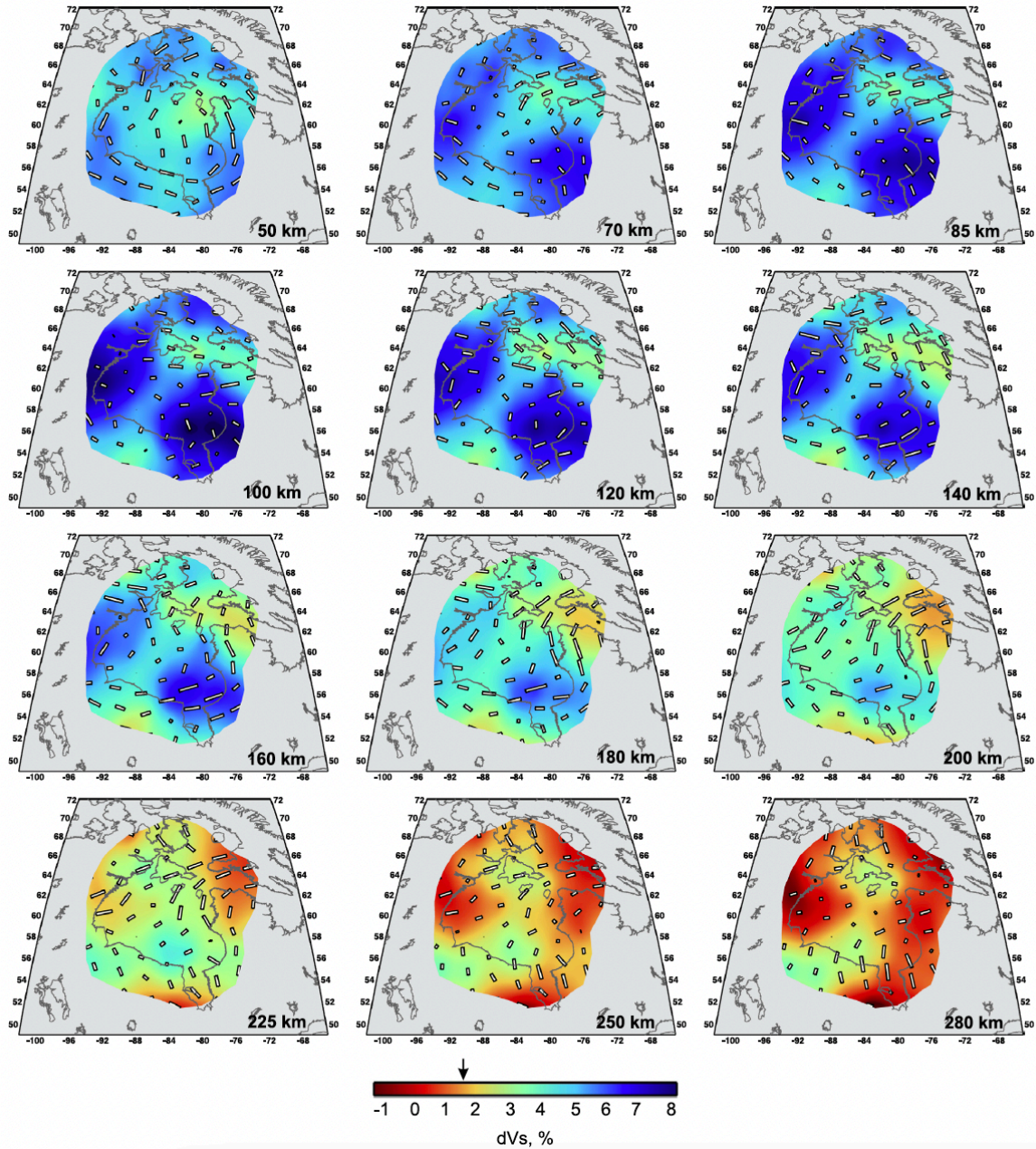


Figure 3.6. Horizontal slices through 3D model of the structure of the upper mantle of the Hudson Bay region (Darbyshire et al., 2013).

inversions were conducted in this study. Due to the nature of Rayleigh wave phase velocity, its relationship with depth sensitivity is non-linear and model-dependent. The depth inversion in this study, though, was performed making use of the linearized seismic package (a least squares method) from Herrmann (2013) using the phase velocity from the tomographic maps, at a range of periods. This study found similar results to those discussed above, low Cordilleran upper

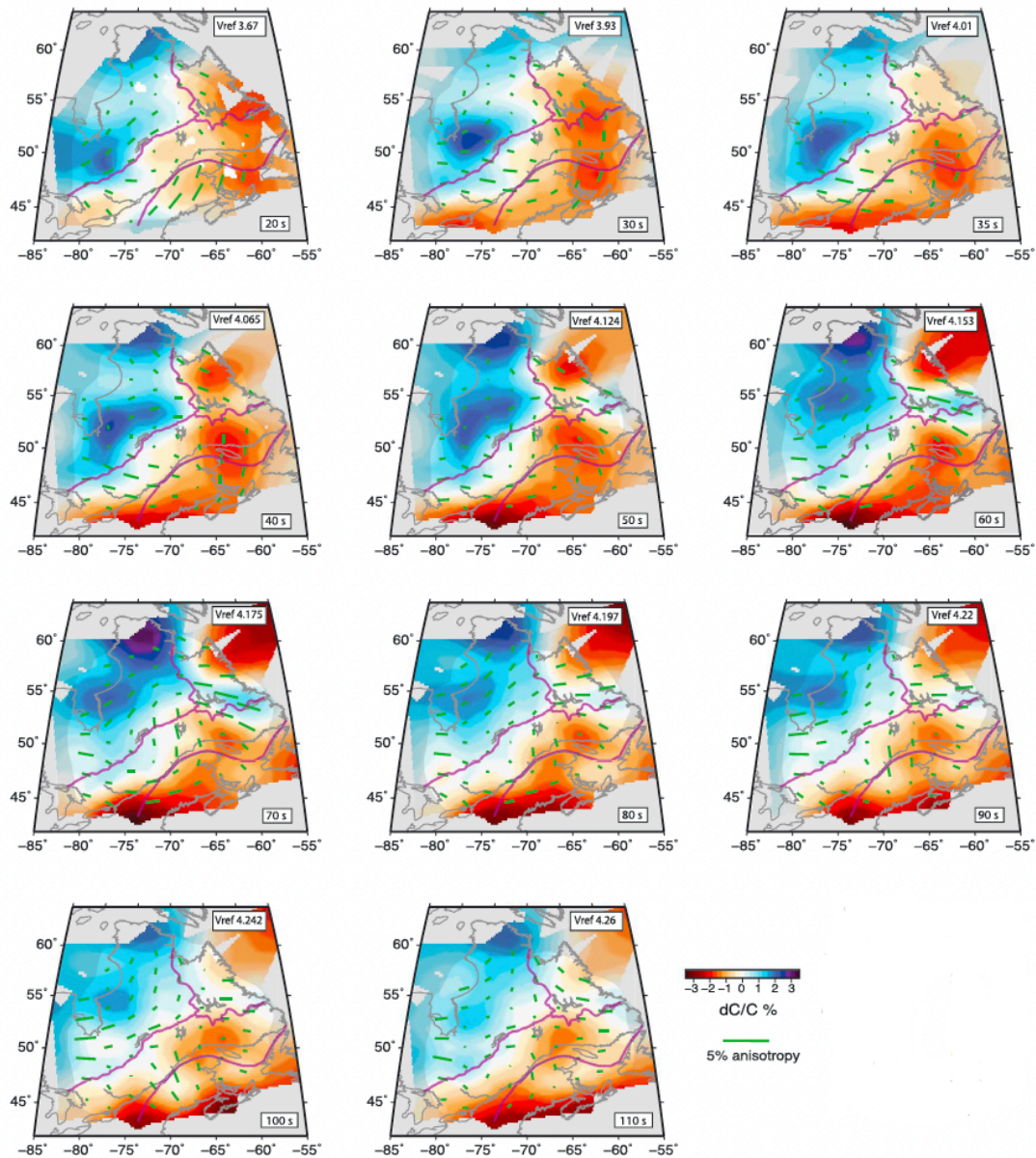


Figure 3.7. Phase velocity maps depicting the eastern Superior, for periods ranging from 20 to 110 seconds (modified from Petrescu et al., 2017).

mantle velocities (as observed by Schaeffer and Lebedev (2014); van der Lee and Frederiksen (2005)), as well as a high velocity cratonic lithosphere which was previously detected by the array method of Chen et al. (2007). Notably, a change in dip in the mantle contact was also observed - the dip direction changes from northeast in the north to southwest, moving south along the Cordillera-craton boundary. This study also finds evidence to support that the lithosphere below the Cordillera is quite thin, approximately 60 km.

A study by McLellan et al. (2018) examines the Northern Canadian Cordillera using fundamental mode Rayleigh wave dispersion measurements with the use of two different two-station approaches: ambient noise cross-correlation (Ritzwoller et al., 2011) and teleseismic two-station interferometry (Meier et al., 2004). The ambient noise method includes two primary stages (Bensen et al., 2007). Step one involves data preparation, while step two involves the cross-correlation and stacking of noise traces at station pairs. Empirical Green's functions are produced as a result, which serve as a simulated representation of surface waves as though one station were a source, the other a receiver (Shapiro and Campillo, 2004). For this method, curves are produced within the period range of approximately 5 to 30 seconds. The earthquake-based dispersion curves provide a wider bandwidth, producing curves at periods ranging from 20 to >200 seconds. Between 20 and 30 seconds, there is overlap between the dispersion curves produced for both methods. Anisotropic phase-velocity maps (Figure 3.8) were produced by inverting these dispersion curves, using the method described by Darbyshire and Lebedev (2009), discussed earlier. Inversion for depth profiles was also done according to the method of Lebedev et al. (2009). This work found high positive phase velocity anomalies at depth within the crust of the Canadian Shield, which seemingly indicate igneous and metamorphic rock compositions within the craton, and within the Mackenzie mountains, negative velocity anomalies were observed reflecting the sedimentary Selwyn Basin. There

was also evidence to suggest that the Cordillera-craton boundary probably extends to the Tintina Fault below the Cordillera.

Recent work by Foster et al. (2020) studies the anisotropic structure of the craton around the region of the Mid-Continental Rift (MCR), using a Rayleigh wave two-station method. The methodology used for dispersion measurements follows that of Petrescu et al. (2017) outlined above. The smoothed dispersion curves are then inverted to produce phase velocity maps

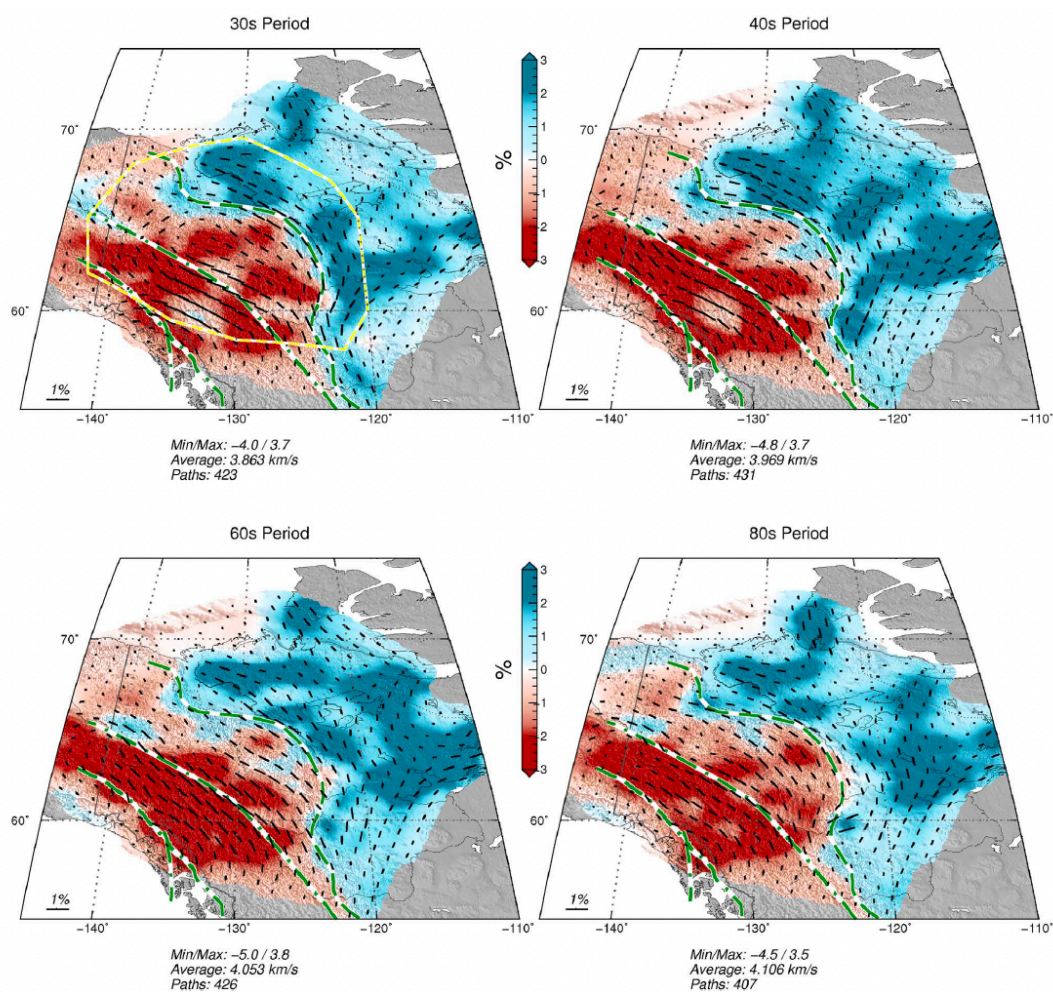


Figure 3.8. Phase velocity maps of the northern Canadian Cordillera for periods ranging from 30 to 80 seconds (taken from McLellan et al., 2018).

(Figure 3.9), following the method described by Darbyshire and Lebedev (2009). This study confirms previous findings of fast seismic velocities within the Superior Craton (Frederiksen

et al., 2013; Boyce et al., 2019; Darbyshire and Lebedev, 2009; Petrescu et al., 2017; Schaeffer and Lebedev, 2014). At longer periods however (100s and longer), high seismic velocities are only exhibited on the western half of the Superior craton, while the eastern side exhibits velocities similar to the neighbouring orogens. The Proterozoic orogens present as relatively low-velocity anomalies. The Mid-Continental Rift is the slowest velocity anomaly within the study area, though the slowest region varies at different periods.

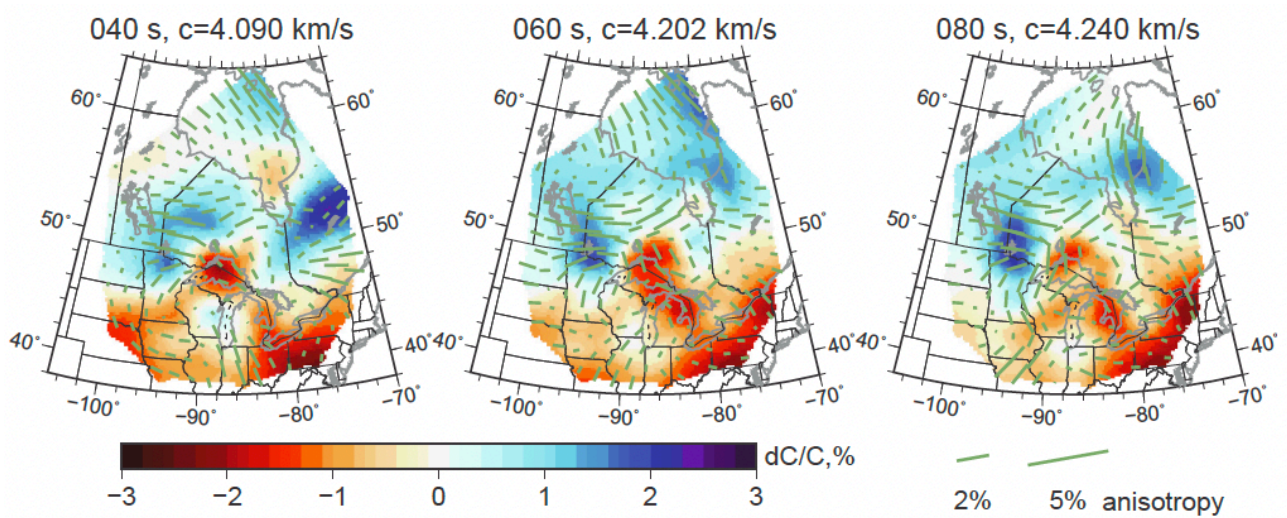


Figure 3.9. Phase velocity maps depicting the craton surrounding the MCR at periods between 40 and 80 seconds (modified from Foster et al., 2020).

3.1.3 Array Methods.

An array method was used by Chen et al. (2007) to provide insight into the velocity structure of the upper mantle of the Slave craton with the use of Rayleigh wave inversion - specifically through the examination and analysis of Rayleigh wave phase and amplitude data in the northwestern Canadian shield. The method used employs fundamental mode Rayleigh wave data, which were inverted at periods ranging from 12 to 142 seconds using the crossing ray paths from teleseismic events recorded by a regional array to obtain phase velocities. It is important to note that array methods require stations to be close together. The array approach

is similar to the two-station method (measuring dispersion curves), but in a way that generates a 1D average dispersion curve for the structure beneath the array, thus requiring dense station coverage, which makes it an unfeasible approach for Canada as a whole. This model recovered high S-wave velocities for the upper mantle, ranging from 4.7 ± 0.2 km/s, that extend to a depth of 220 ± 65 km, which characterises the cratonic lithosphere. The Slave craton also exhibits a high degree of mantle depletion. At all periods, it was found that azimuthal anisotropy was significant, which implies that below the Slave craton, the upper mantle anisotropy is impacted by “both lithospheric fabric and sub-lithospheric flow” (Chen et al., 2007).

3.2 Body-wave Tomography Observations.

Body waves are seismic waves that move through the Earth’s interior, unlike surface waves which travel near the surface of the Earth. There are two types of body waves: primary or pressure wave (P-wave) and secondary or shear wave (S-wave). P-waves travel approximately 40% faster than S-waves in any given material (Shearer, 2009). P-waves are capable of travelling through both solid and liquid media, while S-waves cannot propagate through liquid. Body-wave tomography is a technique used to image the interior of the Earth using these P- and S-waves either separately or together. The studies reviewed in this section use teleseismic earthquakes, for which rays are steeply incident in the upper mantle, and provide information on a depth range of 400 to up to 700 km depth (Youssof et al., 2015).

Work conducted by Villemaille et al. (2012) studied mantle evolution and modification of southeastern Canada and northeastern USA, using a teleseismic P-wave travel-time tomography method. The observations from the resulting model (Figure 3.10) indicated no

systematic variation of the seismic wavespeed between the Archean and Proterozoic provinces, though there is evident lithospheric modification across the terranes. The Grenville Province (Proterozoic) exhibits localized modification while the Appalachian orogen (Paleozoic) exhibits universal modification, and beneath the Superior craton (Archean), there is no modification of the structure of the lithosphere.

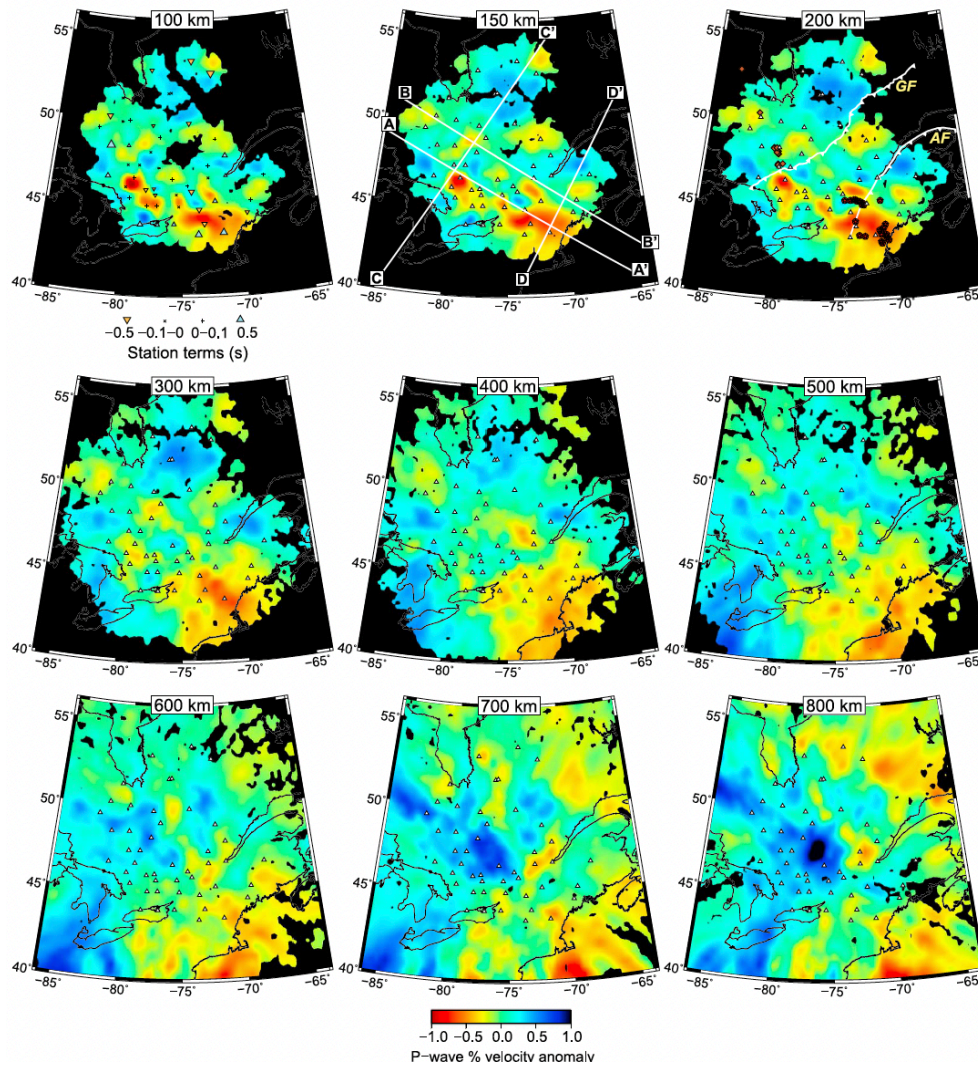


Figure 3.10. Final model depicting slices through the upper mantle, transition zone and mid-mantle (taken from Villemaire et al., 2012).

A study by Frederiksen et al. (2013) examines the central North American continental lithosphere, using P-wave tomography. The resulting model found evidence of a high-velocity anomaly occurring at depth beneath the lithosphere (at approximately 400 km depth), which

was interpreted to be a possible Farallon plate fragment. The model further interprets a large high velocity anomaly existing between a depth range of 150-200 km, to be the Western Superior lithospheric mantle. Furthermore, although mostly poorly resolved due to lack of resolution in this region, the Mid-Continental Rift shows a low-velocity anomaly along the segment that is well resolved (i.e., the western arm of the MCR). The isotropic tomography model also indicates the presence of a low-velocity anomaly beneath Minnesota and the Dakotas, though a conclusive interpretation of this finding was not made.

Another eastern Canada study conducted by Boyce et al. (2016) specifically examined the subduction beneath Laurentia, using both P and S wave tomography. Their models found that there is a consistent increase of the seismic wave speeds from the Appalachian domain (Phanerozoic) through the Grenville Province (Proterozoic), toward the Superior Province (Archean). The models also depict a low velocity anomaly, possibly related to the passage of the Great Meteor hotspot across the major tectonic regions. Additionally, a clear vertical wave speed boundary was detected beneath the Grenville Front, which is interpreted to result from a metasomatized cratonic edge.

Liddell et al. (2018) conducted a body wave (P and S wave) tomographic study of northern Hudson Bay. Their models find complexities in the structure of the Archean Churchill plate, which suggest a complicated accretionary record. The Churchill Province exhibits relatively lower seismic velocities than the Archean Superior Province. The model also demonstrates a distinct velocity contrast at shallow depths surrounding Southampton Island.

Another 2019 study conducted by Chen et al. examines the seismic evidence for a mantle suture along the Canadian Cordillera and consequently, the ramifications for the origin of the

Cordillera. The tomography model (Figure 3.11) shows negative velocities relative to the global reference model (Kennett et al., 1995) below the southern Cordillera, though in the east, beneath the Alberta foreland basin, positive velocities are seen. The lithospheric margin is defined by a structure exhibiting high velocities, which dips steeply to the west, approximately 10° from vertical, beneath the Cordilleran Foreland Belt. The dip of the Cordillera-craton

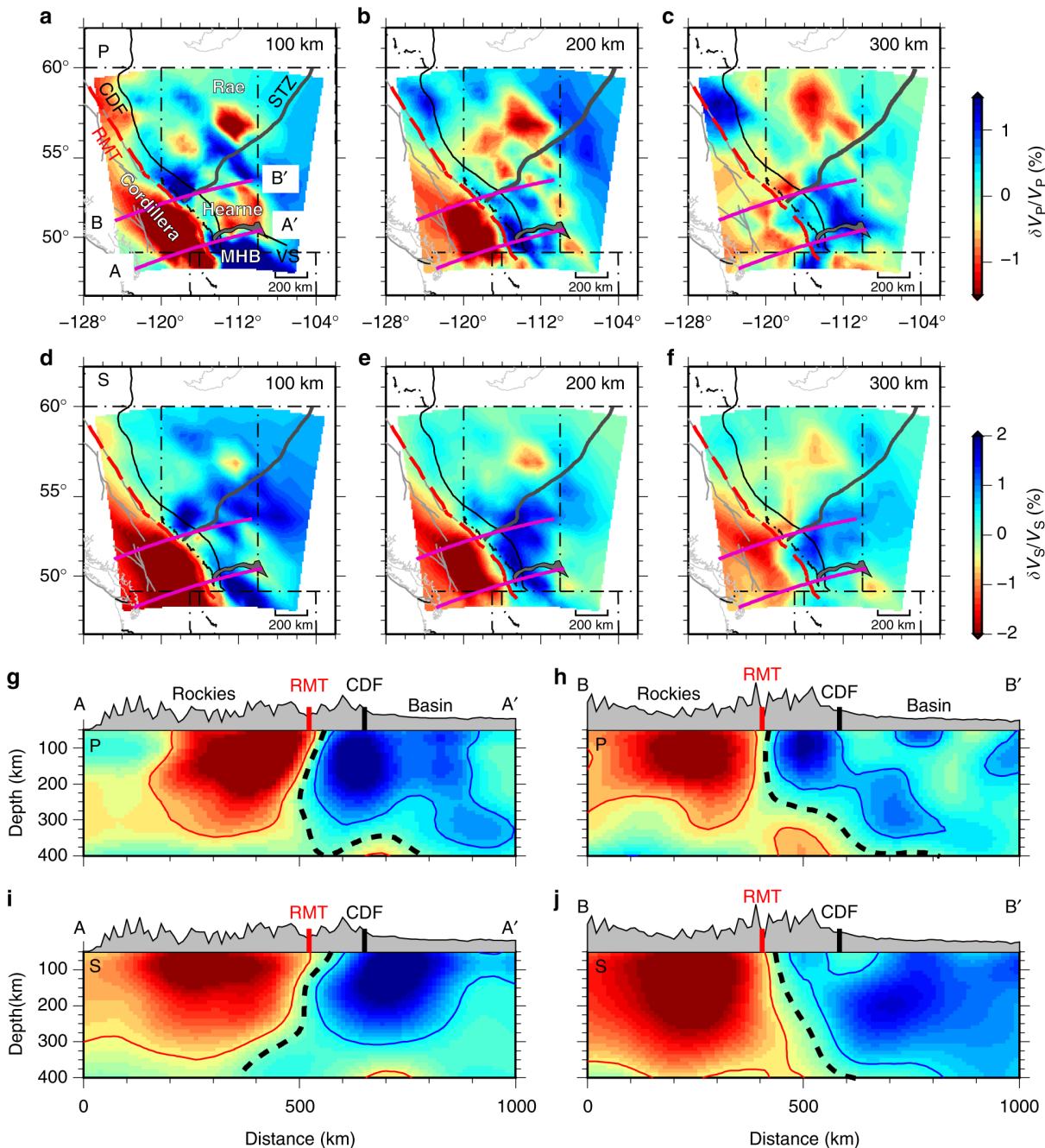


Figure 3.11. P-wave and S-wave models from finite-frequency tomography modelling. **a-c** depict P-wave velocities at depths of 100, 200 and 300 km respectively. **d-f** depict S-wave velocities at depths of 100, 200 and 300 km respectively. The purple lines show the locations of velocity profiles. **g-h** show the P-wave velocity anomalies across the two profiles. **i-j** show the S-wave velocity anomalies across the two profiles (Chen et al., 2019).

boundary changes with latitude beneath the Rocky Mountain Trench, changing from west-dipping to east-dipping, and then sub-vertical across a relatively short distance, approximately 50 km. North of this, the Cordillera-craton boundary fuses with the Rocky Mountain Trench-Tintina Fault system. Chen et al. (2019) interpret the findings from this model to be in favour of a collisional origin theory for the Canadian Cordillera.

A study conducted by Boyce et al. (2019) focuses on North America, in order to present a new P-wave tomography model of the North American continental mantle. This new model (Figure 3.12) depicts high velocities beneath the Superior Province, which continue east into the Grenville Province, at depths of approximately 100-150 km. However, at depths of about 200-250 km, this high velocity below the Grenville Province continues along the Grenville deformation front. A notable observation is the decreasing P-wavespeeds from North to South across the Grenville Province, which was interpreted to be as a result of metasomatism which altered the composition of the lithospheric mantle.

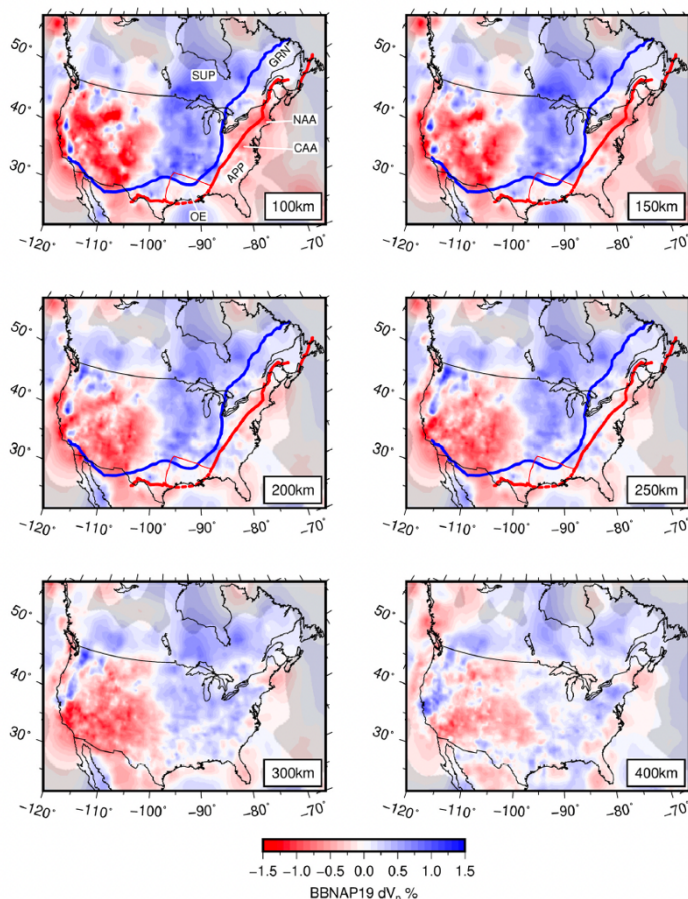


Figure 3.12. Tomography model for the upper mantle of North America (taken from Boyce et al., 2019).

A P-wave teleseismic traveltime tomography study conducted by Bollmann et al. (2019) provides a direct update of the Frederiksen et al. (2013) model. The work finds extensive high velocity anomalies across the study area, with the highest velocities occurring in the vicinity of the transition zone, in addition to notable low velocity anomalies in the vicinity of the MCR. These low velocity anomalies are interpreted to be related to rift-related material underplating the crust, though another possible contributor to this anomaly is the enriched, hydrated and/or warm subcrustal mantle lithosphere below the MCR. Two distinct high velocity anomalies are noted in the mid-mantle which were interpreted as the fragments of the subducted Farallon and Kula plates.

3.3 Shear-wave Splitting.

Shear-wave splitting allows for unambiguous observations of anisotropy, as it is effectively diagnostic of anisotropy and can be detected on a single seismogram. Shear-wave splitting is the phenomenon by which quasi-S waves in anisotropic materials have two possible polarizations with different velocities, which will be separated in time as the wave travels (Stein and Wysession, 2003). This phenomenon of shear-wave splitting has been detected in both the crust and upper mantle. A good example of shear-wave splitting can be observed in teleseismic SKS arrivals, which are quite useful as they have a known linear polarization as they exit the core. Shear-wave splits provide two numbers – one is indicative of the fast qS polarization (the quasi-S wave solution to the anisotropic eigenvalue equation), while the other denotes the time delay between the pulses – which respectively provide information on the directionality/orientation and the strength of anisotropy (Stein and Wysession, 2003). There is a trade-off however, between the thickness of the anisotropic layer and the strength, as a very

anisotropic, thin layer will generate a similar time delay as a weakly anisotropic, thick layer. The fast axis of the shear-wave splits usually provides a good proxy for the detected 2-psi anisotropy recovered from tomographic inversion, allowing for insightful comparisons, with the caveat of depth resolution. The SKS splits are an integrated observation of everything from the core-mantle boundary to the surface, whereas the 2-psi surface wave tomography measurements are associated with a certain depth range, as a function of period. Shear-wave splitting is usually assumed to be dominated by anisotropy above the 410 km discontinuity, as olivine is lost beyond this discontinuity. A database for shear-wave splits (Barruol et al., 2009) will be used for comparative purposes for the results obtained.

3.4 Receiver Function Observations.

Receiver function analysis has also been quite essential to constraining lithospheric properties. Abt et al. (2010) examine the North American lithospheric discontinuity with the use of receiver function analysis. The work finds that there is a negative velocity contrast at the lithosphere-asthenosphere boundary (LAB), as well as a lack of a sharp LAB beneath lithosphere of >130 km thickness. There also appears to be a negative velocity contrast within the lithosphere. Additionally, below western North America, the observed asthenospheric velocity is very seismically slow due to its actively deforming nature.

Audet et al. (2019) examine the northern Canadian Cordillera (NCC) using receiver functions to analyze the lithospheric structure. The work found that the LAB exists at a depth of approximately 50 km. The region also exhibited a thinned lithosphere, which was interpreted to be the result of decompression melting and cooling. Below the eastern NCC, there are

structures that dip to the west, which is not consistent with an ancestral North American mantle lithosphere underlying the region of the NCC to the west of the Tintina fault.

Miller and Eaton (2010) examine a profile from Vancouver Island to Baffin Island using S receiver functions to investigate the lithospheric framework. The results show a relatively shallow lithosphere-asthenosphere boundary below British Columbia at approximately 85 km, which increases sharply at the Cordillera-craton boundary to approximately 260 km, and maintains that estimated depth across the Canadian Shield. The results also provide supporting evidence for the formation of the cratonic lithosphere via arc accretion, due to the presence of mid-lithospheric structures which continue underneath numerous extensive magmatic arcs (at an approximate age of 1.8 Ga).

3.5 Combination Method Observations.

Yuan and Romanowicz (2010) examine the lithospheric layering across the North American craton, using a joint inversion method of both long period seismic waveforms and SKS splitting measurements, building on previous receiver function studies which found a structural boundary below continental cratons at depths that would be too shallow to correspond to the lithosphere-asthenosphere boundary. This study finds variations in the azimuthal anisotropy with depth which is indicative of two defined lithospheric layers continuous along the stable part of the craton. The upper layer is approximately 150 km thick, and thins toward the surrounding Palaeozoic regions. The lithosphere-asthenosphere boundary was found to occur at depths of 180-240 km, coinciding with the thermal conductive root. This study suggests that

the aforementioned relatively shallow structural boundary is consistent with a sharp mid-lithospheric boundary.

A 2011 study by Yuan et al. models the azimuthal anisotropic velocity of the North American mantle, using both long period seismic waveforms and splitting measurements. The model corroborated the findings of the Yuan and Romanowicz (2010) study which detected two layers of continental craton lithosphere due to variations in the orientation of the fast axis of azimuthal anisotropy. At sublithospheric depths below the craton, the anisotropy is interpreted to align with absolute plate motion.

A further 2014 study by Yuan et al. focuses on central and eastern North American using full waveform tomography, modelling both the lithosphere and asthenosphere, aiming to improve upon the resolution provided by the previous long-period waveform modelling (Yuan et al., 2011). This new model depicts a uniformly thick lithosphere for the North American craton, extending to depths of 200-250 km. In this region, the notable Proterozoic tectonic sutures appearing in the surface geology continue to depths of 100-150 km, manifesting as narrow bands of isolated radial anisotropy. The entirety of the upper mantle below the craton exhibits low velocity at depths ranging from 200-300 km. The study put forward the suggestion that repeated tectonism such as rifting episodes that appear at the surface also leave their mark on the continental lithosphere mantle, in the form of isotropic and anisotropic velocity anomalies that can be observed at depths of up to 150 km.

3.6 Magnetotellurics of the Canadian Lithosphere.

Past magnetotelluric (MT) studies of the sub-regions of Canada have confirmed that MT data play a significant role in the understanding of the tectonic history and geology of Canada. For the past 35 years, high-quality broadband MT measurements have been recorded at over 6000 sites throughout Canada at both lithospheric and asthenospheric depths, through projects such as Lithoprobe, as well as other programs such as the Federal Geothermal Energy Programme (GEP) (Jones et al., 2014). A summary of the major observations for Canada and its respective sub-regions is presented here.

3.6.1 Canada as a whole

Jones et al. (2014) studied the electrical resistivity of the lithosphere of Canada using data obtained from the Lithoprobe project and others. Interpretations of resistivity values in terms of the state of the mantle lithosphere indicate that the Cordillera is relatively warm, in most areas, with the exception of southwest British Columbia, where it is much colder in comparison, likely due to the subduction associated with the Juan de Fuca tectonic plate. Within the Cordillera, it is largely dry, i.e. the peridotite present contains very little water. However, this phenomenon does not extend to beneath the Prairies or the Wopmay Orogen, where there is the presence of water is implied. Furthermore, at a depth of 100 km the Canadian Shield was found to be cold and dry, while at depths of 200 km, the majority of Canada is dry (Jones et al., 2014).

3.6.2 Superior Province

The Archean lithosphere of the western Superior Province was examined by Ferguson et al. (2005) using MT data, as part of the Lithoprobe project. As previously indicated (*Section 2.1*),

the Western Superior features a 100-200 km expanse of E-W trending granite-greenstone and metasedimentary belts. The results of this study overwhelmingly point to the conclusion that the western Superior lithospheric mantle has remained unchanged since the accretion of subprovinces circa 2.7 Ga. The North Caribou terrane, contained within the Superior Province, is approximately 3.0 Ga in age. The southern boundary of the terrane exhibits a conductivity anomaly that continues for 600 km in the E-W direction. This anomaly was thought to be the result of the approximately 2.7 Ga metasedimentary rocks found in the English River subprovince, specifically a conductive component contained within. The results of this study found that this North Caribou terrane consists of very resistive crust (Ferguson et al., 2005).

Work done by Hill et al. (2021) investigated the resistivity framework of the central Superior Province at crustal depths, relating to the Archean craton, through the use of 3D MT resistivity models. The findings of their work include low resistivity anomalies in the mid to lower crust, which is indicative of pathways for historic fluid flow which emerged as a result of terrane accretions. Extensive flow in the lower crust was brought about by later orogenic collapse, and allowed for magma refertilization. It is thought that these zones of low resistivity were conserved through time due to the decoupling of the upper crust from the mid-lower crust during the final stages of deformation (Hill et al., 2021). The thickness of the electrical lithosphere was estimated to be 250-270 km below the Superior craton (Hammer et al., 2011).

3.6.3 Trans-Hudson Orogen

MT studies conducted over the past 30 years characterise the Trans-Hudson orogen by a sizeable high-conductivity zone below the North American Central Plains (NACP) (e.g., Jones and Savage, 1986; Jones et al., 1993; Ferguson et al., 2005), through the use of 2-D models. A study conducted by Jones et al. (2005) takes a closer look at the NACP conductivity anomaly.

The anomaly is largely associated with metamorphosed sulphides in areas where there is bedrock exposure, which occurred in response to subduction (and consequently, compression), and were positioned at crustal depth along the boundary of the Hearne. This study found the presence of a conductivity anomaly occurring in the upper mantle, originating at a depth of approximately 80-100 km. The thickness of the electrical lithosphere beneath the THO was found to be approximately 200 km (Hammer et al., 2011). The electrical properties of this lithospheric mantle anomaly bear great similarity to that seen for the central Slave craton anomaly. The properties of the Slave craton mantle anomaly are linked primarily to the diamond-rich Lac de Gras kimberlite located directly below, though this does not perfectly translate for the new NACP mantle anomaly; the Fort à la Corne kimberlite (related to the Sask craton), although located within the vicinity of the anomaly, does not exactly sit below it.

3.6.4 Rae and Hearne

The Rae and Hearne cratons form part of the Churchill Province within the Canadian Shield. An MT study conducted by Spratt et al. (2014) examines the deep lithosphere below the central Rae craton, via 2-D resistivity models, in order to obtain insights into its structure and composition. The models show a conductive zone just below 35 km depth within the crust, that does not seem to be continuous at or within the vicinity of the main geological boundaries. The conductive layer is interpreted to consist of Archean-aged metasedimentary material. The resistivity values obtained for the upper mantle indicate that it becomes more resistive across the Rae craton from north to south, ranging from $\sim 3000 \Omega\text{m}$ in the northeast (below Baffin Island) to $> 50,000 \Omega\text{m}$ in the south (closer to the Hearne). At depths of 180-220 km, there is a consistent drop in resistivity to $\sim 500 \Omega\text{m}$ which was thought to be due to the presence of the lithosphere-asthenosphere boundary (Spratt et al., 2014).

Work conducted by Jones et al. (2002) studied the Snowbird Tectonic Zone (STZ) within the Western Churchill Province of the Canadian Shield. The Archean-aged STZ is approximately 2000 km in length, and is characterised by Hoffman (1988) as a suture reactivated during the Palaeoproterozoic, which divides the Rae and Hearne domains. The subsurface model obtained indicates that the STZ had an important role in the transformation of the western Canadian Shield. The model further reveals the existence of neighbouring mantle zones with strikingly different geophysical characteristics, which demonstrates that the Hearne domain overlies the leading edge of the Rae domain (Jones et al., 2002).

3.6.5 Slave Craton

The Slave craton is located in the northwestern region of the Canadian Shield, which hosts some of the Earth's oldest dated rocks - the Acasta gneisses - aged at 4.027 Ga (Stern and Bleeker, 1998). An MT study conducted by Jones et al. (2001) investigated the electrical properties of the Slave craton. The findings show a significant conductivity anomaly at 80-100 km depth, located in the vicinity of the Lac de Gras kimberlite field, which is thought to be in response to dissolved carbon (or hydrogen). The base of the electrical lithosphere across the Slave was found to occur at approximately 260 km depth (Jones et al., 2003). The upper mantle of the Slave craton exhibits highly anomalous geophysical characteristics and shows spatial correspondence to a >3.0 Ga ultra-depleted harzburgite-rich layer, underlain by a mildly depleted Iherzolite layer. The anomaly is interpreted to be related to ancient tectonic processes (subduction and consequently, trapping), rather than associated with the Eocene-aged kimberlite phenomenon (Jones et al., 2001).

A further study examining the electrical properties of the Slave craton was conducted by Jones et al. in 2003. The study focuses on the same conductivity anomaly discussed above, now

named the Central Slave Mantle Conductor (CSMC), which was found to have resistivities between 10 and 15 Ωm , starting at a depth range of 80-120 km (Figure 3.13). The NE-SW striking CSMC is fully encompassed within the NE-SW striking central zone, as described by Grütter et al. (1999). This study confirms the findings of Jones et al. (2001), which suggested that the CSMC anomaly is due to the presence of carbon, though this 2003 study qualifies, carbon in its graphite state, or “carbon on grain-boundary films” (Jones et al., 2003).

3.6.6 Grenville Orogen

Boerner et al. (2000) review the EM response of the Abitibi-Grenville Lithoprobe transect. The findings have shown numerous conductivity anomalies that originate from Precambrian tectonism. Additionally, the results indicate the presence of a homogeneous layer of high conductivity within the mid-crust of the Grenville Front. This was determined to be due to present properties of the lower crust (e.g. pressure, temperature and fluid saturation). Electrical

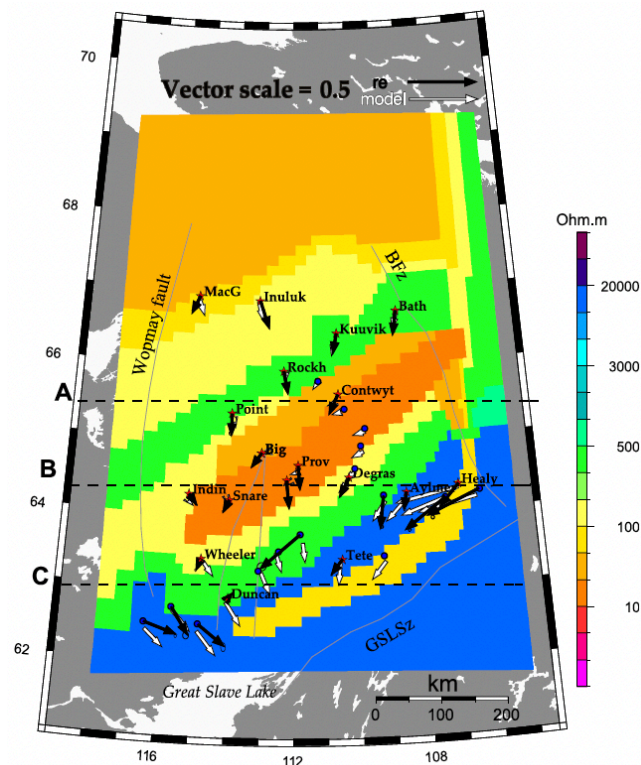


Figure 3.13. Horizontal slice at 140 km through 3D resistivity model (Jones et al., 2003).

anisotropy is noted across the transect, though it was unclear if this finding was representative of the lower crust or upper mantle. Despite this, the results linked to electrical anisotropy suggest that the lithosphere has been intact since the Neoarchean. The electrical lithosphere extends to a depth of 150 km. Seismic tomography results suggest the lithosphere is highly rigid, which could indicate that the source of the anisotropy is ancient.

Work conducted by Adetunji et al. (2015) examines the large-scale resistivity structure of the lithosphere under the Proterozoic Grenville Province, via MT profiles. The results show the presence of a considerable lithospheric conductor within the Grenville Province, below the Central Metasedimentary Belt (CMB), between 100 and 150 km depth. The findings suggest that the conductivity anomaly is likely due to the presence of grain boundary graphite films, though the source of this carbon is unclear. It is speculated to be linked to the Great Meteor hotspot, or alternatively, associated with tectonic processes that are pre-Grenvillian in age. The results also indicate that the entirety of the Grenville Province in southern Ontario features pervasively resistive lithosphere. The resistivity anomaly was determined to be a segment of the Superior lithosphere – Archean-aged root – which is encompassed within the Grenville orogen. This study finds that the electrical lithosphere is layered in a manner as described by Yuan and Romanowicz (2010), with the occurrence of a mid-lithospheric discontinuity (MLD), which puts the thickness of the electrical lithosphere at approximately 180-240 km.

3.6.7 Wopmay Orogen

Wu et al. (2005) investigated the geoelectric response of the Proterozoic Wopmay Orogen, as well as its adjacent terranes. The results show a conductor at the Fort Simpson and Hottah terrane boundary, which is attributed to the collision of the involved terranes. Conductive crust is present below the western edge of the Great Bear magmatic arc. The model also indicates

conductive mantle beneath the Hottah terrane which is interpreted to be the result of graphite-rich rocks that were emplaced during subduction. The base of the electrical lithosphere is mainly located at approximately 100 km depth, deepening to approximately 120 km in the northwest (Wu et al., 2005).

3.6.8 Cordillera

3.6.8.1 Northern Cordillera

A study by Habibian Dehkordi et al. (2019) re-analysed the Lithoprobe Slave-Northern Cordillera Lithospheric Evolution (SNORCLE) Corridor 2 MT data, using updated tensor decomposition methods to investigate the northern Cordillera tectonic framework. The results of this work (Figure 3.14) find a significant east-dipping conductivity zone within the vicinity of Stikinia. The results also show a metasedimentary wedge of Ancestral North American rocks within the mid to lower crust, which is overlain by numerous relatively small high-conductivity zones which can be explained by the deformation and metamorphic activity associated with the thrust faulting during the Mesozoic, as well as conductors within the wedge (east of the Tintina Fault), that also dip to the east. Hammer et al. (2011) finds that the electrical lithospheric thickness below the northern Cordillera is quite thin, approximately 55-70 km.

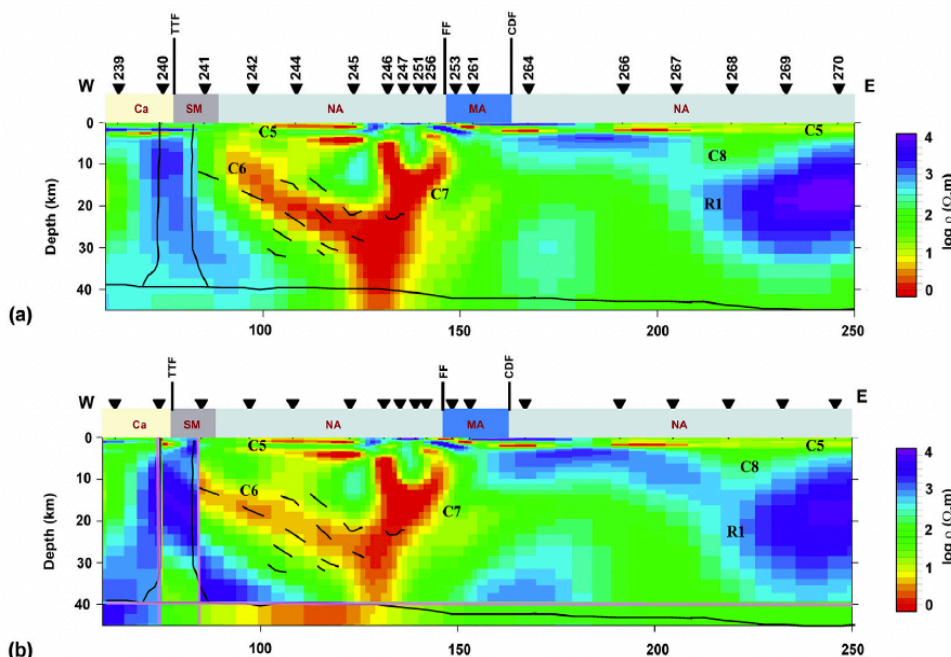


Figure 3.14. Inversion models for Corridor 2b. (a) Unconstrained model & (b) Model with regularization tears (i.e. where the conductivity contrasts across a postulated boundary and does not contribute to the overall roughness) at Moho and Tintina Fault Zone margins (modified from Habibian Dehkordi et al., 2019). Abbreviations: TTF: Tintina Fault, FF: Forcier Fault, CDF: Cordillera Deformation Front.

3.6.8.2 *Southern Cordillera*

Ledo and Jones (2001) examined and interpreted the resistivity of the southern Cordillera using MT data collected along several profiles, with attention given to the Intermontane and Omineca Belts. The model obtained shows an overall conductive lower crust underlying a relatively resistive upper crust. The conductivity of the Intermontane Belt lower crust remains consistent across all profiles, while the conductivity of the lower crust of the Omineca Belt is highly variable, differing by up to two orders of magnitude along strike, with higher conductivities in the south, and lower conductivities in the northern unextended region of the belt. The lower crustal conductivity is attributed to the presence of fluids, specifically brines, but there may also be an interplay with partial melts in the lowermost crust in certain areas, such as the southern Omineca Belt region.

A study conducted by Rippe et al. (2013) examines the fluid content within the upper mantle of the southern Cordillera with the use of long-period MT data along two profiles. The model shows, in the forearc, a conductive zone that dips to the east, as well as a conductive mantle wedge, thought to exist due to the fluid emissions from the subduction of the Juan de Fuca tectonic plate. The model is also indicative of a conductive region within the upper mantle of the back arc, at a depth of <60 km which is interpreted as a shallow lithosphere-asthenosphere boundary. The conductivity here is interpreted to be the result of the presence of aqueous fluid, and to a lesser extent partial melts.

3.7 Commonalities observed for the North American Lithosphere.

A summary of the common major observations for the lithosphere (and sublithospheric mantle as applicable) of the various sub-regions of Canada is presented here.

3.7.1 Cordillera-craton Boundary.

The Cordilleran upper mantle exhibits low seismic velocities (e.g., Frederiksen et al., 2001; van der Lee and Frederiksen, 2005; Schaeffer and Lebedev, 2014; Zaporozan et al., 2018), while the cratonic lithosphere exhibits high velocities (e.g., Frederiksen et al., 2001; Chen et al., 2007; Zaporozan et al., 2018; McLellan et al., 2018). A change in dip in the mantle contact is observed. The dip direction changes from northeast at the north to southwest, moving south along the Cordillera-craton boundary (Zaporozan et al., 2018; Chen et al., 2019). Furthermore, observations of the lithospheric Cordillera-craton boundary indicate that the boundary extends to the Tintina Fault system (McLellan et al., 2018; Chen et al. 2019). The sharpness of the Cordillera-craton boundary is quite strong, specifically in the western Cordillera, transitioning from slow to fast velocity over a distance of approximately 150 km (McLellan et al., 2018, Bedle and van der Lee, 2009). The thickness of the seismic and electrical lithosphere below the Cordillera compares quite well, as both seem to indicate that the lithosphere is quite thin at approximately 60 km (Zaporozan et al., 2018; Hammer et al., 2011).

3.7.2 The Superior Craton.

Many studies have confirmed that the Superior Craton exhibits fast seismic velocities, compared to the surrounding areas (e.g., Frederiksen et al., 2001; Frederiksen et al., 2013; Boyce et al., 2019; Darbyshire and Lebedev, 2009; Petrescu et al., 2017; Schaeffer and Lebedev, 2014; Foster et al., 2020). All of the Superior is fast, though there are internal

variations; the western Superior in Canada is particularly fast (Frederiksen et al., 2007; Frederiksen et al., 2013). The lithospheric thickness across the craton was constrained by Darbyshire et al. (2007) to approximately 140 km and 200 km in central and western Superior respectively, with the thickest lithosphere occurring in northeastern Ontario (220-240 km). This contrasts slightly with the thicker electrical lithosphere of approximately 250-270 km (Hammer et al., 2011).

3.7.3 Hudson Bay.

Lithospheric thicknesses below the Hudson Bay region were found to be between 185 km in the northeast and about 260 km in the central-southern area (Darbyshire and Eaton, 2010; Darbyshire et al., 2013). This compares well with the thickness of the electrical lithosphere, which was found to be approximately 200 km (Hammer et al., 2011). Within the upper mantle, below Central Hudson Bay at the intracratonic basin, the presence of a relatively low velocity zone was detected, surrounded by Archean aged regions exhibiting high seismic velocities (Darbyshire et al., 2013; Liddell et al., 2018; Foster et al., 2020).

3.7.4 Farallon Plate Fragment.

A distinctive high velocity anomaly is imaged near the transition zone in the expected location of the subducted fragments of the Farallon plate (van der Lee and Nolet, 1997; Schaeffer and Lebedev, 2013; Frederiksen et al., 2013). Bedle and van der Lee (2009) also find supporting evidence for the Farallon plate fragment, though the associated high velocity anomaly was found to be wider. The Schaeffer and Lebedev (2013) study found that these fragments appear to extend eastward beneath the continental US, extending toward the Great Lakes.

Overall, many advancements have been made over the years, in our understanding of the North American lithosphere and mantle, and it is likely that there is still much yet to be refined, as methods and techniques evolve. This work seeks to improve upon the currently most cited model (Schaeffer and Lebedev, 2014), and further, to add to the understanding of lithosphere in the Canadian context.

4. DATA AND METHODOLOGY.

4.1 Instrumentation.

The aim of this work is to obtain the best possible two-station coverage of Canada. The Geological Survey of Canada manages a network of seismographs (high-gain instruments) and accelerographs (low-gain instruments) across Canada, which comprise the Canadian National Seismic Network (CNSN) (Bent et al., 2019). In western and northern Canada, there is additional station coverage, due to the Portable Observatories for Lithospheric Analysis and Research Investigating Seismicity (POLARIS) (Eaton et al., 2005), which was initially a temporary project that has since concluded, though some of the stations were made permanent. Furthermore, the Earthscope USArray Transportable Array (TA) (Williams et al., 2010) stations (specifically those located in areas close to the USA-Canada border) also provided increased coverage.

For the purposes of this project, existing raypaths (Petrescu et al., 2017; McLellan et al., 2018; Zaporozan et al., 2018; Foster et al., 2020; Bryksin (pers. comm.)) using stations from the above listed networks were first collated. Additional paths were added to this dataset, using stations across Canada, to increase the raypath coverage in sparsely covered regions, primarily the Prairie provinces and the southern Cordillera, using the CNSN network. Importantly, the number of active stations across Canada was not constant, as some of the stations were from temporary deployments.

4.2 Data Acquisition and Pre-processing.

The Incorporated Research Institutions for Seismology (IRIS) catalogue was used to retrieve data for relevant events. Two criteria were used to ensure that the events selected showed a strong signal to noise ratio.

1. Relevant events were filtered for earthquakes with a minimum magnitude of 6, to ensure that the emitted energy from the earthquake was adequate for long-distance analysis.
2. These compiled events were further filtered for station-pair paths lying within 5° of the same great-circle path as the source (Figure 4.1). This requirement is based on the assumption that the waveform changes which occur between the two stations being considered are dictated by the subsurface structure along the great-circle path between the two stations. Consequently, the larger the angle between the stations, the less this assumption holds.

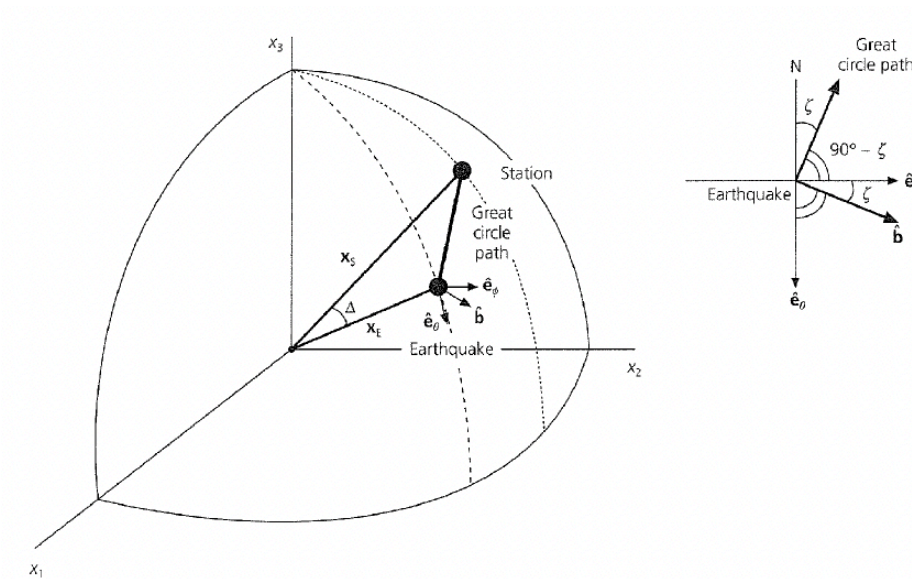


Figure 4.1. Schematic of the relation of the great circle path between the source (earthquake epicenter) and seismic station (*left*), and the convention for the azimuth, ζ (*right*) (Stein and Wysession, 2003)

This data retrieval process (from IRIS) is fully automated using Andrew Schaeffer's 2-Station Phase and Group Velocity Measurement Software (2StnPvPy) (Schaeffer, pers. comms.) based on the teleseismic interferometry method of Meier et al. (2004), translated to Python. The procedure involves an initial query detailing a station pair (or multiple station pairs) within an indicated network (or networks), which produces an output pickle file, which is then used to source relevant events between the station pair/s from a selected start-time (usually 1993-10-01, to recover data from seismic stations from the earliest date of deployment), with a minimum magnitude of 6. All the events meeting these requirements are collected, and are automatically filtered for station-pair paths existing within 5° of the same great-circle trace as the source. Only the events that fit both criteria are compiled for further processing.

4.3 Dispersion curve measurement.

In order to calculate the phase velocity between both seismic stations, it is necessary to identify the same peaks in both of the recorded measurements. This is somewhat difficult due to the nature of seismic waves, as phase can differ by a factor of $2n\pi$ radians (n is an integer). This factor can change the measured velocity to a significant degree. To combat this issue, the waveform is compared to the fundamental mode phase velocity at low frequency, where 2π radians manifests as a major, unrealistic velocity change. A smooth dispersion curve is then assumed. This measurement must be performed for all events selected for a given station-to-station path. Per path, the number of phase velocity curves measured changes based on several factors including but not limited to, earthquake location, observation time and instrumentation malfunction during data recording.

Full execution of the previous step launches an interactive window to facilitate dispersion curve picking. Within this interactive window, the waveforms of the two stations are shown in separate plots, as well as an additional plot of the cross-correlation (Figure 4.2). The frequency-phase velocity plot is the interactive window, where the curve picking of a frequency window for each curve actively occurs. The black dashed line illustrates the reference curve based on global values (Figure 4.2). Based on how well the curve representing a singular event matches the reference curve, the option to skip curve picking for the current event is available. Below this plot, there are 2 separate plots that show the directionality of the event, i.e. if the ray travelled from station 1 to station 2 or vice-versa (Figure 4.2).

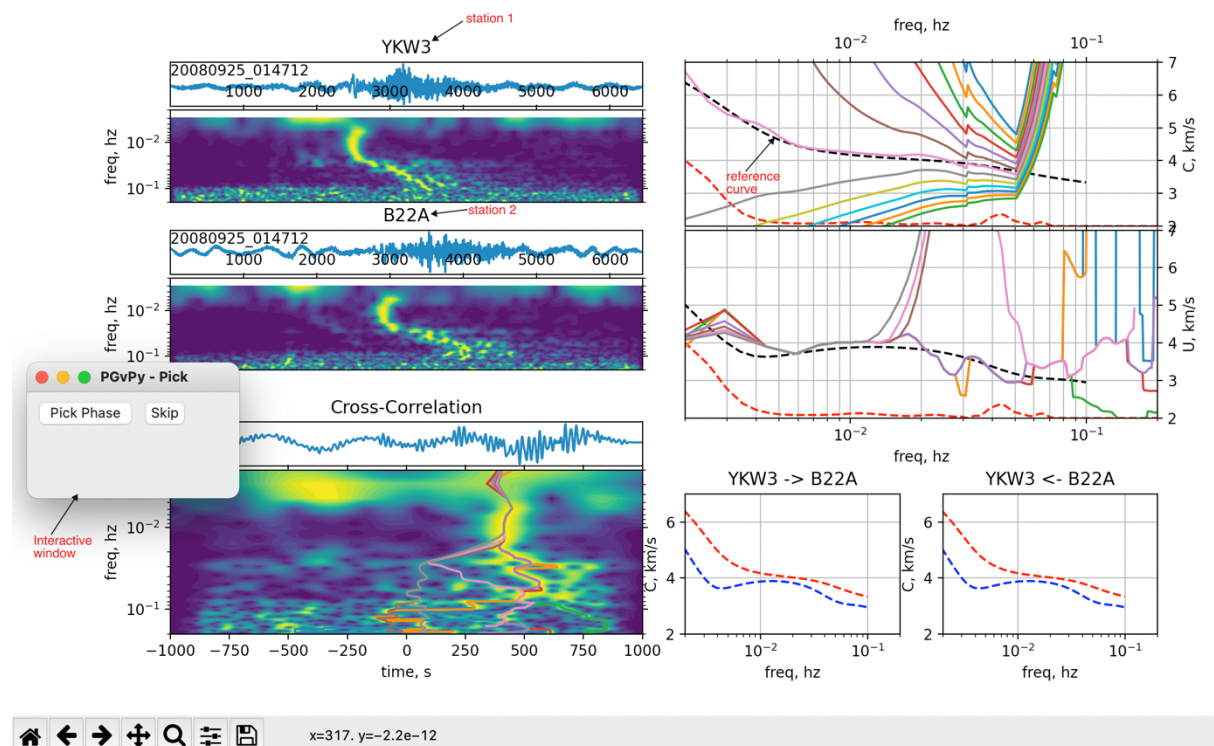


Figure 4.2. Interactive window allowing for curve picking. The dashed black line represents the reference model.

Once the curve picking is completed, the selected path segment is shown on the directional plots (which shows all previous picks for the station pair) (Figure 4.3). Before continuing to the next event, the software allows the user to reattempt or discard the pick. This curve-picking

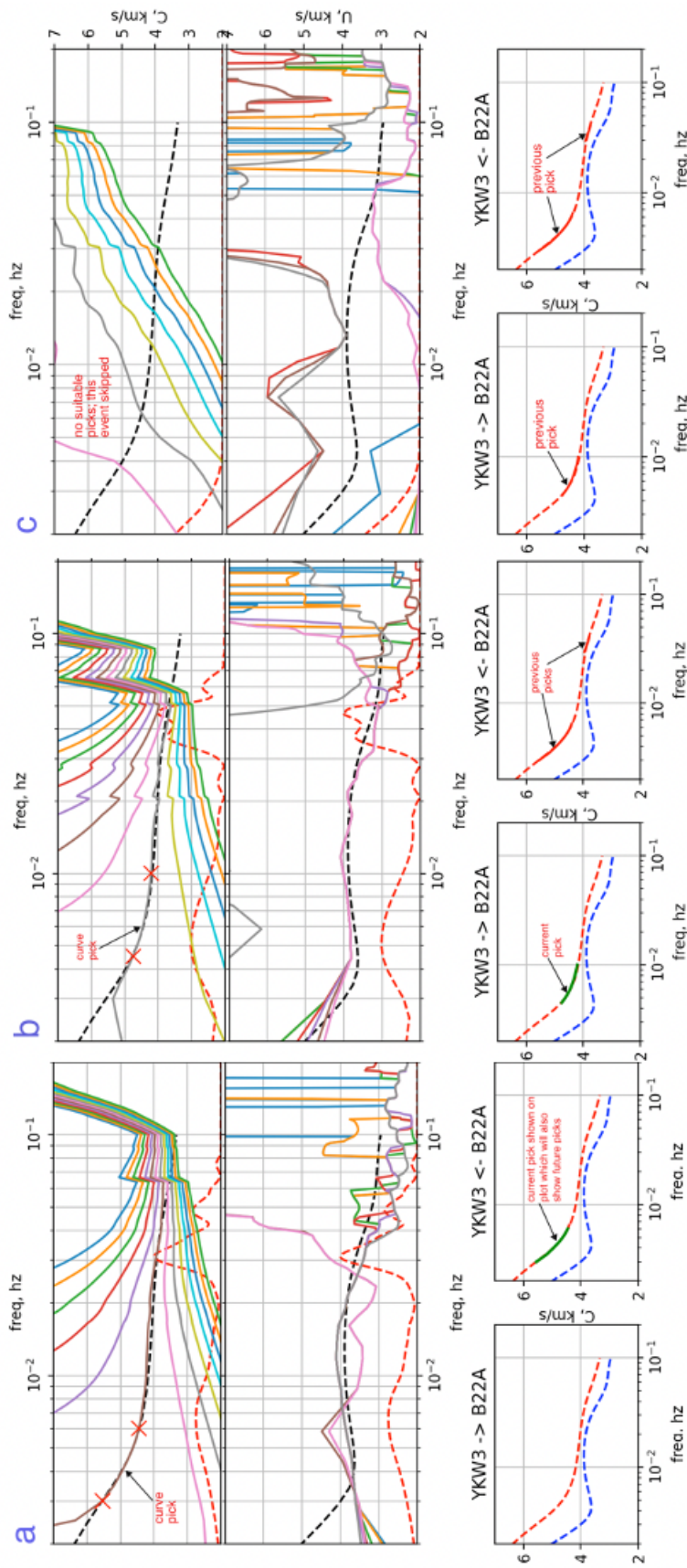


Figure 4.3. Sample of the curve picking process. (a) depicts the first pick between station pair YKW3 & B22A – which appears in green on the directionality plot (bottom) showing that the signal travelled from B22A to YKW3. (b) depicts curve picking of a later event. The current pick is highlighted in green along the directionality plots, while the previous picks are shown in red. (c) depicts an example of an event with no suitable picks. This event is skipped.

process is repeated until all relevant events for the station pair have been examined. The software automatically calculates the measurement error for each curve, which is a measure of the standard deviation of the picks for each curve. This measurement error is used in the outlier analysis process (*sub-section 4.5.2*). The selected data (from this curve-picking process) are collated in a .dat file, which is then further examined for the purposes of quality control.

4.4 Quality Control.

Once all the events for a particular station pair have been examined and the relevant curves selected, it is necessary to ensure that they are all both reasonable and coherent. This process involves selecting the generated .dat file (from the previous step) and examining all of the picked curves. The software prompts the user to refine the curve, by omitting certain segments and keeping others, through user selections, or to delete the curve entirely based on how much the curve deviates from the average. The aim is to maximise the amount of useable data, i.e. deleting as little as possible, while also maintaining consistency. The verified curves and curve segments are then averaged to produce a dispersion curve for the station pair (Figure 4.4), and this data is output to a new .dat file. This resulting dispersion curve is then added to the catalogue of collated existing raypaths, for use in the production of phase velocity maps. This quality control process was repeated for every station pair being considered, and each resulting dispersion curve was added to the collated catalogue.

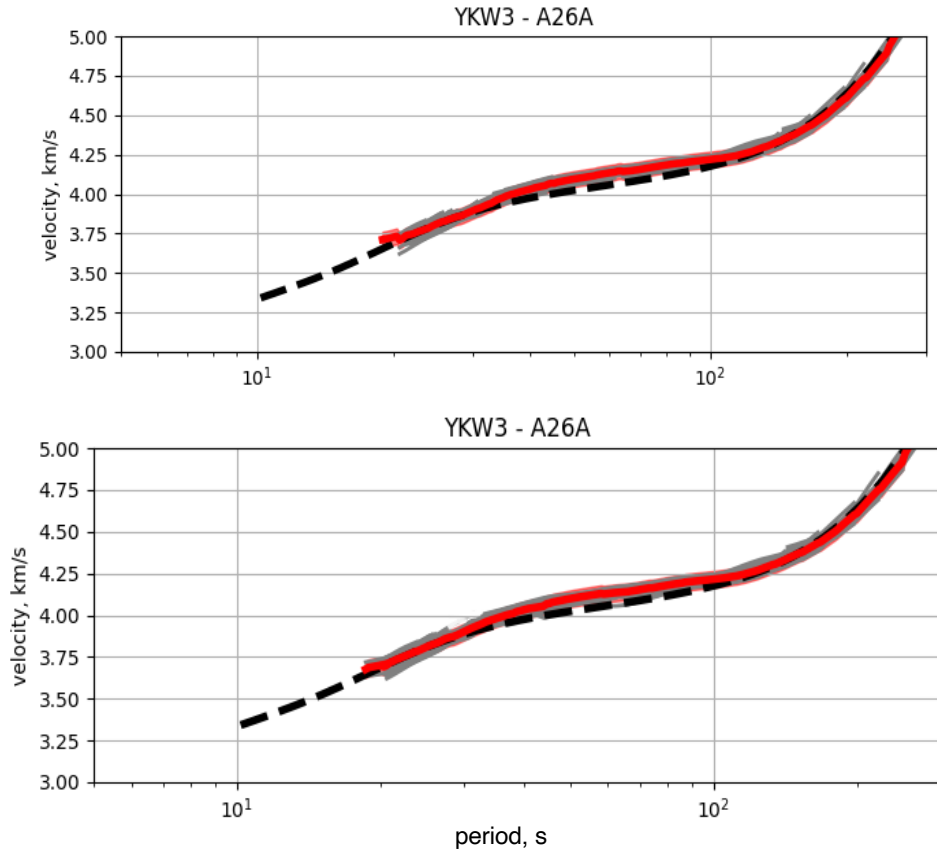


Figure 4.4. The plots depict a dispersion curve for a sample station pair (YKW3 and A26A). The first plot (top) depicts the average curve (red) before quality control, and the second plot (bottom) shows the refined average curve (red). The bottom plot shows that the kink seen in the top plot is largely smoothed.

4.5 Inversion and Phase Velocity Maps.

4.5.1 Inversion Code

At this stage, all curves being considered have been quality controlled. In order to perform the inversion process, the tomography code of Sergei Lebedev (Darbyshire and Lebedev, 2009) was used. Inversion was carried out for isotropy, 2ψ azimuthal anisotropy and 4ψ azimuthal anisotropy, at seventeen discrete periods between 15 and 400 seconds. In a smoothly varying medium, surface wave anisotropy is usually represented as small perturbations from the

isotropic velocity given by a series of π -periodic and $\pi/2$ -periodic terms (Smith and Dahlen, 1973) as:

$$C(\omega, \psi, \theta, \phi) = A_0(\omega, \theta, \phi) + A_1(\omega, \theta, \phi) \cos 2\psi + A_2(\omega, \theta, \phi) \sin 2\psi + A_3(\omega, \theta, \phi) \cos 4\psi + A_4(\omega, \theta, \phi) \sin 4\psi,$$

Equation 4.1. Surface wave anisotropy equation as defined by Smith and Dahlen (1973).

where $C(\omega, \psi, \theta, \phi)$ denotes the phase velocity at the azimuth ψ (evaluated with respect to local meridian), ω is the angular frequency, θ and ϕ indicate the longitude and latitude respectively, $A_0(\omega, \theta, \phi)$ refers to the isotropic-average phase velocity, and (A_1, A_2) and (A_3, A_4) refer to the coefficients of the π -periodic (2ψ anisotropy) and $\pi/2$ -periodic (4ψ anisotropy) azimuthal variations respectively.

For this procedure, rough value estimates for the smoothing and damping parameters were used for the initial run, which were later finalized (*Sub-section 5.5.3*). The inversion is linear, and uses ray theory for sensitivity analysis. For the tomographic model, a triangular model-grid with 200-km grid-node spacing and a 40 km integration-grid spacing was constructed, based on the station and path density. Once the inversion was successfully completed, initial phase velocity maps were produced using Generic Mapping Tools (GMT) (Wessel et al., 2019).

4.5.2 Outlier Exclusion Analysis

Outlier analysis was carried out to ensure there were no spurious paths that are impossible for the inversion scheme to fit. Such paths may have data issues, which would effectively bias the variance-reduction statistics. The outlier analysis procedure involved repeating the tomographic inversion with extremely low smoothing and damping values. The idea behind this is to produce a rough model that fits the data as well as possible. Phase velocity maps for

this non-regularized inversion were also plotted. The outlier exclusion process itself is an automated procedure which omits outlier measurements at each period, i.e. paths within the model with high misfits even with very little regularization. The threshold factor used to determine what constitutes an outlier, for the purposes of this work, is a proportion of the measurement error, i.e. paths with a misfit below the product of the threshold and the measurement error, are retained. The threshold value selected in this case was 0.95.

The script which automates this process takes as input the two previously processed datasets – the rough model inversion (the non-regularised inversion), as well as the initial, regularized inversion data. The script outputs a directory containing the accepted paths, and an updated input file for the inversion code, with the outlier exclusion process applied.

4.5.3 Regularization Testing.

Following the production of these initial maps and the outlier exclusion analysis, regularization tests were carried out to ascertain appropriate smoothing and damping coefficients. This procedure was also automated via a Python script (Bagherpur Mojaver, pers. comms.) that utilised Sergei Lebedev's tomography code.

Smoothing is the most widely used form of regularization for ray-based tomography, as it encourages long-wavelength structures, effectively biasing the model to simple structures which are easier to interpret (Parker, 1994). A smoothed model which exhibits regions of sharp changes indicates that the sharp change is required solely by the data. Smoothing tests were carried out at several different periods, to ensure consistency in the final values chosen. For the smoothing tests, all other input parameters were kept constant (e.g. the damping value). Additionally, for the isotropic smoothing tests, smoothing parameters for the anisotropic

components were fixed, and the reverse was true for the running of the anisotropic smoothing tests. The smoothing tests involved running the inversion for various smoothing values (37 discrete smoothing values in this case), which equated to 74 tests, as smoothing tests were done for both the isotropic and anisotropic components.

Damping tests were carried out for the same periods as those selected for the smoothing tests. In a similar fashion, damping tests were also performed for both the isotropic and anisotropic components, using another Python script (Bagherpur Mojaver, pers. comms.), that ran the inversion while keeping all other parameters fixed; isotropic damping values were fixed while the anisotropic tests were performed and vice-versa. Thirty-three different damping values were used for this testing process, equating to 66 tests, in aggregate.

Following the completion of the smoothing and damping tests (at various periods – 30s, 60s, 75s, 90s and 100s), a Python script (Bagherpur Mojaver, pers. comms.) was used to plot trade-off curves for both the smoothing (Figure 4.5) and damping (Figure 4.6) at each period. These periods were selected for regularization tests as they provided varying levels of path coverage, while maintaining relatively good resolution. Trade-off curves are used to determine at which point the regularization method being considered balances the model roughness and the misfit. This value usually occurs at the “knee” of the trade-off curve – the change in slope represents a transition from fitting the signal to fitting the noise. In order to ensure that suitable damping and smoothing parameters were chosen, phase velocity maps were plotted for the various damping and smoothing values at which testing occurred. Generated maps that appeared either too rough, or overly smoothed and damped, immediately caused certain regularization values to be discarded. Finalization of the smoothing and damping values was done using the trade-off curves, in conjunction with the produced phase velocity maps.

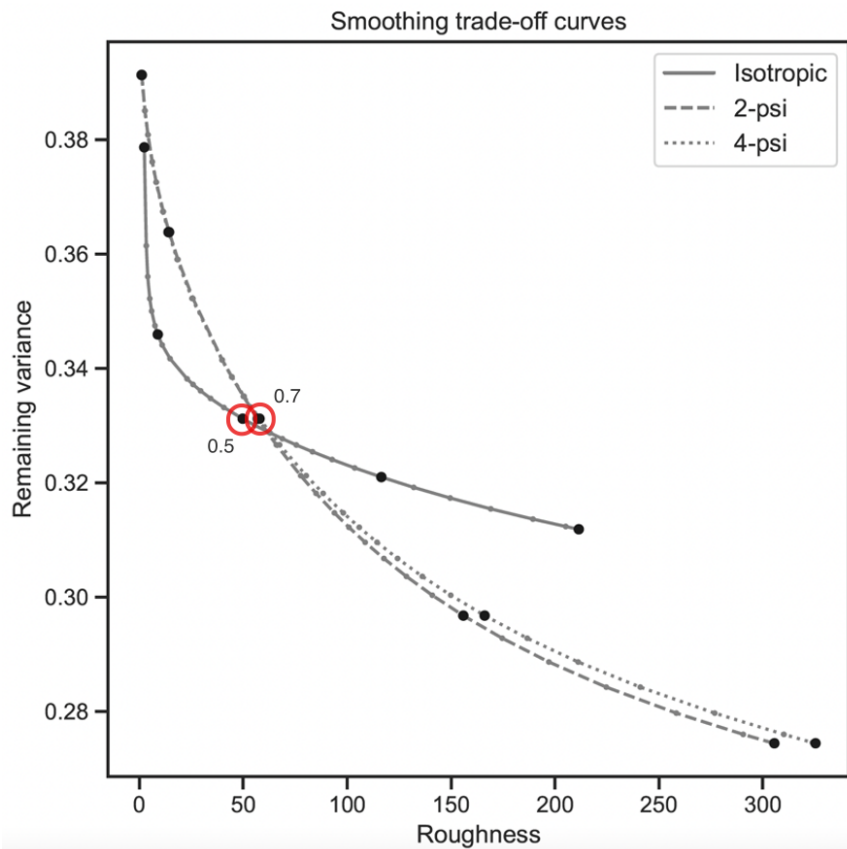


Figure 4.5. Smoothing trade-off curves for isotropic, 2ψ and 4ψ components at 90s period.

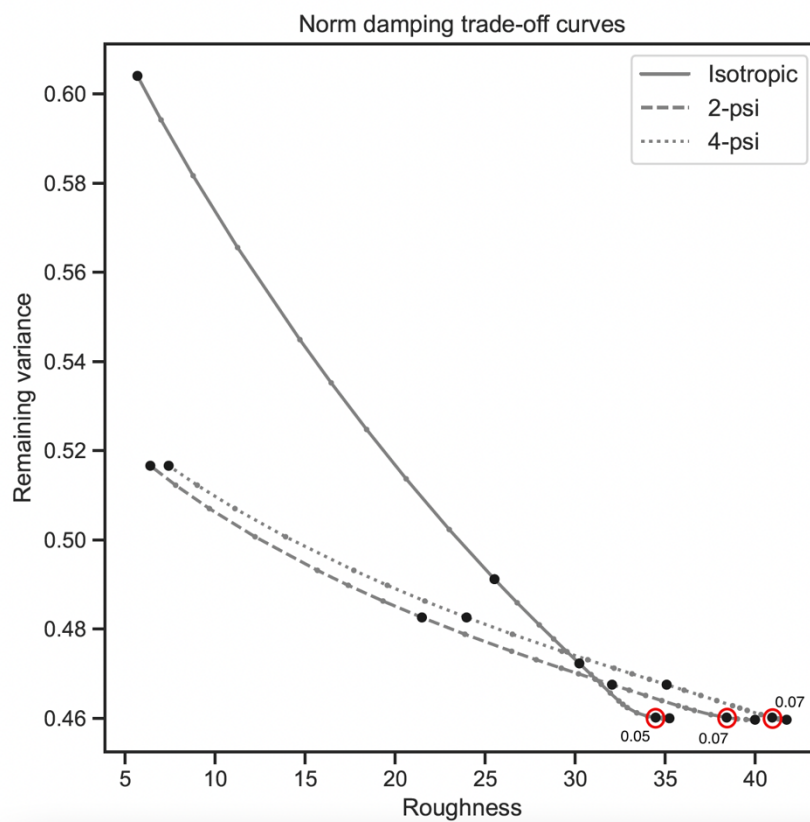


Figure 4.6. Damping trade-off curves for isotropic, 2ψ and 4ψ components at 90 s period.

The regularization parameters chosen for smoothing were 0.5 for isotropy and 0.7 for both 2ψ and 4ψ anisotropy, while for damping 0.05 was chosen for isotropy and 0.07 was chosen for both 2ψ and 4ψ anisotropy. The inversion process was then carried out again with these new parameters, and new phase velocity maps were produced.

4.5.4 Adjustment of Reference Velocity.

When the inversion procedure is first carried out, the average of all the phase velocities across the study area, for a given period, is used for initial reference velocity. In principle, this should stabilize the phase velocity maps so that the average of the phase velocity perturbation is 0.0. In reality however, this is very rarely the case, so an adjustment of the initial reference velocity is required to get an appropriate balance. This process requires some trial and error for the reference velocity (Table 4.1) in order to obtain an average isotropic phase velocity of 0.0, for each period. Once this has been achieved, new phase velocity maps are again generated. These are the final versions of the phase velocity maps.

Period (s)	Reference Velocity (km/s)
015	3.490
025	3.788
030	3.900
045	4.075
050	4.104
060	4.145
075	4.178
090	4.210
100	4.233
125	4.298
150	4.381
200	4.618
225	4.761
250	4.915
300	5.287
350	5.735
400	6.030

Table 4.1. Reference velocity for each period.

4.6 Resolution Tests.

Following the completing of the inversion procedure, resolution tests were carried out to ascertain areas of limited resolution. Three different types of resolution tests were carried out – leakage tests, ‘reversed’ anisotropy, and spike tests (similar to traditional checkerboard tests) (see *Chapter 5*).

4.6.1 Leakage Tests

Leakage is the term used to describe variation in one parameter appearing as false variation in another parameter. Leakage tests are important because they test for the possibility of false structure in one parameter resulting from true structure in another. In the event that there is strong leakage, the parameters in question cannot be interpreted separately due to the cross-contamination. Leakage tests take as the input the result of the inversion for a given period, where a single component of the phase velocity map (isotropic-only, 2ψ -only, or 4ψ -only) is assigned as the input model for the resolution test. This is used to generate synthetic data which are then inverted using the same path distribution and regularization parameters as in the original inversion. A small amount of Gaussian noise is added to the system, prior to the inversion. The ideal result of the leakage test is for the chosen component to be well-recovered, while negligible amplitudes for the remaining 2 components are recovered. Leakage tests are carried out for each of the 3 components at various periods.

4.6.2 ‘Reversed’ Anisotropy Tests

‘Reversed’ anisotropy resolution tests are very similar to leakage tests. The input model for the test is the original model (i.e., the final tomographic model from the data), with the 2ψ anisotropy rotated through 90° . This test is quite useful as it gives an indication of the effect a

bias in azimuthal coverage has on the fast direction of anisotropy. For the purposes of this type of resolution test, the rotated direction should be well-recovered. If instead, the rotated direction is not recovered, but rather the anisotropy direction rotates back to its original position, this will likely suggest that the anisotropy is an artefact.

4.6.3 Spike Tests

Spike tests are used to determine how well localized anomalies are recovered; a checkerboard test is a spike test in which a grid of anomalies with alternating polarities are used. The procedure to design the pattern is automated using a python script authored by Omid Bagherpur Mojaver (pers. comms.), simplifying Sergei Lebedev's initial procedure. The script launches an interactive map that allows the user to design the input model for the spike tests.

All three types of resolution tests are executed within the same script for all periods, and the various results are output in separate sub-directories. Phase velocity maps were then generated for these results, at different periods, to investigate how well the input models for each type of resolution test were recovered. This resolution testing process was repeated several times to determine the regions where the resolution was lacking, as well as how well structures seen were resolved.

4.7 Overlay Maps.

Overlay maps were generated as a last step, to determine how the recovered model aligned with tectonic boundaries (Whitmeyer and Karlstrom, 2007) within Canada. This process was completed with the use of an open-source geographic information system application, QGIS.

The software took as input the shapefile of the tectonic boundaries (Whitmeyer and Karlstrom, 2007), the Atlas of Canada data for province and territory boundaries, as well as the gridfile used to create the phase velocity maps during the plotting of the inversion data. The Canada Atlas Lambert projection was used to set up the maps, since this projection is ideal for maps of Canada. Maps for each period were produced and exported (see *Section 6.2*).

5. RESOLUTION TESTING

Resolution tests introduced in *section 4.6* were carried out to determine the reliability of the maps obtained from the tomographic inversions. As previously stated, three different types of resolution tests were carried out – “spike” tests, leakage tests and “reversed” anisotropy tests – which are examined below.

5.1 Ray Path Maps.

Following data processing, ray path maps were produced to provide an indication of data coverage, which influences the recovered resolution. Maps were produced for the 17 distinct periods examined, and as expected, at the shortest and longest periods, the ray path coverage was quite low, resulting in poor resolution. Surprisingly, for the 225s period, a relatively long period, the ray path coverage remains quite decent (Figure 5.1). The densest ray path coverage occurred at the 100s period, with 2110 rays, while the sparsest coverage occurred at 400s, with 51 rays. In comparison, the shortest period, 15s, made use of 872 rays.

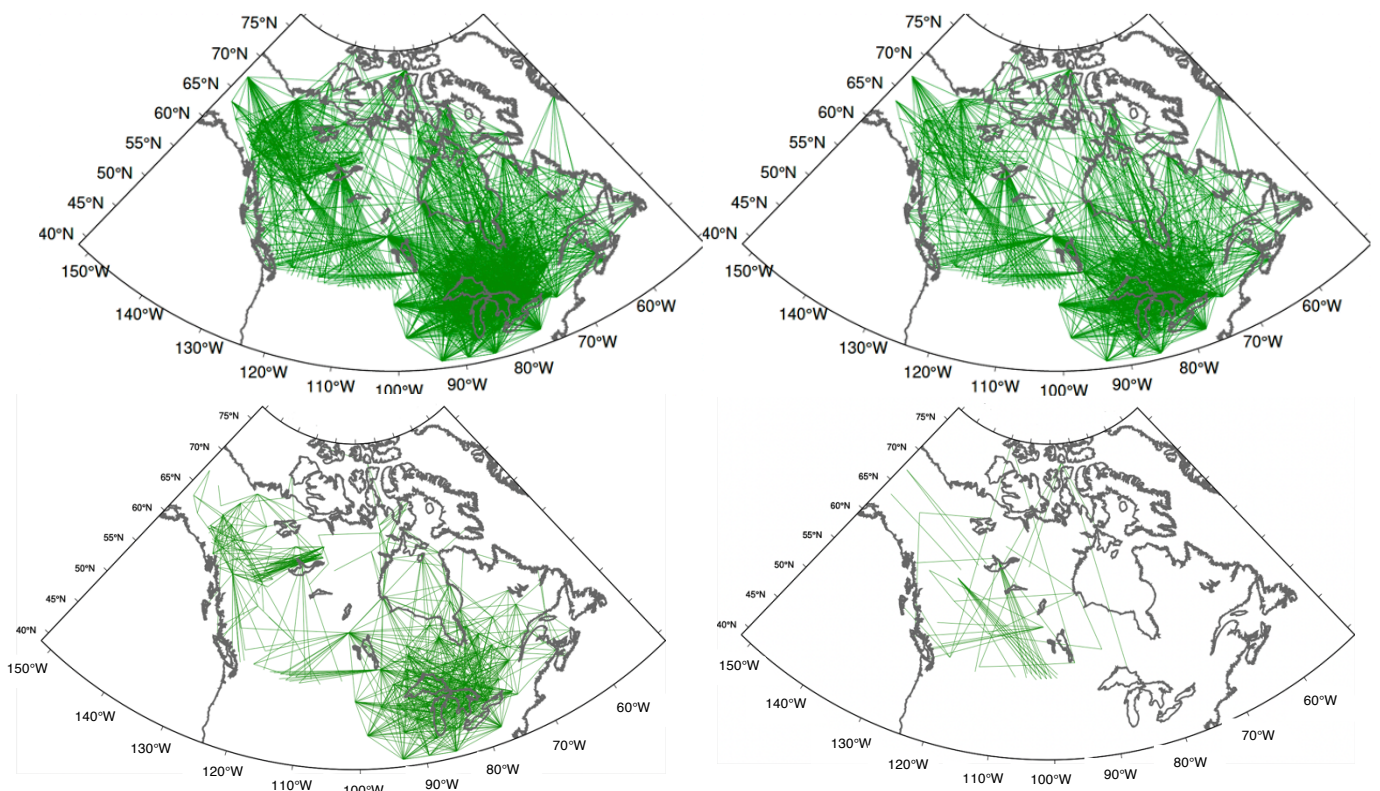


Figure 5.1. Comparison of ray path density at 100s (top left), 225s (top right), 15s (bottom left) and 400s (bottom right).

5.2 Structural Recovery and “Spike” Tests.

Broad swaths of low velocity across the country (Figure 5.2) were used to test the overall resolution and reliability of the dataset and the results. The resolution recovery is strong across the region, with some minor edge effects apparent, which suggests that the tomographic inversion model is quite robust. Importantly, for the third input model (Figure 5.2), the recovery of the anomaly’s amplitude varies. Spurious anisotropy of less than 1% is apparent in the recovered models.

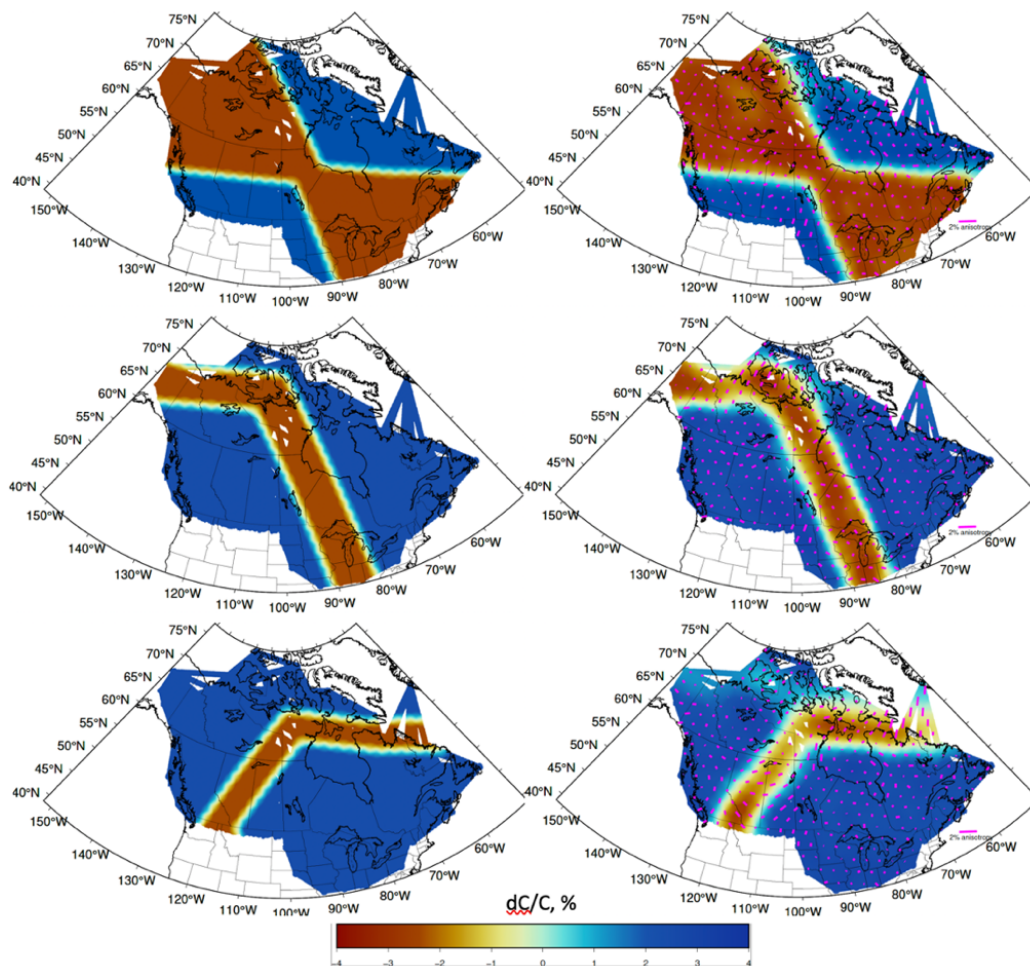


Figure 5.2. Spike tests for period 100s. Input models are created by selecting grid nodes to represent fast or slow phase velocity anomalies, which are inverted for a new model with the same ray path density and inversion parameters as used for the actual dataset. Input models are on the left and recovered models are on the right.

Classic checkerboard-style tests were carried out (Figure 5.3), with anomalies becoming increasingly smaller, to determine where the resolution broke down, and consequently, the reliability of the derived interpretations in specific areas. Figure 5.3(a) with large anomalies (~810 km across) shows strong resolution recovery across the country, though slightly less in Alberta and Nunavut, which is representative in terms of path density in these regions. There is no smearing at this scale. The resolution recovery with smaller anomalies (~630 km across), as seen in Figure 5.3(b) is still quite robust, though the variation in path density across the country makes more of a pronounced difference. Small amounts of smearing are apparent around Nunavut and northern Manitoba. With increasingly smaller anomalies (~435 km across) (Figure 5.3(c)) the resolution is completely lost in some areas – along the Saskatchewan-Manitoba border, the northern Cordillera, and the Northwest Territories-Nunavut border – highlighting areas of dense and sparse path coverage. In these areas where resolution is lost, there is more leakage into the anisotropic parameters. In sparsely-sampled areas (e.g. northern Saskatchewan) the size of the anomaly represents the resolution limit, while in regions of dense path coverage (e.g. southeastern Ontario), the resolution limit is smaller than the smallest checkerboard anomalies.

The resolution along the Cordillera-craton tectonic boundary was also tested to determine the robustness of the sharpness of the velocity transition, for the purposes of interpretation. Figure 5.4 shows robust resolution recovery. The test depicting the boundary alone however (Figure 5.4, bottom), shows comparatively worse resolution recovery at the northernmost tip of the Cordillera. Additionally, to the west of the low velocity region in British Columbia, there appears to be smoothing of the small-scale details, which is indicative of the scale of resolvable heterogeneity.

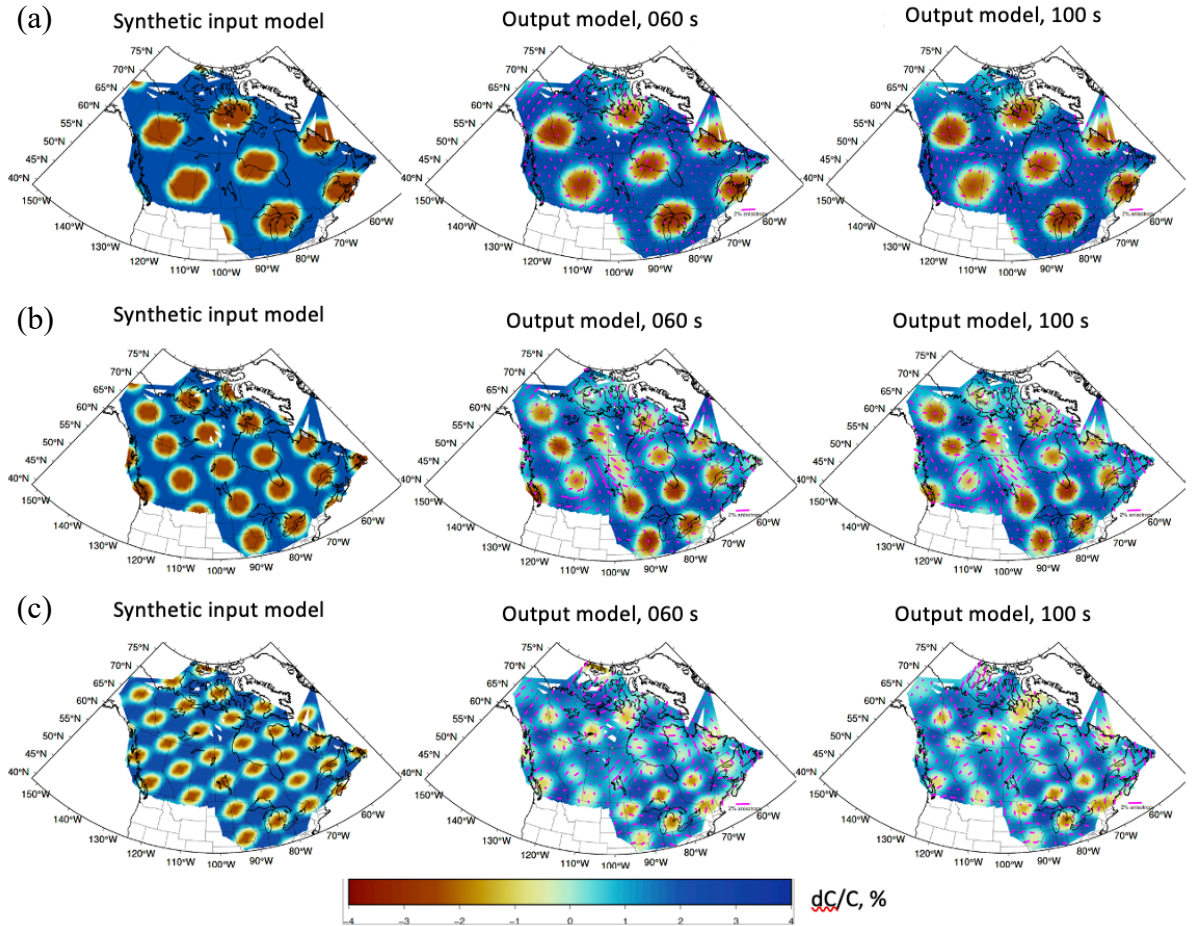


Figure 5.3. Traditional checkerboard-style tests for periods 060 and 100s. The synthetic input model is inverted for a new model with the same ray path coverage and inversion parameters as used for the real data. The resolution for the largest anomalies (top), medium-sized anomalies (middle) and the smallest anomalies (bottom) is shown.

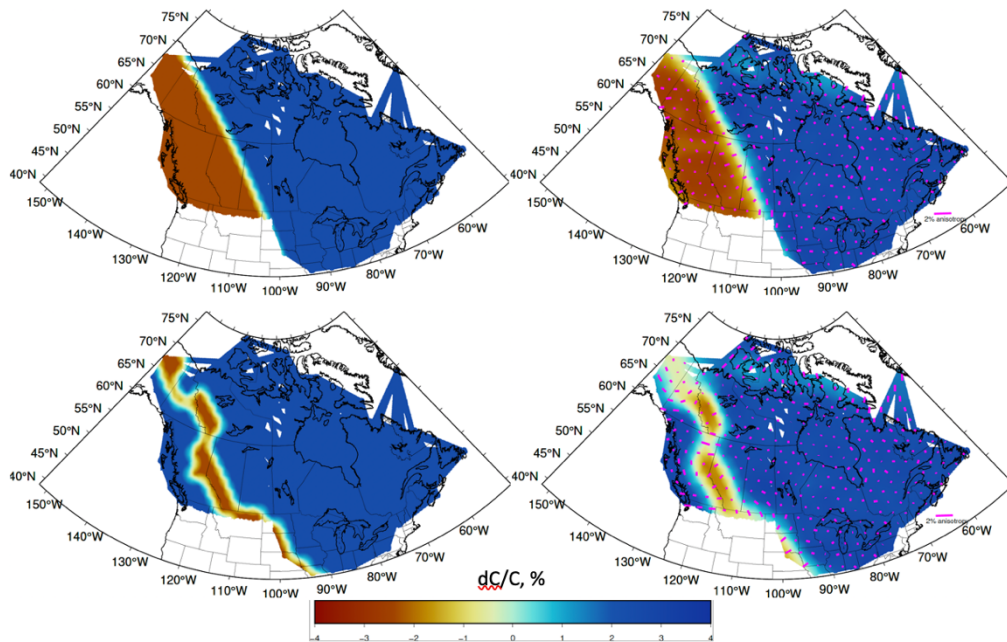


Figure 5.4. Spike tests for 100s for the Cordillera (top) and the Cordillera-craton boundary (bottom). The input model is inverted for a new model with the same path coverage and inversion parameters as used for the real data.

5.3 Leakage Tests.

If there is leakage of one parameter into another this means there is cross-contamination of the components and therefore the separate components are not safe to interpret. The isotropy-only test depicts good recovery of isotropic heterogeneities, while both the 2ψ and 4ψ artefacts are small (0.1-0.3%) (Figure 5.5), relative to the recovered anisotropy from the tomographic inversion (see Figures 6.3-6.11; approximately 1.7%-3.5%). The largest leakage occurrences can be seen at the model edges (Figure 5.5), which is expected. Based on the 2ψ -only leakage test, there is minimal leakage into isotropic anomalies with low amplitudes apparent from the muted colours ($\pm 0.5\%$). It can be concluded then, that this leakage is safe to ignore. The 2ψ recovery appears to be reasonable. The leakage from isotropic structure to anisotropy is more of a concern, and appears to be coverage-dependent (Figure 5.5(a)). The spurious 2ψ anisotropy does not exceed 0.6%, which suggests that stronger values are safe to interpret. The 4ψ artefacts are small in amplitude relative to the original tomographic inversions. The 4ψ -only test depicts a very similar scenario compared to the 2ψ -only test, though more leakage into 2ψ is apparent, at the edges mostly, than the other way round. Overall, this suggests that robust interpretations can be made from the isotropy and 2ψ anisotropy tomography models (see *Chapter 6: Results*), provided that the edge effects seen here are taken into consideration. The 4ψ anisotropy results will not be specifically interpreted (in *Chapter 7: Interpretation*) as the 4ψ pattern is not as well-constrained as the 2ψ pattern, and the 4ψ pattern is much harder to interpret as there is no clear relationship with olivine crystal alignment, as is the case with the 2ψ anisotropy pattern.

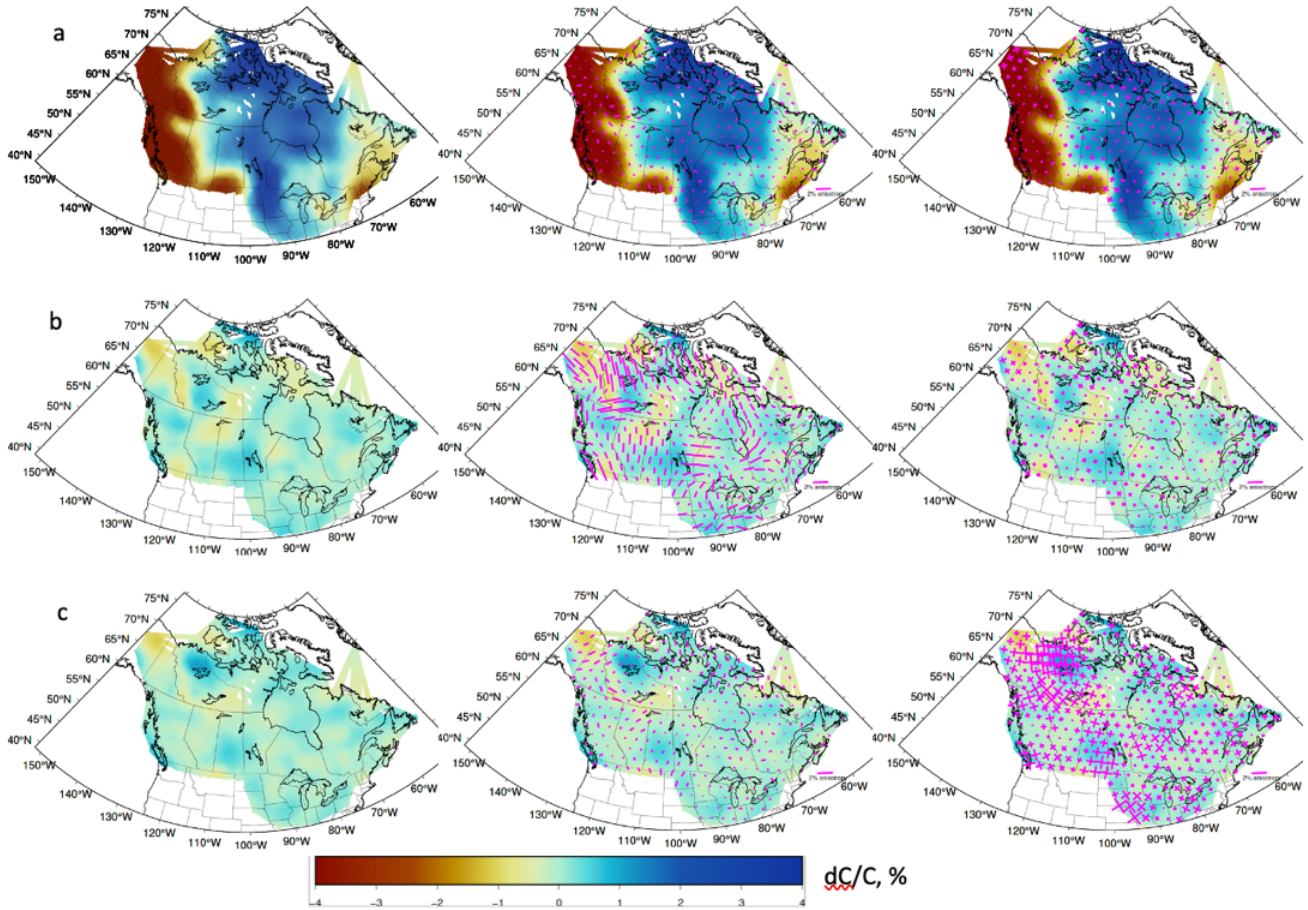


Figure 5.5. Leakage tests for 100s – (a) isotropic only, (b) 2ψ only and (c) 4ψ only.

5.4 “Reversed” Anisotropy Tests.

To further understand the extent of the leakage, the 2ψ anisotropy bars (from the previously derived phase velocity maps) were rotated through 90° and the 4ψ anisotropy values were set to zero, and this was used as the input model and reinverted. The recovered resolution model (Figure 5.6) depicts a smoothed version of the input model, and there are no systematic rotations that are cause for concern, indicating there is negligible azimuthal bias within the dataset. The recovered model also indicates nominal leakage into 4ψ anisotropy. This resolution test then, provides good evidence that the 2ψ anisotropic patterns are robust.

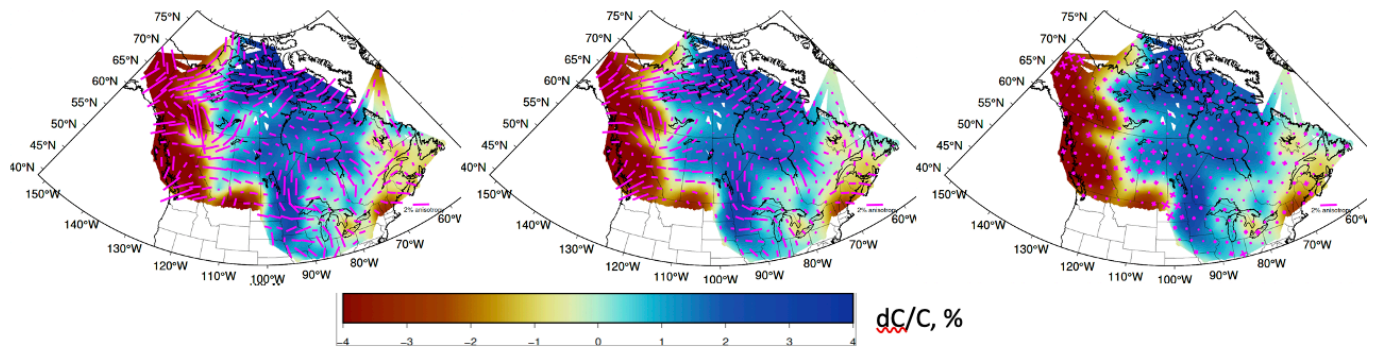


Figure 5.6. “Reversed” anisotropy tests at period of 100s. A synthetic input model for anisotropy rotated through 90 degrees is used (left) and inverted for a new model – 2-psi (middle) and 4-psi (right).

6. RESULTS

6.1 Dispersion Curves.

New dispersion curves (Figure 6.1) were obtained, as previously mentioned, to fill in the gaps in coverage within Canada, specifically in the Prairie provinces, with the use of USArray (TA) stations situated near to the Canadian-USA border (Figure 6.2). Furthermore, station pairs for these new curves were strategically picked to limit azimuthal bias in the tomographic models, which, if present, could lead to the presence of spurious anisotropy in the final model. A total of 2602 dispersion curves was used for this analysis – 117 of which were new dispersion curves.

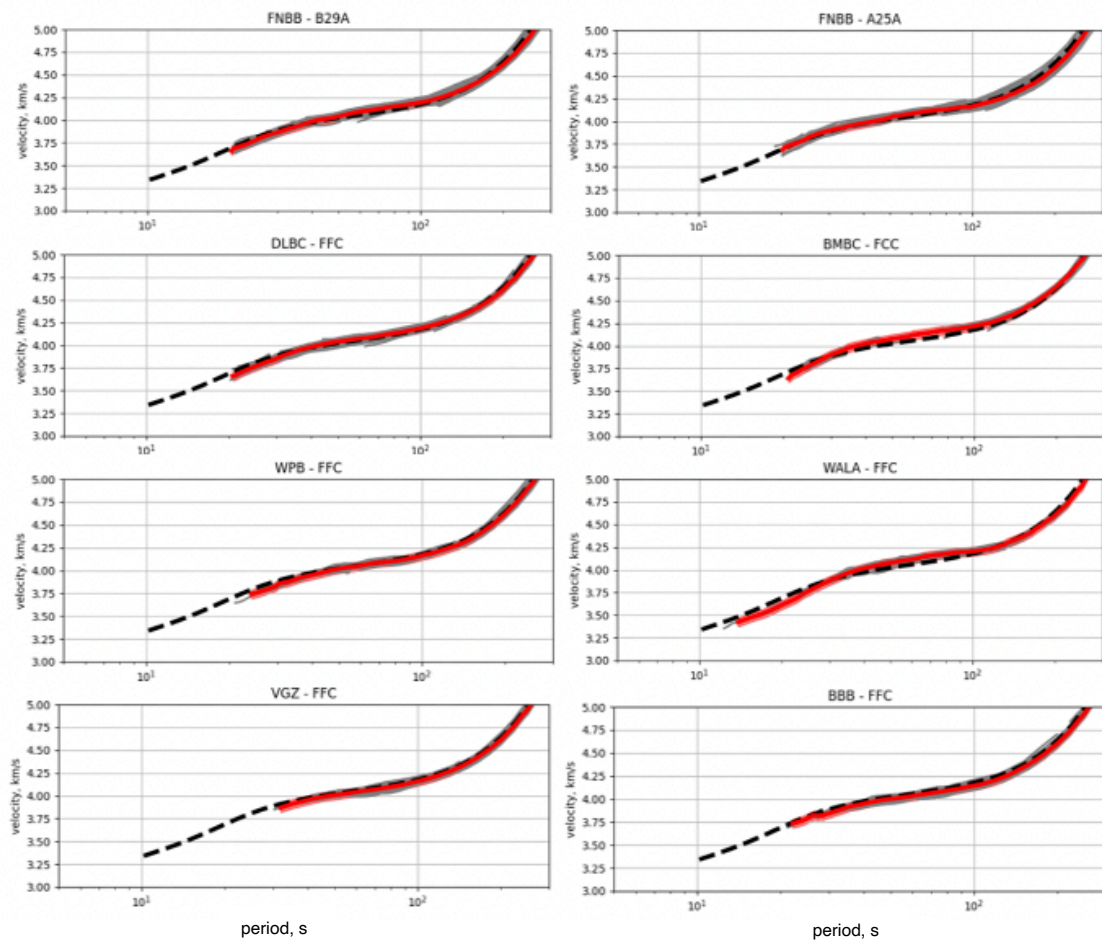


Figure 6.1. Sample of 8 of the 117 dispersion curves, utilizing stations from the CNSN and the USArray TA seismic network. The black dashed line represents the reference model, the grey lines represent each individual curve selection, and the red line represents the averaged curve (based on the grey curves).

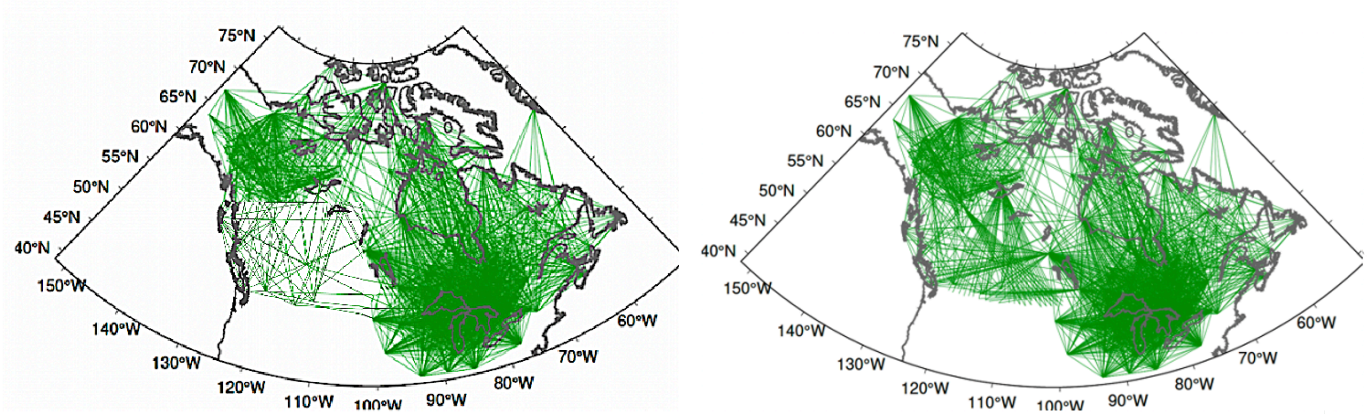


Figure 6.2. Comparison of the raypath coverage prior to (left) and following (right) the addition of new dispersion curves in regions of the southern Cordillera and Prairie provinces.

6.2 Phase Velocity Maps.

The individual two-station pair ray paths shown in *section 5.1*, and their associated dispersion curves, were used in the aforementioned tomographic inversion method (see *section 4.5*) to create phase-velocity maps. Seventeen maps were produced at periods of 15, 25, 30, 45, 50, 60, 75, 90, 100, 125, 150, 200, 225, 250, 300 and 400 seconds. A wide range of periods was selected to effectively capture variations within the subsurface. The maps were plotted using the Generic Mapping Tools (GMT) mapping software (Wessel et al., 2019).

The 15-second period map (Figure 6.3) shows very limited coverage. Consistently low velocities are seen within the northern Cordillera, with slightly higher low velocity continuing along the western edge of the country. Low velocities extend into southern and central Alberta. Low velocity is also seen within the vicinity of the Mid-continental Rift (MCR). The anisotropy reflects some spurious or artificial anisotropy, usually where there is azimuthal bias or leakage due to the presence of sharp transitions. This effect can be observed at the northwestern corner of the map where unrealistic anisotropy amplitudes (amplitudes significantly larger than those elsewhere on the map) are noted. The unrealistic amplitudes noted in this 15s map are attributed

to bias due to the lack of azimuthal coverage in the area. There are also large 4ψ values at the edge where coverage is non-uniform.

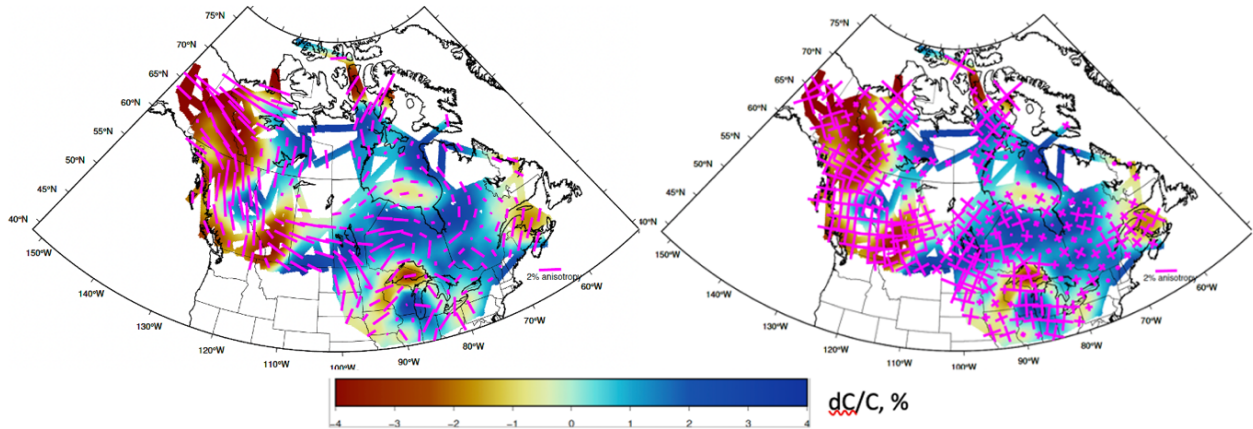


Figure 6.3. Phase velocity maps for 15s ($c=3.490$ km/s) – 2ψ anisotropy (left) and 4ψ anisotropy

At 25 seconds (Figure 6.4), isotropic anomalies of $\pm 2.7\%$ are observed. The low velocity of the Cordillera becomes more pronounced, as does the tectonic distinction between the Cordillera and the craton, with low velocity concentrating in southern Alberta. Low velocities also continue to persist within the vicinity of the MCR. A low velocity zone is also apparent beneath northernmost Nunavut. The majority of the craton exhibits high velocity with several low velocity anomalies scattered across the region (e.g. Saskatchewan, Hudson Bay). The anisotropy is decidedly stronger relative to the 15-second period map – there is strong SW-NE 2ψ anisotropy across Ontario and Quebec. There are also evident spurious patterns (e.g., unrealistic amplitudes and odd changes in anisotropy direction observed at the southwestern edge of the map) and large 4ψ edge values. At 30 seconds (Figure 6.4) however, the localized low velocity anomalies (present at 25s) are fewer, while the transition from low to high velocity at the Cordillera-craton boundary continues to sharpen. Interestingly, the low velocities of the Cordillera continue into southern Alberta. The anisotropy here is very similar to the 25s map, with some inevitable spurious patterns and large 4ψ values along the edges.

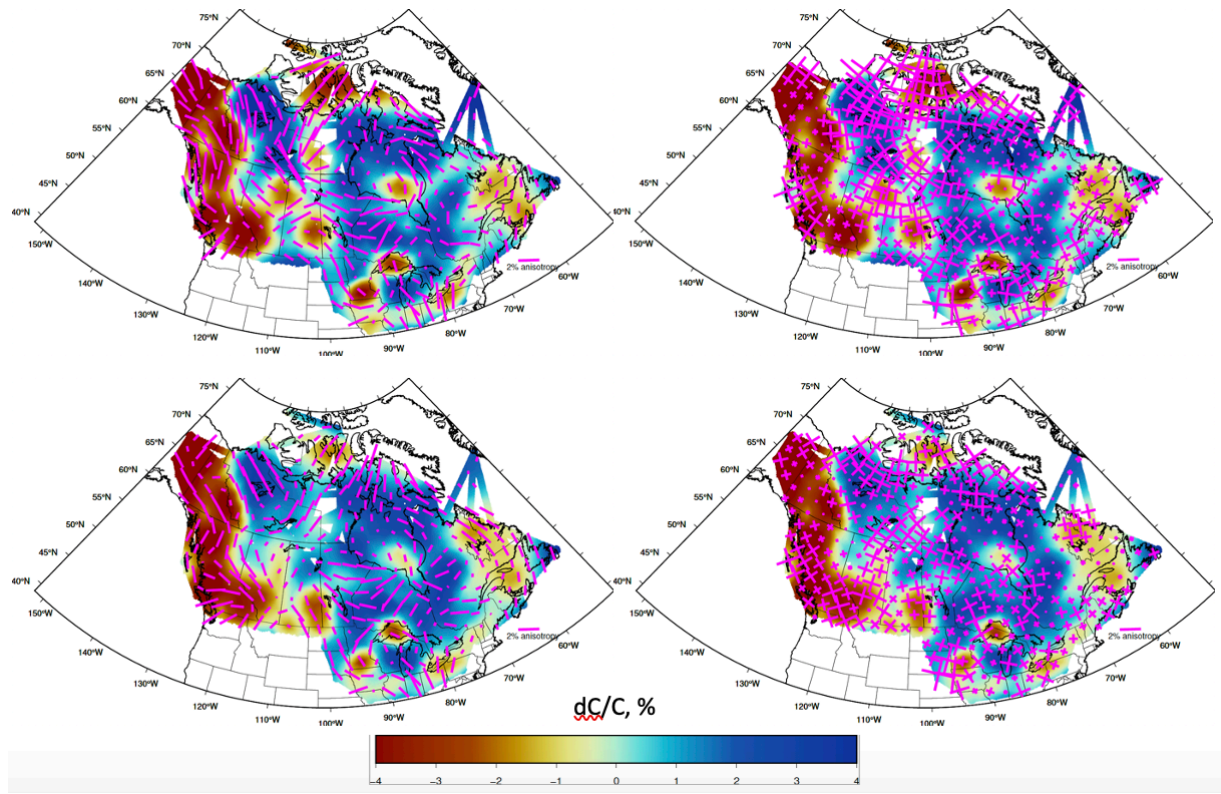


Figure 6.4. Phase velocity maps for 25s ($c=3.788$ km/s) (top) and 30s ($c=3.900$ km/s) (bottom) – 2ψ anisotropy (left) and 4ψ anisotropy (right).

At 45 seconds (Figure 6.5), low velocities cover most of Alberta, and penetrate into southern Saskatchewan. The low velocity anomalies present across the Shield at shorter periods do not appear, with the area having high velocity, with some smaller areas of very fast velocity. The anisotropy becomes stronger, and consequently, so do the 4ψ edge effects. At 50 seconds (Figure 6.5), isotropic velocity anomalies range up to $\pm 3.5\%$. The low velocity across most of Alberta persists, with the lowest velocities in southern Alberta. The isolated regions of faster velocity within the craton merge. The Cordillera-craton boundary becomes more distinct. The sharpness of the tectonic boundary transition – the shift from low to high velocity – increases. The sharpest region appears in the Northwest Territories, with the shift from low to high velocity occurring over approximately 150 km, while the boundary appears to be least sharp along the British Columbia-Alberta provincial border, where the velocity transition occurs over approximately 700 km. The anisotropy remains quite similar to the 45-second map, though there is more pronounced N-S 2ψ anisotropy in Nunavut.

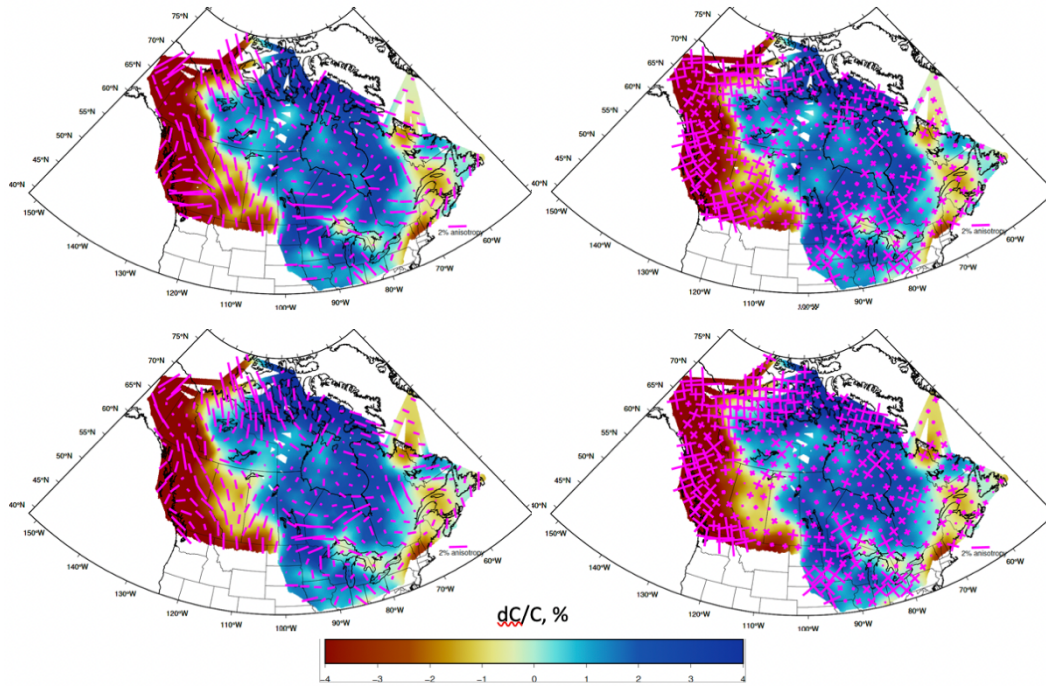


Figure 6.5. Phase velocity maps for 45s ($c=4.075$ km/s) (top) and 50s ($c=4.104$ km/s) (bottom) – 2ψ anisotropy (left) and 4ψ anisotropy (right).

The 60-second period map (Figure 6.6) shows a pronounced distinction between the Shield and the Appalachians in the southeastern region, and again, a strong Cordillera-craton tectonic boundary. The low velocities below southern Alberta and Saskatchewan, extending from the Cordillera, continue. The anisotropic patterns seem more uniform over larger areas – especially in areas where there is even azimuthal coverage – though there remain some 4ψ edge effects, such as the large 4ψ values along the southwestern Cordillera. The eastern Cordillera exhibits NW-SE 2ψ anisotropy that shifts E-W to NW-SE at the coast. In the craton, the strongest 2ψ anisotropy corresponds to regions of particularly high velocity.

The 75-second and 90-second maps (Figure 6.6) show very similar traits to those of the 60-second map. There is a noticeable shift in the Cordillera-craton boundary as the period increases. At 75 seconds, the sharpness of the Cordillera-craton boundary in the Northwest Territories increases slightly, with velocities transitioning over a 140 km distance, while along the British Columbia-Alberta provincial border, the sharpness of the Cordillera-craton tectonic

boundary increases significantly, with velocities transitioning over just 200 km. There is also a prominent decrease in the fast velocities in areas surrounding the Shield as the period increases. The anisotropy patterns are relatively similar across all three periods, with negligible variation where azimuthal coverage is robust.

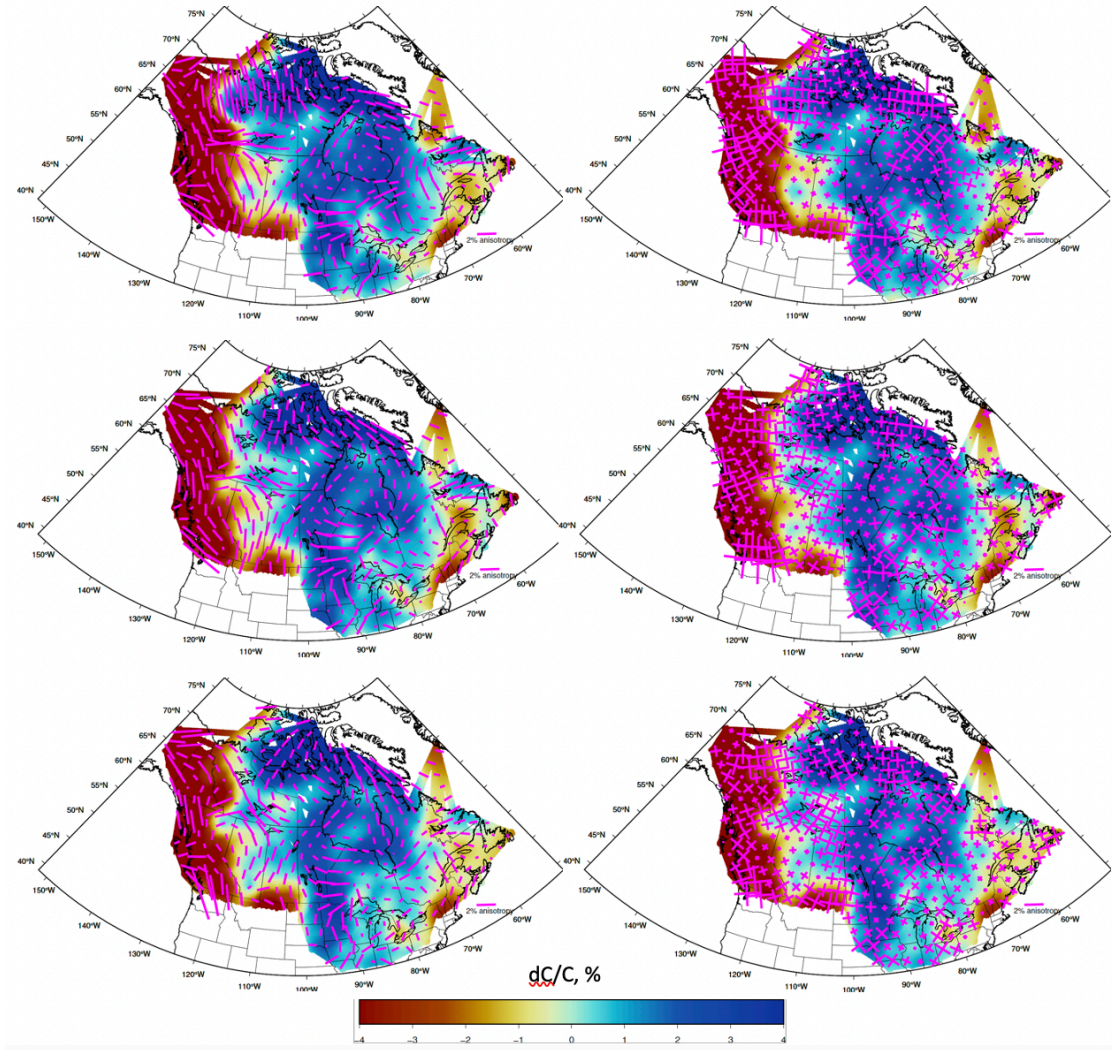


Figure 6.6. Phase velocity maps for 60s ($c=4.145$ km/s) (top), 75s ($c=4.178$ km/s) (centre) and 90s ($c=4.210$ km/s) (bottom) – 2ψ anisotropy (left) and 4ψ anisotropy (right).

At 100 seconds (Figure 6.7), the isotropic anomalies observed range between $\pm 4\%$. The evolution of the Cordillera-craton transition is striking, becoming quite irregular in northeastern British Columbia and west of Great Bear Lake. The sharpness of the transition continues to increase in the regions where a sharp boundary was previously observed, but remains the same where the transition was previously noted to be less sharp. The Northwest

Territories region of the boundary shifts to relatively high velocity over approximately 130 km, while where the boundary appears to be least sharp (in the vicinity of the British Columbia-Alberta provincial border), the transition occurs over 200 km. Additionally, the low velocities extend eastward across southern Alberta and Saskatchewan. The anisotropy exhibits relatively consistent patterns across various zones, which will be useful for interpretation, while taking into consideration the 4ψ edge effects.

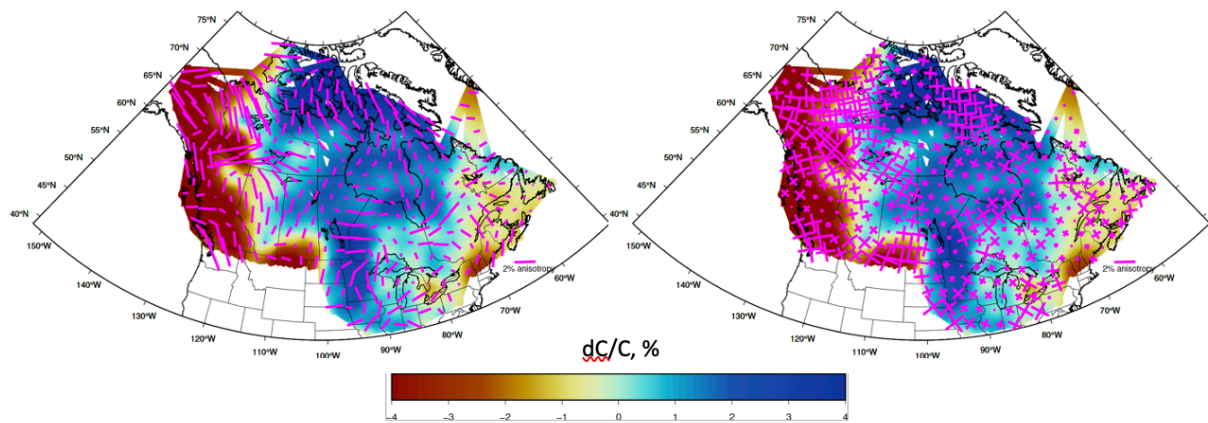


Figure 6.7. Phase velocity maps for 100s ($c=4.233$ km/s) – 2ψ anisotropy (left) and 4ψ anisotropy (right).

At 125 seconds (Figure 6.8), isotropic velocity variation reduces to $\pm 2.7\%$. While the overall patterning of high and low velocity regions is maintained, the boundaries become less sharp, most noticeably along the Cordillera-craton boundary. The anisotropy becomes less pronounced but exhibits similar patterns to that of the 100-second map (Figure 6.7). At 150 seconds, there is a pronounced shift in the spatial distribution of low and high velocities across the country, likely tied to the associated depth range. Along the Cordillera-craton boundary there appears to be a strong signal from lithospheric thickness variations below the craton, superposed on the variation between the Cordillera and craton. For this reason, interpretation of anisotropy patterns at longer periods such as these becomes complicated, as there is likely a combination of frozen lithospheric fabric and active mantle flow patterns.

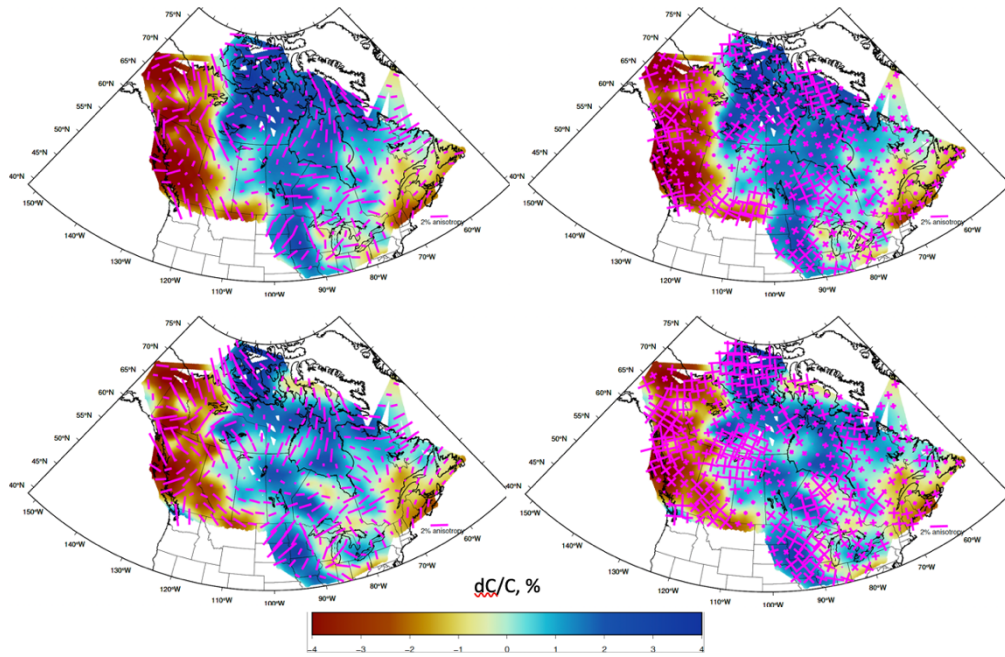


Figure 6.8. Phase velocity maps for 125s ($c=4.298$ km/s) (top) and 150s ($c=4.381$ km/s) (bottom) – 2ψ anisotropy (left) and 4ψ anisotropy (right).

At 200 seconds (Figure 6.9), the ray path coverage remains quite good, despite the long period, though the map is very heterogeneous, which makes interpretations challenging. The velocity variation likely reflects some lithosphere, in the thickest parts of the craton, but largely interaction with the asthenosphere, producing this spatial arrangement. At 225 seconds (Figure 6.9), the path coverage is still very good. The velocity variations across Canada are quite consistent with the 200-second map, though the low velocity zones become more pronounced.

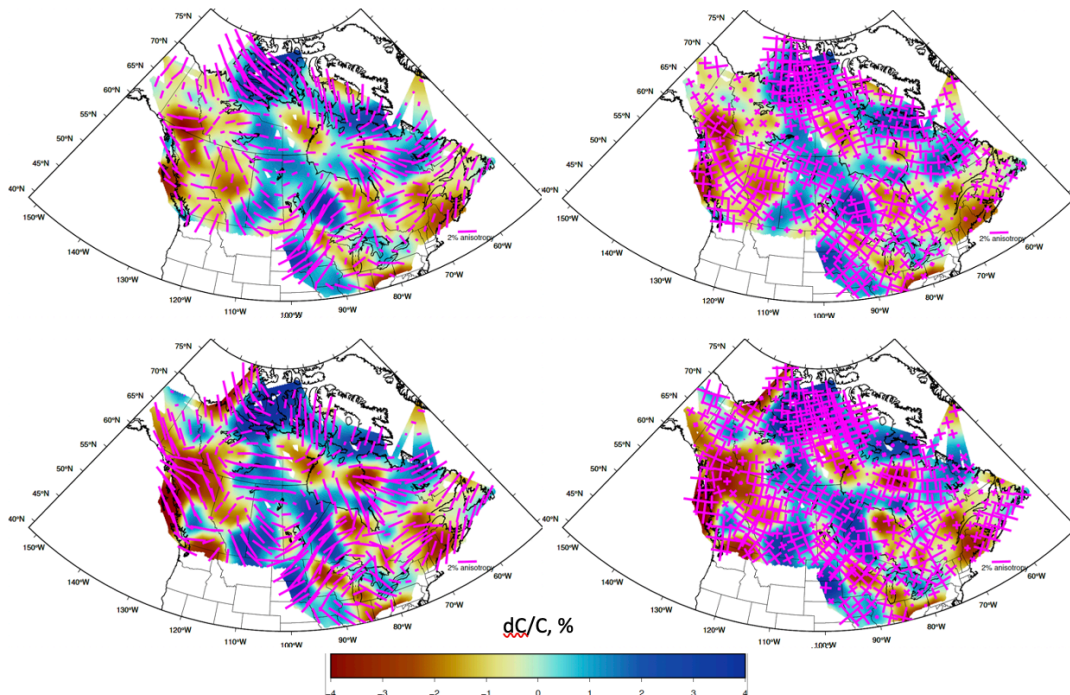


Figure 6.9. Phase velocity maps for 200s ($c=4.618$ km/s) (top) and 225s ($c=4.761$ km/s) (bottom) – 2ψ anisotropy (left) and 4ψ anisotropy (right).

The 250-second map (Figure 6.10) maintains good ray path coverage, and exhibits a relatively similar picture to that of the maps at 200 and 225 seconds (Figure 6.9), though there is likely more asthenosphere interaction. At 300 seconds (Figure 6.10) however, there is a dramatic shift in the velocity variation across the country, with a low velocity zone across BC and southern Yukon and NT narrowly connected to another low velocity zone across most of Quebec, southernmost Ontario, and parts of Atlantic Canada (Nova Scotia and Prince Edward Island), surrounded by consistently high velocities. It is important to note that the ray path coverage across the middle of the country (Manitoba, Ontario and southern Nunavut) becomes sparse, while remaining relatively dense in the surrounding areas. At these long periods the anisotropy becomes unstable and unsafe to interpret.

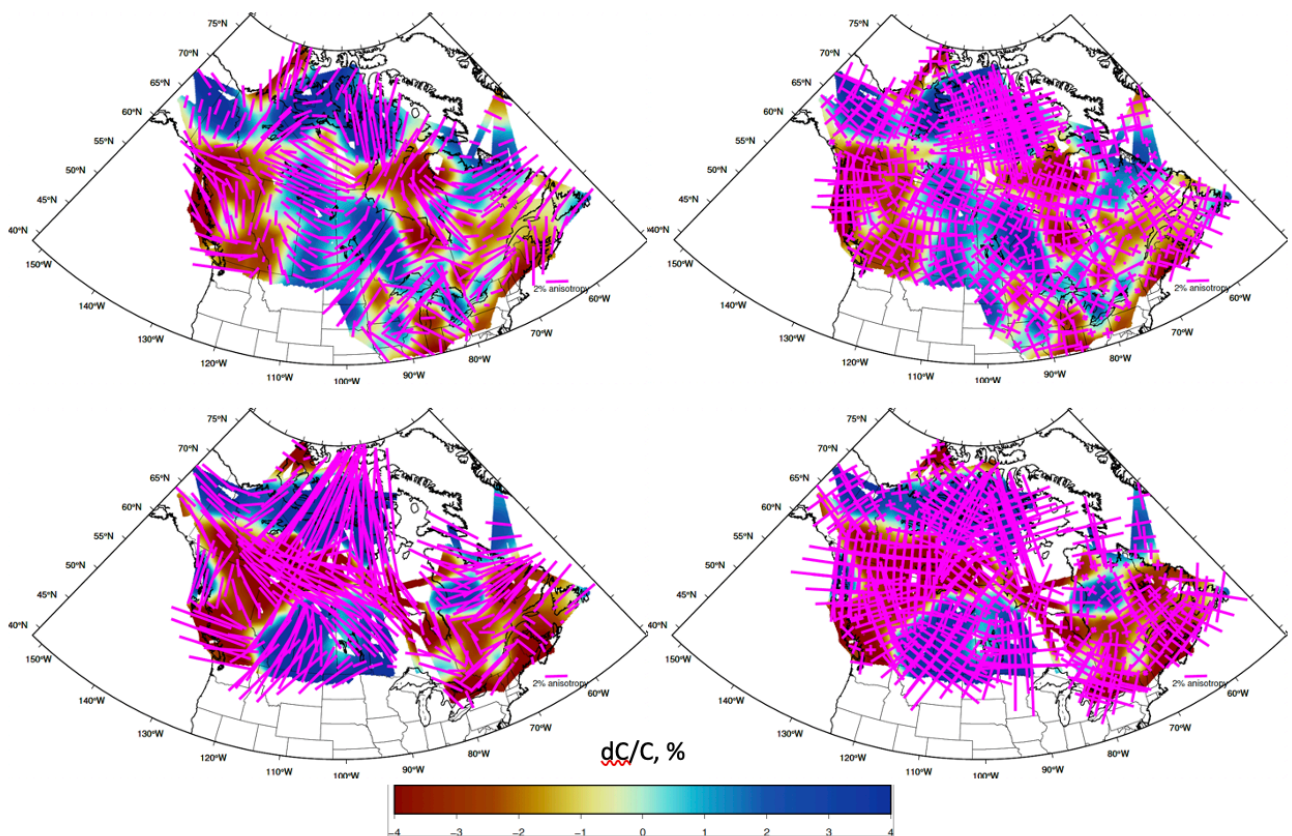


Figure 6.10. Phase velocity maps for 250s ($c=4.915$ km/s) (top) and 300s ($c=5.287$ km/s) (bottom) – 2ψ anisotropy (left) and 4ψ anisotropy (right).

The maps at 350 and 400 seconds (Figure 6.11) are very similar, though the reduction in ray path coverage between the periods makes a substantial difference. Ray path coverage is completely lost in eastern Canada due to a lack of measurements from previous studies that covered this region. The velocity variation reflects a continuous spatially diagonal low velocity zone spanning southern British Columbia to mid-Nunavut, surrounded by relatively high velocity.

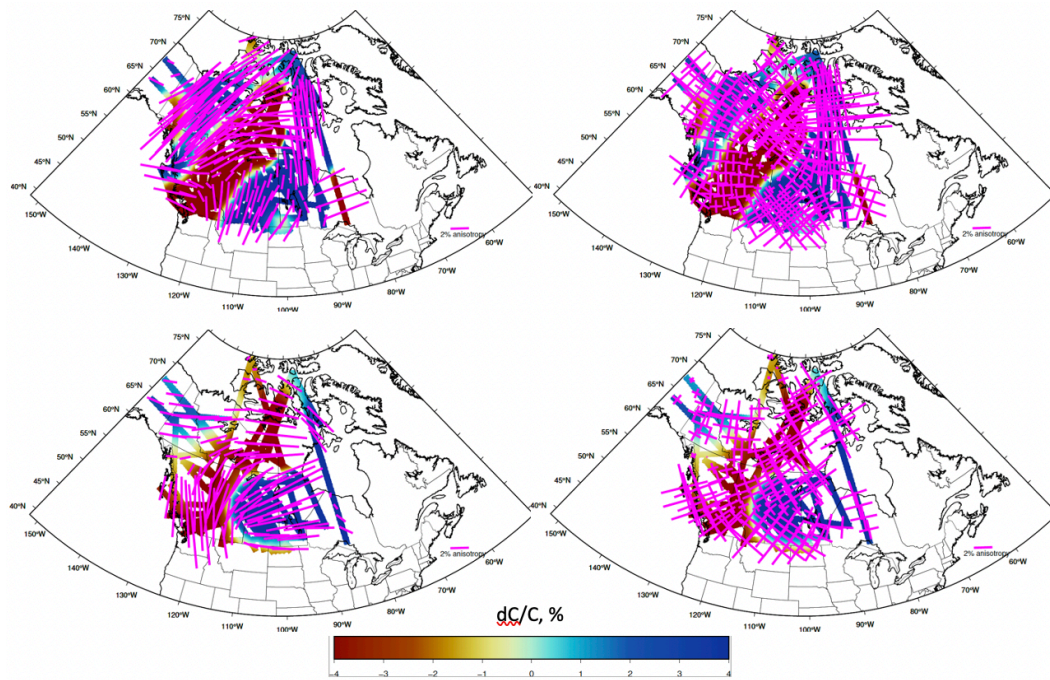


Figure 6.11. Phase velocity maps for 350s ($c=5.735$ km/s) (top) and 400s ($c=6.030$ km/s) (bottom) – 2ψ anisotropy (left) and 4ψ anisotropy (right).

6.3 Summary.

For most of the sampled periods (25-150 seconds), the difference between Precambrian and younger terranes distinctly stands out, with a sharp Cordillera-craton boundary. The sharpness of this transition changes across these periods, becoming most pronounced at 100 seconds. There is also great definition of the transition between the Shield and the Appalachians in the

southeast at 50-90 seconds. The anisotropic patterns are relatively consistent from approximately 45-125 seconds, and become more complex at 150 seconds. 4ψ edge effects persist across all periods.

Notably, across several periods, the penetration of low velocities across southern Alberta (first appearing at 25 seconds) and Saskatchewan (first appearing at 45 seconds) continues until the 150-second period. There are also several interesting anomalies that present within the Shield at various periods.

Major interpretable features coming out of the results are as follows and are further examined in *chapter 7* (Figures 7.1-7.6):

1. Southwestern Cordilleran low velocity and associated anisotropy pattern (A)
2. Southeastern Cordilleran low velocity and associated anisotropy (B)
3. Cordillera-craton boundary (C)
4. Extent of low velocities beneath southern Alberta and southern Saskatchewan (D)
5. Nunavut high velocity feature and associated anisotropy (E)
6. Variations within the Shield, including the MCR (F)
7. The Appalachians and their associated anisotropy (G)

7. INTERPRETATION.

The dispersion maps for representative periods at lithospheric and sublithospheric depths were selected and common features as described in *section 6.3* were highlighted across the selected maps. Possible interpretations are put forward in *sections 7.3-7.8*. The depth relation with period is outlined in *section 7.1*.

7.1 Depth Sensitivity Summary.

The periods (Table 7.1) represent averages over varying depth ranges, and as a result, these variations between periods can be used to make interpretations about depths. The periods featured in Table 7.1 were selected as representative periods, as they provide an excellent overview of the anomalies identified in *section 6.3*. These periods showcase how the identified anomalies present at various depths. The depth ranges were determined through the use of depth sensitivity kernels based on the AK135 1-D model (Kennett et al., 1995). There are overlaps in the depth ranges associated with each period, which require the differences noted at longer periods to be related to the greater end of the depth range.

Period (s)	Depth Range (km)	Depth of Peak Sensitivity (km)
30	20-70	30
50	40-120	40
75	50-150	85
100	80-200	110
125	80-250	140
150	80-330	160

Table 7.1. Selected periods and their corresponding depth range and depth of peak sensitivity.

Very short periods are generally associated with the crust and upper lithosphere, while intermediate periods are generally associated with the deeper lithosphere, and longer periods are associated with sublithospheric mantle. This relationship of course, varies depending on the location within Canada. For instance, in regions where the lithosphere is especially thick (e.g., cratonic lithosphere) both very short and long periods will represent lithospheric structure, with only the longest periods possibly representing sublithospheric mantle. This is not the case for regions such as the Cordillera and the Appalachians where the lithosphere, in contrast, is much thinner, so longer periods will reflect the sublithospheric mantle. Additionally, the depth sensitivity is model-dependent – large deviations from the AK-135 model will alter these sensitivities. This phenomenon is due to the nonlinearity of surface-wave depth dependence.

7.2 Major Features.

For the periods listed in *Table 7.1*, corresponding dispersion maps were divided into regions with common properties, as indicated in *section 6.3*. There are many major features of interest across Canada, identified as Features A-G (Figures 7.1-7.6) for the purposes of simplicity.

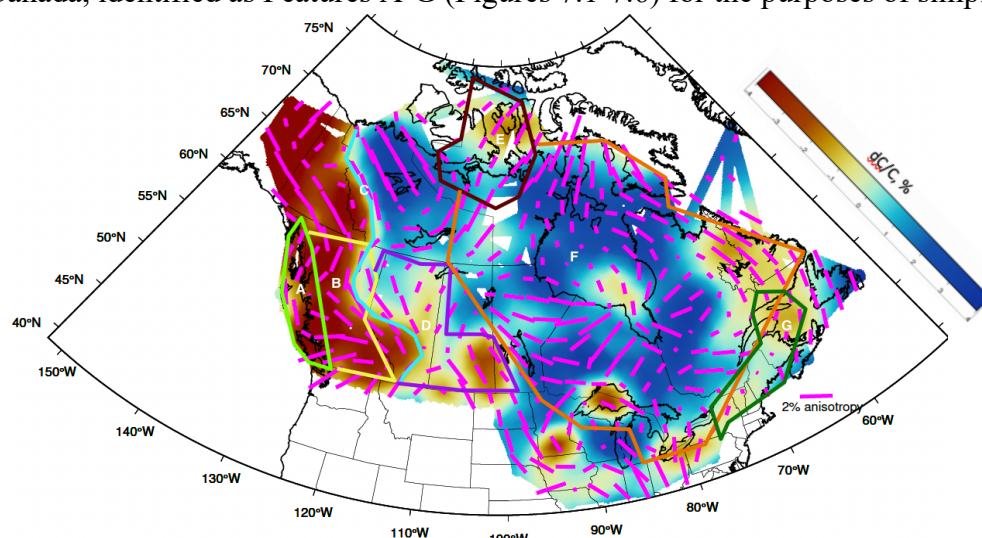


Figure 7.1. Phase velocity map for 30s highlighting features to be interpreted – Feature A (neon green), Feature B (yellow), Feature C (teal), Feature D (purple), Feature E (brown), Feature F (orange), Feature G (dark green). Reference phase velocity, $c = 3.900$ km/s.

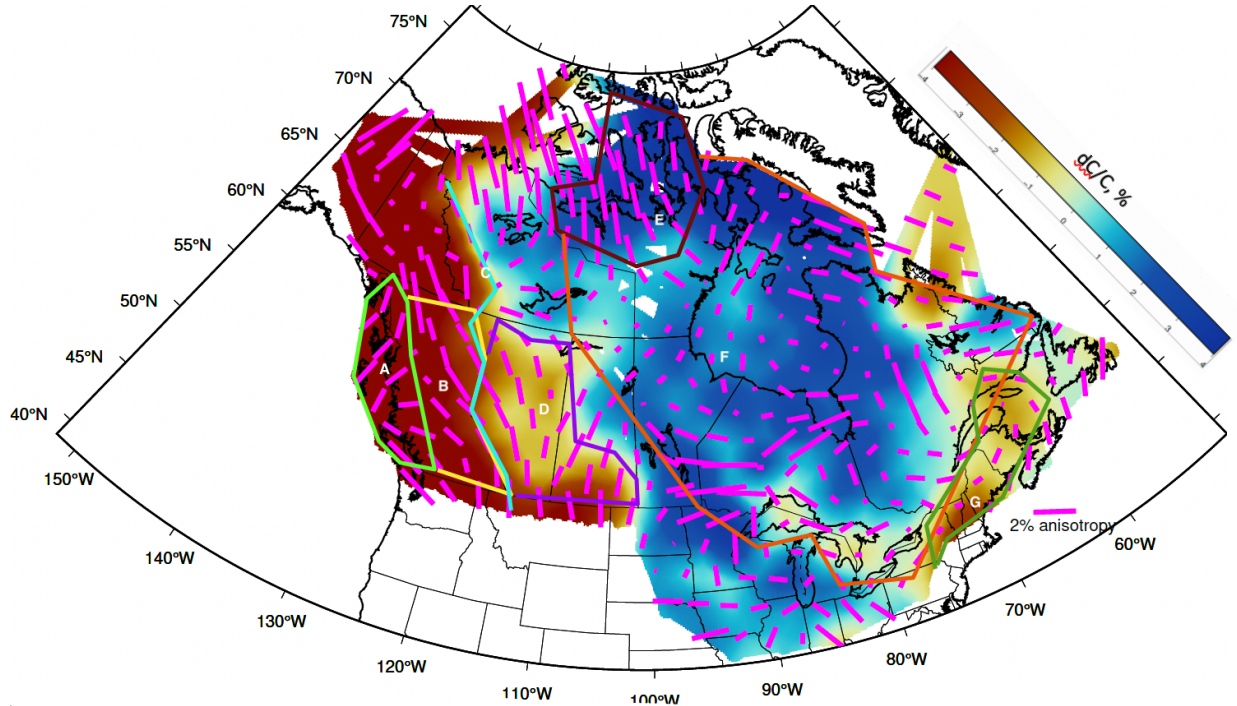


Figure 7.2. Phase velocity map for 50s highlighting features to be interpreted – Feature A (neon green), Feature B (yellow), Feature C (teal), Feature D (purple), Feature E (brown), Feature F (orange), Feature G (dark green). Reference phase velocity, $c = 4.104$ km/s.

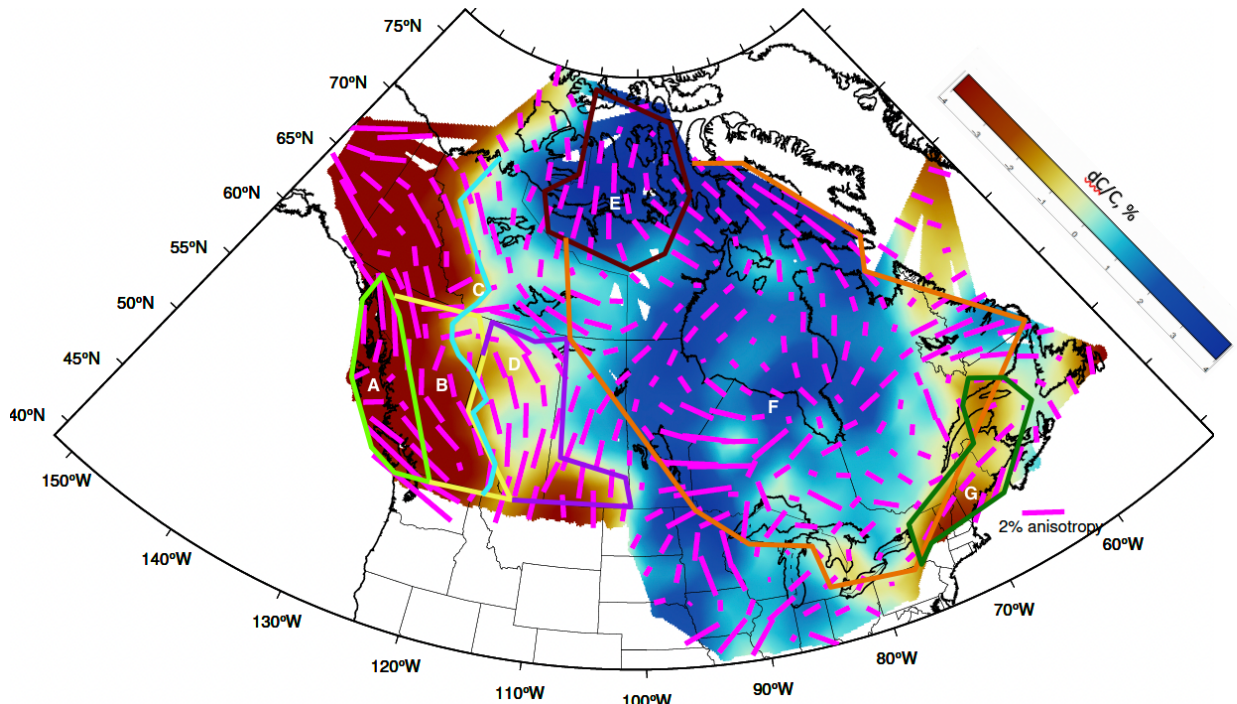


Figure 7.3. Phase velocity map for 75s highlighting features to be interpreted – Feature A (neon green), Feature B (yellow), Feature C (teal), Feature D (purple), Feature E (brown), Feature F (orange), Feature G (dark green). Reference phase velocity, $c = 4.178$ km/s.

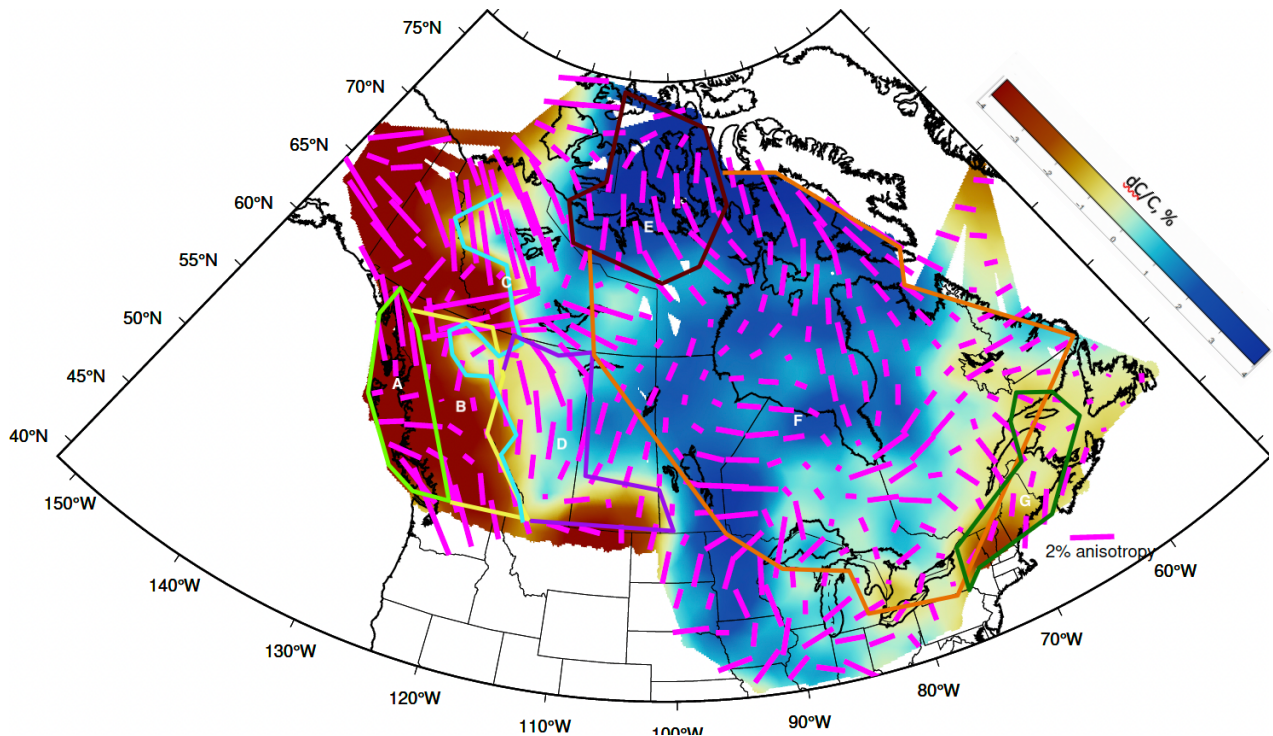


Figure 7.4. Phase velocity map for 100s highlighting features to be interpreted – Feature A (neon green), Feature B (yellow), Feature C (teal), Feature D (purple), Feature E (brown), Feature F (orange), Feature G (dark green). Reference phase velocity, $c = 4.233$ km/s

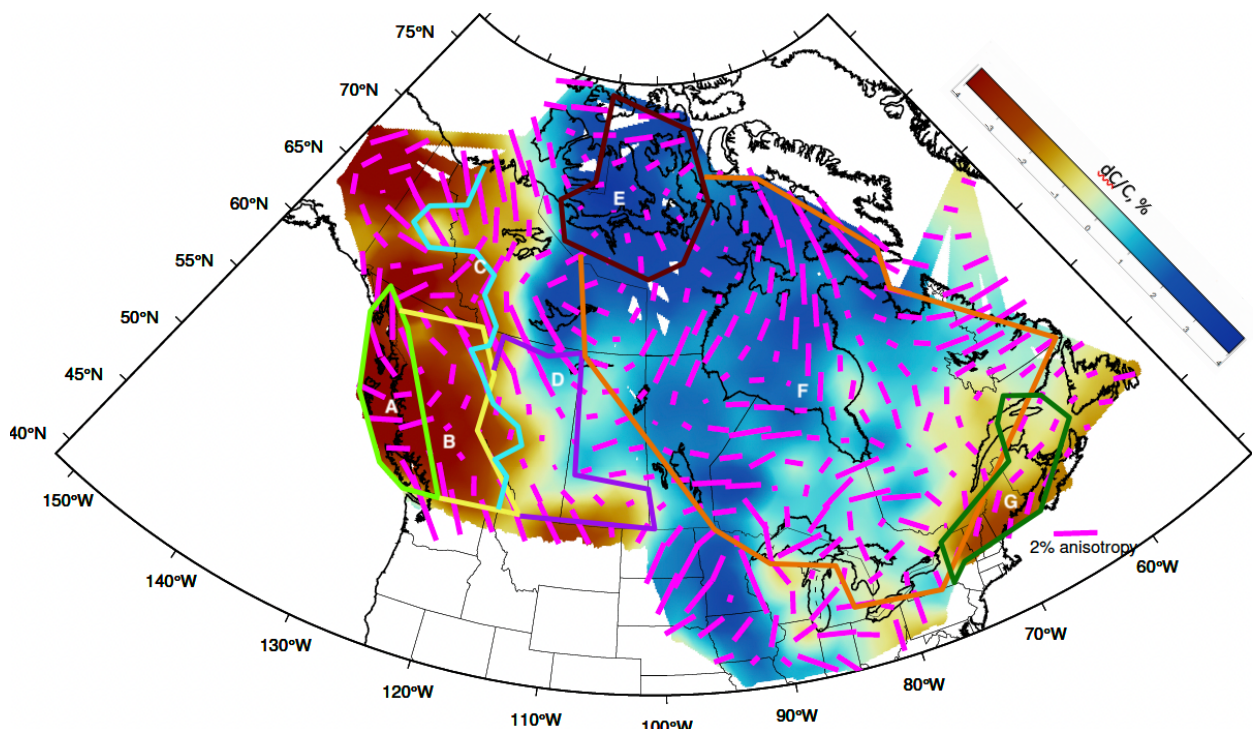


Figure 7.5. Phase velocity map for 125s highlighting features to be interpreted – Feature A (neon green), Feature B (yellow), Feature C (teal), Feature D (purple), Feature E (brown), Feature F (orange), Feature G (dark green). Reference phase velocity, $c = 4.298$ km/s

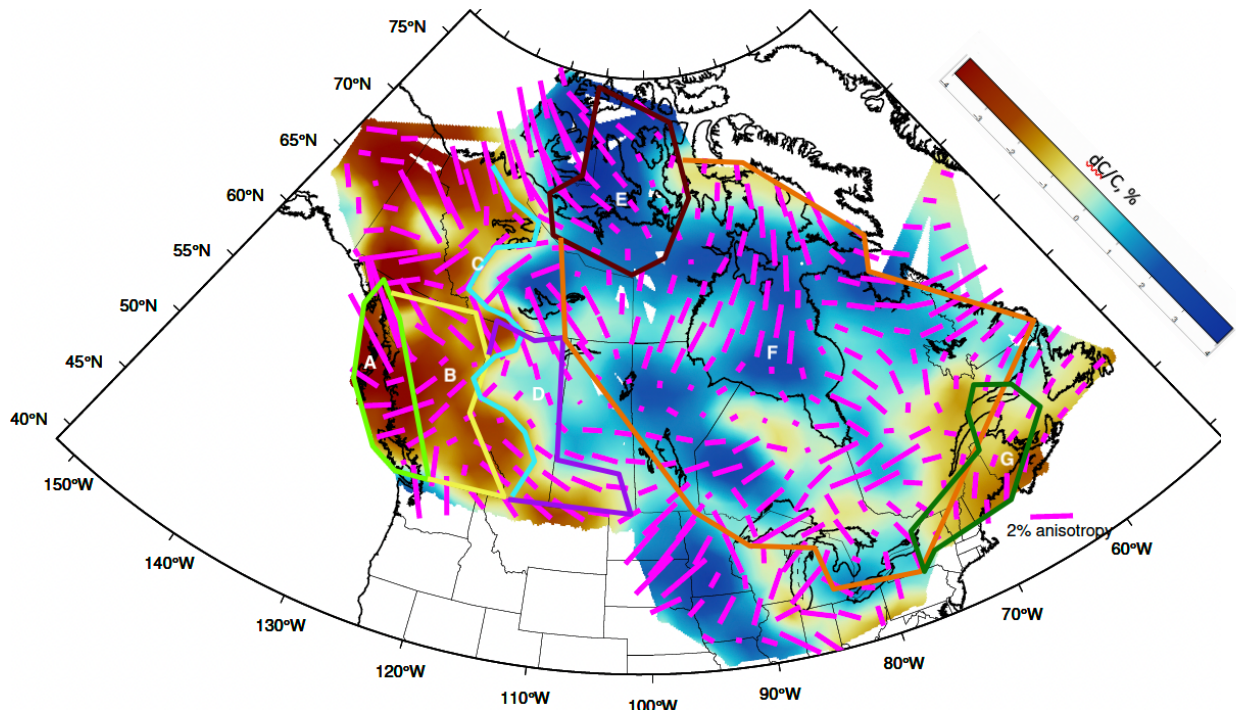


Figure 7.6. Phase velocity map for 150s highlighting features to be interpreted – Feature A (neon green), Feature B (yellow), Feature C (teal), Feature D (purple), Feature E (brown), Feature F (orange), Feature G (dark green). Reference phase velocity, $c = 4.381$ km/s.

Feature A highlights a low velocity anomaly and its associated anisotropy beneath the southwestern Cordillera. Just east of Feature A, Feature B is a region with very similar velocity values, but very different anisotropy. Feature C highlights the Cordillera-craton boundary, separating the low velocity Cordillera from the high velocity Shield, which becomes sharper as periods increase to 100 seconds, as the model resolution improves. Feature D refers to the moderate to very low velocity across Alberta and southern Saskatchewan, respectively. Feature E depicts a high velocity anomaly across northern Nunavut and its associated anisotropy. Feature F encompasses the Shield, which contains several smaller scale anomalies. Feature G highlights the Appalachian region and its associated anisotropy. Figures 7.1 through 7.6 show these notable features which will be interpreted in the following sections (*sections 7.3-7.9*).

7.3 Southern Cordilleran Velocity Anomalies and Associated Anisotropy.

The entirety of the Cordillera exhibits a relatively uniform low velocity anomaly of up to -4%. The anisotropy recovered across its expanse is not consistent, but rather depicts regions of consistent but azimuthally varying anisotropy within the Cordillera. Such features occur along the southern Cordillera, namely Features A and B (Figures 7.1 & 7.2). Both Features A and B in their entirety are low velocity zones, which corroborates the findings of previous studies (e.g., Schaeffer and Lebedev, 2014; Zaporozan et al., 2018). Given that this low velocity appears relatively consistent across all periods being considered, the depth association would indicate that these low velocities persist to a depth of at least 300 km. Low velocities are associated with the Cordillera as it is comprised of younger material (< 180 Ma) and is an actively deforming zone. These low velocities further indicate mantle that is hot, hydrated, or both. Given the history of subduction in North America, it is likely a combination of both.

Although Features A and B share several similar characteristics, they were identified as distinct interpretable features due to the slightly higher velocities noted across Feature B, as well as the change in anisotropy orientation from Feature A to Feature B which is most prominently noted on the 50s map (Figure 7.2). The higher velocities in Feature B become more apparent at longer periods, specifically at 150s (Figure 7.6), while the nonhomogeneous anisotropy noted previously at shorter periods becomes less apparent at intermediate to long periods, starting from periods of 100s and longer. Additionally, at 100s (Figure 7.4), the Cordillera-craton boundary becomes much more defined, allowing for areas of faster velocity, related to the craton, to appear (anomalies of up to +1%).

7.3.1 Feature A Anisotropy

The anisotropy in this region at shorter periods, (Figures 7.1 & 7.2) possesses a NE-SW orientation in the north (at an approximate latitude of 55°N), which then rotates to a NW-SE orientation toward the southern end of Feature A (approximate latitude of 50°N). This change is more discernible at the 50s period (Figure 7.2). Interestingly, at moderate periods, (Figures 7.3 & 7.4), the anisotropy orientation is NW-SE in the north of the feature, rotating to almost E-W at the centre of the region, before returning to a NW-SE orientation. At long periods, the anisotropy across Feature A remains consistent with that at moderate periods.

Within the shallow upper mantle of Feature A, one very likely source of anisotropy in the southwestern region of the Cordillera is the structural fabric resulting from the assemblage of NW-SE trending terranes (Rasendra et al., 2014), where the fast axis represents frozen tectonic fabrics associated with the accretion of terranes throughout the Palaeozoic and the Mesozoic. This region of the Cordillera is characterised by the NW-SE trending Insular superterrane, consisting of several smaller terranes, amalgamated offshore and later accreted to the continent during the Mesozoic. Moving northward within Feature A, the noted distinct change in the fast direction of the anisotropy pattern by approximately 90°, from NW-SE to a SW-NE orientation is likely related to the lithospheric subduction, which would have occurred during terrane accretion. This almost perpendicular shift in anisotropy could be attributed to olivine fast axes being preferentially oriented parallel to flow, but perpendicular to the trench (Faccenda and Capitanio, 2012).

As described above, the anisotropy changes with depth (Figures 7.3-7.6), as the influence of subduction becomes more apparent. The anisotropy in this region at shallow mantle depths is terrane-influenced, and then becomes influenced by flow associated with deeper subduction

processes. The region within which Feature A is located is influenced by the Juan de Fuca subduction zone, and the associated flow pattern affects the anisotropy, which is an asthenospheric phenomenon. Alternatively, this anisotropy may also be influenced by the effects of the Eocene extensional tectonics (Ledo and Jones, 2001), which affected the southwestern region of the Cordillera (south of 52°N). The switching of the fast axis to an almost E-W orientation is possibly explained by a shift in the fabric at depth, likely related to further subduction activity. The NW-SE trending anisotropy toward the south of Feature A is noticeably stronger within the deep upper mantle compared to the shallow upper mantle.

Shear-wave splitting measurements, which measure anisotropy averaged vertically through the upper mantle, were also compared to the surface-wave anisotropy at two selected periods, 50s and 100s. The 50s map overlay (Figure 7.7) indicates that the splitting is not consistent with the recovered anisotropy for the depth range associated with this period. Additionally, while the 100-second overlay (Figure 7.8) shows more consistency in the northwestern region of Feature A, the split measurements in the southern region remain inconsistent with the recovered surface wave anisotropy even at this longer period, correlating to increased depths. This is particularly interesting, given the thin lithosphere beneath the Cordillera, as quite a bit of asthenospheric activity would be expected in the approximately 80-250 km depth range. As such, the anisotropy is largely asthenospheric in nature, and these results are indicative of a depth-dependent flow pattern. It is possible that there is enough variation in the fast anisotropic orientations with depth, that the splitting average will not correspond to the surface-wave results. Alternatively, it could suggest that the Cordilleran splitting measurements may have better lateral resolution, and the surface-wave anisotropy is smeared at the edge of the model.

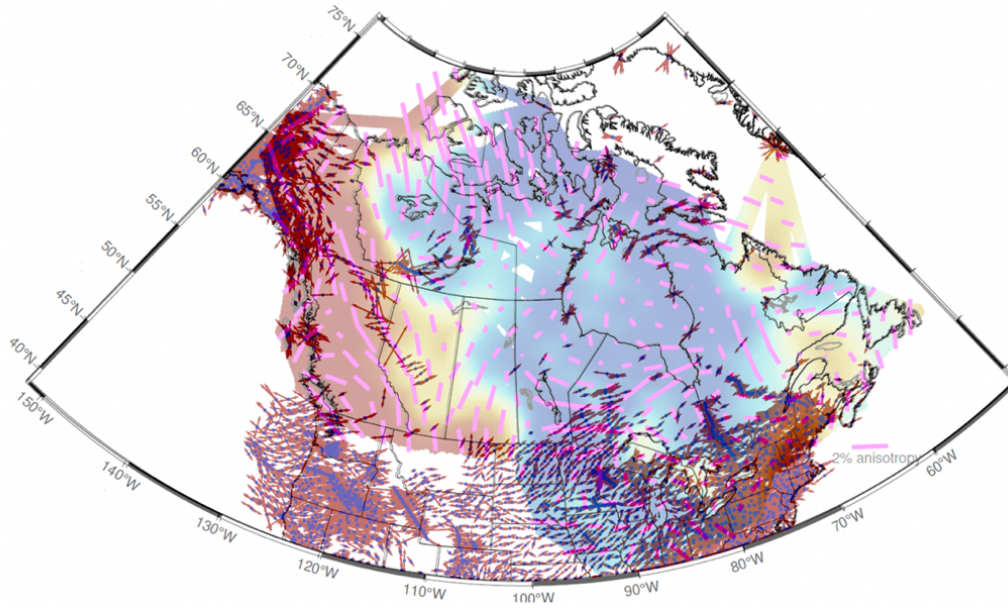


Figure 7.7. Phase velocity maps for 50s overlain with the shear wave splitting database measurements (IRIS DMC, 2012; Barruol et al., 2009).

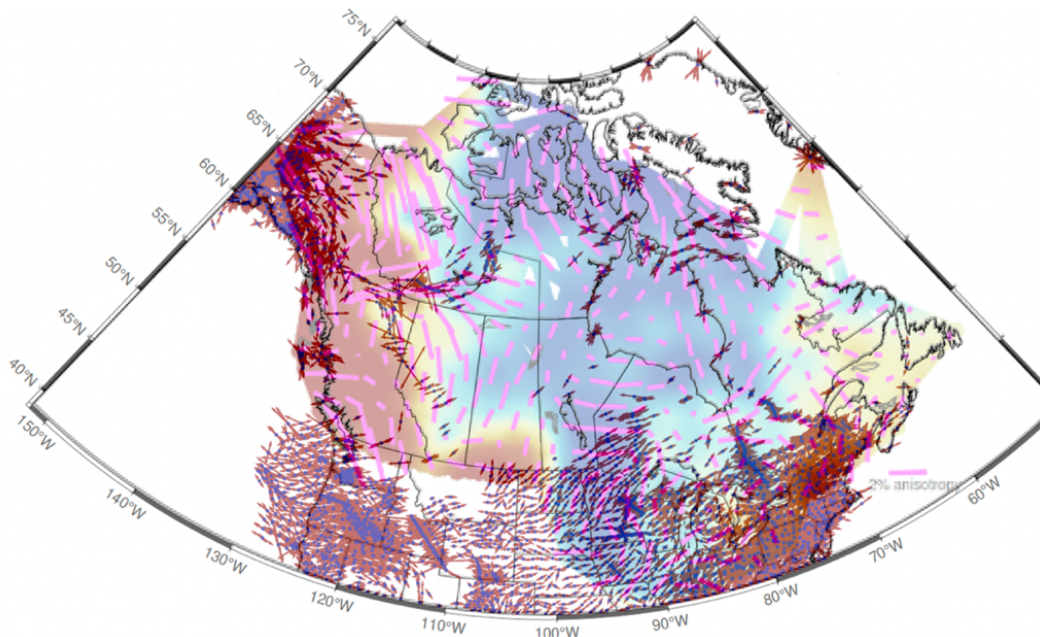


Figure 7.8. Phase velocity maps for 100s overlain with the shear wave splitting database measurements (IRIS DMC, 2012; Barruol et al., 2009).

7.3.2 Feature B Anisotropy

At short periods within Feature B (Figures 7.1 & 7.2) there is notable variability in the anisotropy orientation. At 30s, the anisotropy seems to follow the trends of the anisotropy of Feature A very closely, while at 50s, the anisotropy changes drastically, depicting an area of

NW-SE anisotropy, that rotates toward a more N-S orientation toward the southeastern edge of the highlighted region. Notably, at intermediate periods (Figures 7.3 & 7.4), the anisotropy weakens toward the middle of the feature, at the western margin. In this region, although the anisotropy is not extremely strong, its orientation deviates from the previously noted anisotropy trend of Feature A (associated with the slow velocities in this area). In fact, within the northwestern vicinity of the feature, very strong NE-SW trending anisotropy (values of approximately 4%), originating at the Cordillera-craton boundary, appears to overlap with the NW-SE trending anisotropy of the Cordillera. Generally, the mid to southern area of Feature B exhibits varying intensity of NW-SE trending anisotropy, while the northern region is far more complex as described above. Similar to the 50-second period map (Figure 7.2), the anisotropy becomes noticeably weaker close to the centre of the feature, at the western margin. At long periods (Figures 7.5 & 7.6), the anisotropy of the northern region seems to follow the trends of that noted within Feature A, while the anisotropy of the southern region appears to trend strongly in a NW-SE direction, aligning strongly with the anisotropy noted in Feature D (discussed in more detail in *sub-section 7.5*).

The anisotropy pattern described above shows similarity to that of Feature A at very short periods, suggesting that the recovered anisotropy is strongly terrane influenced. This changes at shallow upper mantle depths (around 50s period), likely explained by the frozen-in NW-SE trending tectonic fabric of the Intermontane, Omineca and Foreland Belts, originating from their island arc settings, i.e. trench-parallel anisotropy. Within the Stikinia terrane of the Intermontane Belt, to the northeast and east of this terrane, subduction associated with volcanism led to the closure of the Cache Creek ocean (Gabrielse and Yorath, 1991), though the anisotropy across this region remains consistent. This large-scale coherency of the anisotropy orientation observed across upper mantle to asthenospheric depths within this

region of the Cordillera is indicative of a history of consistent deformation of the lithosphere and the underlying asthenosphere. The overlapping anisotropy at the northeastern apex of the feature makes interpretation complicated. However, given the depth sensitivities associated with moderate to long periods (Table 7.1), it is possible that some of the anisotropy in the northern region could be caused by perturbations in mantle flow patterns as the lithosphere thickens abruptly to the east, similar to the keel diverted mantle-flow phenomenon occurring in the southeastern Ontario region (Fouch et al., 2000). Furthermore, the weakened anisotropy observed along the west of the feature at various periods may potentially be explained by the competing effects of fossil anisotropy within the lithosphere and the anisotropy produced as a result of present-day plate motion and basal shear, which can create a cancellation effect (Chen et al., 2021). The anisotropy exhibited in the southern region remains relatively consistent across intermediate to long periods, and is coherent with the orientation of the anisotropy noted in Feature D, at similar depth sensitivities, which likely suggests that the anisotropy here is most likely due to active mantle convective flow. Notably, the shear-wave splitting measurements across Feature B (Figures 7.3 & 7.4), were minimal, so were not compared to the anisotropy.

7.4 Cordillera-craton Boundary.

Feature C (Figures 7.1-7.6) corresponds to the Cordillera-craton boundary which characterizes the transition between the thin, slow lithosphere of the Cordillera and the thick, fast lithosphere of the craton. The boundary changes between each period, as its location was chosen by the exceeding of a velocity threshold ($> -2\%$) to the east. Per the 30 and 50-second map (Figures 7.1 & 7.2), it is at short periods not a very sharp boundary, with the transition between the slow

and fast velocities occurring across more than 200 km, though this sharpness increases as depth increases (to 100s period), with the transition occurring over 120 km. This corroborates the findings of Bao and Eaton (2015) which found a “remarkably abrupt” Cordillera-craton transition. Of importance is the variation of sharpness along the boundary at the same period. Due to the moderate to low velocities appearing in Alberta, the exact location of the boundary, and consequently the transition of slow to fast velocity, is unclear. The resolution at this depth is quite robust, where this gradational boundary is noted, indicating that this occurrence is not a smoothing artefact. This phenomenon dissipates at periods of 90 seconds and above, so the location of the southernmost region of the boundary can be ascertained. It becomes clear then, that the most prominently sharp transition from slow to fast velocity occurs at the northern Cordillera-craton boundary. Interestingly, the boundary changes its shape across the various periods, culminating in a very defined, but irregular boundary at 100 seconds (Figure 7.4), which is considered to be most representative as the raypath coverage, and as a result, the resolution, is highest at this period. The shape of the boundary at this period also closely resembles the results obtained by the S-wave velocity model of Schaeffer and Lebedev (2014) in the 150 km slice, which is indicative of the robustness of this recovered model. The Cordillera-craton boundary is still definitely discernible at long periods (up to 150s), which implies that this boundary persists to asthenospheric depths of >300 km, which corroborates the findings of Zaporozan (2017).

The structure of this boundary has significant implications for the origin of the Cordillera and the deformation patterns within, as the Cordillera-craton boundary (CCB) has undergone a minimum of two distinct evolutionary stages – the building through accretion of terranes and lithospheric removal resulting in the current sharp CCB. This is thought to have been achieved via regional lithospheric-scale delamination (Bao et al., 2014). Generally, this boundary

correlates strongly with the Cordilleran Deformation Front (CDF), aligning closely in most places, with some exceptions. The northern region of the boundary, where low velocity transitions to high velocity, is much sharper compared to the south. At upper mantle depths, the mid to northern extent of the Cordillera-craton boundary appears to align more closely with the Tintina Fault, rather than the CDF, or somewhere in the vicinity between the two (McLellan et al., 2018) (Figure 7.9). In the north, high velocities are recovered west of the CDF, a feature also observed in previous studies (e.g., Schaeffer and Lebedev, 2014), where it was attributed to the potential presence of the “Mackenzie Craton” – an Archean-aged fragment with no surface exposure, but instead, geophysical and geochemical evidence for its existence. According to Schaeffer and Lebedev (2014), the westward surface expression of the Rocky Mountain Front in this region of the northern Cordillera-craton boundary is indicative of a mechanically strong lithosphere east of the Rocky Mountain Front. The sharp boundary here, in conjunction with the high seismic velocities to the east of the Rocky Mountain Front, supports the presence of the “Mackenzie Craton”, which these results corroborate. Additionally, results obtained by Estève et al. (2021) put forward the theory that the Tintina Fault (Figure 7.9) in the vicinity of the northern Cordillera-craton boundary is a lithospheric-scale shear zone, which effectively cuts through the mantle resulting in large-scale lateral displacement of cratonic mantle material, which directly impacts the location and sharpness of the Cordillera-craton boundary in this region. A lithospheric-scale shear zone would also produce anisotropy aligned with the shear zone, consistent with these results.

South of this, at approximately 57°N, as depth increases at moderate to long periods, the shape of the boundary, where the velocity transitions from slow to fast, changes, and the Cordillera-craton boundary becomes more complicated to discern. Chen et al. (2019) put forward evidence for the determination of the location of the southern Cordillera-craton boundary within the

upper mantle, that was adapted for two possible hypotheses for the origin of the Cordillera. They propose a sharp and complex boundary which is interpretable as a reworked craton margin in the case of the accretionary model, or a collisional boundary occurring in the Late Cretaceous, in the case of the collision model (Chen et al., 2019). Furthermore, the velocity of the deeper upper mantle becomes fast east of the boundary, approaching craton-like velocities, where at shorter periods it was previously slow. This is most notably observed at 100s period. It follows then, in accordance with the findings of Chen et al. (2019), that the observed less-slow velocities east of the CCB, can be attributed to the presence of reworked cratonic material, providing an indication that the eastern Cordillera overlays cool lithosphere associated with continuation of the craton to the west, beneath the Cordillera. These results also appear to be generally consistent with the westward dip direction of the CCB noted by Chen et al. (2019).

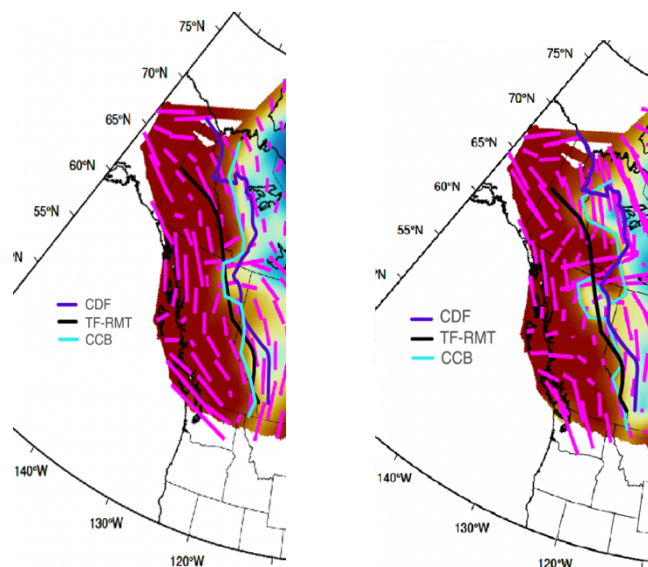


Figure 7.9. Zoomed-in map of the Cordillera, showing the Cordillera-craton boundary (CCB), overlain with the Cordilleran Deformation Front (CDF) and the Tintina Fault-Rocky Mountain Trench (TF-RMT), for periods 75s (left) and 100s (right).

7.5 Alberta and Southern Saskatchewan Velocity Anomaly.

Feature D (Figures 7.1-7.6) extends across most of Alberta, exhibiting an expansive region of slightly low velocity, between 1-2.5% faster than the Cordillera, which transitions to a very

low velocity anomaly toward southern Alberta, extending eastward through southern Saskatchewan. This southern transition is the most novel observation in this thesis. At short periods (30 and 50-seconds), Feature D (Figures 7.1 & 7.2), is initially relatively fast at the northwestern vertex of the feature, which transitions to moderately low velocities at 50s period, though the anisotropy is coherent across both periods. The anisotropy appears to be largely consistent across all of Alberta, trending in an NW-SE direction. The anisotropy orientation approaches an almost N-S alignment moving towards southeastern Alberta as the velocities become lower. Across the entirety of the eastern neighbouring province, Saskatchewan, the anisotropy trends in a NE-SW orientation, which rotates to an almost N-S direction, moving northwest to southeast across the province. The anisotropy patterns across both Alberta and Saskatchewan converge along the southern area of the border between the two provinces, within the vicinity of the lowest velocities of Feature D.

As intermediate periods (Figures 7.3 & 7.4), while the orientation of the anisotropy is largely the same as described above at the shorter periods, the point at which the anisotropy converges moves toward the southwestern corner of Alberta. Importantly, the slightly low velocities dissipate in central Alberta, becoming slightly high velocities, and the transition between low and high velocities in southern Saskatchewan becomes noticeably sharper. The anisotropy in this period range in the region of southern Saskatchewan becomes weaker, and the orientation rotates from that of an almost N-S direction to an E-W direction to the west and an abrupt change to almost N-S to the east.

At long periods (Figures 7.5 & 7.6), low velocities once again persist at the centre of the feature where at shorter periods it was relatively fast. To the north however, velocities in this region are now fast, where they were previously slow. At the 150s period, the northern vertex of the

feature once again exhibits low velocity, but a rather low velocity anomaly, up to -3%. Just south of this remains fast. The anisotropy at 125s across the entire feature is quite weak, with the exception of the northwestern vertex. Despite the weak anisotropy, the orientation remains largely consistent with previous periods. For the 150s period, the anisotropy becomes quite pronounced once again in the southern Saskatchewan area, and the northwestern region remains strong.

The low velocities appearing relatively shallowly (at very short to short periods) across the entire region can be largely attributed to the basins contained therein; namely the Foreland Basin within Alberta (with a basin thickness ranging from 4-6 km) and the Williston Basin (with a basin thickness of 5 km) located in southeastern Saskatchewan. The geometry of the Alberta Foreland Basin takes the form of a retroarc foreland basin, forming behind the mountain belt. The continued basin evolution due to continued mountain building resulted in continued incorporation of early foreland basin strata into the mountain front (Leckie and Smith, 1992; Cant and Stockmal, 1989).

This low velocity anomaly extending into southern Saskatchewan has been noted before in global models (e.g., Schaeffer and Lebedev, 2013; Moulik and Ekström, 2014), with much lower resolution, with not quite as low velocities. This model captures the anomaly more robustly – sharpening its boundaries as well as its location. When compared to the global models, the anomaly recovered here extends further north than previously observed, though its width remains relatively unchanged. Schaeffer and Lebedev (2013) note the relatively low velocity anomaly below Saskatchewan across their 100-200 km depth slices, which they attribute to reworked or altered Archean lithosphere, related to regional metamorphism during the Trans-Hudson Orogen (THO). The anomaly noted in this model is much lower velocity

(similar velocities to that noted across the Cordillera), which cannot be explained fully by reworked Archean lithosphere. Another potential possibility is put forward by Németh et al. (2005) who propose a lithospheric tectonic model of the THO from seismic refraction wide-angle reflection studies, which provides evidence of mantle suturing below the Sask craton trending in a NW-SE direction towards the Superior craton (Figure 7.10). This study samples the upper mantle via the arrivals of mantle phases with sufficient signal penetration (Németh et al., 2005). The presence of this mantle suture could explain the low velocity anomaly existing below southern Saskatchewan, at upper mantle depths. This mantle suture is indicative of the initiation of subduction of the microcontinent beneath the Flin-Flon Glennie complex (Figure 7.10), which could produce the very low velocities noted. The mantle suturing associated with the collision of the Wyoming craton and the Medicine Hat Block as described by Whitmeyer and Karlstrom (2007) may also be responsible for the low velocities prevalent in southwestern Saskatchewan. Another plausible explanation for this low velocity feature could be one similar to that of the well-documented significant lithospheric erosion of the Wyoming craton (e.g., Whitmeyer and Karlstrom, 2007; Dave and Li, 2016; Foley, 2008), though it is unclear whether this phenomenon could extend into southern Alberta and Saskatchewan, as this noted erosion is associated with the Yellowstone hotspot. It is also possible that there may be some lithospheric modification associated with the hotspot in this region that could result in these very low velocities, but there is no evidence supporting this hypothesis. It is somewhat plausible then, that if there is lithospheric erosion, this could create space to allow Cordilleran mantle to have followed eastward. Interestingly, Jones et al. (2014) provide Canada-wide hydration maps which show a large zone of increased hydration in the vicinity of the Saskatchewan low velocity zone (Figure 7.11), which could contribute to the very low velocities noted here. It should be noted though, that the Canada-wide hydration map does not agree with the seismic results everywhere.

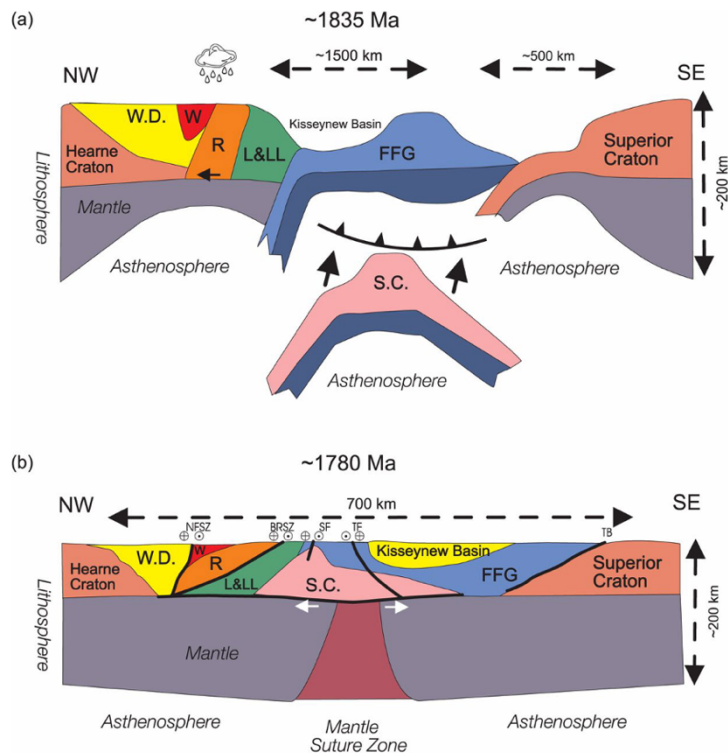


Figure 7.10. Schematic of the evolution of the lithospheric tectonic model beneath the southern THO, with (a) depicting before final collision, and (b) showing after final deformation, as well as the location of the Mantle Suture Zone. S.C.- Sask craton, FFG- Flin Flon-Gennie complex, BRSZ- Birch Rapids shear zone, L&LL- La Ronge-Lynn Lake complex, NFSZ- Needle Falls shear zone, R- Rottenstone domain, SF- Stanley Fault, TB- Thompson Belt, TF- Tabbornor Fault, W- Wathaman batholith, W.D.- Wollaston Domain (taken from Németh et al., 2005).

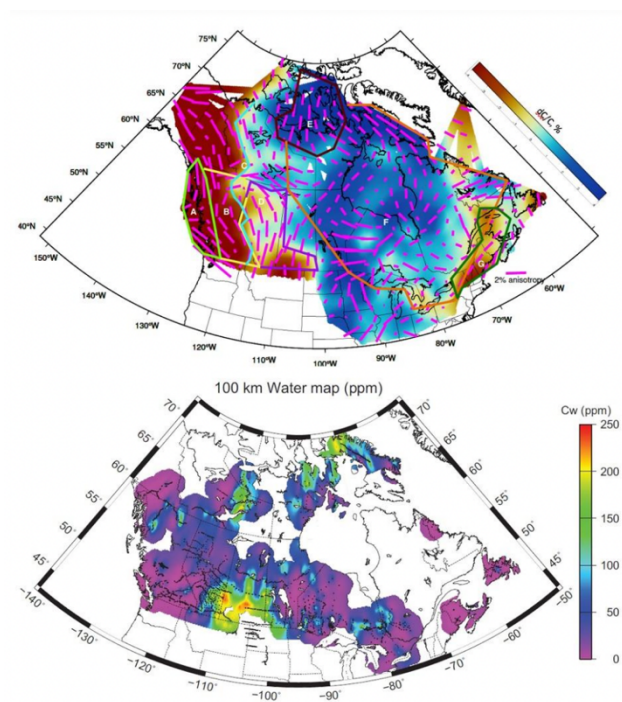


Figure 7.11. Comparison of 75s seismic velocity response and MT-derived hydration map for 100 km depth (Jones et al. 2014).

At short periods, the western anisotropy appears to follow the orientation of the thrust belts, while the eastern anisotropy would follow that of the deformed strata resulting from the THO collision. At increased periods, the relatively low velocities across Alberta dissipate, with high

velocities taking their place, with the sole exception of low velocities being maintained at the western provincial border along the Canadian Rockies. The anisotropy direction though, while largely remaining the same, shifts further to the west, and becomes noticeably weaker along the southern region, before becoming pronounced again at deeper mantle depths. Németh et al. (2005) suggest that anisotropy in this southern Saskatchewan region is likely the result of alignment of olivine associated with “frozen-in” fabric, parallel to the most recent plastic flow, thought to be due to an approximately NNW-SSE extensional deformation within the region. They do note that the anisotropy patterns could have been slightly disturbed due to the tectonic collisions which resulted in the suture.

7.6 Nunavut Velocity Anomaly and Associated Anisotropy.

Feature E (Figures 7.1-7.6) is a region which exhibits very high velocities and highly consistent N-S anisotropy across the entirety of the feature, though this phenomenon does not present at the shortest period being considered. The feature extends across most of northwestern Nunavut, maintaining the very high velocities of this anomaly (+4%) across periods 50s to 150s, from upper lithosphere to sublithospheric mantle. While these high velocities are maintained at depth, the anisotropy becomes slightly weaker, and the orientation of the anisotropy changes. At intermediate periods, the anisotropy orientation still reflects a relatively N-S orientation in the northwestern region, but the southeastern region now exhibits NW-SE trending anisotropy. At long periods, specifically 125s (Figure 7.5), the anisotropy is very weak to the south of the feature, but the northern anisotropy trends in an almost E-W orientation. At 150s (Figure 7.6), the anisotropy remains weak to the east of the feature, but is oriented in a NW-SE direction in the western region.

The very high velocities across Feature E are largely attributed to the very old cratonic lithosphere in this region – namely the Slave craton (Chen et al., 2007) and the northwestern Rae craton (Yuan et al., 2011), both >2.0 Ga. The Thelon Tectonic Zone (TTZ) is a 2.0-1.8 Ga magmatic arc, which separates the Slave and the northwestern Rae cratons. The anisotropy within this region (the Slave, northwestern Rae, and the TTZ) is thought to be representative of layered anisotropy, as opposed to homogeneous fabric, determined via SKS splitting analysis (Snyder and Bruneton, 2007). The anisotropy occurring at relatively shallow depths (above 120 km; in the period range of 30 to 50s) in the western region of Feature E is largely associated with the large-scale folding observed at the surface as described by Snyder and Bruneton (2007), while to the east, the shallow anisotropy can be attributed to compression along the convergent margin of the Rae craton (Snyder et al., 2013).

Beyond this, at depths of > 150 km, the orientation of the anisotropy across Feature E becomes slightly less consistent, with the deviations seemingly separated by the TTZ. The anisotropy at mantle depths (at the 125s period), for the Rae craton appears to agree with the findings of Snyder et al. (2013), which describes E-W anisotropy aligning with the plate motion. Slightly deeper at the 150s period (Figure 7.6), the anisotropy rotates to an orientation almost perpendicular to that described at the 125s period (Figure 7.5). In the case of the Slave craton though (western Feature E), the anisotropy at depth appears to align with kimberlite eruptions associated with the Lac de Gras kimberlite field, which is a strong indicator that the anisotropy is partially related to the preferred fabric orientation of hypothesized dykes at depths of 200 km (Snyder and Bruneton, 2007).

7.7 Variations within the Shield.

Feature F (Figures 7.1-7.6) extends across most of the Canadian Shield, which is the geological core of North America, consisting of Precambrian crust. Generally, this region exhibits fast to very fast velocity perturbations of approximately +1% to +4% due to the presence of various very old cratonic provinces within the Shield (Bollman et al., 2019; Frederiksen et al., 2013; Petrescu et al., 2017; Schaeffer and Lebedev, 2014). Across the feature though, there are small scale anomalies in velocity as well as variations in the strength and orientation of the recovered anisotropy, over relatively short length scales.

At crustal depths – periods 50s (Figures 7.1 & 7.2) and below – the areas of lower velocity along the eastern border of the Superior may be attributed to lithospheric thinning approaching the Appalachians. However, the area of lower velocity along the Churchill Province appears to correlate closely with the presence of sedimentary basins, namely the Foxe and Lancaster Sound Basins occurring in this region. Toward the south, the region of low velocity in southern Ontario correlates with the Keweenaw MCR/Great Lakes region. The low velocity region beneath Hudson Bay similarly corresponds to the Hudson Bay Basin. These lower velocities are also associated with the THO, which is most apparent at the 25-30s period range (Figure 7.1). The low velocity anomaly along the Quebec-Labrador border can be potentially linked to the variations in local or regional crustal thicknesses. Notably, all of the Superior craton at this depth exhibits fast velocities.

At intermediate (Figures 7.3 & 7.4) and longer (Figures 7.5 & 7.6) periods, only the western Superior continues to show extremely high velocities (e.g., Bollmann et al., 2019). The majority of the eastern Superior depicts velocities similar to the surrounding Proterozoic

orogens, which corroborates the findings of previous body-wave studies (e.g., Frederiksen et al., 2013), though the northeastern Superior remains relatively fast. The notable difference between the western and eastern Superior is associated with lithospheric root differences. The western Superior exhibits characteristics of a conventional cratonic root, whereas the root of the eastern Superior is proposed to have undergone metasomatism to various extents, due to the Great Meteor Hotspot (Frederiksen et al., 2013). It follows then, that there does not appear to be a distinct systematic difference between Shield components of different ages, across the entirety of Feature F, with Archean-aged Shield components to the east no longer exhibiting the very high velocities associated with cratonic lithosphere at depth, while the western regions (both Superior and Churchill) continue to exhibit fast cratonic velocities at depth. At periods corresponding to depths associated with the uppermost mantle, there is a strong distinction between the high velocities below the Superior craton and the relatively lower velocities linked to the Trans Hudson Orogen and the MCR.

Similar to Feature E (*section 7.6*), the Superior Province appears to exhibit layered anisotropy (Darbyshire and Lebedev, 2009; Darbyshire et al., 2013; Petrescu et al., 2017). Shallowly, at upper lithospheric depths, corresponding to periods 75s through 100s (Figures 7.3 & 7.4), the western Superior depicts strong E-W anisotropy, the central Superior NE-SW anisotropy, while the eastern Superior exhibits both E-W and NW-SE anisotropy orientations. This observed variation can be explained by responses to tectonic stress during the accretion stage of the terranes which the Superior is comprised of, such as fracturing and layering of anisotropic minerals. At lithospheric depths (periods 50-125s), the strongest anisotropy is the E-W anisotropy observed below the western Superior, which Babuska and Plomerová (2006) propose can be explained by the layering of subducted oceanic lithosphere, within which the frozen anisotropy of the original plates persists. The shear wave splitting measurements

(Figures 7.7 & 7.8) corresponding to the western Superior generally agree with the surface wave anisotropy findings. Furthermore, the splitting measurements largely match the pattern of variations in anisotropy observed in the central and eastern Superior at the 50s and 60s periods. In the central and eastern Superior, the terrane boundaries are relatively more complex (compared to the west), which explains the shifting between strong and weak anisotropy across this region of the Canadian Shield. The anisotropy across the central-eastern Superior at lower lithospheric depths (Figures 7.3 & 7.4) is thought to have been somewhat compromised by the presence of large volumes of melt material related to plume activity, which would have disrupted the aligned olivine in the lithosphere, weakening the anisotropy in this region (Foster et al., 2020).

According to Foster et al. (2020), in the vicinity of the activity of the Great Meteor hotspot, the lithosphere appears to have undergone modification, by effectively lowering seismic phase velocities, and disrupting the anisotropy. The recovered results are consistent with this theory, as the azimuthal anisotropy is weaker and generally correlates with the orientation and directionality of the terrane boundaries within the craton, and the phase velocity is estimated to be >2% lower than the western Superior at lithospheric depths due to this lithospheric modification. Notably, the shear wave splitting measurements in this area (Figures 7.7 & 7.8), do not correlate, which can be explained by the differences in the depth sensitivity of both methods, where the Rayleigh wave results are more sensitive to lithospheric depths, while the shear wave splitting measurements are sensitive to the cumulative effect of both lithospheric and asthenospheric anisotropy. It is very likely that frozen fabrics have a major impact on the split measurements, given the thick lithosphere (approximately 200 km).

The Churchill Province exhibits very weak, heterogenous shallow anisotropy, which is not specifically interpreted. At depth however, the anisotropy is much stronger, with NW-SE fast orientations, which can be linked to Rae underthrusting the Hearne, as well as convergence due to northwestern thrusting along the Chesterfield fault zone west of Hudson Bay (Berman et al., 2007). This convergence phenomenon, present in this region, implies the presence of Rae terranes within the Hearne (Snyder et al., 2013).

7.8 The Appalachians Velocity Anomaly and Associated Anisotropy.

Feature G (Figures 7.1-7.6) extends across the Appalachian region, defined by a low velocity zone. The nature of this low velocity zone is age-dependent as the orogenic lithosphere cools over time – the Appalachians are younger than the Shield, but older than the Cordillera. As such, the Appalachians exhibit higher heat flow than the Shield (Mareschal et al., 1989), corresponding to low velocities. At shallow depths (periods 50s and below), there is a clear distinction between the cratonic lithosphere and the Appalachian region. Notably, the lithosphere in this region has been reported to be very thin – between approximately 80 km (Evans et al., 2019) and 100 km (van der Lee, 2002). Along the southern region of Feature G, there exists a very low velocity zone, which also persists to asthenospheric depths. It has been put forward by Menke et al. (2016) that this low velocity zone is not due to the Great Meteor Hotspot track that cuts across it, but rather due to a Northern Appalachian Anomaly (NAA) linked to asthenospheric upwelling, more specifically, an eddy in the flow field occurring at the continental margin.

The orientations of the anisotropy at both lithospheric and asthenospheric depths are very similar, though the deeper anisotropy is slightly stronger. The northern region exhibits NW-SE orientations, while the southern region exhibits broadly N-S orientations. Additionally, the shear wave splitting measurements (Figures 7.7 & 7.8) are not very coherent with the anisotropy orientation recovered. Work by Chen et al. (2018) also finds shear wave splitting results in this region that align with Figures 7.7 and 7.8., from which they conclude a multi-layered anisotropic structure. The noted inconsistency with the recovered anisotropy and the splitting measurements is likely due to the cumulative nature of splitting measurements. Barruol et al. (1997) suggest that the anisotropy may be explained by the fabrics maintained during tectonic accretion of the region, which these findings corroborate.

7.9 Summary.

The resolution tests (see *Chapter 5*) indicate that this tomography model of Canada is well resolved at the lithospheric scale, with features as small as 400 km being recoverable. In areas where the ray path coverage is less dense there is some smearing, as expected. Smearing is strongest at the model edges and in regions where there is lower raypath density, such as the Prairie provinces, specifically northern Saskatchewan, while smearing is weakest in regions where the ray path density is high, such as across the Superior craton and the Cordillera. The anisotropy results are quite robust as well, as is evident from the leakage resolution tests which depict leakage only at the model edges, and “reversed” anisotropy resolution tests which indicate no apparent azimuthal bias.

The prominent low velocity anomaly occurring across the Cordillera features two distinct regions of homogeneous anisotropy in the vicinity of the southern Cordillera. This difference

is interpreted as a response to the influence of the subduction zone associated with the Juan de Fuca slab in the west. Additionally, the sharp transition from the slow Cordillera to fast cratonic lithosphere is captured quite well at short, intermediate and long periods. The rapid change in the shape of this Cordillera-craton boundary at increasing depths is noteworthy. At the highest resolution period (100s), the boundary is noted to be most irregular, changing direction drastically, especially along the northern and central region of the boundary.

The most novel velocity feature recovered is the low velocity anomaly detected in southern Saskatchewan, which has been detected before in global models at much lower resolutions. This model indicates that the anomaly extends further northward than previously thought, with much sharper boundaries. Adjacent to this low velocity anomaly is a moderately low velocity region across the majority of Alberta. At the shortest periods, these low velocities are attributed to the presence of sedimentary basins across this expanse, namely the Alberta Foreland Basin and the Williston Basin in southeastern Saskatchewan. The low velocity anomaly occurring in Saskatchewan persists to mantle depths, though this is not the case in Alberta. This persistent low velocity is interpreted to be linked to possible lithospheric erosion which would create space for Cordilleran mantle to move eastward, or alternatively, the mantle suture beneath the southern THO.

Very high velocities were noted in northern Nunavut, exhibiting consistent N-S trending anisotropy shallowly, which is not maintained across the entire region at depth. The orientation of the anisotropy becomes slightly less consistent at periods corresponding to mantle depths, seemingly separated by the TTZ. The anisotropy here supports previous findings concluding layered anisotropy.

Several small-scale anomalies were noted across the Canadian Shield, such as the correlation of lower velocity regions with sedimentary basins. Additionally, the low velocity anomaly observed in southern Ontario correlates with the Keweenawan MCR/Great Lakes region. Most of the eastern Superior exhibits velocities similar to those of the surrounding Proterozoic orogens, while the western Superior remains quite fast. The root of the eastern Superior appears to have undergone metasomatization to varying degrees, due to the influence of the Great Meteor Hotspot, accounting for the lower velocities in this area.

The Appalachians are defined by a low velocity zone, which is characteristic of orogenic belts. To the south of this region, a very low velocity zone is observed that extends to asthenospheric depths. Although there are two possible interpretations of this feature, this low velocity anomaly is likely not due to the Great Meteor Hotspot track that geographically coincides with the anomaly, but is instead due to the NAA associated with asthenospheric upwelling at the continental margin.

This work presents a tomography model of Canada with greater resolution than that of previous global models, which corroborates the findings of several previously conducted geophysical studies. Optimally, accompanying depth inversions through these aforementioned notable anomalies would have aided with the interpretation process.

8. CONCLUSION.

8.1 Conclusions.

The aim of this study was to examine the subsurface of Canada as a whole, rather than in isolated segments, at the lithospheric scale, by integrating new data with existing data. New dispersion curves were obtained via the CNSN and USArray (TA) stations to populate regions of sparse ray path coverage, thereby increasing overall resolution. Tomographic inversion of this collated dataset was carried out using the approach described by Darbyshire and Lebedev (2009), for both isotropic velocity and anisotropy. Seven major, distinct features were observed across the maps which were interpreted across several periods.

The model shows a low velocity anomaly in the Cordillera, with distinct regions of consistent anisotropy to the south. This deviation in anisotropy is interpreted to be due to the influence of the subduction of the Juan de Fuca slab to the west. The highest resolution map (100s period) highlights a very irregular Cordillera-craton boundary that displays a rapid transition from the slow Cordillera to fast cratonic lithosphere. Though the shape of this boundary changes across different periods, the boundary itself persists to sublithospheric mantle depths. The low velocities of the southern Cordillera appear to extend to the east across Alberta and Saskatchewan – where there exists a very low velocity anomaly in southeastern Saskatchewan, which is interpreted to be as a result of a mantle suture in this region.

In northern Nunavut, the maps show a very high velocity anomaly with very consistent N-S trending anisotropy, that transitions to two distinct regions of consistent anisotropy, seemingly separated by the TTZ. Across the Canadian Shield, although the velocity recovered here ranges from fast to very fast velocity, there appears to be a correlation between lower velocity regions

and the occurrence of sedimentary basins. There is a clear distinction between the Shield and the Appalachians, defined by a rapid transition in velocity from fast to slow, respectively. In the south of the Appalachian region, a very low velocity anomaly is noted, which is interpreted to be due to the NAA, despite its intersection with the Great Meteor Hotspot track.

8.2 Recommendations.

1. Include more station paths across central Canada to further increase the resolution of regions where resolution completely breaks down - along the Saskatchewan-Manitoba border, the northern Cordillera, and the Northwest Territories-Nunavut border. It may be necessary to install more seismic stations to increase the resolution in some areas.
2. More detailed interpretations can be achieved via a depth inversion of this tomography model.

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APPENDIX.

Station Pairs and Number of Events

Table 10.1. List of station pairs and number of events for new dispersion curves.

First Station	Second Station	Number of Events
BBB	FFC	181
BMBC	FCC	185
BMBC	DAWY	70
BNAB	DLBC	12
BNAB	FFC	181
BPEB	FFC	187
BTB	FFC	195
CBB	FFC	87
DAWY	EDB	198
DAWY	FNSB	179
DLBC	FFC	184
DLBC	FSJB	183
DLBC	HOPB	182
DLBC	LLLB	198
DLBC	SHB	91
DLBC	SLEB	124
DLBC	UBRB	189
DLBC	VDEB	119
DLBC	WSLR	115
Continued on next page		

EDB	FNSB	177
FFC	A13A	188
FFC	A15A	157
FFC	A16A	165
FFC	B14A	150
FFC	B17A	158
FNBB	A21A	166
FNBB	A22A	173
FNBB	A23A	184
FNBB	A24A	172
FNBB	A25A	196
FNBB	A26A	198
FNBB	A27A	198
FNBB	A28A	199
FNBB	A29A	199
FNBB	B21A	180
FNBB	B22A	178
FNBB	B23A	156
FNBB	B25A	198
FNBB	B26A	184
FNBB	B27A	183
FNBB	B28A	200
FNBB	B29A	199
Continued on next page		

JEDB	FFC	181
LLLB	FFC	176
LZB	FFC	153
MAYB	FFC	166
MGMB	FFC	174
MGRB	FFC	195
MYRA	FFC	73
NLLB	FFC	182
PCLB	FFC	160
PTRF	FFC	173
PNT	FFC	174
RUBB	FFC	156
SHB	FFC	197
SLEB	FFC	121
SNB	FFC	177
SYMB	FFC	193
UBRB	FFC	179
ULM	A11A	156
ULM	A12A	159
ULM	A13A	162
ULM	A14A	160
ULM	A15A	149
ULM	A16A	147
Continued on next page		

ULM	B13A	159
ULM	B14A	153
ULM	B15A	156
ULM	B17A	116
ULM	VDEB	171
VDEB	FFC	180
VGZ	FFC	196
WALA	FFC	182
WALA	YKW3	176
WPB	FFC	190
WSLR	FFC	199
YKW2	A20A	79
YKW2	A25A	104
YKW2	A29A	141
YKW3	A07A	178
YKW3	A08A	139
YKW3	A09A	183
YKW3	A10A	151
YKW3	A11A	163
YKW3	A12A	169
YKW3	A13A	166
YKW3	A14A	184
YKW3	A15A	197
Continued on next page		

YKW3	A18A	177
YKW3	A19A	170
YKW3	A20A	190
YKW3	A21A	173
YKW3	A22A	166
YKW3	A23A	164
YKW3	A24A	172
YKW3	A25A	184
YKW3	A26A	197
YKW3	A27A	196
YKW3	A28A	181
YKW3	B07A	154
YKW3	B08A	138
YKW3	B09A	164
YKW3	B10A	182
YKW3	B11A	161
YKW3	B12A	192
YKW3	B13A	174
YKW3	B14A	173
YKW3	B19A	175
YKW3	B20A	184
YKW3	B21A	184
YKW3	B22A	179
Continued on next page		

YKW3	B23A	173
YKW3	B25A	186
YKW3	B26A	174
YKW3	B27A	200
YKW3	B28A	184
YOUB	FFC	178

Dispersion Curves.

Dispersion curve measurements for all station pair paths are included in the electronic supplemental file all-curves.zip. Each dispersion curve is listed in the following format:

STA1 STA2	(names of stations)
LHZ LHZ	(not important, but must be there)
lat1 lon1 lat2 lon2	(lats & longs of stations)
period(s) phasevel(km/s) error(km/s)	(phase velocity data)
period(s) phasevel(km/s) error(km/s)	
period(s) phasevel(km/s) error(km/s)	