

The Proterozoic building of North America: insights from broadband seismic tomography

PhD Thesis

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Declaration

Declaration of Originality

This dissertation is submitted for the degree of Doctor of Philosophy at Imperial College London and Associateship of the Royal School of Mines. Unless explicitly stated in the text, this thesis is the result of my own work.

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Publications arising from this thesis

Chapter 3 Boyce, A., I. D. Bastow, F. A. Darbyshire, A. G. Ellwood, A. Gilligan, V. Levin, and W. Menke (2016), Subduction beneath Laurentia modified the eastern North American cratonic edge: Evidence from P wave and S wave tomography, *J. Geophys. Res.*, *121*(7), 5013–5030, doi: 10.1002/2016JB012838

Chapter 4 Boyce, A., I. D. Bastow, S. Rondenay, and R. Van der Hilst (2017), From relative to absolute teleseismic travel-times: the Absolute Arrival-time Recovery Method (AARM), *Bull. Seis. Soc. Am.*, *107*(5), 2511–2520, doi: 10.1785/0120170021

Gilligan, A., I. D. Bastow, A. Boyce, L. Petrescu, M. Liddell, F. A. Darbyshire, D. Hawthorn, V. Lane, D. Daly, D. Simpson, and D. Heffler (2016a), Peering beneath the Canadian crust, *Astron. Geophys.*, *57*(6), 6.24–6.27, doi: 10.1093/astrogeo/atw221

Bastow, I., A. Boyce, E. Caunt, J. Guilloud De Courbeville, S. Desai, R. Kounoudis, E. Golos, S. Burdick, and R. van der Hilst (2017), African Cenozoic hotpot tectonism: new insights from continent-scale body-wave tomography, *AGU Fall Meeting Abstracts*

Notations, Definitions and Conventions

Symbols

V_p P-wave velocity

V_s S-wave velocity

Δ Epicentral distance

m_b Body-wave earthquake magnitude

Abbreviations

AF Appalachian Front

BB Broadband

CNSN Canadian National Seismograph Network

Ga Giga annum

GF Grenville Front

GMT Generic Mapping Tools

LAB Lithosphere-Asthenosphere Boundary

Ma Mega annum

MLD Mid-Lithospheric Discontinuity

MCCC Multichannel Cross Correlation

POLARIS Portable Observatories for Lithospheric Analysis and Research Investigating Seismicity

QM-III Québec-Maine Across Three Sutures

RMS Root Mean Square

SAC The Seismic Analysis Code

SNR Signal-to-Noise Ratio

SP Short Period

TA Transportable Array Seismograph Network

THO Trans-Hudson Orogen

Conventions

Backazimuth The angle in degrees measured clockwise from north from a seismic station to the great circle arc between a seismic station and an earthquake.

Slowness, p The reciprocal of the apparent velocity of a wave (horizontal slowness), equivalent to the ray parameter p

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‘Never let the facts get in the way of a good story’

Anon

Abstract

Cratons, the Archean ($\geq 2.5\text{ Ga}$) cores of the continents are the longest-lived features of Earth's surface. They are underlain by cold, iron-depleted, thick ($> 250\text{ km}$), seismically fast lithospheric mantle roots or keels that have survived thermal and chemical erosion over multiple Wilson cycles. Recently however, numerous studies have shown that cratonic mantle lithosphere can undergo modification and may be removed entirely. I use passive-source seismology to investigate this further for the Superior Craton of North America and its abutting Proterozoic platform.

Initially, I use P- and S-wave relative arrival-time tomography to investigate the regional seismic structure of southeast Canada. Here, three broad zones of decreasing seismic wavespeed span the Archean Superior, Proterozoic Grenville and Phanerozoic Appalachian provinces, respectively. A vertical boundary in wavespeed beneath the Grenville Front is interpreted as evidence for subduction-driven metasomatic enrichment of the Laurentian margin.

Due to the loss of the background mean velocity structure, relative arrival-time datasets are not easily combined and are also unsuitable for estimation of the physical properties of the mantle. To address this, I develop the Absolute Arrival-time Recovery Method (AARM) to retrieve absolute arrival-times from temporary seismograph networks, whose data are often noisy. Tests indicate that AARM picks are accurate to $\leq 0.25\text{ s}$, akin to uncertainties in ISC bulletins. As a result, small aperture, temporary deployments now represent an exploitable resource with which to fill gaps in global seismic tomographic studies.

Finally, I incorporate multiple regional networks from eastern Canada into a continental P-wave absolute arrival-time tomographic inversion to produce the most up-to-date compressional wavespeed image of the North American lithosphere. These results show along-strike variability in upper lithospheric velocity structure within the Proterozoic Grenville Orogen ($1.3 - 1.0\text{ Ga}$). When reconciled with the geological record, these results may provide the first seismological evidence for delamination of subcontinental lithospheric mantle in a Proterozoic large, hot, orogen.

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Chapter 1

Introduction

1.1 Overview

‘Shields’, the Precambrian cores of the continents, are underlain by thick, cold, chemically depleted and therefore buoyant, lithospheric roots or keels (*Griffin et al.*, 2003; *Hawkesworth et al.*, 1990; *Jordan*, 1988). Existing models for keel formation generally invoke Archean processes, such as the extraction of komatiite magmas (e.g., *Griffin et al.*, 2003; *Hawkesworth et al.*, 1990; *Jordan*, 1988; *Pearson and Wittig*, 2008; *Pearson et al.*, 1995) to explain their intrinsically low density. However, keels may not be restricted to Archean domains (e.g., *Griffin et al.*, 2004; *Schaeffer and Lebedev*, 2014; *Simons et al.*, 2002; *Yuan et al.*, 2011) and may have also formed in multiple stages (*Aulbach*, 2012; *Darbyshire et al.*, 2013; *Yuan and Romanowicz*, 2010) in a variety of ways (e.g., *Gray and Pysklywec*, 2010; *Lee et al.*, 2011). Investigation has also shown keels may not have remained entirely tectonically stable since their formation and have been modified by subduction (e.g., *Pedersen et al.*, 2013; *Youssof et al.*, 2015), erosion (e.g., *van Wijk et al.*, 2010; *Villemaire et al.*, 2012), mantle flow (e.g., *Eaton and Frederiksen*, 2007; *Kaban et al.*, 2015) and lithospheric delamination (e.g., *Dave and Li*, 2016; *Gao et al.*, 2004; *Johnson et al.*, 2013; *Morency et al.*, 2002; *Rey and Houseman*, 2006; *Ueda et al.*, 2012), amongst other processes. Further, cratonic regions have been involved in numerous tectonic collisions and Wilson cycles, but the impact of protracted orogeny at cratonic margins is debated. Fundamental to resolving this issue is a better understanding of Precambrian lithospheres response to such collisions, and to what extent it is resistant to modification processes (*Chiarenzelli et al.*, 2010; *Poudjom-Djomani et al.*, 2001).

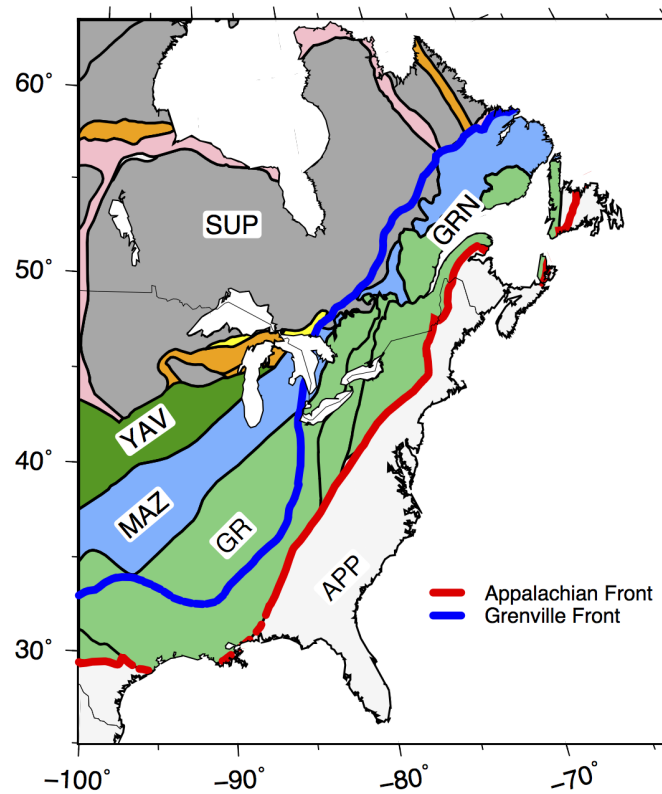


Figure 1.1: **Figure to show simplified Proterozoic geology of eastern North America** (adapted from *Whitmeyer and Karlstrom, 2007*). Archean rocks of the Superior (SUP) underthrust the Grenville Province (GRN) comprising Palaeoproterozoic Yavapai (YAV - 1.80 – 1.68 Ga), Mazatzal (MAZ - 1.70 – 1.60 Ga) and Granite-Rhyolite (GR - 1.55 – 1.35 Ga) material. Further to the southeast is the Appalachian terranes (APP). The Grenville (blue) and Appalachian (red) tectonic deformation fronts are shown as thick lines.

The geological record of eastern North America spans > 3 Ga (e.g., *Hatcher, 2010; Hoffman, 1988; Li et al., 2008c; Whitmeyer and Karlstrom, 2007*), including the Archean formation of the Superior province, the Proterozoic Grenville Orogen, and the Phanerozoic Appalachians (Figure 1.1). All three terranes were affected by hotspot tectonism $\sim 190 - 110$ Ma: the Great Meteor Hotspot (*Heaman and Kjarsgaard, 2000*). Today, the remnants of the Grenville orogen, which acted for an unusually-protracted period of time (1.3 – 1.0 Ga), extends ≥ 4000 km in eastern North America (Figure 1.1), far from the influence of present day active tectonics and hotspot activity. The region is thus an ideal natural laboratory for the study of Precambrian lithospheric formation and evolution, including the potential impact of protracted subduction-related metasomatism, and hotspot tectonism. These

issues can all be investigated seismically: cool mantle lithosphere depleted in heavy elements (e.g., iron), exhibits fast wavespeeds (e.g., *Duffy and Anderson, 1989; Jackson et al., 2005; Lee, 2003; Li et al., 2008a; Ritsema et al., 2011; Vacher et al., 1998*), but metasomatic modification will serve to reduce anomalies in seismic tomographic images (e.g., *O'Reilly and Griffin, 2006*).

Capitalising on new and existing seismograph networks deployed in southeast Canada (Figure 1.2), I first conduct relative arrival-time P- and S-wave regional tomographic inversions to image the seismological structure of the lithosphere. I investigate the implications for keel formation and the lithospheric impact of the Great Meteor hotspot across the three tectonic domains. Additionally, I focus on whether lithospheric variations in seismic velocity between Archean and Proterozoic terranes, are reconciled best with formation derived variation in chemical depletion or lithospheric modification processes during the Grenville Orogeny, prior to keel stabilisation.

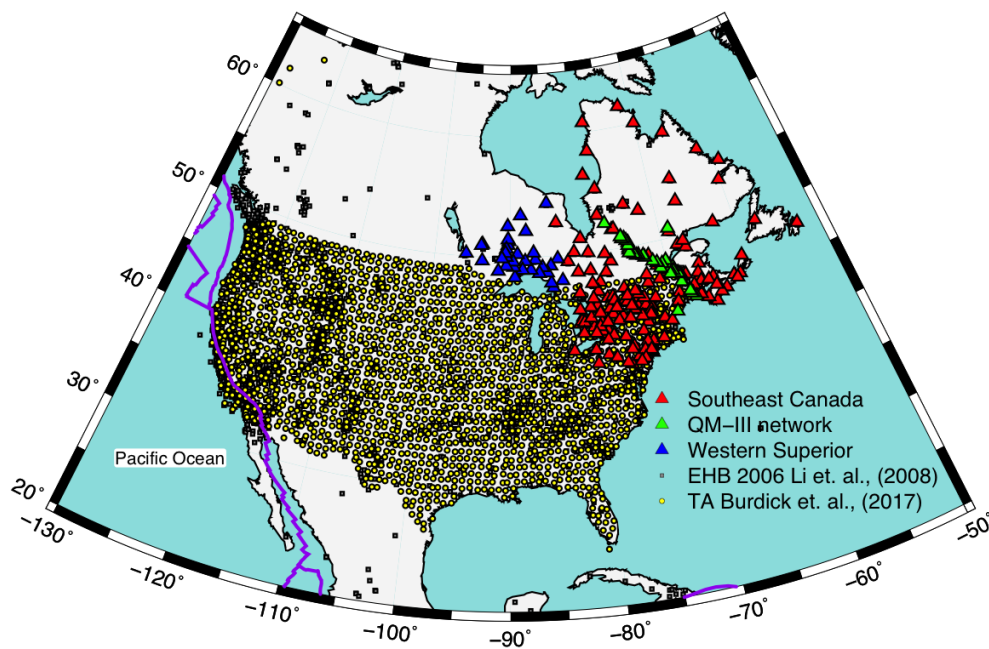


Figure 1.2: **Figure to show seismograph deployments in the U.S. and eastern Canada** for which datasets are freely available. In southeast Canada, datasets from predominantly temporary seismic deployments are available from the eastern Superior, Grenville and Appalachians (red triangles), western Superior (blue triangles) and the QM-III network (green triangles). In the U.S., numerous permanent and temporary seismograph deployments (e.g., *Li et al., 2008a*, - grey squares) have been supplemented by the extensive USArray TA network (e.g., *Burdick et al., 2017*, - yellow circles). Purple lines indicate active plate boundaries.

Whilst useful for conducting regional lithospheric studies (e.g., *Bastow et al., 2008; Frederiksen et al.,*

2013; *Rawlinson et al.*, 2006; *Rondenay et al.*, 2000; *VanDecar et al.*, 1995), relative arrival-time datasets cannot be combined with other datasets from the continent or across the globe. This is because analysis of often-noisy waveforms from temporary seismograph deployments and the subsequent calculation of relative arrival-times, irreversibly removes the background mean velocity structure (e.g., *Rawlinson and Kennett*, 2004; *VanDecar and Crosson*, 1990), and may vary dramatically region to region. Therefore, to combine seismograph networks from southeast Canada with larger datasets from the U.S. (e.g., the USArray - Figure 1.2), I develop a method to extract absolute arrival-times by recovering the background mean velocity structure, thus datasets can be combined on the continental-to-global scale.

Exploiting this technical advancement, I incorporate smaller temporary network deployments from southeast Canada into a continental scale absolute arrival-time inversion for North America. When combined with USArray data (Figure 1.2), this model provides the first P-wave tomographic model capable of imaging lithospheric heterogeneity along the entire Grenville Orogen. Using these results, I discuss whether the observed along strike heterogeneity within the Grenville Orogen (*Corrigan and Hanmer*, 1997; *McLelland et al.*, 2013; *Whitmeyer and Karlstrom*, 2007) can be explained by variations in the extent of lithospheric delamination during orogenesis. This model thus presents the opportunity to investigate whether the age of cratonic constituents controls how stable lithosphere evolves during protracted continental collision.

1.2 Cratons worldwide

Cratons are the longest surviving constituents of the Earth's continents (Figure 1.3). These coalesced Archean nuclei (prior to 2.5 Ga), often surrounded by younger Proterozoic mobile belts and collisional orogens, together form the Precambrian shields. As a consequence of their longevity, cratons are assumed to have resisted thermomechanical erosion throughout their history, over multiple Wilson cycles (e.g., *Griffin et al.*, 2009; *Hoffman*, 1990; *Lee et al.*, 2011). Paramount to the survival of cratons are their underlying lithospheric keels or roots. Keels are mechanically strong, highly viscous (perhaps $20 \times$ higher than normal mantle: *Burov*, 2011; *Cooper et al.*, 2004; *Lenardic and Moresi*, 1999; *Sleep*, 2003) and exhibit low heat flow (*Artemieva and Mooney*, 2001; *Davies*, 2013; *Mareschal and Jaupart*, 2004). Furthermore, they are frequently linked to Archean tectonic processes (e.g. the extraction of komatiite magmas) which depleted the mantle in heavier elements such as Al, Ca and Fe

(e.g., *Griffin et al.*, 2003; *Hawkesworth et al.*, 1990; *Jordan*, 1988; *Pearson et al.*, 1995; *Simon et al.*, 2007) resulting in low density and thus increased buoyancy. Mechanisms to achieve thickening of depleted cratonic lithosphere typically fall into three end-member models (e.g., *Gray and Pysklywec*, 2010; *Lee et al.*, 2011); the residue of melting above a plume (e.g., *Arndt et al.*, 2009; *Griffin et al.*, 2003), accretion/stacking of oceanic lithospheric slabs (e.g., *Pearson and Wittig*, 2008; *Simon et al.*, 2007) and collisions of segments of arc lithosphere.

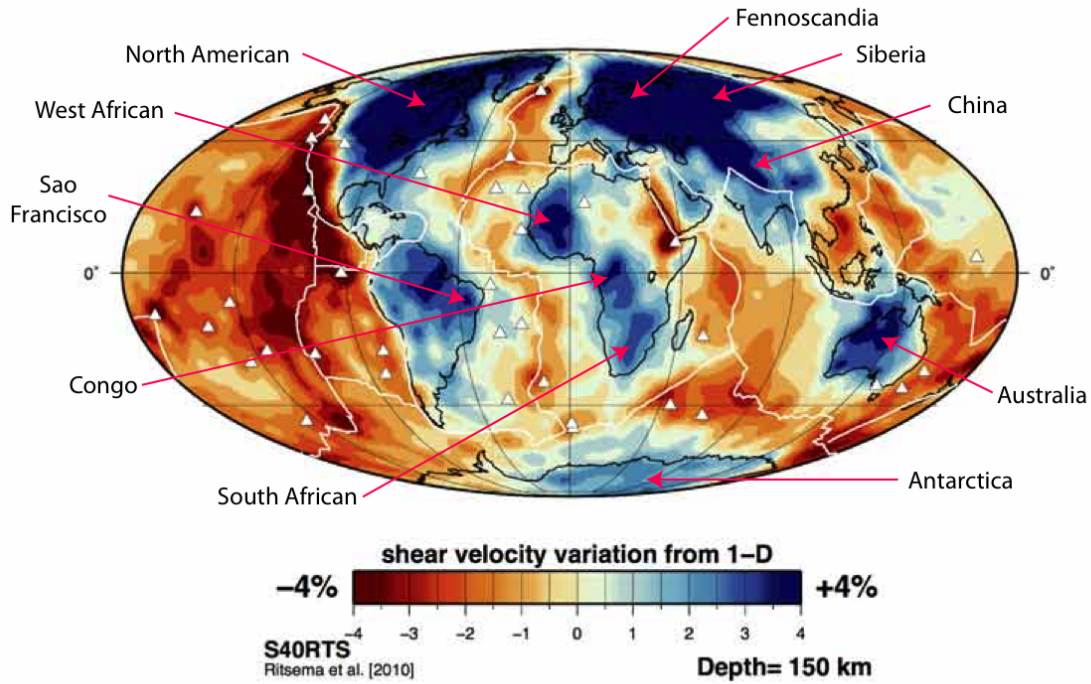


Figure 1.3: **Inferred locations of cratons** worldwide outlined by fast wavespeeds ($\delta V_s > +3\%$) using a depth slice at 150 km through the shear wave model of *Ritsema et al.* (2011). White triangles are hotspot locations (*Sleep*, 1990).

Chemical depletion and low temperatures mean that cratonic keels manifest as high seismic velocity anomalies in mantle tomographic images (Figure 1.3), when compared to the oceans and younger continents (e.g., *Li et al.*, 2008a; *Ritsema et al.*, 2011). Lithospheric keels can also be extremely thick, ≥ 250 km (e.g., *Lebedev and Van der Hilst*, 2008; *Lekić and Romanowicz*, 2011), in contrast to young lithosphere that is typically ≤ 100 km. Keels are thought to be layered (e.g., *Abt et al.*, 2010; *Darbyshire et al.*, 2013; *Fischer et al.*, 2010; *Yuan and Romanowicz*, 2010) with cold Archean cores of highly depleted material, underlain by a younger, thermal boundary layer in the lower lithosphere (e.g., *Lee et al.*, 2011). In this model, the surrounding Proterozoic domains are only underlain by the younger less depleted lithosphere and consequently are less buoyant (*Griffin et al.*, 2004; *Poudjom-Djomani*

et al., 2001). However more complex internal cratonic structures may also exist (e.g., *Aulbach*, 2012; *Cooper and Miller*, 2014; *Eaton and Perry*, 2013; *Ecken et al.*, 2018; *Snyder*, 2002). In Australia and North America, a thick lithospheric keel exists beneath both Archean and Proterozoic domains (e.g., *Schaeffer and Lebedev*, 2014; *Simons et al.*, 2002; *Yuan et al.*, 2011). The keels below southern Africa are confined to Archean domains (e.g., *James et al.*, 2001). Some cratonic regions lack thick lithospheric keels all together (e.g., North China and Wyoming: *Dave and Li*, 2016; *Kusky et al.*, 2007), so the extent of lithospheric keels beneath cratons is debated.

1.3 Modification of Precambrian lithosphere by tectonic processes

The disparity in observations regarding the extent of lithospheric keels beneath cratons globally, questions whether they can be modified, post-formation, by tectonic processes. Intuitively, isolation from active plate boundaries may protect the cratons and their associated keels from modification, but throughout Earth's history, these regions are known to have experienced large scale orogenesis (e.g. the Proterozoic Trans Hudson and Grenville Orogens of North America *Hoffman*, 1988). The impact of orogenesis at cratonic margins is poorly understood however. Physical evidence for keel modification, of which only some is long-lived, may include the removal of basal lithospheric material (e.g., *Dave and Li*, 2016; *Gao et al.*, 2004; *Kusky et al.*, 2007), surface uplift (e.g., *Weeraratne et al.*, 2003), offset of lithospheric boundaries in seismic velocity with respect to their surface expressions (e.g., *Frederiksen et al.*, 2013) and geochemical analyses of outcropping rocks (e.g., *Chiarenzelli et al.*, 2010; *Yang et al.*, 2008). The spatial extent of modification also varies. In China and Wyoming, for example, there is evidence for entire cratonic keel modification/removal (e.g., *Dave and Li*, 2016; *Gao et al.*, 2004; *Kusky et al.*, 2007), whilst in Europe and Africa, modification is restricted to the margins of cratons (e.g., *Pedersen et al.*, 2013; *Youssof et al.*, 2015). Further, the timing of lithospheric modification throughout the history of a craton may also vary from the present (e.g., *Weeraratne et al.*, 2003), through recent past (e.g., *Dave and Li*, 2016), to deep time (e.g., *Frederiksen et al.*, 2013). Because not all physical evidence of lithospheric modification is long-lived (e.g., *Harig et al.*, 2010), a bias may exist towards evidence for more recent modification episodes, where lithospheric remnants can easily be imaged in the mantle. The farther back in time one wishes to look, the greater variety of constraints must be sought, thus geophysicists are required to exploit available geological evidence as well.

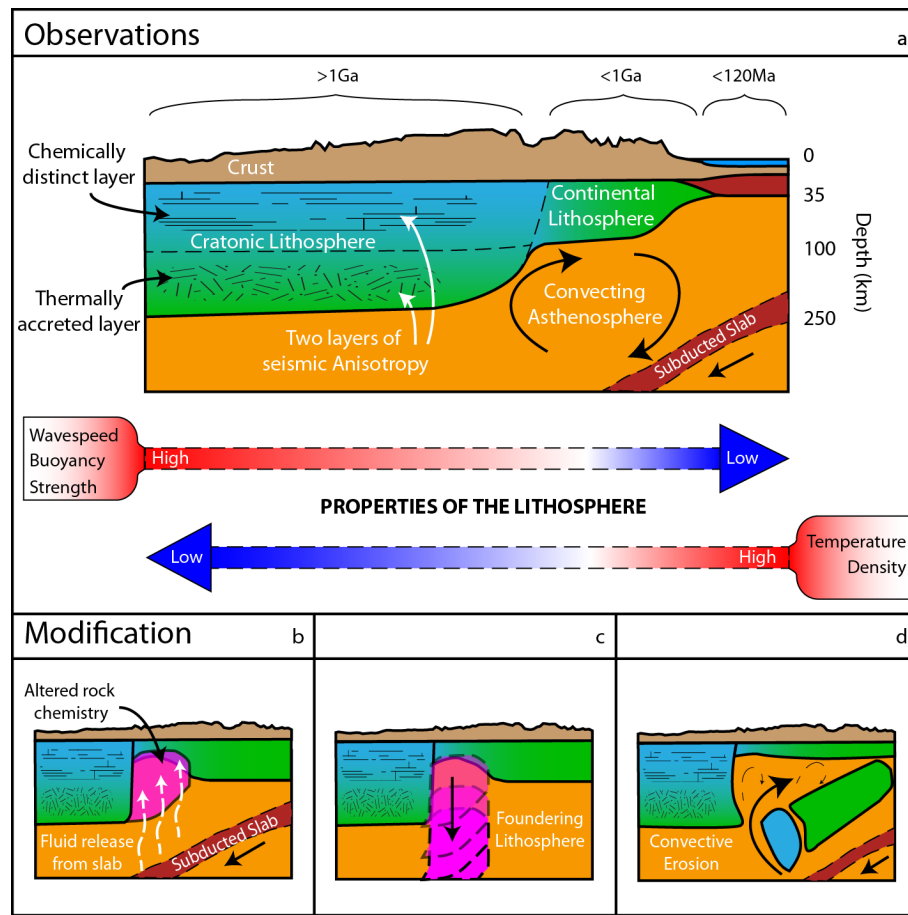


Figure 1.4: **Summary of cratonic margins** (adapted from *Gilligan et al., 2016a*) with properties and processes indicated relative to other parts of continental plates. Red to blue indicates high to low seismic velocity, buoyancy, strength, temperature and density (a). b-d show 3 processes responsible for modification at cratonic margins; metasomatism from a down going slab (b), lithospheric foundering/delamination (c), convective erosion at a lithospheric basal step.

Tectonic and geodynamic mechanisms previously invoked to explain modification of cratonic lithosphere are legion (Figure 1.4; also see *Gerya, 2014*). Firstly, subduction beneath the margin of a craton, causing release of fluids from the down going plate (dewatering) can cause metasomatism of the mantle lithosphere above it (e.g., *Chiarenzelli et al., 2010; Fischer and Gerya, 2016; Nolet and Zielhuis, 1994; Pedersen et al., 2013; Wang et al., 2015; Wenker and Beaumont, 2017; Windley et al., 2010; Youssof et al., 2015*), resulting in a detectable reduction in seismic velocities (e.g., *O'Reilly and Griffin, 2006*). Convection in the mantle, associated with subduction or a lithospheric step, can lead to the modification of cratonic lithosphere (e.g., *Currie and Wijk, 2016; Menzies et al., 2007; van Wijk et al., 2010*) and adjacent younger continental lithosphere (e.g., *Bao et al., 2014; Kaislaniemi and van Hunen, 2014*). Hotspot tectonism is thought to impart a long-lived impact on cratonic lithosphere (e.g., *King and Ritsema, 2000; Nyblade et al., 1996; Rondenay et al., 2000; Villemaire et al.,*

2012; Wang *et al.*, 2015), but its impact may not be consistent across tectonic domains of varying age (Villemaire *et al.*, 2012). Continental lithosphere may also be modified by underplating, erosion or thinning (e.g., Currie *et al.*, 2008; Foley, 2008; Rondenay *et al.*, 2000; Villemaire *et al.*, 2012; Yang *et al.*, 2008; Zheng *et al.*, 2007), rifting (e.g., Frederiksen *et al.*, 2013; Weeraratne *et al.*, 2003) and deformation when subject to basal drag from viscous mantle flow (Eaton and Frederiksen, 2007; Kaban *et al.*, 2015).

In some instances, cratonic lithosphere can delaminate, sink into the deeper mantle (e.g., Dave and Li, 2016; Gao *et al.*, 2004; Johnson *et al.*, 2013; Kusky *et al.*, 2007; Morency and Doin, 2004; Morency *et al.*, 2002; Rey and Houseman, 2006; Ueda *et al.*, 2012; Wang *et al.*, 2017) and is replaced by less depleted material or a thinner lithosphere remains. This can result from subduction, hotspot interaction, global mantle overturn or continent-continent collision (e.g., Dave and Li, 2016; Gao *et al.*, 2004; Kusky *et al.*, 2007). During continental collision, overthickening of both the crust and lithospheric mantle can result in gravitational instability (e.g., England, 1993; Houseman *et al.*, 1981; Li *et al.*, 2016; Sacks and Secor, 1990; Schott and Schmeling, 1998) causing delamination. This may be initiated by thermal (Nelson, 1992) or fluid-derived (Leech, 2001) processes. Unusually high temperatures at the base of the lithosphere often lead to crustal re-melting and subsequent emplacement of A-type granitoid intrusives (e.g., Harig *et al.*, 2010; Kay and Kay, 1993; Platt and England, 1994; Whalen *et al.*, 1987), which are notably voluminous at times of supercontinent formation (Kay and Kay, 1993). Subsequent extensional collapse (Dewey, 1988), thins the crust allowing a sag basin to form (Nelson, 1992). Less depleted asthenosphere accretes to the base of the lithosphere reducing buoyancy (Zheng *et al.*, 2007). Geological indicators of such an orogenic cycle are thus: Crustal thickening and subsequent thinning, uplift followed by subsidence and basin formation, emplacement of A-type granitoid intrusives, delaminated lithospheric blocks and preserved mantle lithospheric structures (e.g., Dewey, 1988; England, 1993; Harig *et al.*, 2010; Houseman *et al.*, 1981; Kay and Kay, 1993; Nelson, 1992).

1.4 Seismological strategy

1.4.1 Arrival-times

Much of modern seismology relies on the measurement of seismic phase arrival-times. Assuming the source time is known, this is the time taken for earthquake energy to travel from the hypocentre to a

given station. Seismologists often measure travel-time residuals, the difference between observed and predicted travel-times ($\mathbf{time}_{\text{residual}} = \mathbf{time}_{\text{observation}} - \mathbf{time}_{\text{predicted}}$), as this contains information on the wavespeed structure between source and receiver. Negative arrival-time residuals indicate earlier arrivals than expected, indicative of fast wavespeed structure along the ray path; positive residuals result from later arrivals than expected (Figure 1.5). Typically, travel-time tables or 1-D reference Earth models, such as iasp91 (Kennett and Engdahl, 1991) or ak135 (Kennett et al., 1995), are used to calculate predicted times for each ray. The choice of model is important since earthquake location, ray paths and delay times are all dependent upon it (Van der Hilst and Spakman, 1989). Ray theory, as mentioned above, is successful at modelling first order wavefield propagation characteristics since Fermats principle states that when a ray path is perturbed, the effect on travel-time is second order (Snieder and Sambridge, 1992). This relies on the asymptotic high frequency approximation of limited band-width data.

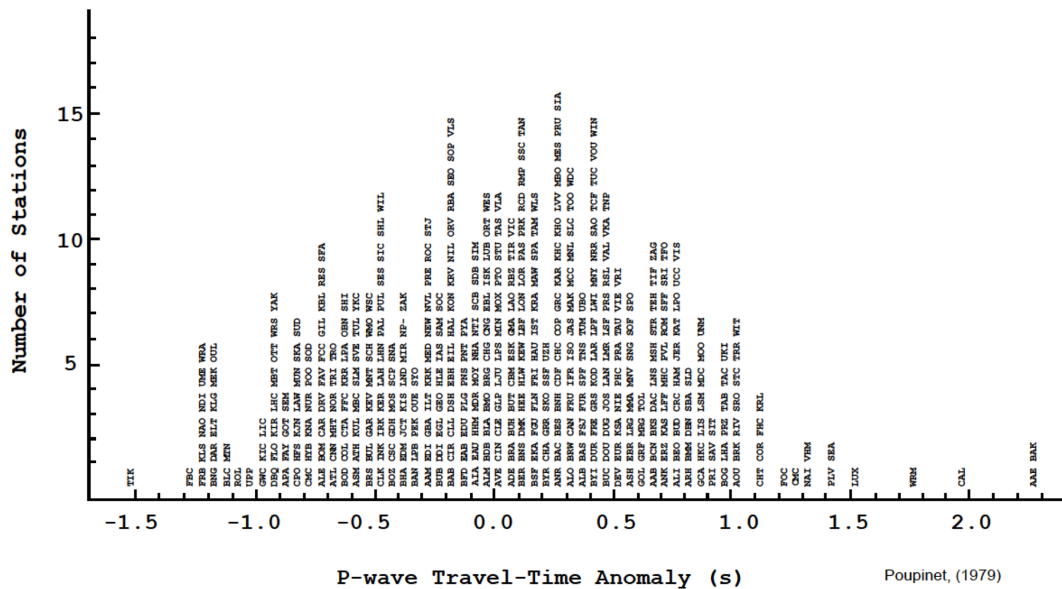


Figure 1.5: **Histogram of mean P-wave travel-time residuals** for named continental permanent seismograph stations. Stations TIK and AAE are located in cratonic Siberia and Ethiopia, above the African Super plume, respectively. Redrawn after Poupinet (1979).

Relative arrival-times

Relative arrival-time residuals present a deviation in arrival-time with respect to the average arrival-time across a seismic network. Small regional networks often record low signal-to-noise seismograms due to sub-optimal deployment conditions, but at teleseismic distances, once filtered, these waveforms are largely coherent across networks of $< 1500 \text{ km}$ in aperture (e.g., Rawlinson and Kennett, 2004;

VanDecar and Crosson, 1990). The calculation of relative arrival-times, $\mathbf{T}_{\text{rel}}$, capitalises upon this through alignment of coherent peaks or troughs of a particular phase, for a given earthquake recorded across a network. This can be achieved through cross correlation (e.g., *VanDecar and Crosson*, 1990) and stacking (e.g., *Rawlinson and Kennett*, 2004) amongst other methods (e.g., *Chevrot*, 2002).

Relative arrival-time residuals $\mathbf{RES}_{\text{rel}}$ (Equation 1.1) are retrieved from the difference between the relative arrival-time, $\mathbf{T}_{\text{rel}}$, and the deviation of the expected arrival-time, $\mathbf{T}_{\text{exp}}$, with respect to the average expected time across the network $\bar{\mathbf{T}}_{\text{exp}}$, computed by ray-tracing through a 1-D reference model. The relative arrival-times for each station, $\mathbf{T}_{\text{rel}}$, are small time-shifts that align the coherent phase of each waveform around an arbitrary point in time, whose distribution is forced to be zero.

$$\mathbf{RES}_{\text{rel}} = \mathbf{T}_{\text{rel}} - (\mathbf{T}_{\text{exp}} - \bar{\mathbf{T}}_{\text{exp}}), \quad (1.1)$$

By removal of the background mean velocity structure, $\bar{\mathbf{T}}_{\text{exp}}$, we assume rays follow a similar path until they are in close proximity to the station network, thus arrival-time heterogeneity now results from wavespeed structure directly below the network (e.g., *Bastow*, 2012). Inversions of relative arrival-time datasets are legion (e.g., *Bastow et al.*, 2008; *Frederiksen et al.*, 2013; *Rawlinson et al.*, 2006; *Rondenay et al.*, 2000; *VanDecar et al.*, 1995) contributing major insight into global upper mantle seismic structure. However, due to the loss of the background mean velocity structure in Equation 1.1, which may vary substantially from region-to-region, such datasets cannot be combined or compared easily to other datasets worldwide. Thus absolute arrival-times are required at the continental to global scale.

Absolute arrival-times

When seismic noise does not obscure the onset of incoming energy on a given seismogram, the absolute arrival-time, $\mathbf{T}_{\text{abs}}$, may be picked directly. The difference between $\mathbf{T}_{\text{abs}}$ and $\mathbf{T}_{\text{exp}}$ gives the absolute arrival-time residual $\mathbf{RES}_{\text{abs}}$ in the equation below:

$$\mathbf{RES}_{\text{abs}} = \mathbf{T}_{\text{abs}} - \mathbf{T}_{\text{exp}}. \quad (1.2)$$

The requirement for highly accurate earthquake catalogues, and the aforementioned noise constraints

imposed upon variably emergent teleseismic waveforms limit the availability of these datasets. Automatic arrival-time pickers have been developed and applied with varying success in an attempt to combat this however (e.g., *Amaru et al.*, 2008; *Di Stefano et al.*, 2006; *Diehl et al.*, 2009). Absolute arrival-times and their respective residuals, contain information regarding the entire ray path from source to receiver and thus are tied to average global wavespeed structure (Figure 1.5). Such data are typically reported to global pick databases such as the ISC (*International Seismological Centre*, 2016) and utilised in global tomographic inversions (e.g., *Li et al.*, 2008a).

1.4.2 Seismic tomography

Seismic tomography seeks to recover slowness heterogeneity (reciprocal of velocity) within the Earth’s interior by dividing it into a large number of cells or blocks (e.g., *Spakman and Nolet*, 1988). Aside from discrete block parameterisations, schemes can include; 3-D interpolation using B-splines among a grid of nodes (e.g., *Rawlinson et al.*, 2006; *VanDecar et al.*, 1995), discontinuities embedded into a 3-D grid (e.g., *Zhao*, 2004), spherical blocks of uniform latitude and longitude (e.g., *Van der Hilst et al.*, 1997) or a representation of 3-D structure using a set of basis functions (*Thurber and Ritsema*, 2007). Due to spatially variant data sampling, data-adaptive gridding can often be desirable however (e.g., *Li et al.*, 2008a; *Zhang and Thurber*, 2005) who used an automatic adaptive mesh where the sample density controls mesh density. A detailed review of irregular cell and adaptive mesh techniques is presented by *Sambridge and Rawlinson* (2005).

The geometry of the propagating wavefield in delay time tomography can be obtained by finite frequency differencing (e.g., *Vidale*, 1988, 1990), shortest path methods (e.g., *Moser*, 1991) or ray tracing algorithms (e.g., *Červený*, 1972), where the travel-time of each ray is a path integral, approximated by summing the travel-times through each block of the model. In the simplest formulation of a linear travel-time seismic inversion (Equation 1.3), $\mathbf{G}$ contains the length of the ray in each model block i.e. the linear operator used to predict the data from the parameterised model of Earth’s interior, $\mathbf{m}$. This is compared to $\mathbf{d}$, the travel-time vector or data:

$$\mathbf{d} = \mathbf{G}\mathbf{m}. \quad (1.3)$$

Nonlinear inversions commonly compute the misfit between data and predicted data by using a starting Earth model and calculating the forward solution. Perturbations to the initial model are then

computed which aim to minimise this misfit in subsequent iterations i.e. $\mathbf{m} = \mathbf{G}^{-1}\mathbf{d}$ is replaced with $\delta\mathbf{m} = \mathbf{G}^{-1}\delta\mathbf{d}$. These algorithms are used in instances of strong velocity perturbations (e.g., *VanDecar et al.*, 1995).

Because the model vector may not be the same size as the data vector and some parts of the model may not be sampled by ray paths, tomographic models are non-unique. Direct matrix inversion to solve $\mathbf{m} = \mathbf{G}^{-1}\mathbf{d}$ (i.e. the inverse of Equation 1.3) cannot be achieved, so instead, regularised iterative methods are routinely used to efficiently solve sparse systems of equations such as conjugate gradient and LSQR (*Li et al.*, 2008a; *Nolet*, 1985; *Paige and Saunders*, 1982; *Scales*, 1987; *Spakman and Nolet*, 1988; *VanDecar et al.*, 1995). By controlling amplitudes and gradients of velocity anomalies through regularisation (e.g., *VanDecar and Snieder*, 1994), we can obtain the minimum structure (simplest) model that fits the data most well in an “Occams’s razor” approach (e.g., *Constable et al.*, 1987).

Propagating seismic waves acquire information about the subsurface off the ray path in practice. This is particularly problematic for lower frequency core-phases and surface waves, but for high frequency body waves, ray theory remains an adequate approximation in the upper mantle. Finite frequency sensitivity kernels that appear as banana doughnuts (no sensitivity along the central ray path) in a 3-D Earth, increase in width with decreasing frequency (*Dahlen et al.*, 2000). This results in a more accurate representation of the entire propagating wavefield in a 3-D Earth and has resulted in some intriguing insight into the Earth’s interior, particularly at depth (*Montelli et al.*, 2004; *Sigloch*, 2011; *Sigloch and Mihalynuk*, 2013). Whether these developments outweigh the improvements offered by increased data coverage (spatial sampling) was openly debated (e.g., *Dahlen and Nolet*, 2005; *De Hoop and Van Der Hilst*, 2005; *Montelli et al.*, 2004). More recently however, this method has gained wider acceptance due to a greater quantity of velocity constraints available when multiple phase frequency bands are included (e.g., *Rawlinson et al.*, 2010). Further, full waveform inversion (e.g., *Fichtner et al.*, 2013; *Liu and Tromp*, 2006; *Zhao et al.*, 2005) minimises the misfit between observed and computed seismograms, rather than travel times (*Thurber and Ritsema*, 2007) but this requires complex algorithms for predicting the entire propagated wavefield (e.g., *Komatitsch et al.*, 2002).

Outwith the finite frequency effects on ray-path geometry, other potentially-large sources of error and limitations of tomography are; sparse distribution of source and receivers, poorly constrained sources, errors in travel-time data (e.g., *Romanowicz*, 1991) and a lack of quantitative measure of resolution (*Nolet*, 1985). When these issues are adequately addressed alongside careful interpretation (see; *Nolet*, 1985; *Rawlinson and Spakman*, 2016; *Rawlinson et al.*, 2010; *Romanowicz*, 2003; *Sambridge and*

Rawlinson, 2005), linear relative arrival-time body-wave inversion remains an attractive method to image upper mantle heterogeneity. Linear relative arrival-time inversions are extremely effective in mapping the location of positive and negative polarity anomalies within the model space, in regions away from active plate boundaries, where velocity gradients are not expected to be strong. Capitalising on the alignment of coherent waveforms enables the use of noisier seismograms derived from temporary networks in which the first-break of energy cannot be seen. Body-wave relative arrival-time inversions (e.g., *Villemaire et al., 2012*) are also typically able to image shorter-wavelength heterogeneity than lower-frequency surface wave tomographic methods (e.g., *Petrescu et al., 2017*). Thus the first investigation within this thesis comprises a relative arrival-time tomographic study of southeast Canada.

1.5 Thesis outline

- **Chapter 2:** *Tectonic setting and previous geophysical work* provides a geological history specific to the Archean Superior and Proterozoic Grenville provinces in the context of Precambrian North America. This is followed by a summary of the key geophysical investigations preceding this work which focus on the southeast Canadian/eastern North American upper mantle and their implications for the formation and evolution of the continent.
- **Chapter 3:** *Relative arrival-time tomography of southeast Canada* presents a relative arrival-time study of southeast Canada to propose a mechanism by which the edge of the Superior craton was modified after its formation.
- **Chapter 4:** *The Absolute Arrival-time Recovery Method (AARM)* details the Absolute Arrival-time Recovery Method (AARM) developed to retrieve absolute arrival-times from regional networks in southeast Canada.
- **Chapter 5:** *Absolute arrival-time tomography of North America* presents a new absolute arrival-time P-wave inversion for the North American continent. Variation in seismic wavespeed imaged in the Grenville Province indicates delamination is a mechanism by which the margins of cratons can be modified during Proterozoic orogenesis.
- **Chapter 6:** *Concluding remarks* collates all pertinent results from this thesis, followed by a discussion of future work to develop this study further.

Chapter 2

Tectonic setting and previous geophysical work

2.1 Tectonic setting

2.1.1 The Superior Province

The 2.6 – 3.0 *Ga* Superior craton at the heart of the Canadian Shield (Figure 2.1) is the largest area of exposed Archean rocks on Earth (e.g., *Card*, 1990). It comprises fragments of immature continental crust, volcanic arcs and other material of oceanic affiliation (e.g., *Hoffman*, 1988; *Langford and Morin*, 1976; *Ludden and Hynes*, 2000) and formed a substantial part of the palaeocontinent of Laurentia. At the margin of the Superior craton, radiating dyke swarms such as the Matachewan at 2.45 *Ga* (*Hoffman*, 1990) and others at 2.2 *Ga* and 2.1 *Ga* (*Ernst and Bleeker*, 2010) are linked to failed Proterozoic rifting. To the northwest, the Superior collided with the Rae and Hearne domains during the Palaeoproterozoic Trans-Hudson Orogen (THO - Figure 2.1; *Hoffman*, 1988), the remnants of which now reside beneath Hudson Bay and Palaeozoic sediments (Figure 2.1). Following the THO, the southern margin of the Superior witnessed multiple continent-arc collisions and accretion of juvenile terranes from 1.8 – 1.3 *Ga* (Figure 2.1 adapted from: *Whitmeyer and Karlstrom*, 2007). These included the Yavapai province at 1.80 – 1.68 *Ga*, the Mazatzal province from 1.70 – 1.60 *Ga* and the Granite-Rhyolite province (from 1.50 – 1.30 *Ga*) which lie beneath the Palaeozoic sediments in the south (Figure 2.1). To the east however, these terranes underwent heavy reworking during the subsequent Grenville orogeny (*Hynes and Rivers*, 2010).

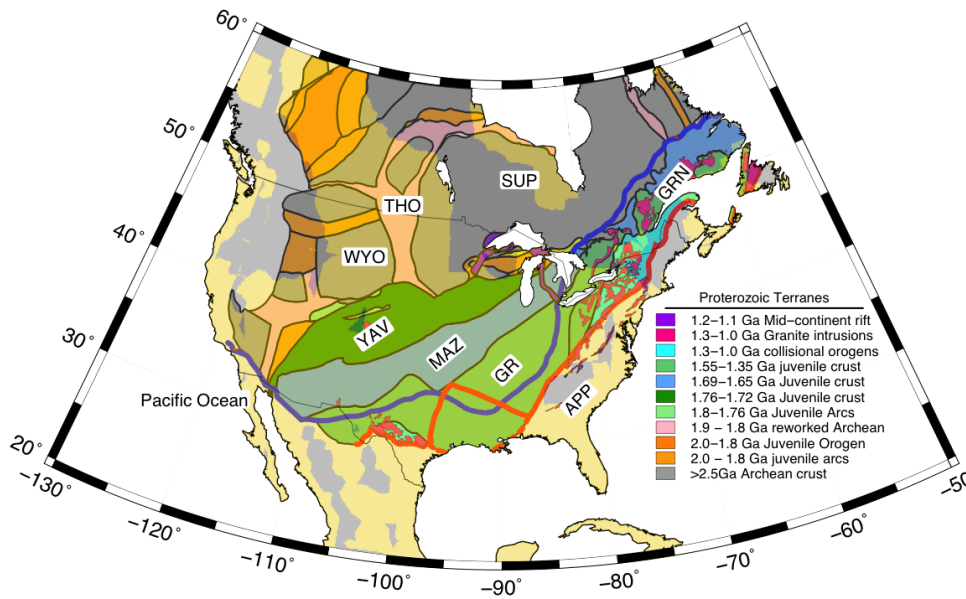


Figure 2.1: **Figure to show simplified Proterozoic geology of North America** (adapted from *Whitmeyer and Karlstrom, 2007*). Rocks outcrop in Canadian Grenville but subcrop south of the Great lakes in the U.S. (*Laske and Masters, 1997*), indicated by yellow transparency. Basement geology is also variant along strike. In the north, Archean rocks of the Superior (SUP) underthrust the Grenville Province (GRN) comprising Palaeoproterozoic Yavapai (YAV - 1.80 – 1.68 Ga), Mazatzal (MAZ - 1.70 – 1.60 Ga) and Granite-Rhyolite (GR - 1.55 – 1.35 Ga) material. In the south, the Grenville Province is formed of mostly younger Granite-Rhyolite material. The eastern Grenville Province was heavily reworked during orogenesis (*Whitmeyer and Karlstrom, 2007*), during the formation of the Rodinian supercontinent (*Li et al., 2008c*). The Grenville (blue) and Appalachian (red) tectonic deformation fronts are shown as thick lines. Emplacement of A-type granitoid intrusives (1.3 – 1.0 Ga) are shown in pink. The location of Appalachian terranes (APP), Wyoming craton (WYO) and Trans-Hudson (THO) orogen are also indicated.

2.1.2 The Grenville Province

At 1.3 – 0.9 Ga, the southeastern margin of Laurentia witnessed one of the most protracted and spatially extensive collisions in Earth’s history, the Grenville orogeny (e.g., *Li et al., 2008c; Rivers, 2009, 2012*). It spanned $\geq 4000\text{ km}$ from Labrador to Texas (Figure 2.1). The surface expression is bounded by the Grenville front to the west and the Appalachian front in the east. Grenville terranes are exposed in Canada but south of the Great Lakes in the U.S., the Grenville is concealed by Palaeozoic sedimentary cover (*Laske and Masters, 1997*). As such, consensus on the formation and evolution of the Grenville Province in the U.S. is lacking (see reviews by: *Darbyshire et al., 2017; Davidson, 1995; Rivers, 1997, 2015*). In the north, the Grenville Province comprises an amalgam of provinces of

varying age; Superior, Mazatzal, Granite-Rhyolite constituents and some accreted terranes. However in the U.S. to the south, the Grenville primarily comprises Granite-Rhyolite material which was young $\leq 0.25\text{ Ga}$ at the time of collision (*Whitmeyer and Karlstrom, 2007*).

At the southeastern Superior margin, an ‘Andean-style’ subduction system (*Rivers, 1997*), dipping to the northwest, may have operated for $> 300\text{ Ma}$, metasomatically enriching Grenvillian mantle (*Chiarenzelli et al., 2010*). Other authors argue that an opposing phase of subduction also occurred (e.g., *Hanmer et al., 2000; Ludden and Hynes, 2000; White et al., 2000*) allowing the Proterozoic island arcs and fragments of continental and oceanic affiliation (*Dickin, 2000*) to be accreted to the Archean Superior basement in distinct episodes, particularly in the western Grenville. Multiple collisional phases and subsequent extension have been recognised throughout the $> 300\text{ Ma}$ Grenville Orogeny, namely the Elzeverian ($1.3 - 1.2\text{ Ga}$) and Ottowan ($1.09 - 0.98\text{ Ga}$) orogenic phases (e.g., *Rivers, 2009, 2012*), which terminated with the final assembly of the supercontinent of Rodinia. Canadian Laurentia is thought to have collided with Amazonia during this period (*Li et al., 2008c*), but the entire length of the Grenville collision may have comprised up to four cratonic blocks (N-S: Baltica, Amazonia, Congo, Kalahari), although precise constraints are lacking (*Hoffman, 1988; Hoffman, 1991*).

The farthest westward footprint of the Grenville Orogen, or extent of metamorphism, is the Grenville Front (GF - Figures 2.1, 2.2). Observable in the field in Canada, its precise location beneath the Palaeozoic sediments in the U.S. is debated (e.g., *Stein et al., 2017*). In Canada it manifests as a thrust fault, since reactivated in an extensional setting, dipping gently southeast beneath the Grenville Province. This fault plane is thought to extend $\geq 300\text{ km}$ laterally, therefore penetrating to depths below the Moho (*Ludden and Hynes, 2000; White et al., 2000*). This allows the reworked material from the Grenville collision to be thrust above Superior basement, at least within the crust. Whilst its location may seem somewhat co-incidental, the GF may have been an inherited zone of weakness, re-activated during the Grenville Orogeny (*Martignole and Calvert, 1996*). In the U.S., it is unknown whether basement older than 1.35 Ga extends east of the GF (*McLelland et al., 1996; Whitmeyer and Karlstrom, 2007*).

The Elzeverian and Ottowan orogenic phases were interspersed by the emplacement of anorthosite suites (*Corrigan and Hanmer, 1997; McLelland et al., 1996*) at $1.19 - 1.11\text{ Ga}$ thought to arise from orogenic collapse of over thickened crust. Exposure of anorthosite suites (components of anorthosite-mangerite-charnockite granite/gabbro - AMCG - intrusions) exist from southeast Canada to Georgia in varying abundances (*Davidson, 1995; McLelland et al., 1996; Whitmeyer and Karlstrom, 2007*). It is

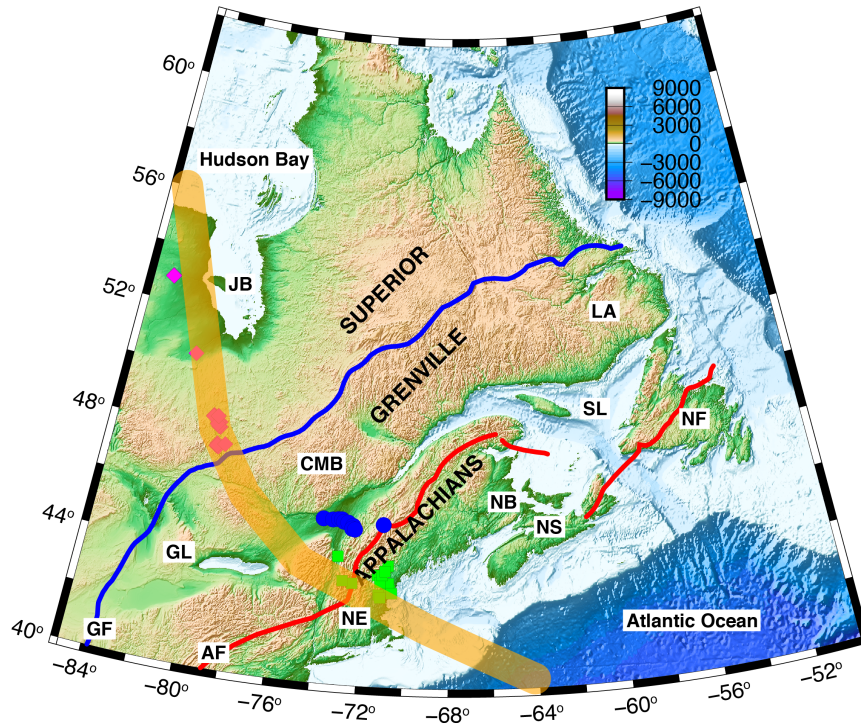


Figure 2.2: **Key tectonic features of southeast Canada and northeast U.S.** The track of the ca. 190 – 110 *Ma* Great Meteor hotspot (*Heaman and Kjarsgaard, 2000*) is marked by the orange band, with subsequent magmatism also indicated: kimberlite pipes (magenta diamonds) and igneous intrusive bodies (blue circles - Monteregian Hills; green squares - White Mountains plutons). Other marked features include: AF - Appalachian Front (red line), CMB - Central Metasedimentary Belt, GF - Grenville Front (blue line), GL - Great Lakes, JB - James Bay, NB - New Brunswick, NE - New England, NF - Newfoundland, NS - Nova Scotia, SL - Gulf of St. Lawrence.

reasonable to expect these intrusives extended beneath the Palaeozoic Appalachian orogen (*McLelland et al., 2013; Whitmeyer and Karlstrom, 2007*) but are relatively less extensive in the Canadian Grenville terranes (*Corrigan and Hanmer, 1997*) (Figure 2.1).

The early stages of Amazonian rifting from the Laurentian core are thought to be associated with the failed Mid-continent rift structure that occurred at 1.1 *Ga* (*Stein et al., 2014*). Breakup of the supercontinent Rodinia initiated after 0.7 *Ga* (*Hoffman, 1991*). The main phase of rifting occurred at 0.62 – 0.55 *Ga* (e.g., *Li et al., 2008c; Whitmeyer and Karlstrom, 2007*) which opened the Iapetus Ocean. Prior to this however, numerous failed rift arms developed, including the St. Lawrence (e.g., *Kamo et al., 1995*). Some eastern Grenvillian terranes may have also rifted from Laurentia and have been subsequently re-attached during Appalachian orogenesis (*Carrigan et al., 2003*).

2.1.3 The Appalachian Province

The closure of Iapetus and north-to-south closure of the Rheic ocean occurred at $0.46 - 0.26 Ga$ (*Hatcher, 2005, 2010; Van Staal, 2005*). There is evidence for both west and east dipping subduction at the Laurentian margin during this time (*Hatcher, 2010*). Terminal continental collision brought together Laurentia and Gondwana to form the supercontinent Pangaea (e.g., *Stampfli et al., 2013*), the modern-day remnants of which are the Appalachian terranes at the southeast flank of the Grenville Province. As in the Grenville collision, little evidence is preserved of the exact location of the continental suture due to terrane overprinting (*Thomas, 2006*), however the farthest westward extent of the Appalachian orogen is marked by the Appalachian front (AF - Figures 2.1, 2.2). Rifting of Pangea caused the opening of the Atlantic Ocean at $\sim 180 Ma$.

2.1.4 Recent tectonic processes

Having remained largely tectonically quiescent since the Appalachian Orogeny, the only further major tectonic events are related to localised magmatism. *Sleep (1990)* recognised that a hotspot track cross cut tectonic boundaries in southeast Canada. This produced dykes and kimberlite pipes within the Superior (*Griffin et al., 2004*) and the alkaline igneous intrusions of the Monteregian Hills and White Mountains and the off shore New England seamount chain (Figure 2.2). Occurring in the Mesozoic (ca. 190-110 Ma), this tectonism (*Bank et al., 1998; Calvert and Ludden, 1999; Heaman and Kjarsgaard, 2000*) is linked to the Great Meteor hotspot (GMH). To the south, radiogenic isotopes and seismic observations have linked the 48 Ma volcanism in Virginia to lithospheric delamination *Mazza et al. (2014)* or a hidden hotspot (*Chu et al., 2013*), respectively.

2.2 Previous geophysical studies

2.2.1 Global and continental observations of the North American lithosphere

Global seismic models show dominantly high seismic velocities ($\delta V_S > 4\%$, $\delta V_P > 1\%$) throughout the Canadian shield at lithospheric depths (e.g., *Lebedev and Van der Hilst, 2008; Li et al., 2008a; Ritsema et al., 2011; Schaeffer and Lebedev, 2013*), stretching from east of the Cordillera to the Grenville Province in Canada and much of the U.S. These high velocities extend to $\sim 250 km$ depth

(e.g., *Lebedev and Van der Hilst*, 2008). This contrasts with the tectonically active, younger, western regions that manifest as low velocity anomalies.

Continental P-wave imaging (*Burdick et al.*, 2017; *Schmandt and Lin*, 2014), also shows consistent high velocities throughout “stable” North America but are not well resolved northwards into Canada (e.g., *Burdick et al.*, 2017). *Schmandt and Lin* (2014) show the highest lithospheric wavespeeds to stretch from the Precambrian rifted margin at $\sim 105^\circ\text{W}$ to the Appalachian front, east of which is dominated by low velocity anomalies, most notably the central and eastern Appalachian anomalies. However, *Burdick et al.* (2017) show high velocities to be confined to the unmodified Mesoproterozoic domains of the central U.S., the younger eastern terranes are dominated by average and localised low velocity anomalies.

Early shear velocity models of North America (e.g., *Bedle and van der Lee*, 2009; *Frederiksen et al.*, 2001; *Grand*, 1987; *Nettles and Dziewonski*, 2008; *van der Lee and Frederiksen*, 2005) show a high velocity lithosphere to extend to depths 250 – 400 km. There is less agreement spatially however as some models show a link between lithospheric wavespeeds and surface tectonic provinces (e.g., *Nettles and Dziewonski*, 2008) whilst others do not (e.g., *Frederiksen et al.*, 2001). *van der Lee and Frederiksen* (2005) also indicate that higher velocities are seen in the Canadian Shield than the North American platform to the south. Capitalising on vast the improvement of data availability afforded by the USArray network, recent surface wave models also show the high velocity North American shield region but with greater resolution (*Clouzet et al.*, 2018; *Schaeffer and Lebedev*, 2014; *Yuan et al.*, 2011, 2014, see Figure 2.3). In Canada, elevated wavespeeds exist throughout the lithosphere (to ~ 250 km depth) in the Archean Superior province and extend into the Proterozoic Grenville at upper lithospheric depths (≤ 100 km *Clouzet et al.*, 2018; *Schaeffer and Lebedev*, 2014; *Yuan et al.*, 2014). Consequently elevated velocities shallow with distance from the cratonic core towards the coast. In the U.S. to the south however, high lithospheric velocities of the Mesoproterozoic, stable continental interior display a vertical wavespeed boundary approximately co-incident with the Grenville Front, therefore intermediate velocities are prevalent throughout the Grenville in the U.S. The lowest velocities are confined to the coastal Appalachian terranes where the lithosphere is thinner (*Grand*, 1987; *Schaeffer and Lebedev*, 2014). The model of *Schaeffer and Lebedev* (2014) also shows a large lower velocity region beneath the Great Lakes (Figure 2.3), a feature not seen in other tomographic models discussed here.

The faster velocities in Archean domains of eastern North America are thought to be confined to the upper lithosphere ($\leq 50 - 160$ km), and are underlain by slightly lower velocities that extend out

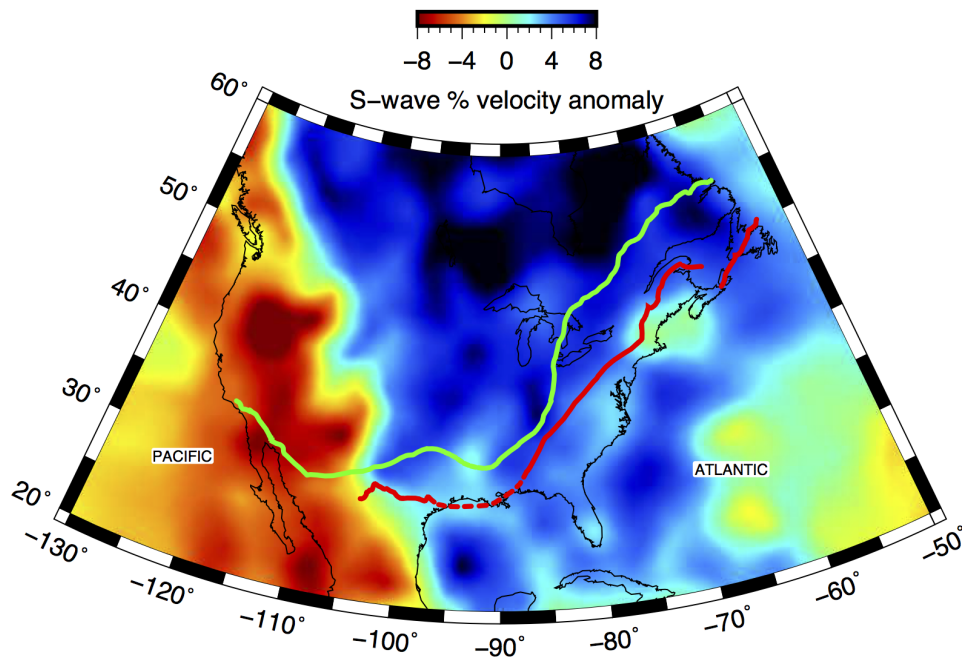


Figure 2.3: **S-wave velocity model of North America, of *Schaeffer and Lebedev (2014)* at 100 km depth** plotted in percent with respect to the mantle reference velocity of 4.38 km/s. The Grenville Front (green line) and Appalachian Front (red line) are indicated as in Figure 2.1.

beneath the Proterozoic domains as well (e.g., Figure 2.4, adapted from *Yuan and Romanowicz, 2010*). From a corresponding change in anisotropic fast direction within the lithosphere *Yuan and Romanowicz (2010)* were able to propose a two-layered, vertically stratified, lithosphere beneath cratonic North America (Figure 2.4). These findings were corroborated by *Darbyshire et al. (2013)*, suggesting cratonic keels form in multiple stages. By combining of gravity, topography, crustal structure and seismic tomography, *Kaban et al. (2015)* have since argued for a shift in the mantle keel beneath the Superior Province by the current mantle flow regime developed after rifting of Pangea.

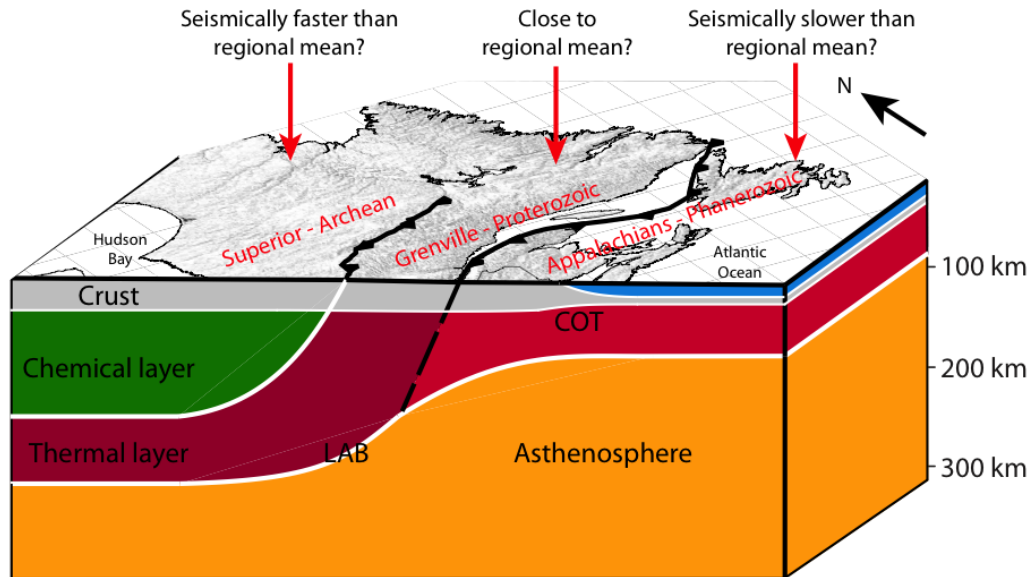


Figure 2.4: **3D projection of a model of the North American keel** (adapted from *Yuan and Romanowicz, 2010*) onto the edge of the Canadian Shield below the study area from the centre of the Superior craton out to the continent ocean transition (COT), beneath which the lithosphere-asthenosphere boundary (LAB) shallows greatly.

Numerous models recognise a region of low velocity extending northwest from the coast in New England into the cratonic interior (e.g., *Schmandt and Lin, 2014*; *van der Lee and Frederiksen, 2005*). This anomaly follows the proposed track of the Great Meteor hotspot and the New England Seamount chain (*Heaman and Kjarsgaard, 2000*) and is thought to terminate at the boundary between the Grenville and the Superior provinces. The anomaly decreases in amplitude to the northwest and widens with depth, but is also displaced from the surface expression by ~ 200 km, effectively dipping to the southwest (*Eaton and Frederiksen, 2007*). To the northeast, the low velocity anomaly below the Gulf of St. Lawrence, northern Grenville Province and Labrador may result from volcanism associated with a failed rift arm (*van der Lee and Frederiksen, 2005*).

2.2.2 Regional observations of the southeast Canadian and eastern U.S. lithosphere

In southeast Canada, the regional body wave imaging of *Villemaille et al. (2012)* shows a marked difference in seismic velocity between the Archean Superior Province and the Phanerozoic Appalachians. A clear distinction is not seen between Archean and Proterozoic domains, suggesting Archean and Proterozoic lithospheric stabilisation processes are similar. This may be due to a lack of station

coverage however. By contrast, the surface wave imaging of *Petrescu et al.* (2017) shows a velocity decrease between the Archean and Proterozoic domains in the lithosphere.

The northwest trending, low velocity ‘divot’ that extends inland from the coast of New England, aligned with the Great Meteor hotspot track, is also identified by these regional tomographic studies (e.g., *Aktas and Eaton*, 2006; *Villemaire et al.*, 2012). The hotspot tectonism is inferred to have extensively modified the Phanerozoic Appalachians but the Archean Superior is largely unaffected, thus the resistance to modification increases with age. It is suggested that the non-vertical nature of the low velocity anomaly (as noted by *Eaton and Frederiksen*, 2007) can be reconciled with the site of a Precambrian failed rift acting as a stress guide (*Aktas and Eaton*, 2006; *Rondenay et al.*, 2000; *Villemaire et al.*, 2012). Within the Central Metasedimentary Belt (CMB) of the Grenville Province (Figure 2.2), *Aktas and Eaton* (2006) recover a high velocity anomaly dipping to the southeast, interpreted a southeast-dipping subduction system that existed prior to the Grenville terminal collision. This is supported by magnetotelluric observations (*Adetunji et al.*, 2014). In the western Superior, *Frederiksen et al.* (2013) show the lithospheric boundary with the Trans-Hudson orogen is offset by ~ 200 km with respect to the surface expression. They infer the lithospheric root of the Superior was likely eroded or modified during the THO.

In the U.S., regional body wave imaging (*Biryol et al.*, 2016) shows high wavespeeds of the Mesoproterozoic core of North America to extend further eastwards in the upper lithosphere (~ 130 km) $\sim 85 - 82^\circ$ W than in the deeper lithosphere (≤ 250 km) $\sim 88 - 84^\circ$ W. Along strike in the Grenville Province, *Biryol et al.* (2016) also note high velocities extend to greater depths in the north ($\sim 38^\circ$ N) than the south ($\sim 34^\circ$ N). The regional surface wave study of *Shen and Ritzwoller* (2016) shows little variation in upper mantle shear velocities across the GF but documents a generally decreased crustal thickness, west-to-east, across the GF in the U.S. This contrasts with the surface wave model of *Pollitz and Mooney* (2016) which displays a marked upper lithospheric boundary in wavespeed coincident with the GF in the U.S. Hence, high velocities are confined to the Superior, Granite-Rhyolite, Mazatzal, and Yavapai provinces, west of the GF.

Shear wave splitting analyses of eastern North America both from Canada (*Darbyshire et al.*, 2015; *Frederiksen et al.*, 2006) and the U.S. (*Long et al.*, 2016; *White-Gaynor and Nyblade*, 2017) do not find mantle fast directions solely attributable to preserved lithospheric fabrics and as such, do not show a distinct correlation to surface tectonic boundaries. However, lithospheric basal topography is known to influence splitting directions beneath the continent (*Eaton et al.*, 2004; *Fouch et al.*, 2000),

interpreted as low velocity asthenospheric flow pulled into the “divot” in the mantle keel (*Fouch et al.*, 2000).

In Canada, at the crustal scale, the LITHOPROBE program (e.g., *Clowes*, 2010) obtained seismic reflection and refraction profiles crossing the GF. These image southeast dipping listric thrust faults beneath the Grenville Province, since reactivated in an extensional fashion (e.g. *Ludden and Hynes*, 2000; *White et al.*, 2000). Reworked Grenvillian-age accretionary material is thus thought to overlie Superior basement within the crust. The extent of the under-thrusting is ≥ 300 km laterally from the surface expression of the Grenville Front (*Ludden and Hynes*, 2000; *White et al.*, 2000), thus reaching depths well below the Moho. Receiver function profiles in the northern Grenville show a thickened crust (*Petrescu et al.*, 2016) when compared to the adjacent Superior. In the U.S. to the south, overthrusting is visible in the crust, but involves post-Grenvillian sedimentary formations and the reworked rocks of the Grenville Orogen (*Santos et al.*, 2002), contrasting the dominantly basement overthrusting in the north (*Ludden and Hynes*, 2000). A transition from unmodified Granite-Rhyolite terranes to the Grenville in the south is accompanied by a crustal thinning in the south (e.g., *Abt et al.*, 2010).

2.2.3 Observations from below the lithosphere of continental North America

Several low velocity anomalies have also been imaged in the deeper mantle beneath eastern North America. At depths just below the lithosphere, *Savage et al.* (2017) use ambient noise data to image multiple low velocity anomalies parallel to the Appalachian Province, argued to result from asthenospheric upwelling due to edge driven convection at the lithospheric step in thickness. *van der Lee et al.* (2008) image a low velocity anomaly extending to 660 km depth, again parallel with Appalachians and largely invariant along strike, terminating around U.S. - Canadian border. This is thought to arise from deep slab dewatering processes derived from remnant lithosphere in the deep mantle (*van der Lee et al.*, 2008). *Biryol et al.* (2016) image a pronounced high velocity anomaly at ~ 600 km depth (at $\sim 37^\circ\text{N}$, 85°W). They relate this to Cenozoic-to-present lithospheric downwelling, but it has also been interpreted as Farallon-related (*Schmandt and Lin*, 2014) or vertical sinking slab piles (*Sigloch and Mihalynuk*, 2013). The present-day downwelling hypothesis also conflicts with a coherently resistive lithosphere observed in recent magnetotelluric studies beneath this region (*Murphy and Egbert*, 2017). Furthermore, density modelling of *Kaban et al.* (2015) infers a dense slab remnant below the Grenville and Appalachians provinces.

At depths $\geq 650\text{ km}$ numerous studies inferred the presence of the Farallon slab as a high velocity anomaly, subducted beneath the western North American continental margin (e.g., *Bedle and van der Lee*, 2009; *Burdick et al.*, 2017; *Grand*, 1987; *Schmandt and Lin*, 2014; *Schmid et al.*, 2002; *Sigloch*, 2011; *Villemaire et al.*, 2012). More recently, deep fast velocity anomalies beneath the eastern U.S. have been reinterpreted to result from a west dipping subduction system. In this model these anomalies thus do not represent Farallon lithosphere (*Sigloch and Mihalynuk*, 2013, 2017).

Chapter 3

Relative arrival-time tomography of southeast Canada

The material in this chapter can be found in an article published in the Journal of Geophysical Research, Solid Earth (*Boyce et al.*, 2016).

3.1 Chapter overview

The cratonic cores of the continents are remarkably stable and long-lived features. Their ability to resist destructive tectonic processes is associated with their thick (~ 250 km), cold, chemically-depleted, buoyant lithospheric keels that isolate the cratons from the convecting mantle. The formation mechanism and tectonic stability of cratonic keels remains under debate. In this chapter I address this issue by using P- and S-wave relative arrival-time tomography to constrain upper-mantle structure beneath southeast Canada and the northeast U.S., a region spanning three quarters of Earth's geological history. In light of these seismic results, I discuss the lithospheric impact of the Great Meteor hotspot that cross-cuts the major tectonic domains and the implications for keel formation. Cognisant of subduction events prior to and collisional tectonics during the Grenville Orogeny an alternative interpretation is also investigated in which the Laurentian margin may have been modified prior to keel stabilisation.

3.2 Introduction

Cratons, the ancient cores of the continents, are the longest-lived tectonic features on Earth. They are thought to have resisted thermomechanical erosion for ≥ 1 Ga over multiple Wilson cycles (e.g., *Griffin et al.*, 2009; *Hoffman*, 1990; *Lee et al.*, 2011). Key to their survival are their thick (~ 250 km) lithospheric keels (e.g., *Lebedev and Van der Hilst*, 2008; *Lekić and Romanowicz*, 2011) that possess a range of attributes that promote longevity; buoyancy, high strength and viscosity (e.g., *Lenardic and Moresi*, 1999; *Sleep*, 2003) and extensive chemical depletion in heavier elements such as Fe, Al and Ca (e.g., *Griffin et al.*, 2003; *Jordan*, 1988). The combination of low temperatures and chemical depletion significantly increases seismic wavespeed, consequently cratonic keels are readily identifiable in tomographic images (e.g., *Li et al.*, 2008a; *Ritsema et al.*, 2011).

While the mechanism for craton formation is debated (e.g., *Lee et al.*, 2011), it is often recognised, particularly in North America, that their keels are vertically stratified (e.g., *Abt et al.*, 2010; *Darbyshire et al.*, 2013; *Yuan and Romanowicz*, 2010). Keels are often thought to comprise a cold core of highly depleted Archean material bounded by a less-depleted, deeper layer of younger affinity (e.g., *Lee et al.*, 2011). Laterally there is more disparity in observations. Thick keels often extend beneath both Archean and (less depleted) Proterozoic terranes (e.g., Australia and North America: *Schaeffer and Lebedev*, 2014; *Simons et al.*, 2002; *Yuan et al.*, 2011) but some cratonic regions show thick keels to be restricted to the highly depleted Archean provinces (e.g., southern Africa: *James et al.*, 2001).

Preserving a geological record spanning over 3 billion years (*Hoffman*, 1990), the Laurentian core of North America has witnessed some of the largest tectonic events in Earth's history (e.g., the Grenville Orogeny). Recent studies (e.g. *Eaton and Frederiksen*, 2007; *Frederiksen et al.*, 2013; *Kaban et al.*, 2015) have suggested that the keel has been modified by tectonic processes, implying keels may not remain as stagnant features after their formation. Globally, subduction (e.g., North China craton, southern Africa and Europe: *Begg et al.*, 2009; *Kusky et al.*, 2007; *Nolet and Zielhuis*, 1994; *Santosh et al.*, 2010; *Youssof et al.*, 2015), underplating or erosion (e.g., *Foley*, 2008; *Rondenay et al.*, 2000; *Villemaire et al.*, 2012), rifting events (*Frederiksen et al.*, 2013) and basal drag from viscous mantle flow (*Eaton and Frederiksen*, 2007; *Kaban et al.*, 2015) have all been invoked to explain keel modification.

Combining new and existing seismograph networks in southeast Canada and the northeast U.S., we present relative arrival-time P- and S-wave regional isotropic tomographic models at higher resolution than previously achieved (e.g., *Aktas and Eaton*, 2006; *Rondenay et al.*, 2000; *Villemaire et al.*, 2012).

We pay particular attention to the question of whether upper mantle wavespeed variations between Archean and Proterozoic domains are explained best by variations in chemical depletion, inherent to the plate formation process, or subsequent metasomatic enrichment during the Proterozoic Grenville Orogeny.

3.3 Methods

3.3.1 Seismic data and networks

Data were sourced from the IRIS Data Management Center and Canadian National Data centre from the period 2007 – 2014, and more recently, the Imperial College Maritimes SEIS-UK deployment of the QM-III experiment (e.g., *Petrescu et al.*, 2016) between 2013 and 2014. Within the $\sim 3000 \times 2000$ km study area (Figure 3.1), 160 stations were used, of which 13 are short-period instruments (Teledyne-Geotech S-13); the remainder are broadband instruments; Wielandt-Streckeisen STS-1 or STS-2, Gralp CMG-3ESP, CMG-3T, CMG-3TP or CMG-40T, or Nanometrics Trillium 120PA. Data from short-period stations are only included in the P-wave dataset, all traces are normalised to the shortest period response to allow consistent phase picks across the network (See Appendix B). Within the epicentral distance range 30° - 120° , 238 earthquakes of $m_b > 5.5$ were used for P (Figure 3.1) and 166 for S yielding a total of 10042 and 9806 rays respectively. The P-wave measurements of *Villemaire et al.* (2012) contribute ~ 6000 rays to our dataset. The backazimuthal coverage of earthquakes is good (Figures 3.2 and 3.3) but dominated by northwestern (Western Pacific) and southern (South America) backazimuths. The distribution in epicentral distance is also good (Figures 3.2 and 3.3).

3.3.2 Relative arrival-times

Initially, data were filtered using a zero-phase, two-pole Butterworth filter (0.4-2.0 Hz for P-waves, 0.04-0.2 Hz for S-waves). Such filters form a trade-off between the infinite frequency approximation required for ray-tracing and limiting high frequency noise in the dataset, an approach also used in previous studies (e.g., *Bastow et al.*, 2005, 2008; *Villemaire et al.*, 2012, see Appendix B). The manual alignment of traces is improved using the multichannel cross-correlation (MCCC) method of *VanDecar and Crosson* (1990) on the filtered data. This allows accurate computation of relative arrival-time

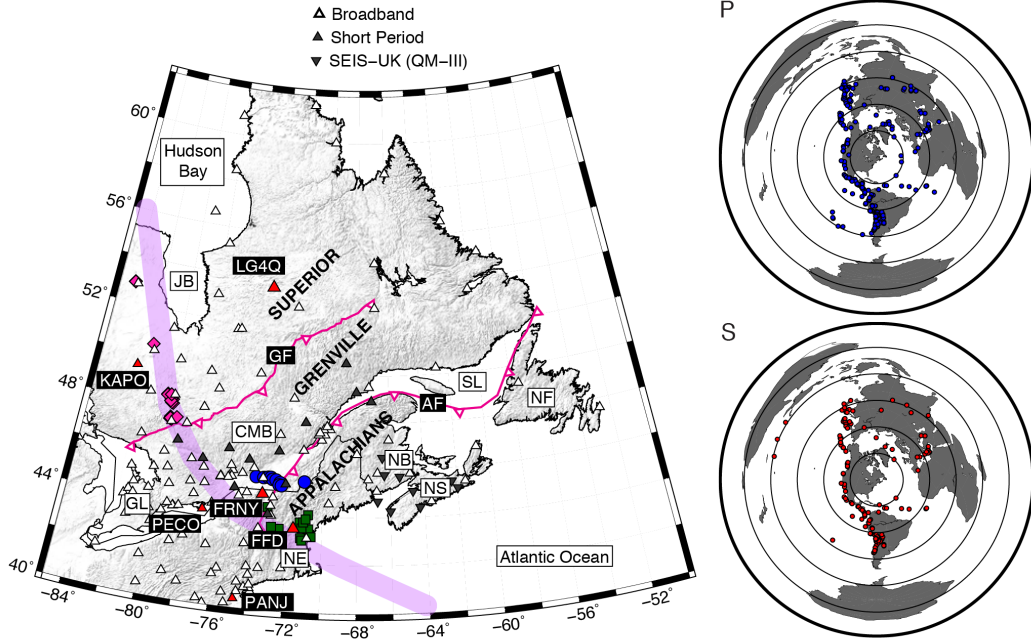


Figure 3.1: **LEFT - Tectonic features and station locations** in eastern Canada and northeast U.S. White triangles represent broadband stations, grey triangles indicate short period stations (only used in P-wave study) and inverted grey triangles, the Imperial College Maritimes SEIS-UK contribution to the QM-III network. Red triangles show the locations of stations highlighted in the P-wave and S-wave residual analysis (Figures 3.2 and 3.3). The lilac transparent band marks the track of the Great Meteor hotspot (*Heaman and Kjarsgaard, 2000*), with associated magmatism: kimberlite pipes (magenta diamonds) and igneous intrusive bodies (blue circles - Monteregian Hills; green squares - White Mountains plutons) AF - Appalachian Front, CMB - Central Metasedimentary Belt, GF - Grenville Front, GL - Great Lakes, JB - James Bay, NB - New Brunswick, NE - New England, NF - Newfoundland, NS - Nova Scotia, SL - Gulf of St. Lawrence. **RIGHT - Global maps of magnitude $m_b > 5.5$ earthquakes** between 2007 and 2014 shown by the blue (P) and red (S) dots, recorded by 160 stations in southeast Canada with high SNR used in this study. The concentric rings mark 30° increments in epicentral distance away from the centre of the seismograph network.

residuals for both datasets. Manual first-break picks are windowed (3 s and 12 s respectively for P and S) to search for the cross correlation maximum position, $\tau_{ij}^{\max}$, between each pair of traces. This is used to define the relative delay time, Δt_{ij} , for each station pair (*i* and *j*) for each earthquake,

$$\Delta t_{ij} = t_i^P - t_j^P - \tau_{ij}^{\max}, \quad (3.1)$$

where t_i^P is the preliminary arrival estimate for trace *i*, picked by hand on the first coherent peak or trough on each waveform. Noise within the waveforms causes a lack of coherency across the traces resulting in an inconsistency in delay-times i.e. $\Delta t_{12} + \Delta t_{23} \neq \Delta t_{13}$ (*VanDecar and Crosson, 1990*). Therefore a least squares minimisation routine is required to compute optimised relative arrival-times, t_i , for each station, where the mean is forced to be zero. t_i are small time-shifts that align the coherent

phase of each waveform around an arbitrary point in time. Errors associated with the relative arrival-times can also be used to approximate error in the MCCC results.

Relative arrival-times have a mean standard deviation of 0.01 s and 0.07 s for P- and S-waves respectively, which appear overly optimistic given the data sampling interval of ≥ 0.01 s. The relative arrival-time residuals $\mathbf{t}_{\text{RES}_i}$ (Equation 3.2) are then calculated by comparison to predicted arrivals, $\mathbf{t}_{\text{e}_i}$, based on the IASP91 travel-time tables (*Kennett and Engdahl, 1991*),

$$\mathbf{t}_{\text{RES}_i} = \mathbf{t}_i - (\mathbf{t}_{\text{e}_i} - \bar{\mathbf{t}}_{\text{e}}), \quad (3.2)$$

where $\bar{\mathbf{t}}_{\text{e}}$ is the mean expected travel-time for an earthquake across the network. Further details can be found within Appendix B.

3.3.3 Analysis of relative arrival-time residuals

Inherent in multichannel cross-correlation is the relative nature of the residuals. Absolute travel-time tomography allows calculation of absolute seismic wavespeeds within the Earth. Relative arrival-time methods, such as that used here, allow for calculation of anomalies with respect to the regional mean of the dataset. The background mean ($\bar{\mathbf{t}}_{\text{e}}$ in Equation 3.2) is completely removed from the output model meaning fast and slow anomalies are relative to the regional average arrival-time. It is recognised on the global scale (e.g., *Li et al., 2008a; Ritsema et al., 2011*), that the Canadian shield is particularly fast and that the coastal terranes within the Appalachians are generally slow.

Analysis of backazimuthal variation of residuals from individual stations within each tectonic domain (Figures 3.2 and 3.3) indicates a general trend of late arrivals (slower, positive residuals) from the SE and early arrivals (faster, negative residuals) from the NW. Within the Superior craton (station LG4Q), the majority of residuals are negative with the exception of the SE direction, away from the cratonic interior. The Appalachians (station FFD) show predominantly late arrivals, except to the NW, where rays pass through faster lithosphere.

Across both P- and S-wave mean relative arrival-time plots (Figure 3.4) there is a marked trend of earlier-than-average arrivals within the Superior craton, close-to-average arrivals within the Grenville and later arrivals within the Appalachians. The shift in the peak of the residual histograms (where residuals are grouped by tectonic province) across the three domains, and the steady increase in

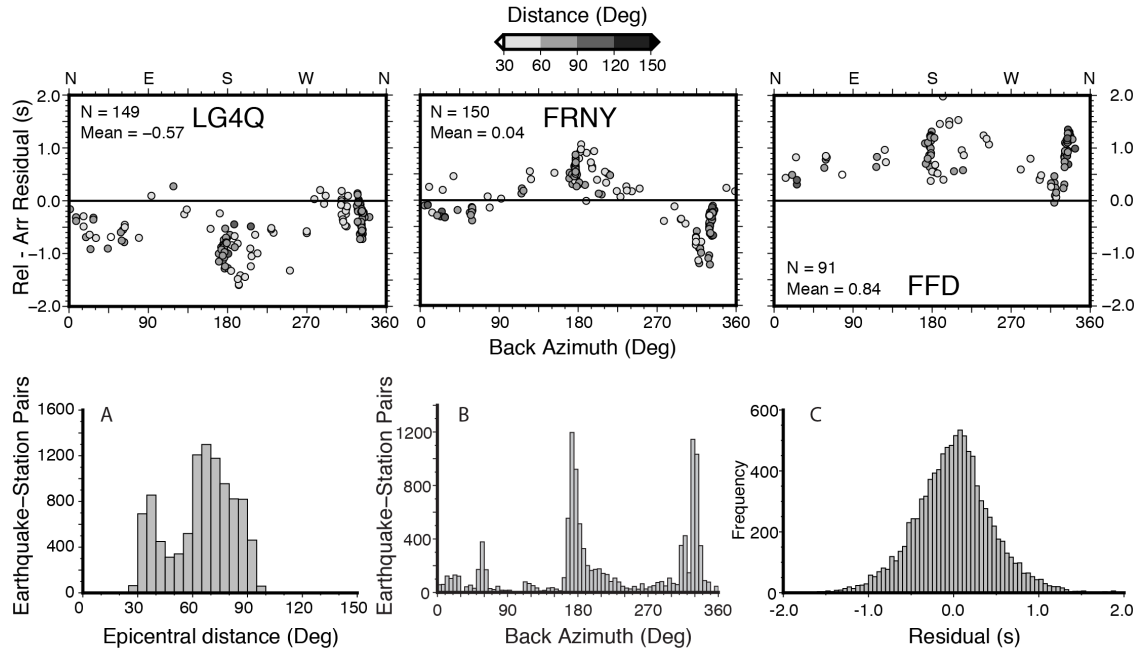


Figure 3.2: **ABOVE - Distribution of relative arrival-time residuals** with backazimuth for three stations (LG4Q, FRNY, FFD labeled in Figure 3.1) within the Superior, Grenville and Appalachians for the P-wave dataset. See Table C.2 for details. Negative values are early arrivals (faster) and positive values are later arrivals (slower). **BELOW - Histogram** of event distribution with epicentral distance (A), event frequency with backazimuth (B) and frequency distribution of residuals (C) for the P-wave dataset. The S-wave dataset shows similar results but with more scatter (Figure 3.3).

residuals along transect F-F' (Figure 3.4) also confirms this trend. Coastal areas, particularly in Newfoundland, appear to suffer from scattered arrival-times, likely due to wave noise. It is also unclear whether the lithospheric impact of the GMH affects the relative arrival-times or whether the lack of ray coverage between 80° and 150° backazimuth hides other trends. Figure 3.5 also confirms the consistency in the distribution of relative arrival-time residuals between both P- and S-wave datasets.

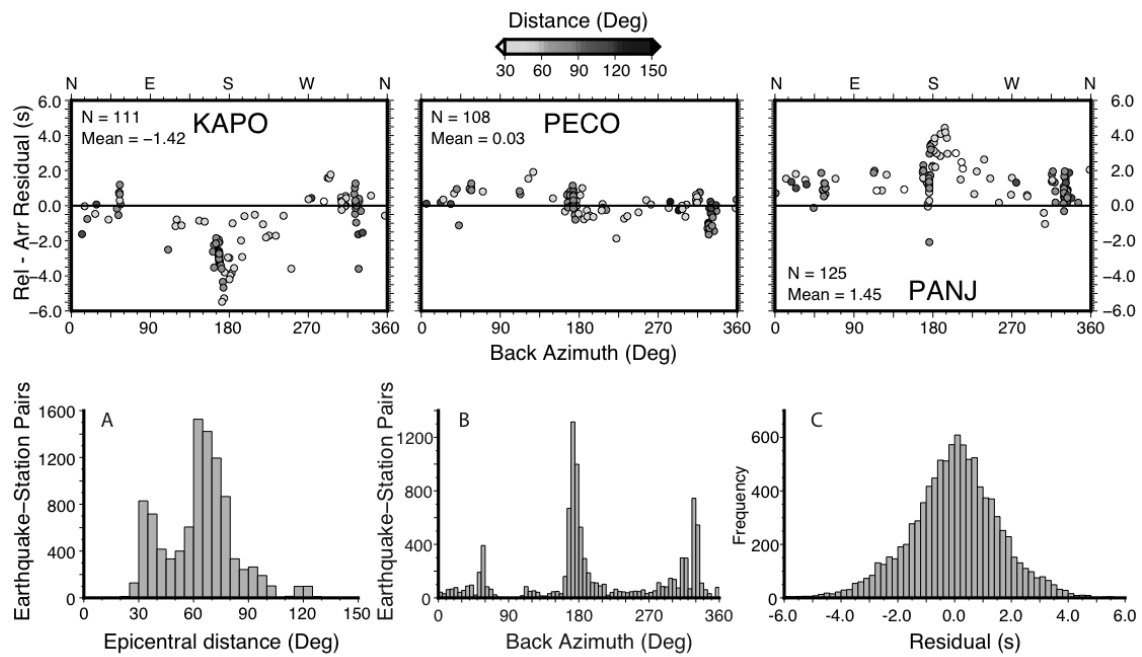


Figure 3.3: **ABOVE** - Distribution of relative arrival-time residuals with backazimuth for three stations (KAPO, PECO, PANJ labeled in Figure 3.1) within the Superior, Grenville and Appalachians for the S-wave dataset. See Table C.2 for details. Negative values are early arrivals (faster) and positive values are later arrivals (slower). **BELOW** - Histogram of event distribution with epicentral distance (A), event frequency with backazimuth (B) and frequency distribution of residuals (C). This dataset shows similar results to the P-wave dataset (Figure 3.2) but with more scatter.

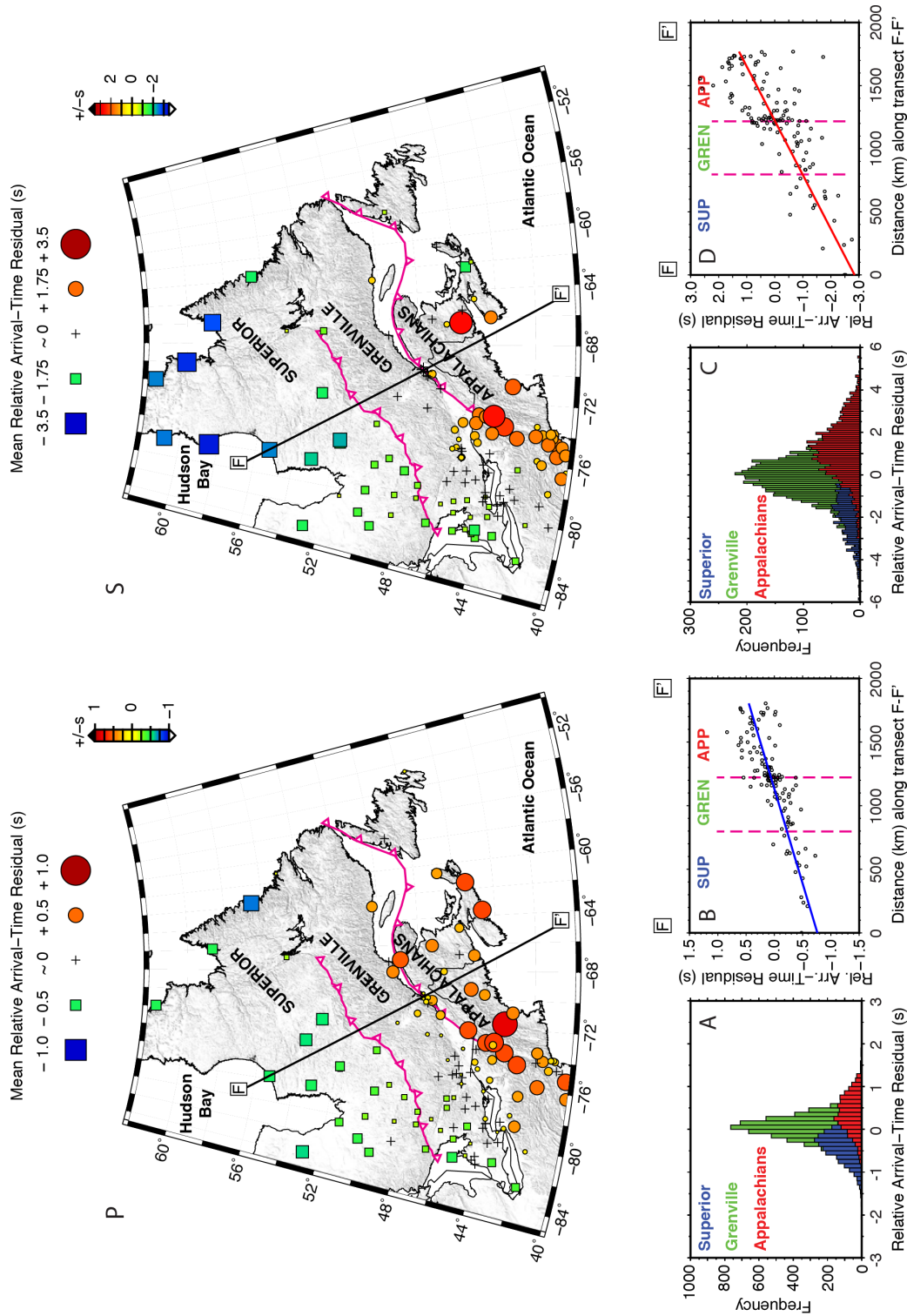


Figure 3.4: **ABOVE** - Plot of mean relative arrival-time residuals at each station for the P-wave (left) and S-wave (right) studies. Negative values represent early arrivals (faster) and positive values represent later arrivals (slower). **BELOW** - A, C - Residual histograms for P and S respectively, grouped by tectonic domain: Superior (SUP - blue), Grenville (GREN - green), Appalachians (APP - red). B, D - relative arrival-time residuals for P and S respectively for all stations plotted with distance along transect F-F' that runs perpendicular to the tectonic boundaries (57°N, 80°W - 41°N, 65°W). A linear regression through each dataset is plotted in blue (P) and red (S).

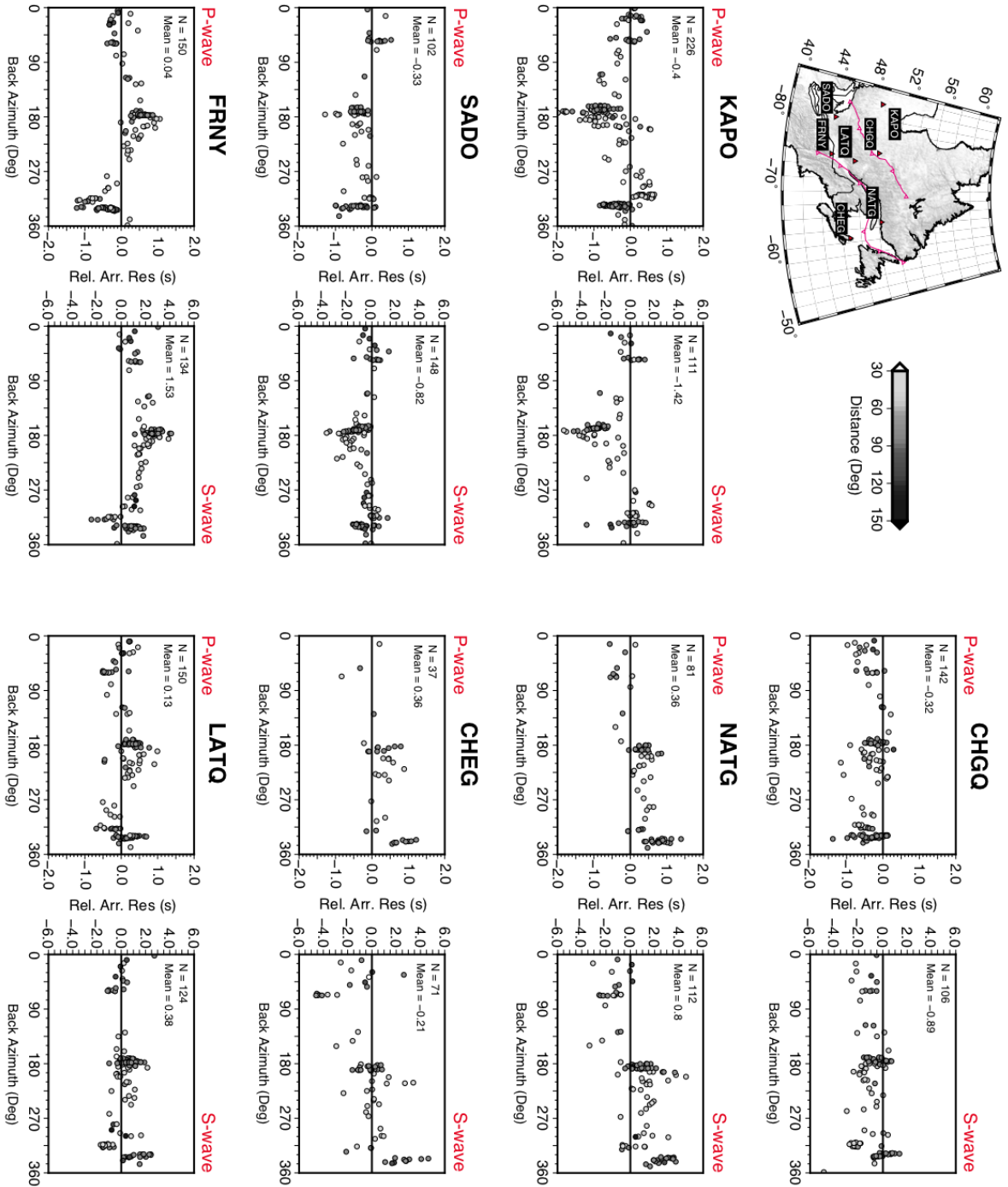


Figure 3.5: Comparison of relative arrival-time residuals against backazimuth for selected stations in the Superior, Grenville and Appalachian terranes for both P- and S-wave datasets. Station locations are shown on the inset map. Negative values are early arrivals (faster) and positive values are later arrivals (slower).

3.3.4 Inversion procedure

Using the regularised, isotropic, least-squares inversion method of *VanDecar et al.* (1995), relative arrival-time residuals are inverted simultaneously for slowness perturbations, source terms and station statics, accounting for unresolvable structure and distant heterogeneity. Station statics for both P- and S-wave inversions are shown in Figure 3.6. Away from active plate boundaries where ray paths may be strongly dependent on velocity structure (e.g., *Sleep*, 1973; *VanDecar*, 1991), a linear inversion scheme, as used here, is adequate (e.g., *Villemaire et al.*, 2012). Two thousand iterations of the conjugate gradient procedure (*Scales*, 1987; *VanDecar et al.*, 1995) are performed, although convergence is reached far sooner (with little extra computational expense). Both P- and S-wave models are parameterised identically with B-spline functions under tension (*Cline*, 1981), allowing velocity to vary smoothly between a regular grid of 176715 knots spanning 35-62°N and 42-96°W and 0-1300 km in depth. The central portion of the model is sampled at 33.3 km intervals in depth, 0.25° in latitude and 0.5° in longitude (Figure 3.7). Coarser spacing is used at the extremities to prevent mapping of unwanted structure into the region of interest from the deeper mantle. Ray-tracing between nodes through interpolation allows for the generation of smoothed velocity models, which are independent of knot position.

Table 3.1: P-wave trade-off curve values

Inversion	Flattening	Smoothing	Residual Reduction
A	125	3750	98.8
B	250	7500	98.4
C	500	15000	97.7
D	1000	30000	96.7
E	1500	45000	95.8
F	2000	60000	95.1
G	4000	120000	92.8
H	8000	240000	90.1
I	16000	480000	87.2
J	32000	960000	84.2
K	64000	1920000	81.4

Over-interpretation of fine-scale structure and inherent non-uniqueness of the models is addressed through the use of regularisation, minimising primary and secondary spatial gradients. These parameters control model complexity (roughness) that trades off against data fit (RMS residual reduction). Models with heavy regularisation experience a decrease in residual reduction and vice-versa (Tables 3.1, 3.2). Because the Canadian shield is one of the most anomalously fast regions on Earth (*Poupinet*, 1979, Figure 1.5), any average 1-D Earth model would not closely approximate velocity structure in

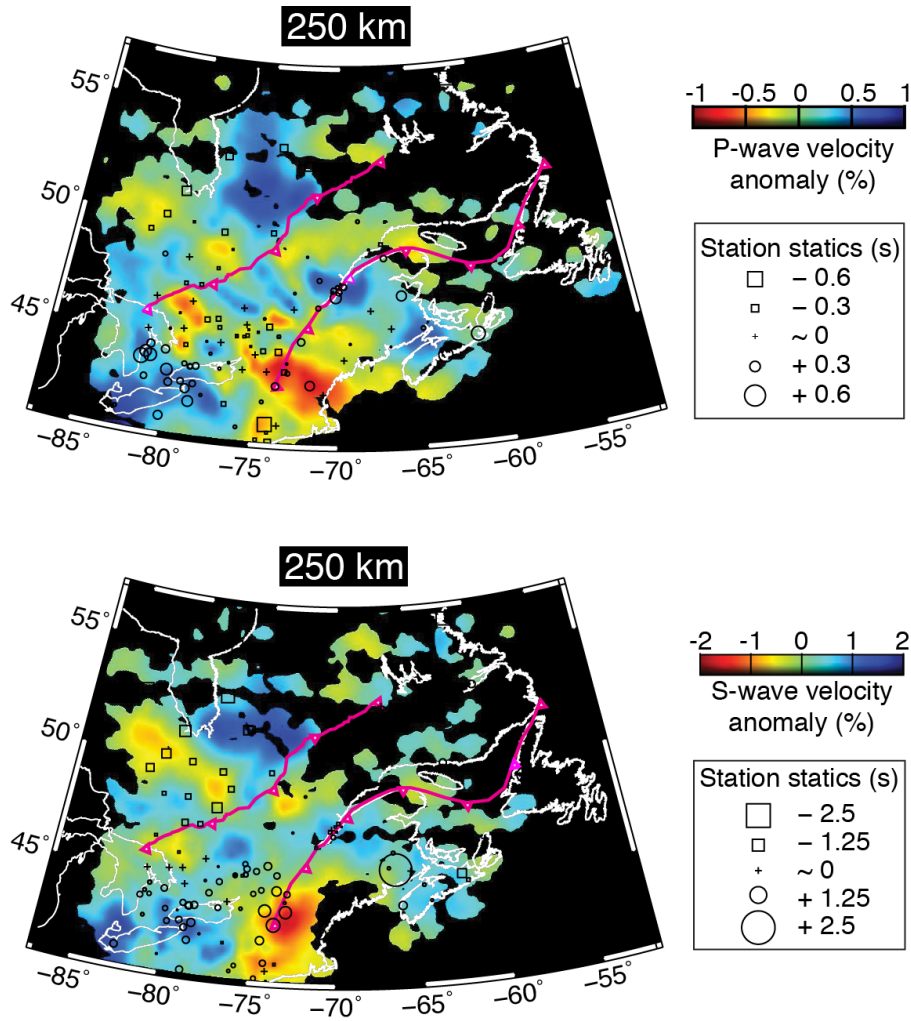


Figure 3.6: **Station statics for P-wave and S-wave inversions** plotted over 250 km model depth slices. The Grenville and Appalachian fronts are shown in magenta. Regions with low (<3) ray hits are muted due to lack of crossing rays and poor resolution.

Table 3.2: S-wave trade-off curve values

Inversion	Flattening	Smoothing	Residual Reduction
A	125	3750	96.9
B	250	7500	95.5
C	500	15000	93.7
D	1000	30000	91.4
E	1500	45000	89.7
F	2000	60000	88.5
G	4000	120000	85.0
H	8000	240000	81.3
I	16000	480000	76.8
J	32000	960000	71.8
K	64000	1920000	66.8

southeast Canada, so regularisation by amplitude damping is not used in this inversion. The MCCC-derived error estimate acts as a optimistic lower bound on data fit so the ideal model sits above this

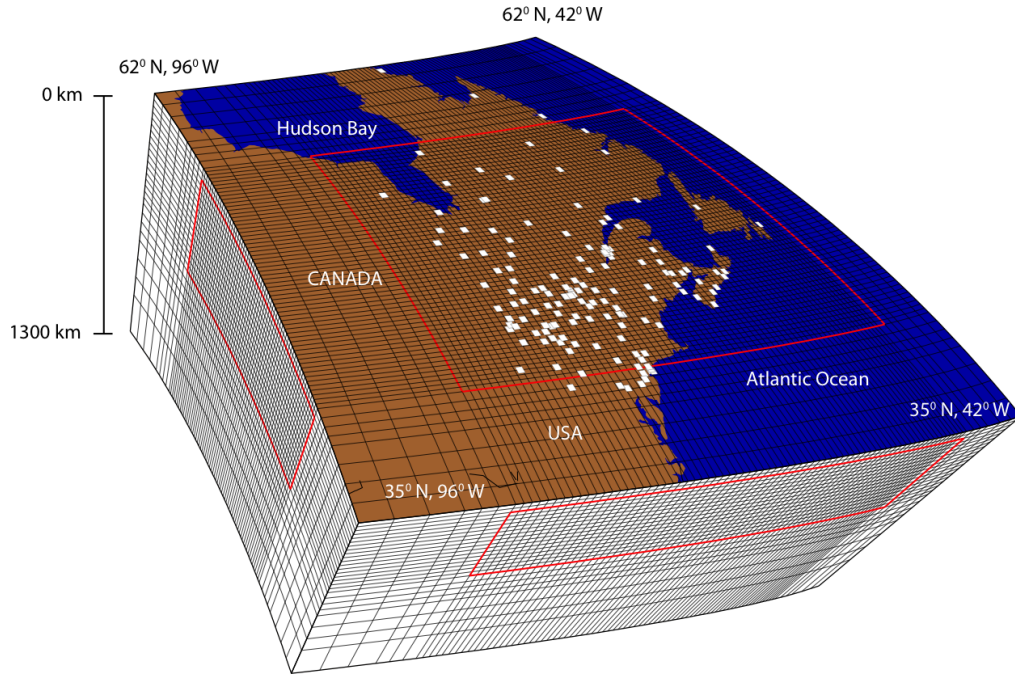


Figure 3.7: **Model parameterisation for the relative arrival-time model.** Knots are separated by 0.5° longitude and 0.25° latitude and 33.3 km depth in the central part of the model in the upper 600 km . The grid is defined by 27 knots from 0 to 1300 km in depth, 77 knots in latitude between $35^\circ - 62^\circ\text{ N}$ and 85 knots in longitude between $96^\circ - 42^\circ\text{ W}$ resulting in 176715 knots. White squares show station locations.

at the knee of the trade-off curve (Figures 3.8 and 3.9), following an Occam's Razor approach. Our final models fit 95% and 89% of the RMS measured delay-time residuals for P and S respectively (F and E in Figures 3.8 and 3.9).

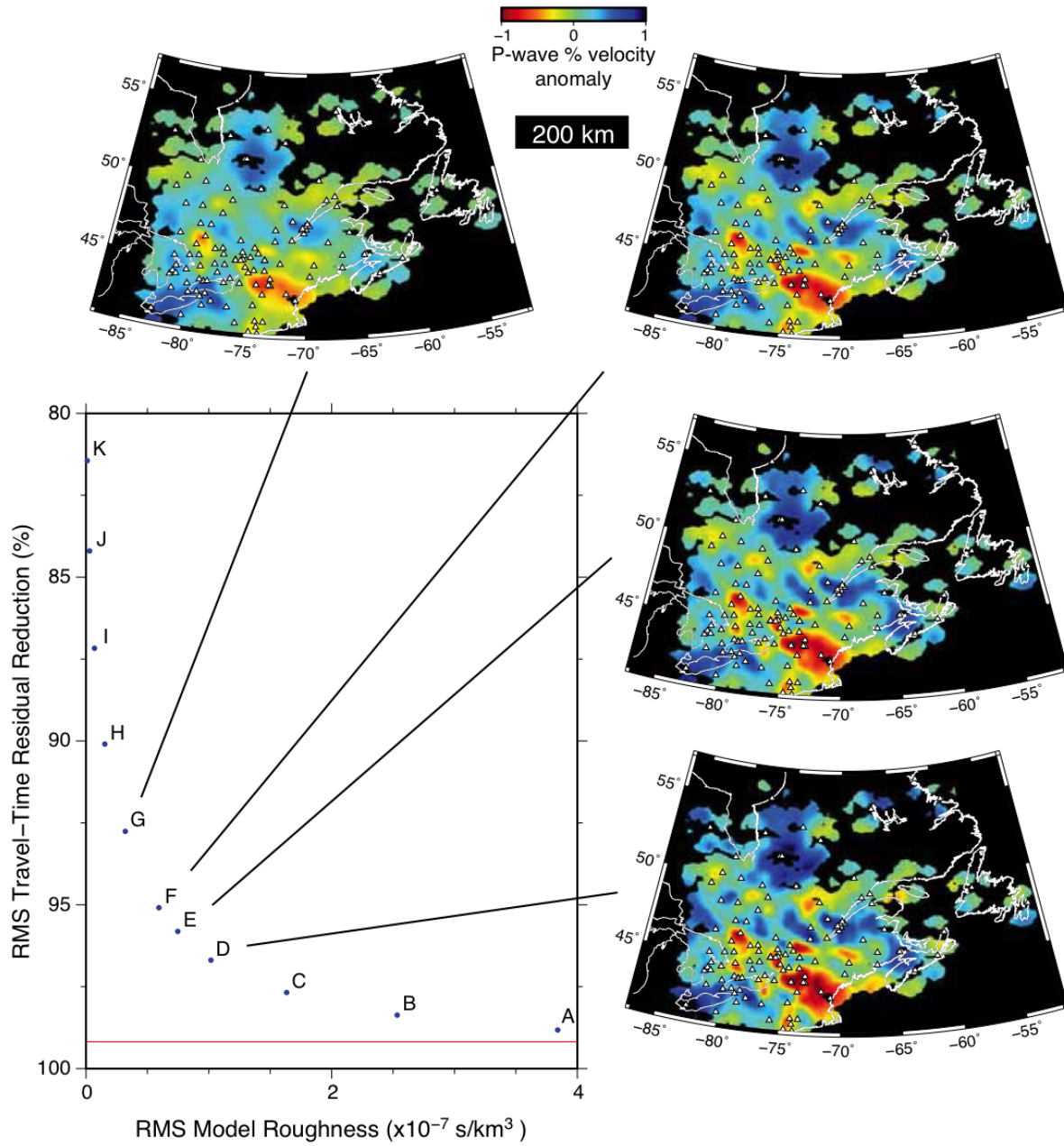


Figure 3.8: **P-wave model trade-off curve** between model roughness and data fit. The corresponding weightings used in the inversions are given in Table 3.1. Depth slices are taken at 200 km, station locations are shown as white triangles and regions with low (<3) ray hits are muted due to lack of crossing rays and poor resolution. The red line is an MCCC derived estimate for data error, the ideal model sits above this line. Model F was picked in this case.

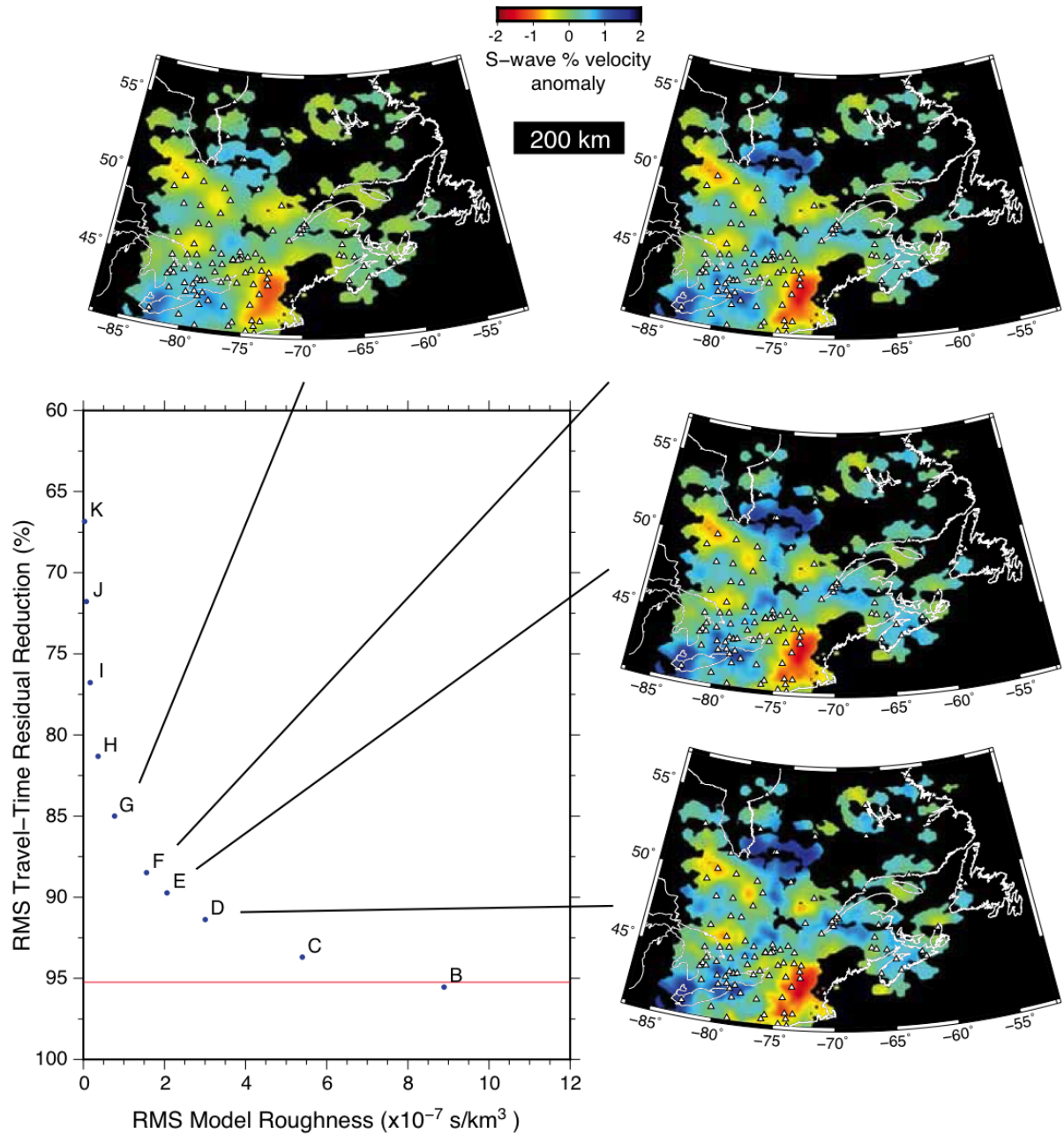


Figure 3.9: **S-wave model trade-off curve** between model roughness and data fit. The corresponding weightings used in the inversions are given in Table 3.2. Depth slices are taken at 200 km, station locations are shown as white triangles and regions with low (<3) ray hits are muted due to lack of crossing rays and poor resolution. The red line is an MCCC derived estimate for data error, the ideal model sits above this line. Model E was picked in this case.

3.4 Resolution testing

We analyse the recovery of synthetic velocity anomaly models to test the resolving power of both P- and S-wave datasets. We model a cratonic core of Superior lithosphere defined by a $\delta V_P = \delta V_S = -3\%$ velocity anomaly from 50 to 250 km depth, the Grenvillian edge of the Canadian shield as $\delta V_P = \delta V_S = -1.5\%$, 50 - 250 km depth and the Phanerozoic Appalachians as a slow velocity anomaly ($\delta V_P = \delta V_S = +3\%$, 50 - 250 km depth). Each of these synthetic features extends approximately 500 km in a NE direction (Figure 3.10). Additionally, we input a narrower, yet longer, slow anomaly ($\delta V_P = \delta V_S = +3\%$, 100 - 200 km depth) representing a proposed thermochemical signature of the GMH track (*Heaman and Kjarsgaard, 2000*).

A Gaussian residual time error component with a standard deviation of 0.02 and 0.1 s for the P- and S-wave data respectively is added to the theoretical travel-times. We then perform inversions using identical parameterisation and regularisation to our observed data. The P-wave model accurately recovers the position of the synthetic anomalies with $\sim 50\%$ amplitude recovery (Figure 3.10). The recovery increases with depth beneath the tectonic blocks but decreases beneath the GMH track. In cross section (Figure 3.10), the anomalies recover the vertical boundaries of the input model, but are vertically smeared, particularly at the edges of the model space. The Superior and Appalachian anomalies are both smeared to 400-650 km apparent depth so the anomalies appear to extend to $\sim 150\%$ of their actual depth. The Grenvillian anomaly is smeared to a lesser extent (350-500 km apparent depth) due to higher density of crossing rays. The GMH anomaly is well constrained, with vertical smearing of only ~ 100 km (Figure 3.10).

The S-wave model also tightly constrains the anomalies spatially, with $\geq 50\%$ amplitude recovery. The extent of vertical smearing (Figure 3.10) is nearly identical to that in the P-wave model with the vertical boundaries preserved at lithospheric depths. Further structural resolution tests show that input anomalies with dipping interfaces, for both P- and S-waves, could be recovered, if present (Figure 3.11). Therefore features in the centre of the model with vertical boundaries are assumed to be reliable. In Figure 3.11 the P-wave model is used; the S-wave model produces similar results. The short wavelength features are also well retrieved in traditional checkerboard tests (Figures 3.12, 3.13).

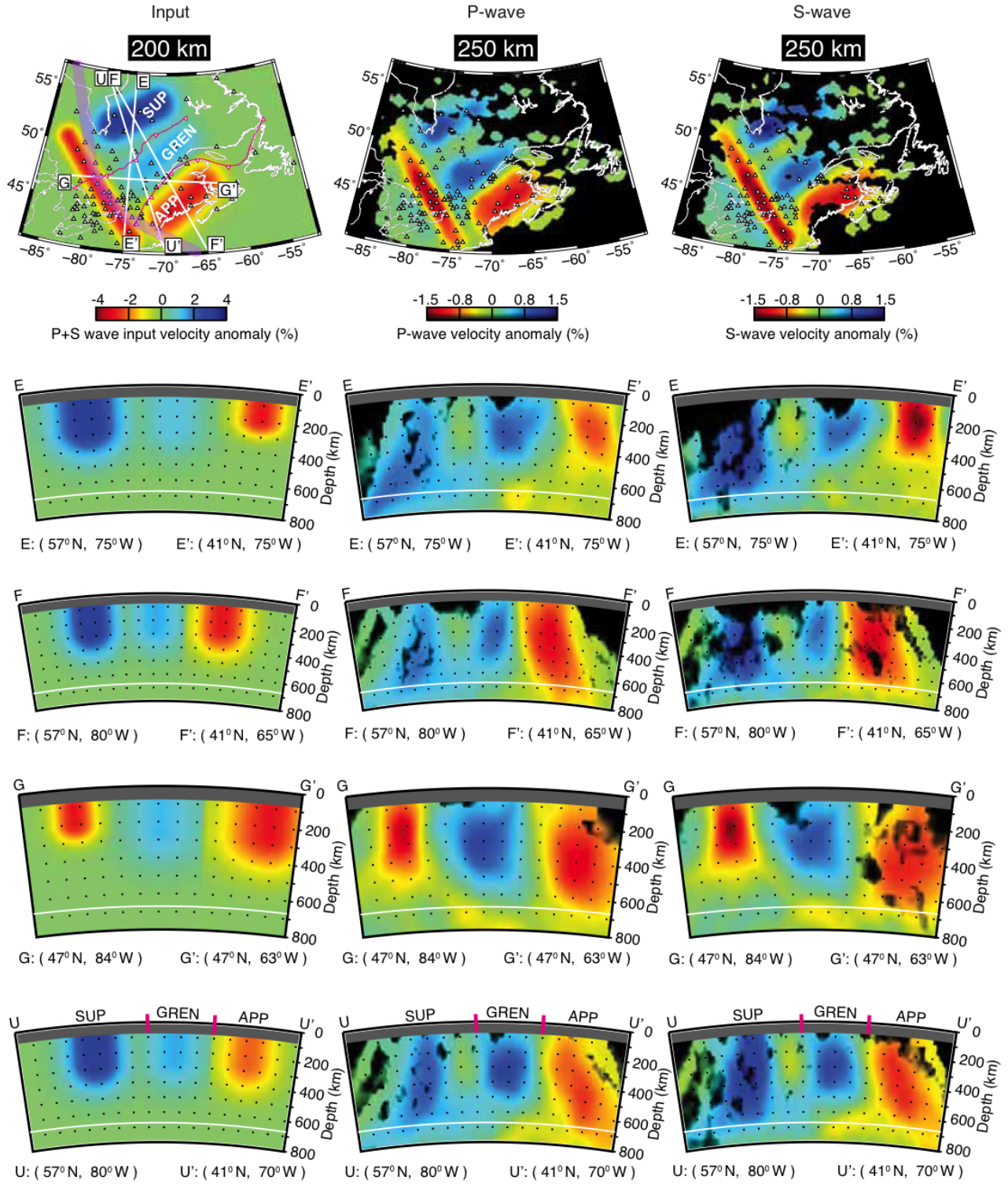


Figure 3.10: **Structural resolution tests** showing input anomalies (left), P-wave model (centre) and S-wave model (right). Synthetic anomalies are defined by Gaussian functions across their widths. Station locations (white triangles), regional tectonics (magenta) and the GMH track (lilac band) are also marked. Regions with low (<3) ray hits and the top 75 km in cross-section are muted due to lack of crossing rays and poor resolution.

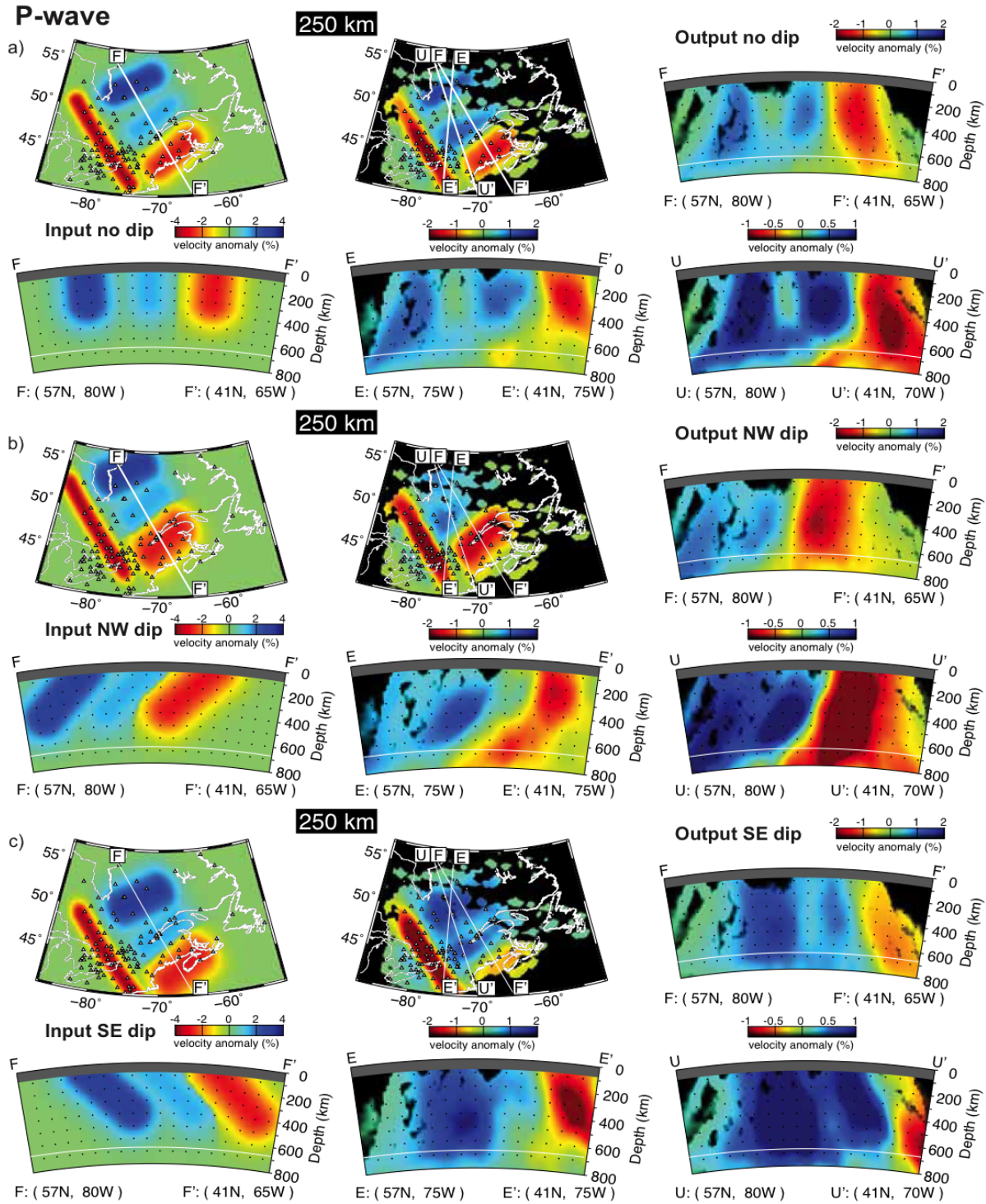


Figure 3.11: **P-wave structural resolution test** for dipping tectonic interfaces. Anomalies are defined as in Figure 3.10 but instead have 45° dip to the northwest (b) and southeast (c). Station locations are shown as white triangles. Regions with low (<3) ray hits and the top 75 km in cross-section are muted due to a lack of crossing rays and poor resolution.

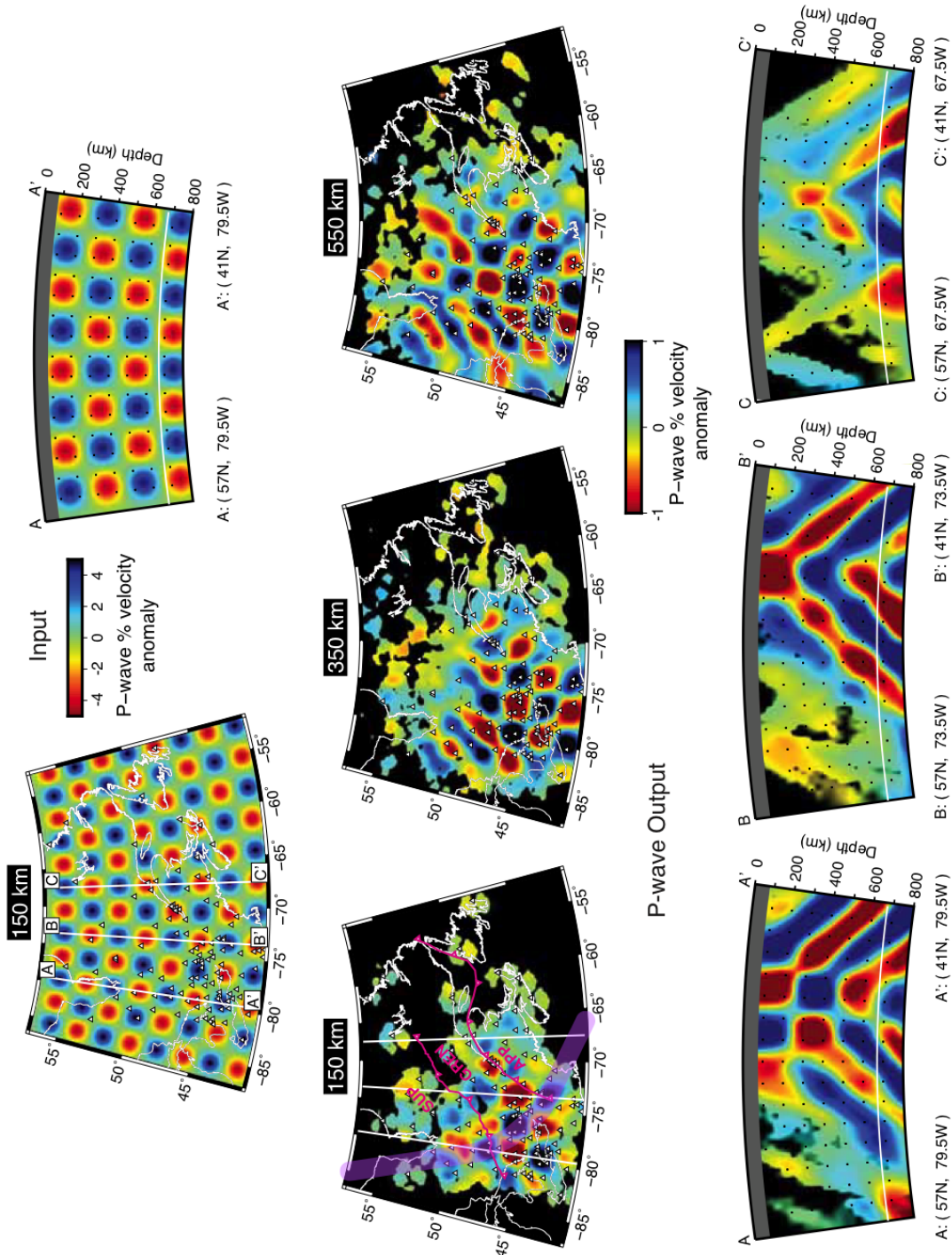


Figure 3.12: **P-wave checkerboard resolution tests** with 75 km wide input anomalies of $\delta V_P = \pm 5\%$ defined by Gaussian functions across their widths, placed at 150 km, 350 km, 500 km, 750 km, 950 km depths. Also shown are station locations (white triangles), tectonics (magenta lines) and the GMH track (lilac band). Regions with low (< 3) ray hits and the top 75 km in cross-section are muted due to a lack of crossing rays and poor resolution.

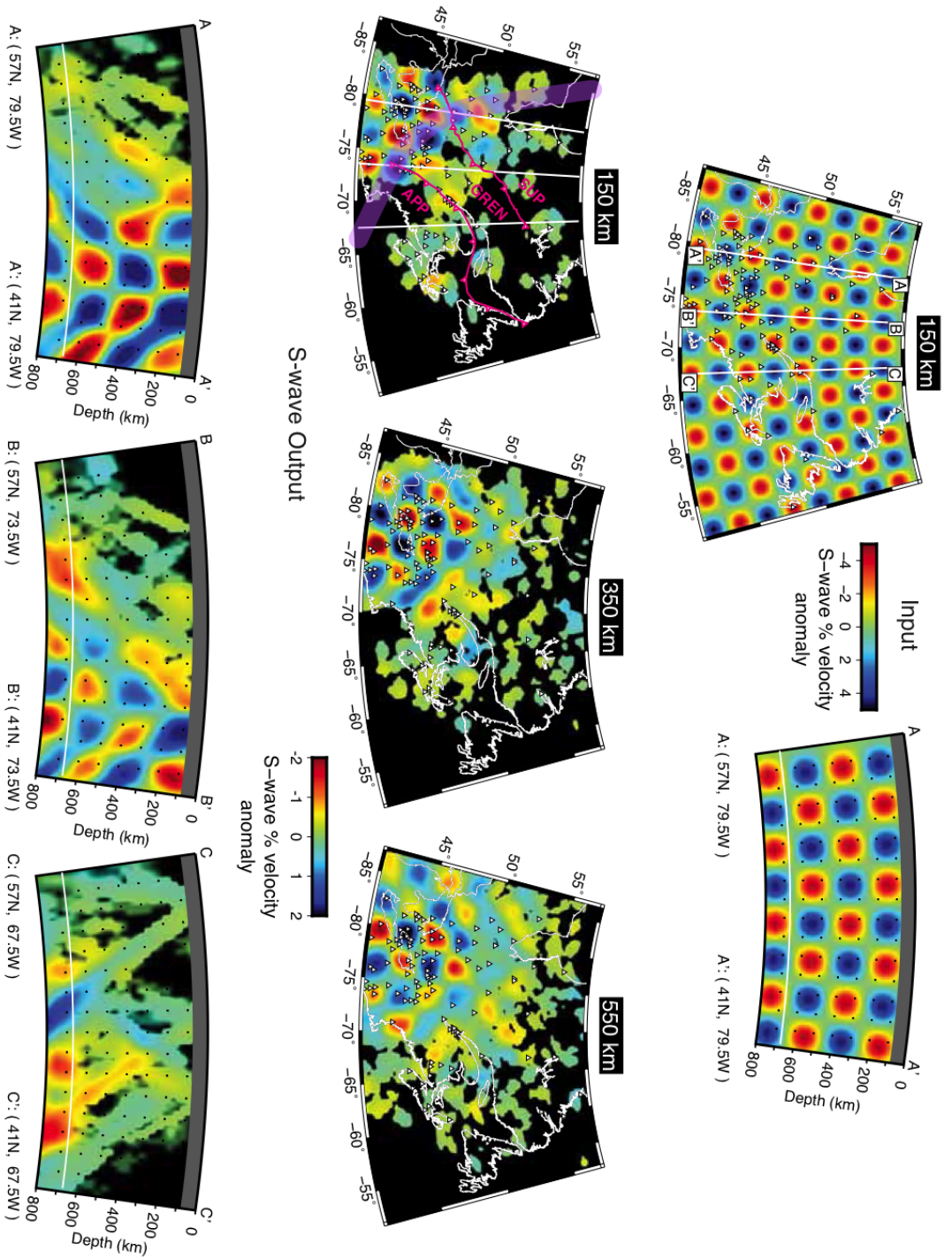


Figure 3.13: **S-wave checkerboard resolution tests** with 75 km wide input anomalies of $\delta V_S = \pm 5\%$ defined by Gaussian functions across their widths, placed at 150 km, 350 km, 500 km, 750 km, 950 km depths. Also shown are station locations (white triangles), tectonics (magenta lines) and the GMI track (ilac band). Regions with low (< 3) ray hits and the top 75 km in cross-section are muted due to a lack of crossing rays and poor resolution.

3.5 Tomographic models

We present relative arrival-time tomographic models in which velocity perturbations are relative to the regional mean (e.g., *Bastow, 2012*), however our wide-aperture network ($\sim 3000 \times 2000$ km) spans regions of both genuinely fast and slow mantle wavespeeds (e.g., *Li et al., 2008a*; *Ritsema et al., 2011*). Relative arrival-time studies can be benchmarked on the global scale with the aid of global tomographic models (e.g., *Bastow, 2012*). In the upper 400 km of the S40RTS global model *Ritsema et al. (2011)*, a mean fast wavespeed anomaly of $\delta V_S = +1.24\%$ exists below -76°E , 46°N , where our stations have near-zero relative arrival-time residual. Peak-to-peak variation within our relative arrival-time dataset (Figure 3.4) is ± 0.85 s and ± 3.25 s, for P- and S-wave respectively. Thus, while the mantle in SE Canada is generally fast compared to the global mean, some of the slowest anomalies in our model could also be slow compared to the global average.

3.5.1 P-wave model

The P-wave tomographic model (Figure 3.14) shows heterogeneous wavespeed structure beneath SE Canada. East of James Bay, within the Superior craton, a pronounced, laterally robust (Figure 3.10), fast wavespeed (A - $\delta V_P = 0.8\%$) structure extends to ~ 350 km apparent depth showing a vertical southern boundary (Figure 3.14). A series of faster wavespeed anomalies (B - $\delta V_P = 0.6\%$), trending approximately W-E extends ≥ 300 km from the Central Metasedimentary Belt (CMB) to the Appalachian Front (Figure 3.14). This fast wavespeed signature widens and moves to the north with depth, and appears to extend to ~ 450 km apparent depth. Fast wavespeeds also exist beneath Nova Scotia to ≥ 350 km apparent depth. At ≥ 550 km model depth, a pervasive fast wavespeed structure is seen throughout the entire western portion of the study area (Figure 3.14). A pronounced, well resolved (Figure 3.10), slow wavespeed anomaly (C) stretches from the southern margin of the study area at the Appalachian coast into the Grenville Province, trending WNW-NW. This anomaly has wavespeed perturbations of $\delta V_P \leq 1.0\%$ that broaden and migrate to the SW with depth (Figure 3.14). At model depths ≥ 250 km, this transitions into a broad slow wavespeed signature beneath the Appalachian terranes. The lithospheric anomalies at model depths ≤ 150 km arguably stretch into the Superior craton, shown at approximately 50°N , 82°W . At model depths of 350-450 km, faster wavespeeds are seen in the Superior, average in the Grenville and generally slower wavespeeds in the Appalachians (Figure 3.14).

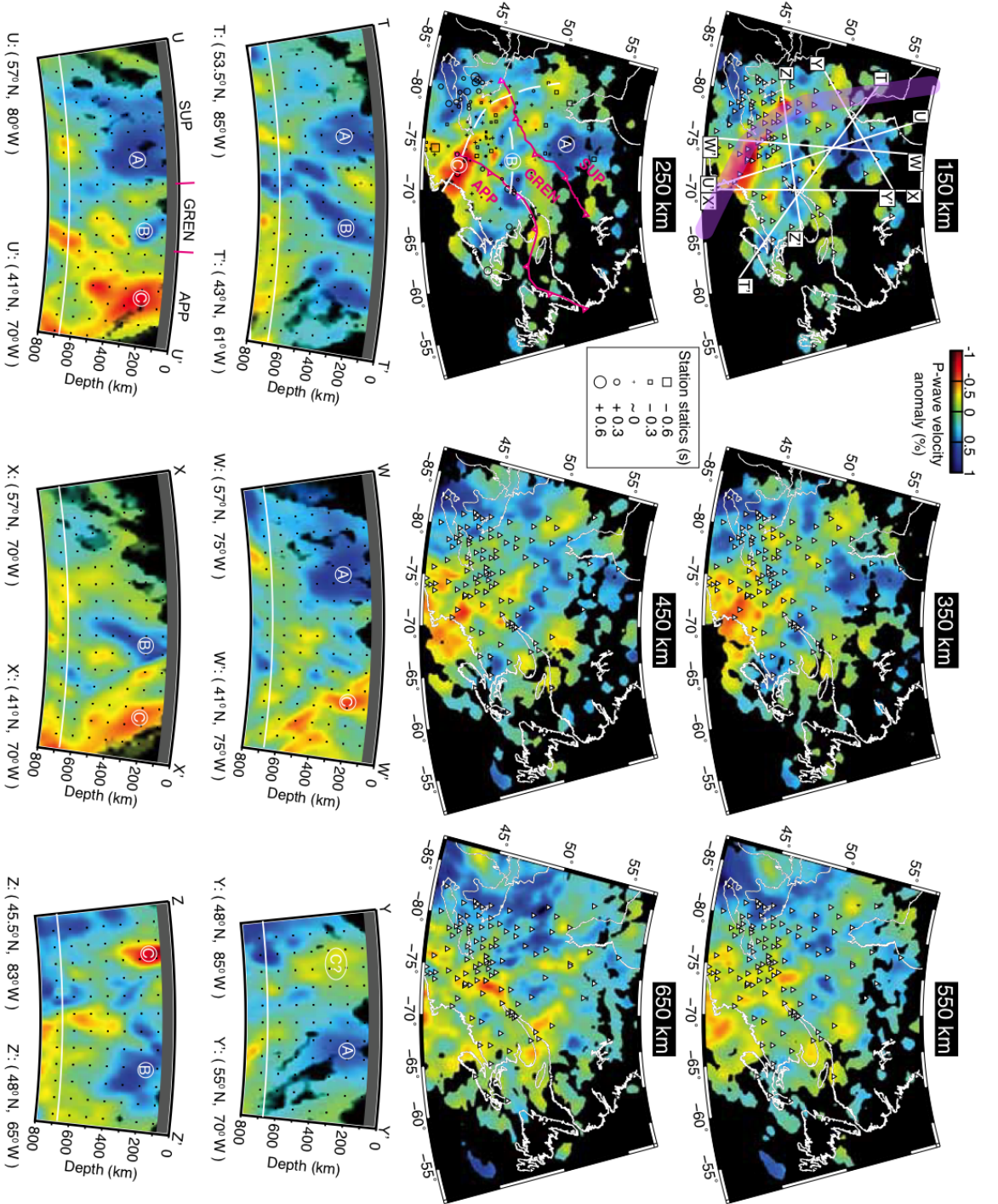


Figure 3.14: **P-wave model depth slices and cross-sections.** Station locations are shown as white triangles and the GMH track as a lilac band. The 250 km map also shows station statics, regional tectonics (magenta) and specific anomalies (A, B, C) discussed in the text. Regions with low (<3) ray hits and the top 75 km in cross-section are muted due to lack of crossing rays and poor resolution.

3.5.2 S-wave model

The S-wave model (Figure 3.15) displays similar features to the P-wave model, despite lower ray hits in the eastern portion. Again, east of James Bay, within the Superior craton, a fast wavespeed anomaly (A - $\delta V_S \sim 1.5\%$) extends to ≤ 350 km apparent depth with a vertical southern boundary (Figure 3.15). Within the Grenville, a strong fast wavespeed structure (B) exists beneath the CMB (150-350 km apparent depth). In contrast to the P-wave model, strong fast wavespeed anomalies stretching to the east, towards the Appalachian Front, are not seen in the S-wave model (Figure 3.15). In the west, a pervasive fast wavespeed ($\delta V_S \sim 1.5\%$) anomaly appears below 600 km model depth (Figure 3.15).

A series of slow wavespeed anomalies ($\delta V_S = -0.8\%$) is seen at ~ 250 km apparent depth, within the Superior craton, at around 50°N , 82°W , and south of the Grenville boundary, at 46°N , 80°W (Figure 3.15) collocated with the P-wave model. To the SE, especially at ≤ 350 km model depth, a $\delta V_S = -1.0\%$ anomaly (C) terminates abruptly at the Appalachian Front, but does not transition into a broad slow wavespeed structure below this depth.

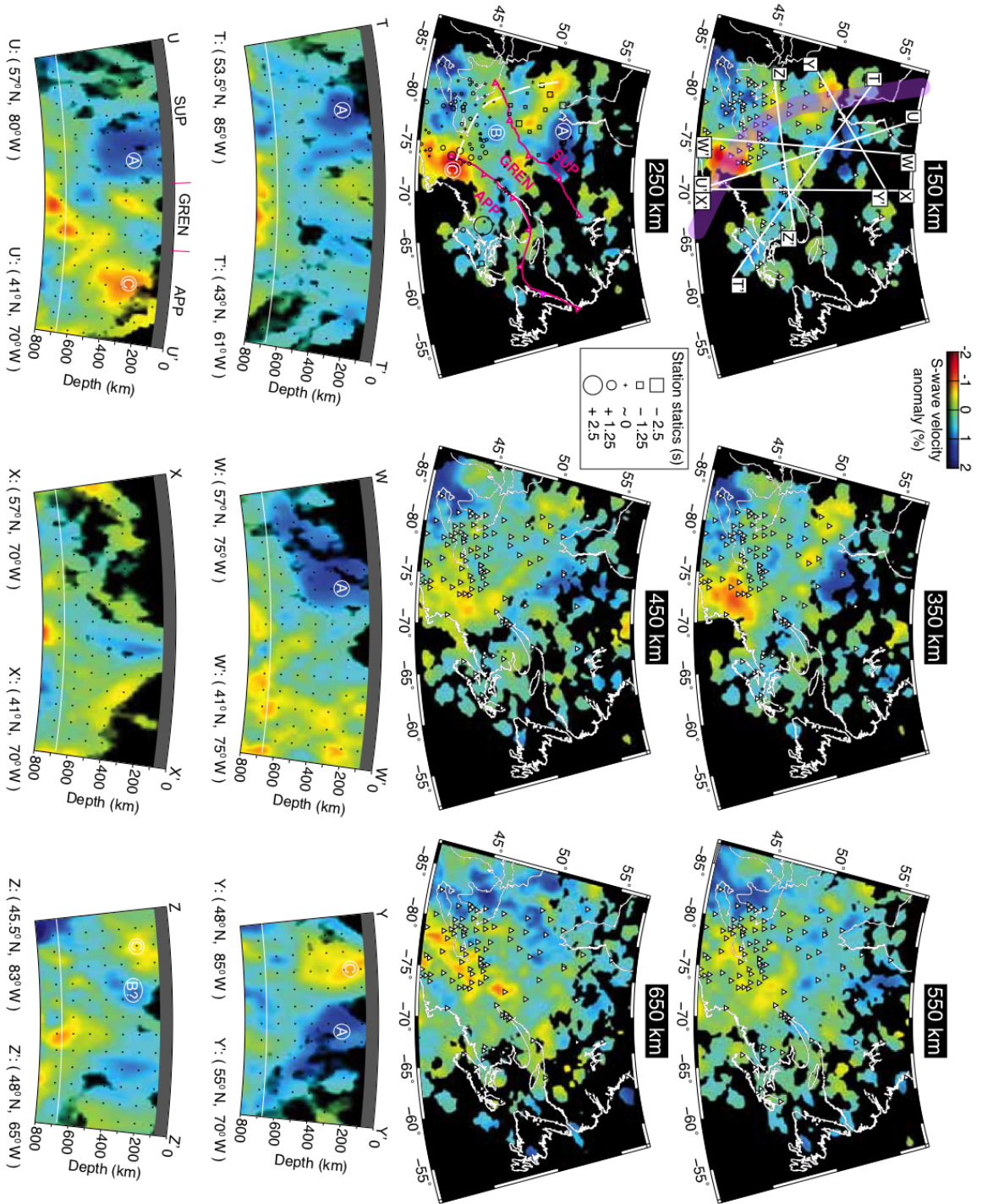


Figure 3.15: S-wave model depth slices and cross-sections. Plotting conventions as for Figure 3.14.

3.6 Discussion

The North American continent has been shaped by some of the largest tectonic events seen on Earth (e.g., the Grenville Orogeny). The features of our models allow us to address some fundamental questions concerning these events. We limit our interpretations to the lithospheric-scale, upper mantle above ~ 400 km depth where our resolution is best (Figure 3.10).

3.6.1 Age-progressive lithospheric subdivisions

The Superior craton is broadly dominated by early arrivals, the Grenville Province by approximately average arrival-times and the Appalachian terranes by generally later arrivals when compared to the regional mean (Figure 3.4). The residual histograms (Figure 3.4), grouped by tectonic domain, also point towards this pattern as the peaks shift towards slower arrivals with decreasing age. Further evidence is provided by the plots of residual against backazimuth (Figures 3.2, 3.3) and the increase in residuals with distance along transect F-F' (Figure 3.4).

In our tomographic models (Figures 3.14 and 3.15), the Superior craton is dominated by faster wavespeeds, while the Grenville Province contains weaker anomalies of both positive and negative amplitudes. Beneath the Appalachians, slower wavespeeds dominate, particularly in the south. Sampling the S-wave model (P-wave model produces similar results) at 200 km model depth along a transect U-U' (Figure 3.16) also confirms an age dependence of seismic wavespeed across the region.

Variations in lithospheric thickness (e.g., *Bedle and van der Lee, 2009*) and strong station statics beneath Nova Scotia and New Brunswick (Figures 3.14 and 3.15), for example, may influence the seemingly clear age-wavespeed trend. Furthermore, recent tectonism, largely associated with the passage of the GMH ca. 120 Ma ago and the subsequent eruption of kimberlite magmas (Figure 3.1 *Heaman and Kjarsgaard, 2000*), may have altered the lithosphere's Precambrian composition. A series of pervasive low wavespeed anomalies in the western part of the models coincides with the GMH track, cross-cutting the Appalachian and Grenville Fronts (Figures 3.14 and 3.15 - anomaly C). This agrees with lithospheric observations made by many previous authors (e.g., *Aktas and Eaton, 2006; Rondenay et al., 2000; van der Lee and Frederiksen, 2005; Villemaire et al., 2012*).

Globally, much variation is seen in the Proterozoic domains that bound Archean lithospheric blocks. In Asia (*Lebedev et al., 2009*) and southern Africa (*Fouch, 2004; James et al., 2001; Ritsema and Van*

Heijst, 2000), a clear distinction in Archean and Proterozoic lithospheric wavespeeds shows a direct link to surface geological provinces. In Australia and Fennoscandia however, Archean mantle lithosphere is thought to underlie the Proterozoic domains (*Bruneton et al.*, 2004; *Fishwick et al.*, 2005; *Sandoval et al.*, 2004; *Simons et al.*, 2002), since no systematic wavespeed heterogeneity exists. In north China, extensive lithospheric modification perhaps conceals a link between lithospheric wavespeed variation and surface provinces (*Santosh et al.*, 2010).

Results from North America are variable. *Villemaire et al.* (2012) see no clear distinction between Precambrian terranes, whereas *Nettles and Dziewonski* (2008) argue for a correlation of lithospheric wavespeed with surface geological age. Beneath Hudson Bay, *Darbyshire et al.* (2013) also note decreased wavespeeds below the Proterozoic terranes with respect to the Archean domains at lithospheric depths. *Frederiksen et al.* (2013) show that the western Superior lithospheric expression is misaligned with the surficial boundary, perhaps linked to modification during the Trans-Hudson Orogeny at ca. 1.8 Ga ago (*Hoffman*, 1988). While the large time period (~ 0.8 Ga) between the Grenville and Trans-Hudson Orogen means caution must be taken when drawing comparisons, both have characteristics akin to the modern-day Himalayas (e.g., *Hynes and Rivers*, 2010; *St-Onge et al.*, 2006), suggesting similar processes may have operated in these two orogenies. Our observation of three domains of wavespeed (Figures 3.4, 3.14, 3.14 and 3.16) agrees with those seen in Asia, southern Africa and the Trans-Hudson Orogen (e.g., *Darbyshire et al.*, 2013; *James et al.*, 2001; *Lebedev et al.*, 2009). Whether this can be attributed to secular changes in plate-forming processes over time (e.g., *Pearson et al.*, 1995) is debatable, given the variability of worldwide observations and potential influence of post-formation modification (e.g., *Frederiksen et al.*, 2013; *Santosh et al.*, 2010). In the following section, we explore potential causes for the observed seismic heterogeneity further.

3.6.2 Causes of seismic heterogeneity

The long-term tectonic stability of our study region suggests that anelasticity and partial melt have little effect on seismic velocities beneath the Canadian Shield (e.g., *Sobolev et al.*, 1996). The most recent magmatic episode is associated with the passage of the GMH at ca. 190-110 Ma ago. *Eaton and Frederiksen* (2007) performed thermal diffusion calculations for a plume-head to estimate the thermal signature of the GMH within the lithosphere. A 500°C temperature anomaly decays to a diffuse anomaly of 80°C at 200 km depth over 120 Ma. Such a temperature anomaly could account for a significant portion ($>50\%$) of the wavespeed anomalies (e.g., *Goes et al.*, 2000) within our models

(Figures 3.14 and 3.15 - anomaly C). However, even mantle potential temperature anomalies associated with large igneous provinces are generally significantly lower than 500°C (e.g., 170°C for the 30 Ma Ethiopian traps: e.g., *Rooney et al.*, 2012). We thus consider thermal anomalies to have negligible impact on our tomographic images.

Anisotropy can also play a role in wavespeed variations (e.g., *Babuška and Plomerová*, 2006). However, the consistency between our P- and S-wave studies with approximately perpendicular particle motions, in both the tomographic images (Figures 3.14 and 3.15) and residual histograms (Figures 3.2, 3.3, 3.5), and our good backazimuthal coverage of earthquakes (Figures 3.2, 3.3), implies that our models are dominated by isotropic wavespeed variations. Moreover, shear-wave splitting studies in the region (e.g., *Darbyshire et al.*, 2015; *Fouch et al.*, 2000; *Frederiksen et al.*, 2006; *Levin et al.*, 1999; *Yuan and Levin*, 2014) calculate a mean delay time (δt) between orthogonally polarised shear-waves of ≤ 1.0 s. This is low relative to more tectonically active areas where splitting measurements, δt , are often ≥ 1.5 s (e.g., *Martin-Short et al.*, 2015).

At the base of the lithosphere, variation in heat flow across the tectonic domain boundaries is negligible (*Mareschal and Jaupart*, 2004) so any resulting long wavelength anomalies (>1000 km) associated with the lithospheric mantle would be too large for this study to capture. Surface heat flow measurements are thought to be dominated by crustal radiogenic heat production (*Jaupart and Mareschal*, 2010). Intense surface erosion of the cratons, and associated loss of radionuclides, means their heat flow is anomalously low (generally $\leq 50 \text{ mWm}^{-2}$; e.g., *Lévy et al.*, 2010). Cold cratonic cores are thus often $>200^\circ\text{C}$ cooler than surrounding, younger terranes (*Mareschal and Jaupart*, 2004). Such thermally derived features would also be expected to be diffuse, in contrast to the abrupt lateral wavespeed boundaries within our models (Figures 3.14 and 3.15).

Eastern North America is thought to require iron depletion to explain the observed wavespeed variations (*Goes and van der Lee*, 2002). Depletion of the mantle at lithospheric depths also decreases the concentration of radionuclides, resulting in lower temperatures and increasing velocities further still (*Hieronymus and Goes*, 2010). This additional compositional effect can cause wavespeed variations of up to $\delta V_P \approx 4\%$ over a few kilometres and provides a better explanation for the steep gradients in our velocity models than thermal heterogeneity (Figures 3.14 and 3.15).

We suggest that the GMH slow wavespeed signature is dominated by a compositional element that has decreasing impact on tectonic domains (Figures 3.14 and 3.15 - anomaly C) of increasing age

(*Villemaille et al.*, 2012). We observe a southwestward offset (cf. *Eaton and Frederiksen*, 2007) of the lithospheric anomaly with depth, collocated with other studies (e.g., *Villemaille et al.*, 2012) and thus we favour the view that the hotspot has exploited rift-weakened lithosphere within the shield (*Aktas and Eaton*, 2006; *Rondenay et al.*, 2000; *Villemaille et al.*, 2012). Our improved resolution over *Villemaille et al.* (2012) also allows us to image the low wavespeed anomaly in the Superior craton (Figures 3.14 and 3.15 - anomaly C) which *Frederiksen et al.* (2013) attributed to cratonic modification by the GMH. We argue that Archean lithosphere is thus more resistant to thermochemical erosion during hotspot tectonism than younger Proterozoic and Phanerozoic lithosphere, but not completely immune to modification - mechanically strong does not mean geochemically immune. Keels may thus not be entirely stagnant features as traditionally thought. Next, we discuss these ideas in the context of keel formation and development, with emphasis on processes that occurred during the Grenville Orogeny.

3.6.3 Formation and evolution of lithospheric keels

The formation mechanism of cratonic lithosphere is debated (e.g., *Lee et al.*, 2011); in particular the development of their deep-seated (~ 250 km), long-lived keels (e.g., *Bedle and van der Lee*, 2009), that may (e.g., *Schaeffer and Lebedev*, 2014; *Simons et al.*, 2002; *Yuan et al.*, 2011), or may not (e.g., *James et al.*, 2001), extend beneath both Proterozoic and Archean terranes. Distinct wavespeed anomalies beneath the Superior craton and Grenville Province are separated by a vertical wavespeed boundary (Figures 3.14 and 3.15) that extends to ~ 400 km apparent depth. This feature is vertically smeared however (Figure 3.10), so the true extent is more likely confined to the region's ~ 250 km thick lithosphere (e.g., *Bedle and van der Lee*, 2009). When considered against the background of absolute wavespeed models (e.g., *Ritsema et al.*, 2011; *Schaeffer and Lebedev*, 2014), the lateral isotropic velocity variation (from fast to average) at the Grenville Front (Figures 3.2, 3.3, 3.4, 3.14, 3.15 and 3.16) may arise from a change in chemical depletion between Archean and Proterozoic provinces, as suggested by *Darbyshire et al.* (2013) in the Trans-Hudson Orogen and other authors worldwide (e.g., southern Africa: *Begg et al.*, 2009; *Youssof et al.*, 2015).

While our tomographic models do not require a layered lithosphere to explain them (body-wave tomography would be a poor way to constrain sub-horizontal interfaces such as an MLD or LAB), they are consistent with the expectation of a lower velocity region beneath a core of highly depleted Superior cratonic lithosphere that is present at shallower depths beneath the Grenville Province (*Yuan*

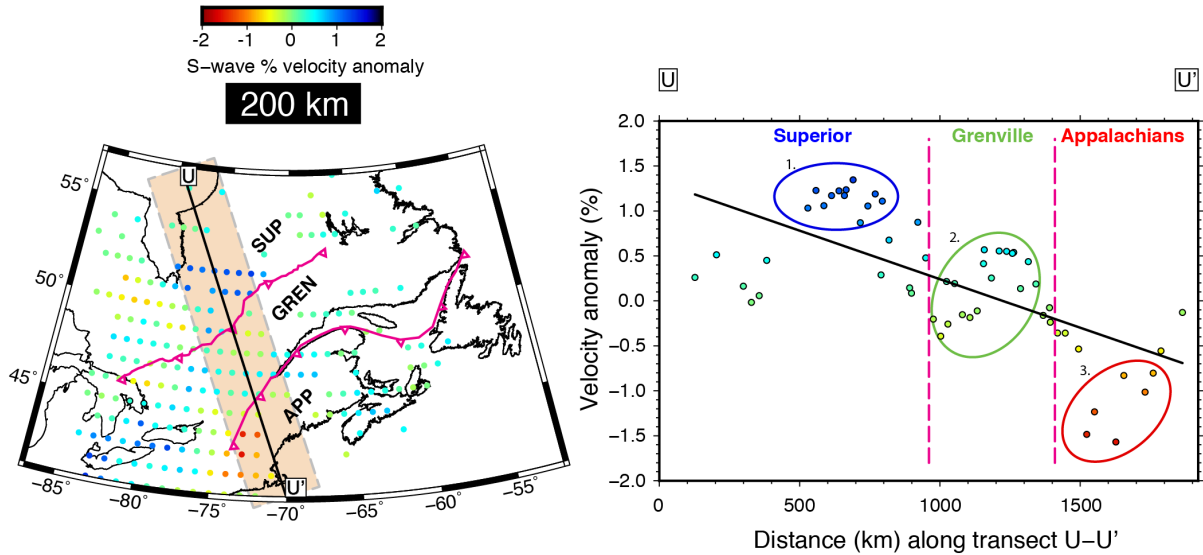


Figure 3.16: **Left - S-wave tomographic model sampled at one degree intervals at 200 km depth showing the transect U-U' cross-cutting the Grenville and Appalachian Fronts with a 400 km wide band overlain. Right - velocity anomaly at all points within 200 km of the transect U-U' (within the orange band) with respect to distance from the northern end of the line.** Points are grouped to aid interpretation where (1) indicates the Superior cratonic interior, (2) the mean wavespeed Grenville and (3) the ‘divot’ within the lithosphere caused by hotspot-driven thermal-mechanical erosion. Black line: linear regression through the data.

and Romanowicz, 2010). This supports the model for two-stage formation of lithospheric keels (e.g., Darbyshire *et al.*, 2013; Yuan and Romanowicz, 2010) that predicts a fast wavespeed Archean core formed, followed by the accretion of slightly slower wavespeed material beneath the Archean and Proterozoic provinces. In a purely lateral sense, we may reliably sample the decrease in chemical depletion across the Grenville Front in the model of Yuan and Romanowicz (2010), suggesting that keel stabilisation was not restricted to Archean times. However, as discussed below, this wavespeed variation may also be interpreted as evidence for metasomatic enrichment of Archean material rather than a less depleted composition derived from Proterozoic keel formation.

3.6.4 Implications for the Grenville Orogeny

Our observed vertical wavespeed boundary between the Superior craton and Grenville Province (Figures 3.14 and 3.15) is not easily reconciled with previous studies spanning the Grenville Front. LITHO-PROBE (e.g., Clowes, 2010) wide-angle seismic images (e.g., Ludden and Hynes, 2000; White *et al.*, 2000) (Figure 3.17A) reveal Superior material under-thrusting the Grenville Province by >300 km to the southeast, a feature thought to intersect the Moho and continue dipping into the mantle. This is

consistent with general models for lithospheric growth at the margins of cratons (e.g., *Snyder, 2002*) and those specific to Grenville Province development (*Rondenay et al., 2000*). Such models invoke a split incoming plate of Proterozoic age in which an upper crustal layer is thrust above the older cratonic Archean material, while the deeper layer is subducted. Furthermore, *Hanmer et al. (2000)* suggested southeast-oriented subduction occurred beneath Amazonia during the Grenvillian collision with Laurentia (*Condie, 2001; Hoffman, 1999*). Alternatively, *Rivers (1997)* and *Rivers and Corrigan (2000)* advocate a protracted period of subduction (~ 300 Ma) in the opposite direction beneath the Laurentian margin, leading to enrichment of Grenvillian mantle material (*Chiarenzelli et al., 2010*). This wealth of often-conflicting research suggests that, drawing from Section 3.6.1, our results may best be interpreted in light of lithospheric modification during, or after, the Grenville Orogeny, rather than secular variation in plate formation processes. In the following sections, we explore the observational and theoretical evidence for craton edge-modification, before presenting a new model for the development of the Grenville Orogen (Figure 3.17).

Mechanisms for craton-edge modification

Growing evidence suggests that under certain conditions, cratonic lithosphere can be modified by subduction zone interaction or hotspot tectonism (e.g., *Nolet and Zielhuis, 1994; Rondenay et al., 2000*). For example, both convection in the mantle wedge (*Menzies et al., 2007*), and metasomatism of the overriding material by fluids derived from the down-going slab (*Kusky et al., 2007*), have been used to explain destabilisation of the North China craton. Such theories invoke processes related to the stagnant slab associated with the subduction of the Pacific plate. In southern Africa (e.g., *Youssof et al., 2015*, and references therein) and Europe (*Nolet and Zielhuis, 1994; Pedersen et al., 2013*), authors have also argued that subduction beneath the craton edge has caused modification of the cratonic lithosphere, through metasomatism. Geodynamical modelling supports the view that metasomatic enrichment of the lithospheric keel is a plausible cause of weakening or modification (e.g., *Wang et al., 2015*). In central Canada, *Frederiksen et al. (2013)* noted that the western margin of the Superior craton at lithospheric depths does not correlate with surface geology, implying the keel was eroded or modified during the Trans-Hudson Orogeny. Additionally, *Eaton and Frederiksen (2007)* and *Kaban et al. (2015)* proposed that sublithospheric mantle flow patterns have modified the keel since the opening of the Atlantic.

Viable mechanisms to produce the vertical wavespeed boundary beneath the Grenville Front (Figures

3.14 and 3.15) are limited, since sub-vertical or shallower plate-scale processes dominate traditional collisional settings. However, craton edge modification may involve small-scale convection (e.g., *Bao et al.*, 2014), gravitational destabilisation of the lithosphere (e.g., *Kusky et al.*, 2007) and broad-scale metasomatism (e.g., *Chiarenzelli et al.*, 2010; *Nolet and Zielhuis*, 1994; *Pedersen et al.*, 2013; *Youssof et al.*, 2015) associated with subduction beneath the cratonic margin.

Bao et al. (2014) documented an abrupt wavespeed transition from craton to Cordillera in western Canada which they attributed to lithospheric modification through subduction zone interaction. They postulated that a protostep in the lithospheric base initiated small-scale convection and thus exploited the zone of weakness between the craton lithosphere and the thickened continental margin. This could produce a comparable signature to that seen beneath the Archean/Proterozoic boundary in our models (Figures 3.14 and 3.15). However, it is unclear why a broadly homogeneous Superior (or Laurentian) lithosphere would have had an existing, exploitable protostep or vertical weak boundary, so this is unlikely to explain our observations. Alternatively, a vertical boundary in wavespeed could result from metasomatism of the cratonic wedge above a subducting plate (e.g., *Chiarenzelli et al.*, 2010; *Kusky et al.*, 2007; *Nolet and Zielhuis*, 1994; *Pedersen et al.*, 2013; *Youssof et al.*, 2015). This process can result in lithospheric gravitational destabilisation (e.g., *Kusky et al.*, 2007), whereby the vertical wavespeed boundary may delineate the detachment front of the foundering lithosphere. However, a foundering lithospheric wedge may interact with the down going slab, and would not manifest in the present-day seismological signature. Lithospheric wavespeeds beneath the Grenville would be expected to be similar to those beneath the Appalachians in this case. We therefore suggest that broad-scale metasomatism (e.g., *Chiarenzelli et al.*, 2010; *Nolet and Zielhuis*, 1994; *Pedersen et al.*, 2013), pervasively lowering seismic wavespeed (e.g., *O'Reilly and Griffin*, 2006), is the most likely mechanism for craton-edge modification in SE Canada.

Model for Grenville Province development

Finally, we propose a new model for Grenville Province development (Figure 3.17), to reconcile our seismological observations (Figures 3.14 and 3.15) with geological constraints (*Chiarenzelli et al.*, 2010; *Hanmer et al.*, 2000; *Hynes and Rivers*, 2010; *Ludden and Hynes*, 2000; *Rivers and Corrigan*, 2000) and studies of tectonic processes at cratonic margins (e.g., *Nolet and Zielhuis*, 1994; *Pedersen et al.*, 2013; *Rondenay et al.*, 2000; *Snyder*, 2002). Convergence between Laurentia and Amazonia (*Condie*, 2001) initiated a >300 Ma Andean style subduction system (*Rivers and Corrigan*, 2000) dipping to

the NW (present day coordinates). As the slab descends, assuming the mantle wedge is sufficiently saturated and corner flow convection is weak (*Iwamori, 1998; Peacock, 1990*), fluids rise approximately vertically (Figure 3.17C). *Schmidt and Poli (1998)* used experimental results to propose that under typical conditions, the majority of fluid release occurs when the slab is ≤ 200 km deep. Thus, depending on slab dip, there is a lateral bound on this process that extends from the trench into the Laurentian margin - this is the metasomatic front.

Fluid infiltration of the cratonic wedge enriches the Grenvillian mantle (*Chiarenzelli et al., 2010; Hynes and Rivers, 2010*) as a result of metasomatism (Figure 3.17C, D) (e.g., *Kusky et al., 2007; Nolet and Zielhuis, 1994*). This lowers the wavespeed (e.g., *O'Reilly and Griffin, 2006*) up to the metasomatic front within the Laurentian margin, directly below the modern-day surface expression of the Grenville Front (Figures 3.14 and 3.15). While the collocation of the Grenville and metasomatic fronts may be slightly fortuitous (due to the inherent control of slab geometry), we speculate that the Grenville Orogeny did not propagate further into the North American continent because it encountered the stronger, unmodified Superior lithosphere. This would provide an explanation for the abrupt change in metamorphic grade across this tectonic boundary (e.g., *Hynes and Rivers, 2010*).

Initially, subduction proceeds beneath Laurentia (e.g., *Rivers and Corrigan, 2000*), but reverses in direction prior to the continental collision (*Hynes and Rivers, 2010*). This positions Amazonia as the overriding plate, allowing the reworked material between the two continental cores to be thrust above the Archean-age, metasomatised Laurentian cratonic wedge or indenter (in agreement with observations of *Hanmer et al., 2000; Ludden and Hynes, 2000; Rondenay et al., 2000; Snyder, 2002*) (Figure 3.17D, E). Although the order of subduction phases could be reversed (e.g., *Snyder, 2002*), due to the timescale (≥ 300 Ma) over which the Laurentian margin formed the overriding plate in the subduction system (e.g., *Rivers, 1997*) and the likelihood that closure between Amazonia and Laurentia would have been initiated with oceanic subduction, it seems that our model (Figure 3.17) presents the most likely option.

Anomaly B in our tomographic images (Figures 3.14 and 3.15) is consistent with a reversal in subduction polarity, although this may be at the limit of our resolution (see checkerboard tests (Figures 3.12, and 3.13) in section 3.4). A reversal in subduction polarity could be caused by buoyant crustal material jamming the trench (*Stern, 2004*) of an existing subduction zone, feasibly provided by significant arc magmatism preceding the continental collision (*Hynes and Rivers, 2010*). A modern example of this is found at the Miocene Solomon arc (*Stern, 2004*). This is supported by the pre-Grenvillian

southeast-dipping feature noted by *Aktas and Eaton* (2006) and *Adetunji et al.* (2014), and closure of a back-arc proximal to the CMB prior to termination of the Grenville Orogeny (*Rivers and Corrigan*, 2000). Additionally, the collision of the two continents and the accreted arcs (*Hynes and Rivers*, 2010), plus the metasomatism from two subduction intervals would allow for the notable heterogeneity in wavespeed within the Grenville Province (Figures 3.14 and 3.15).

After the termination of the Grenville Orogeny, the reduced wavespeed region of depleted Grenvillian material (and perhaps some reworked collisional remnants) begins to cool and may allow the formation of a thermal boundary layer, a common feature of typical convective systems (e.g., *Parsons and McKenzie*, 1978). The lithospheric keel beneath both Superior and the Grenville is thus extended in depth (Figure 3.17F) forming the lower layer in the model of *Yuan and Romanowicz* (2010). Our tomographic models do not require this feature, but keel stratification is widely recognised globally (e.g., *Darbyshire et al.*, 2013; *Griffin et al.*, 2003; *Simons et al.*, 2002; *Yuan and Romanowicz*, 2010). Our model therefore leans towards lateral accretion of terranes as the main driver for formation of the greater Canadian Shield but implies that vertically-driven processes dominate the subsequent pervasive alteration.

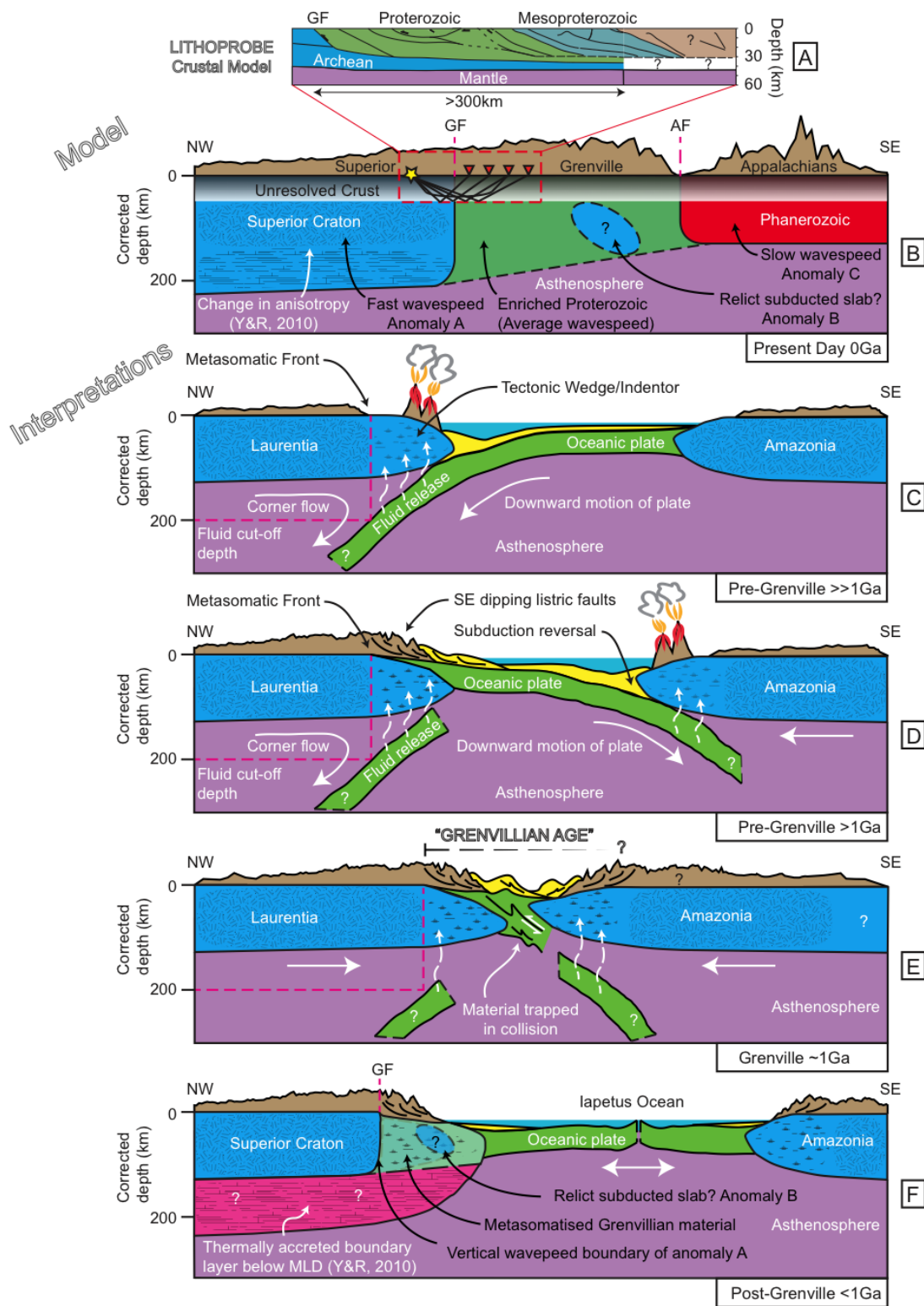


Figure 3.17: **Schematic tectonic interpretation** (panels C,D,E,F) used to explain the presented geophysical and geological datasets (panels A,B). The LITHOPROBE seismic profile (panel A) across the Grenville Front (GF) is adapted from the interpretation of *Ludden and Hynes* (2000), derived from the GLIMPCE-J line, Abitibi-Grenville lines 31, 32, and 33 and Lake Ontario line 71 (after *White et al.*, 2000). Y&R, (2010) refers to *Yuan and Romanowicz* (2010) who proposed a Mid-lithospheric discontinuity (MLD) beneath the craton by imaging a change in anisotropy between 50 and 160 km depth. Anomalies are labeled as in Figures 3.14 and 3.15.

3.7 Conclusion

Using the highest resolution P- and S-wave tomographic models of southeast Canada to date, we show three broad regions of increasing wavespeed, a slow Phanerozoic Appalachian terrane, an average wavespeed Grenville Province and a fast wavespeed Superior cratonic interior. Differing wavespeeds within the Precambrian domains could imply markedly different processes acting in the Archean. A low wavespeed series of anomalies correlates well with the Great Meteor hotspot track. The anomaly amplitudes decrease towards the cratonic interior, implying increased resistance to modification with age of tectonic domain. In a purely lateral sense, the extent of faster wavespeeds within the shield matches that predicted by the two-stage formation model for lithospheric keels: the chemically depleted core formed in the Archean and the thermally accreted, less depleted boundary layer formed in the Proterozoic. Our results, therefore, may support the idea that keel formation was not restricted to Archean times. However, to explain the pronounced vertical wavespeed boundary beneath the Grenville Front and previously published results, we instead propose a model for modification of the Laurentian margin due to subduction beneath the continent. In this model, we invoke Palaeoproterozoic metasomatic fluid release to decrease the wavespeed of the material beneath the Grenville Province; consequently a vertical wavespeed boundary beneath the Grenville Front is seen today. Subduction driven modification of the North American craton edge suggests that cratons may not necessarily be entirely stagnant features over geological time. Apparent variations in mantle wavespeed between Archean and Proterozoic domains worldwide, may thus be explained better by post-formation metasomatic enrichment than by variations in iron depletion levels at the time of plate formation.

Chapter 4

The Absolute Arrival-time Recovery Method (AARM)

The material in this chapter can be found in an article published in the Bulletin of the Seismological Society of America (*Boyce et al.*, 2017).

4.1 Chapter overview

Dense, short term deployments of seismograph networks are frequently used to study upper mantle structure, as seen in Chapter 3. However, these recordings of teleseismic waveforms are often of lower signal-to-noise ratio (SNR) than those recorded at permanent observatory sites. Therefore, waveform coherency across a network is frequently utilised to calculate relative arrival-times between recorded traces, but these measurements cannot easily be combined or reported directly to global absolute arrival-time databases. These datasets are thus a valuable but untapped resource with which to fill spatial gaps in global absolute wavespeed tomographic models.

In this Chapter I develop the Absolute Arrival-time Recovery Method (AARM) to retrieve absolute time picks from relative arrival-time datasets, working synchronously with filtered and unfiltered data. This enables multiple networks from southeast Canada to be combined in a single tomographic inversion later in this thesis. The methodology utilises waveforms aligned for relative arrival-time studies (Chapter 3) and through a weighted stacking approach, enables the calculation of absolute arrival-times and associated estimates of uncertainty. I test AARM on a published dataset (see Chapter

3 - *Boyce et al.*, 2016) from southeast Canada and the available equivalent earthquake-station pairs on the ISC database. The method is also tested on P- and S-wave synthetic data of variable noise levels to assess the performance limits.

4.2 Introduction

Over the past two decades, the number of temporary, regional seismic networks deployed to study upper mantle structure has grown dramatically (*Evans et al.*, 2015). However, due to sub-optimal deployment conditions, teleseismic waveforms of variably emergent nature are generally of lower signal-to-noise (SNR) ratio than those recorded from sparse permanent observatory sites. Fortunately, temporary networks are often small enough in aperture (often $< 1000\text{ km}$) to capitalise on waveform coherency (Figure 4.1) at teleseismic distances (e.g., *Chevrot*, 2002; *Rawlinson and Kennett*, 2004; *VanDecar and Crosson*, 1990). With the aid of zero-phase filtering, which increases seismogram SNR and preserves the relative timing of the peak amplitudes, relative arrival-times can thus be calculated via identification of coherent peaks/troughs of energy for a given earthquake across a network. Relative arrival-time inversions (e.g., *Bastow et al.*, 2008; *Boyce et al.*, 2016; *Frederiksen et al.*, 2013; *Rawlinson et al.*, 2006; *Rondenay et al.*, 2000; *VanDecar et al.*, 1995) are thus our primary source of tomographic images of upper mantle structure beneath many regions of tectonic and geodynamic interest. Despite their high resolution, these models are lacking in one critical respect: they contain no information about the region’s average velocity structure, whether it be fast (e.g., the shields) or slow (e.g., hotspots). Within global tomographic models (e.g., *Li et al.*, 2008a), these small, temporary networks provide a hitherto untapped resource with which to fill gaps in spatial coverage, with important implications for the resolution. The challenge is thus to determine absolute phase arrival-times.

Here, we develop an approach for determining absolute arrival-times (the Absolute Arrival-time Recovery Method - AARM) that exploits the efforts of existing relative arrival-time studies. The first-arrival can be picked from a phaseweighted stack of previously aligned yet unfiltered waveforms allowing absolute arrival-times to be recovered. This is combined with a quantitative pick quality estimate and is rigorously tested on data containing increasing levels of teleseismic background noise. Temporary network deployments can therefore be used to report accurate absolute arrival-time measurements to global pick databases such as the International Seismological Centre (ISC) (*Di Giacomo et al.*, 2014; *International Seismological Centre*, 2016) and be used in global absolute wavespeed tomographic in-

versions.

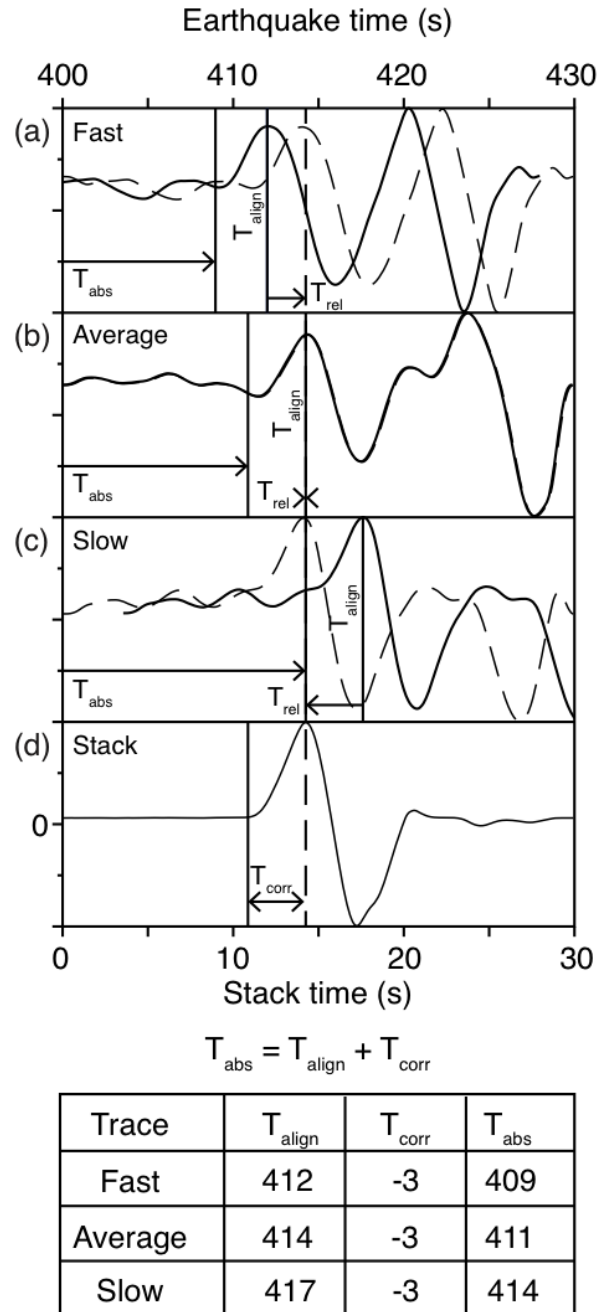


Figure 4.1: **Figure to show the difference between relative (T_{rel}) and absolute (T_{abs}) arrival-times.** Waveform (a) is fast (2 s) relative to the network average (b) and thus arrives first. The slow waveform (c) arrives later (3 s). Using the first major peak, the waveforms are shifted into alignment (dashed waveforms) with the average trace. This point represents the relative arrival-time alignment point T_{align} . The first arrival occurs before this (3 s - indicated by the stacked trace - d) so the correction T_{corr} can be made to find T_{abs} indicated by the vertical lines.

4.2.1 Relative versus absolute arrival-times

Relative arrival-times, $\mathbf{T}_{\text{rel}}$, align the first coherent peak or trough (i.e. maximum or minimum) across a network ($\mathbf{T}_{\text{align}}$ Figure 4.1) by cross correlation (e.g., *VanDecar and Crosson, 1990*), stacking (e.g., *Rawlinson and Kennett, 2004*) or simulated annealing (e.g., *Chevrot, 2002*). The relative arrival-times for each station, $\mathbf{T}_{\text{rel}}$, are small time-shifts that align the coherent phase of each waveform around an arbitrary point in time, whose distribution is forced to be zero. The necessity for waveform coherency during calculation of $\mathbf{T}_{\text{rel}}$, limits the aperture of a station network to approximately $< 1500 \text{ km}$. Relative arrival-time residuals $\mathbf{RES}_{\text{rel}}$ (Equation 4.1) are calculated by comparison to predicted arrivals, $\mathbf{T}_{\text{exp}}$, based for instance on travel-time tables (such as iasp91 and ak135: *Kennett and Engdahl, 1991*; *Kennett et al., 1995*),

$$\mathbf{RES}_{\text{rel}} = \mathbf{T}_{\text{rel}} - (\mathbf{T}_{\text{exp}} - \bar{\mathbf{T}}_{\text{exp}}), \quad (4.1)$$

where $\bar{\mathbf{T}}_{\text{exp}}$ is the mean expected arrival-time for an earthquake across the network. During relative arrival-time analysis, for each earthquake, rays are assumed to follow a similar path until they diverge beneath the regional network.

Resultantly, the contribution from the background mean velocity structure, $\bar{\mathbf{T}}_{\text{exp}}$, is lost because arrival-time variations are assumed to result from local wavespeed structure beneath the network (e.g., *Bastow, 2012*). However, this background may vary substantially, particularly between cratonic (e.g., *Boyce et al., 2016*, see Chapter 3) and active (e.g., *Bastow et al., 2008*) regions. Thus relative arrival-time datasets cannot be combined easily or used to directly recover absolute velocities.

The absolute arrival-time (or onset time), $\mathbf{T}_{\text{abs}}$, marks the start of incoming energy on a recorded waveform (obscured by noise in Figure 4.1a,b,c). Absolute arrival-time residuals $\mathbf{RES}_{\text{abs}}$ are calculated through direct comparison with the arrival-time that is expected for some reference Earth model, $\mathbf{T}_{\text{exp}}$, for a given source-station pair:

$$\mathbf{RES}_{\text{abs}} = \mathbf{T}_{\text{abs}} - \mathbf{T}_{\text{exp}}. \quad (4.2)$$

High signal-to-noise (SNR) waveforms, an accurate earthquake catalogue and often, manual picking by a skilled analyst are required, so absolute arrival-times are mostly limited to sparse permanent

observatory sites.

4.2.2 Conventional, stand-alone phase picking methods

Early work in all of seismology was dominated by manual picking. Increasingly large datasets, particularly in controlled source experiments, motivated the development of automatic picking routines for static correction calculations (e.g., *Cox*, 1999). Cross correlations (e.g., *Hileman et al.*, 1968; *Taner et al.*, 1974), increase in signal energy (*Coppens*, 1985), fractal-based search (e.g., *Boschetti et al.*, 1996) and neural networks (e.g., *Dai and MacBeth*, 1995) have all been used with varying success. In modern controlled-source seismology applications, these datasets are high frequency with known wavelets and source times, thus typical accuracies are on the order of 10 ms (e.g., *Cox*, 1999). However, many of these algorithms require training or some manual intervention.

Teleseismic arrival-time determination has a unique set of challenges. Variably emergent waveforms, diverse noise sources (even for one earthquake recorded across a network), lower frequency datasets and unknown source times and wavelets contribute to a hugely underdetermined problem. *Allen* (1982), *Baer and Kradolfer* (1987) and *Earle and Shearer* (1994) developed the short and long term average ratio (STA/LTA) method to reliably pick phases trace-by-trace. *Amaru et al.* (2008) implemented this regime to pick 85,000 absolute arrival-times for temporary deployments in Europe using filtered data. The automated picking algorithm developed by *Aldersons* (2004) has also been applied to datasets in the Alps (*Di Stefano et al.*, 2006; *Diehl et al.*, 2009). However, these methods can lead to large errors in high noise environments (typical of temporary regional seismic networks) and the algorithms must be carefully trained on a selection of representative reference traces.

During relative arrival-time analysis, when peaks/troughs of a coherent phase across a network are aligned, an opportunity arises to pick the first arrival from a resulting stack (e.g., *Chevrot*, 2002; *Lou et al.*, 2013; *Pavlis and Vernon*, 2010; *Rawlinson and Kennett*, 2004). However, as far as we have been able to determine, no study to date explicitly lays out a theoretical basis for picking $\mathbf{T}_{\text{abs}}$ on unfiltered data; nor do they include an associated quantitative estimate of relative uncertainty (for weighting during subsequent tomographic inversion). It is here that we seek improvement.

4.3 Absolute arrival-time determination

With the aid of five example seismograms recorded at stations in eastern North America (Figure 4.2a) we document our absolute arrival-time recovery method (AARM) below.

4.3.1 Data pre-processing

For a temporary network of stations, relative arrival-times are routinely calculated (Figures 4.1, 4.2b) (e.g., *Boyce et al.*, 2016; *Frederiksen et al.*, 2013), commonly using the methods of *VanDecar and Crosson* (1990) or *Rawlinson and Kennett* (2004). Datasets of filtered waveforms are subject to quality control (QC) measures that remove cycle skips, timing errors, large outlying residuals and waveforms of low SNR.

In order to obtain absolute arrival-times from these datasets, the alignment times ($\mathbf{T}_{\text{align}}$ in Figure 4.1) should optimally be transferred back into the unfiltered data (Figure 4.2c) with a standardised instrument response (here, the broadband velocity response of a standard Blacknest type seismometer available in Seismic Analysis Code, *Goldstein and Snoke*, 2005; *Helffrich et al.*, 2013). Although minimum phase filters will not lead to a shift in the first arrival, the signal peaks or troughs may be distorted (e.g., *Di Stefano et al.*, 2006; *Küperkoch et al.*, 2010; *Leonard*, 2000). Because we rely on alignment of peaks and troughs in the relative arrival-time step and our stacking procedure (described in the next section), best practice for AARM is to avoid filtering during formation of the stack. Indeed *Amaru et al.* (2008) experience a DC shift (~ 0.1 s) towards faster picks relative to a standard database due to filtering or varying instrument responses. Figure 4.3 shows that aligned, filtered velocity seismograms map consistently into the unfiltered velocity and displacement seismograms, despite not necessarily corresponding to a particular peak or trough.

The unfiltered, aligned data are equalised in sample rate, cut to a predetermined length (60 s for P-waves, 120 s for S-waves), normalised and are optionally integrated to displacement. We note that whilst velocity seismograms are typically more impulsive than displacement seismograms, they often contain more high-frequency noise but either (velocity or displacement) should be suitable for obtaining absolute arrival-times. Because the SNR of unfiltered seismograms can contrast greatly from their filtered counterparts, a low-threshold SNR (< 1) is used to remove particularly noisy traces (e.g., *Pavlis and Vernon*, 2010). In the following sections we refer to the signal-to-noise ratio frequently,

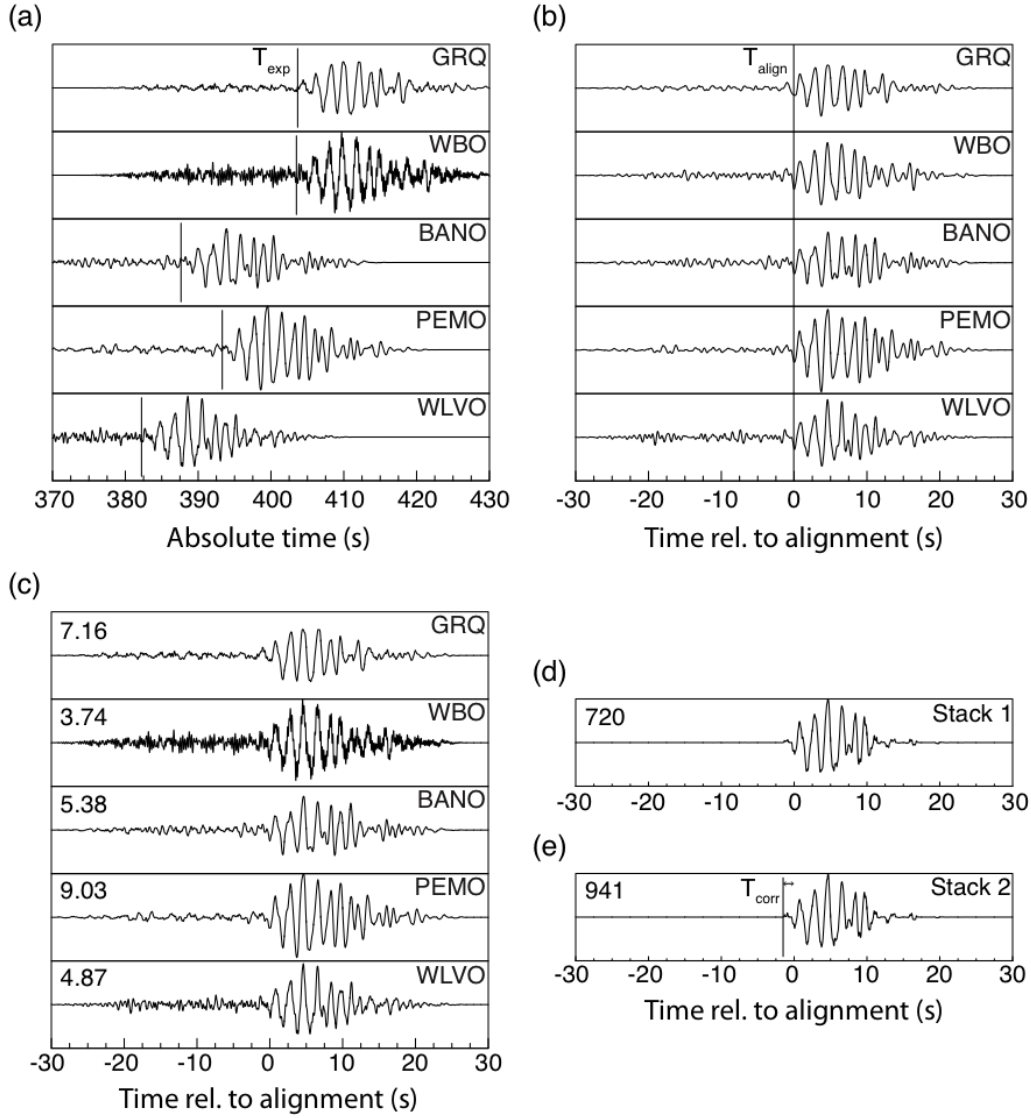


Figure 4.2: **Figure to demonstrate the AARM workflow** using an example earthquake recorded by a selection of 5 broadband instruments in southeast Canada. a) Unfiltered raw traces with the predicted arrival-time, T_{exp} , for the P-wave arrival, shown by the vertical lines. Station names are also shown. b) Traces are filtered (Butterworth two pole between 0.4–2.0 Hz) and aligned on a coherent peak or trough, T_{align} using a relative arrival-time technique (e.g., *VanDecar and Crosson, 1990*). c) Alignments are transferred back to unfiltered velocity seismograms. Numbers within each sub-panel refer to the SNR of each trace. d) An initial phaseweight stack is calculated, this has an SNR of 720. Weightings for a second stack are calculated using SNR or cross-correlation with the initial stack. This results in a higher SNR second stack e). Minor trace adjustments can also be made when using the cross-correlation scheme (T_{adj} in Equation 4.7). A manual first arrival picked by the user (vertical line) is then used to calculate the absolute correction, T_{corr} , to apply to each trace across the network. The absolute arrival-times (T_{abs}) are compared to the predicted arrivals (T_{exp}) from ak135 to give absolute arrival-time residuals, RES_{abs} (See Equation 4.7).

herein defined as:

$$SNR = \frac{A_{signal}}{A_{noise}}, \quad (4.3)$$

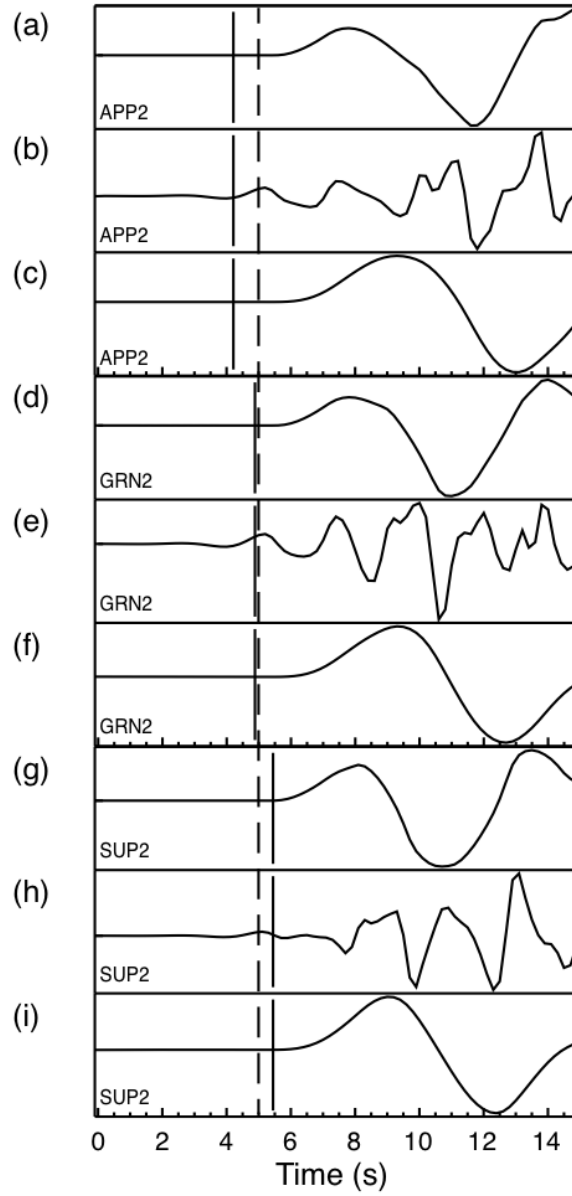


Figure 4.3: **Aligned, unfiltered, synthetic velocity seismograms** (a, d, g), from three stations (APP2, GRN2, SUP2). Filtered velocity seismograms (b, e, h) and unfiltered displacement seismograms (c, f, i) are also shown. The dashed vertical line refers to the initial trace-alignment point (T_{align}) derived from the MCCC code of *VanDecar and Crosson* (1990). The solid vertical lines are the predicted absolute phase arrival-times (T_{exp}) from the 1D velocity model ak135.

where A is the root-mean squared amplitude defined over a pre-set window of 25 s (P-waves) or 55 s (S-waves). The ‘noise’ and ‘noise-plus-signal’ windows are separated by a two second safety gap akin to *Di Stefano et al.* (2006); *Diehl et al.* (2009) around the alignment point ($\mathbf{T}_{align}$). We use P-wave data to describe AARM but note that our testing shows AARM can be reasonably applied to S-waves data whilst accounting for their generally longer periods.

4.3.2 Calculation of arrival-times

Preliminary stacking

The first step in the methodology is to form an initial stack (Figure 4.2d) using the unfiltered aligned waveforms. There are three common stacking methods (e.g., *Rost and Thomas*, 2002; *Schimmel and Paulssen*, 1997). A linear stack is defined as

$$\mathbf{f}(\mathbf{t})_{\text{lin}} = \frac{1}{N} \sum_{j=1}^N \mathbf{s}_j(\mathbf{t}), \quad (4.4)$$

where $f(t)$ is the stacked trace, N is the number of traces and $s_j(t)$ is the waveform. A n^{th} root stack is given by

$$\mathbf{f}(\mathbf{t})_{\text{nth}} = \left(\frac{1}{N} \sum_{j=1}^N \sqrt[m]{s_j(t)} \right)^{\mathbf{m}}, \quad (4.5)$$

where m is the order of the root. Finally, phaseweight stacking is defined as

$$\mathbf{f}(\mathbf{t})_{\text{phase}} = \frac{1}{N} \sum_{j=1}^N \mathbf{s}_j(\mathbf{t}) \times \left(\frac{1}{N} \sum_{j=1}^N \exp(\mathbf{i}\Phi_{\mathbf{k}}(\mathbf{t})) \right)^{\mathbf{v}}, \quad (4.6)$$

where $\Phi_{\mathbf{k}}(\mathbf{t})$ is the instantaneous phase of the waveform. $\mathbf{v}$ defines how readily coherent and incoherent phases are separated i.e. the severity of the phaseweighting. A linear stack is retrieved with $\mathbf{v} = \mathbf{0}$. The recent dual bootstrap resampling stacking method of *Korenaga* (2013) is not considered in this study as we do not require signal recovery of very low SNR (< 1) waveforms.

The three stacking schemes (linear, n^{th} root and phaseweight) were tested systematically on observed data, one example of which is shown in Figure 4.4. The main focus is strong noise suppression without loss of signal in the final stack. Based on this criterion linear stacking is not considered further due to presence of high frequency pre-arrival noise masking the true onset time in the stack (Figure 4.4a). The remaining stack types n^{th} root and phaseweight (Figure 4.4b,c) both achieve a high SNR stacked trace. *Schimmel and Paulssen* (1997) and *Rost and Thomas* (2002) show that n^{th} root stacking tends to produce more impulsive peaks that may lead to lower pick errors (*Douglas et al.*, 1997). This is consistent with our test dataset. However during testing, phaseweight stacking was marginally more

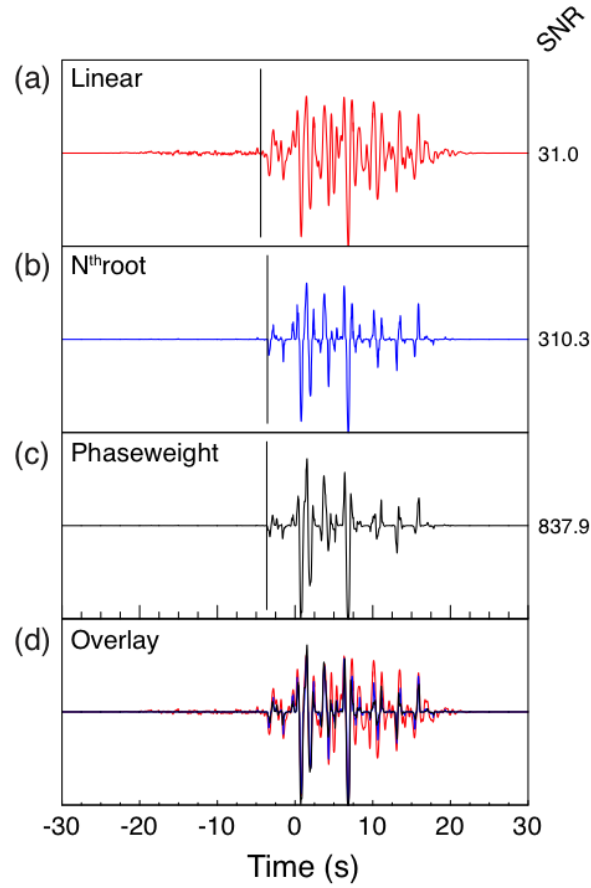


Figure 4.4: **Figure to compare the pre-arrival noise suppression** of linear (a), nthroot (b) and phaseweight (c) stacking for an example earthquake. (d) shows an overlay of the three stack types.

effective in suppressing pre-arrival noise (Figure 4.4c) leading to higher SNR stacks. Therefore, aligned traces are stacked (Figure 4.2d) using phaseweighting ($\mathbf{v} = 4$ in Equation 4.6) across the entire window (60 s).

Weighting and adjustment of traces for final stacking

A signal-to-noise (SNR) or cross-correlation (XC) derived weighting scheme (e.g., *Paulis and Vernon, 2010*) can be used to form a higher SNR second stack (Figure 4.2e) on which the first arrival is picked by the analyst. For each trace, a SNR approximation (Equation 4.3) is made (Figure 4.2c). For each earthquake, these are normalised between 0 and 1 to give a set of SNR derived weights for the second stack. Alternatively, cross correlation of each normalised trace with the preliminary stack can also yield a set of ‘XC’ derived weights for the second stack, again normalised between 0 and 1. This method also allows small adjustments to the initial alignment points if the cross-correlation function maximum (between the trace and the preliminary stack) is offset from zero. In theory this should

result in lower picking uncertainty.

Where traces are very similar across the network, the cross-correlation technique will tend to assign relatively high weights to all traces (Figure 4.5c). However, the SNR weighting scheme may produce a broader spread of weightings and thus the stack will be more robust (Figure 4.6c). In a relatively high-noise environment, the stack may be dominated by one quiet trace when using the SNR weighting but the weights derived from cross-correlation may be more uniformly distributed. In practice, the choice of SNR or XC derived weightings varies from one application to the next. We did not find a strong pattern of improved stacks when using the XC regime on emergent events and the SNR regime on impulsive events (e.g., *Pavlis and Vernon, 2010*).

Our testing showed that linearly distributing cross-correlation or SNR weightings between 0 and 1 is adequate for our purpose. We note that weighting functions have been used to improve signal stacking in some applications (e.g., *Restivo and Helffrich, 1999*) but testing of a variety of weighting functions on both observed and synthetic datasets did not show any significant improvement in absolute arrival-time picks. For the majority of earthquakes, where the number of high SNR waveforms is > 10 , the second stack may appear to offer little improvement in pick accuracy (e.g., Figure 4.2d,e). However, when only a limited number (< 10) of variably noisy waveforms are available, the arrival-time on the initial stack can vary by > 0.3 s when compared to the final stack (Figure 4.7a,b). We therefore choose to compute the second stack in all cases at a minor computational cost.

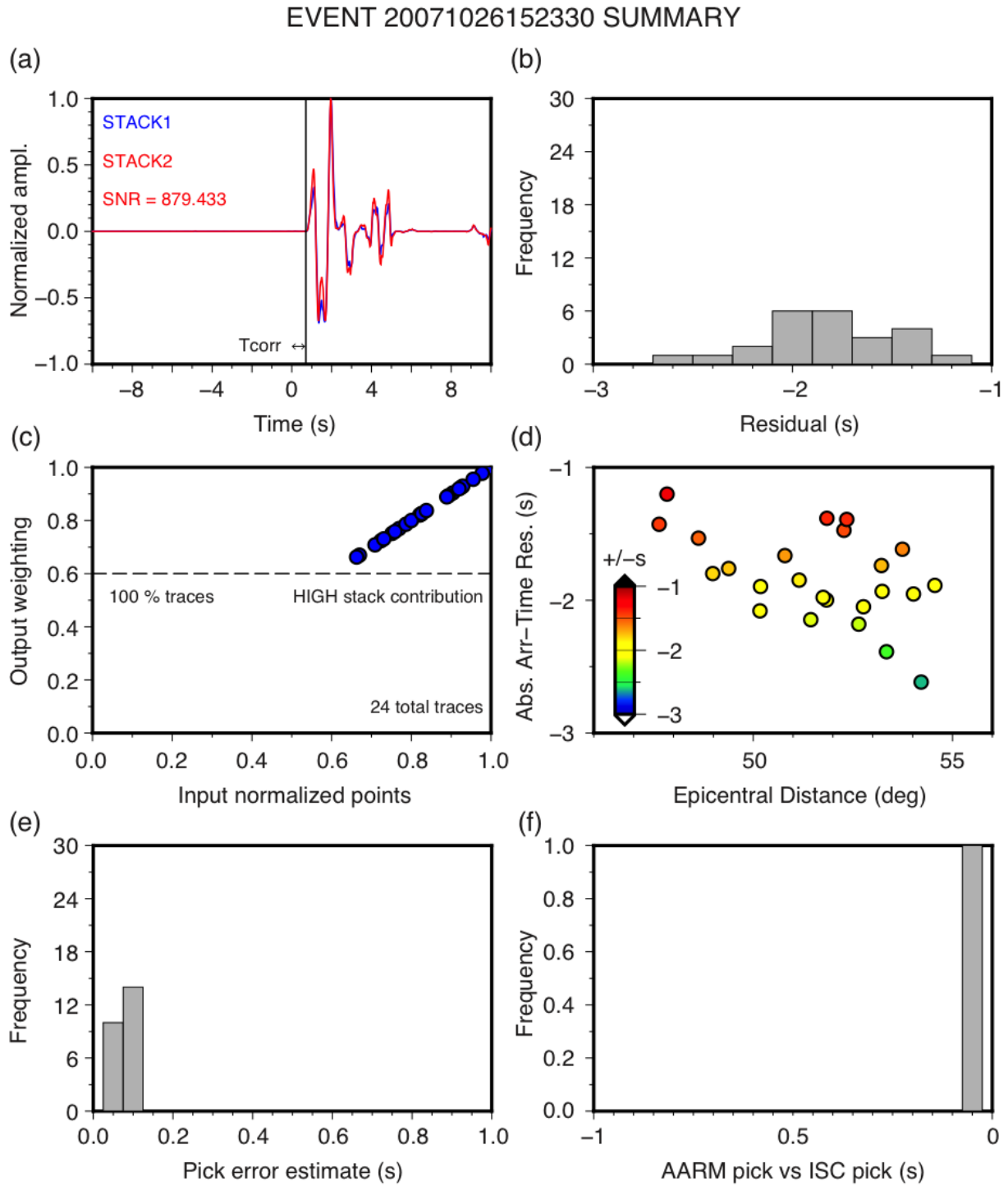


Figure 4.5: **AARM output (using cross-correlation scheme) for an impulsive earthquake** recorded by stations in southeast Canada (*Boyce et al., 2016, Chapter 3*). (a): Output stacks (blue primary and red final) from alignment of traces. The correction time T_{corr} is also shown. (b): Absolute arrival-time residual histogram. (c): Distribution of weights from cross-correlation of individual traces with the primary stack on the linear weighting scheme. (d): Absolute arrival-time residual distribution against epicentral distance. (e): The autocorrelation-derived estimate for picking error. (f): The difference between calculated absolute arrival-times from AARM and ISC picks where available (only one equivalent source-station pair exists for this event).

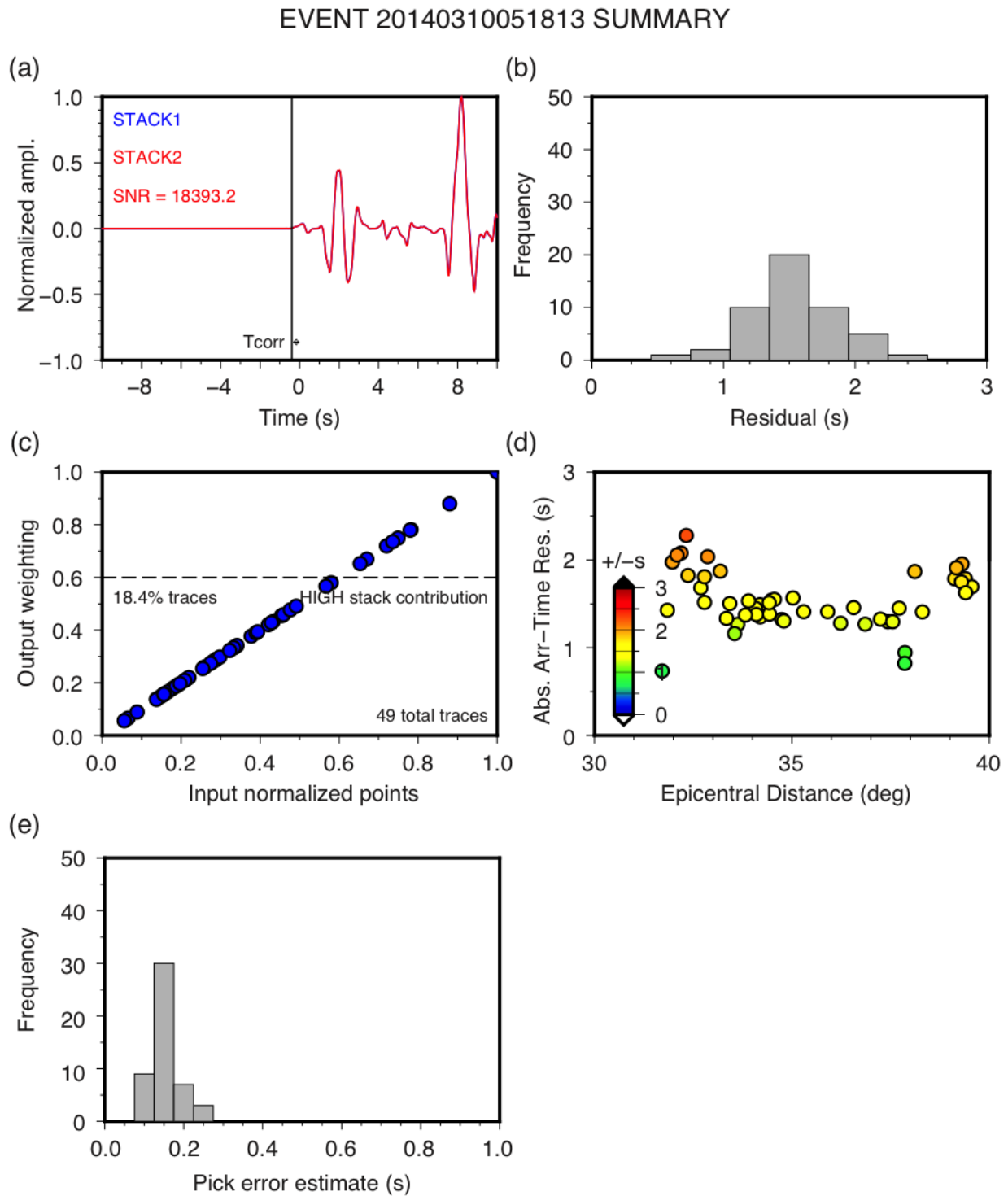


Figure 4.6: **AARM output (using SNR scheme) for an emergent earthquake** recorded by stations in southeast Canada (*Boyce et al.*, 2016, Chapter 3), subplots are the same as Figure 4.5. Note that no corresponding ISC picks were available for this earthquake.

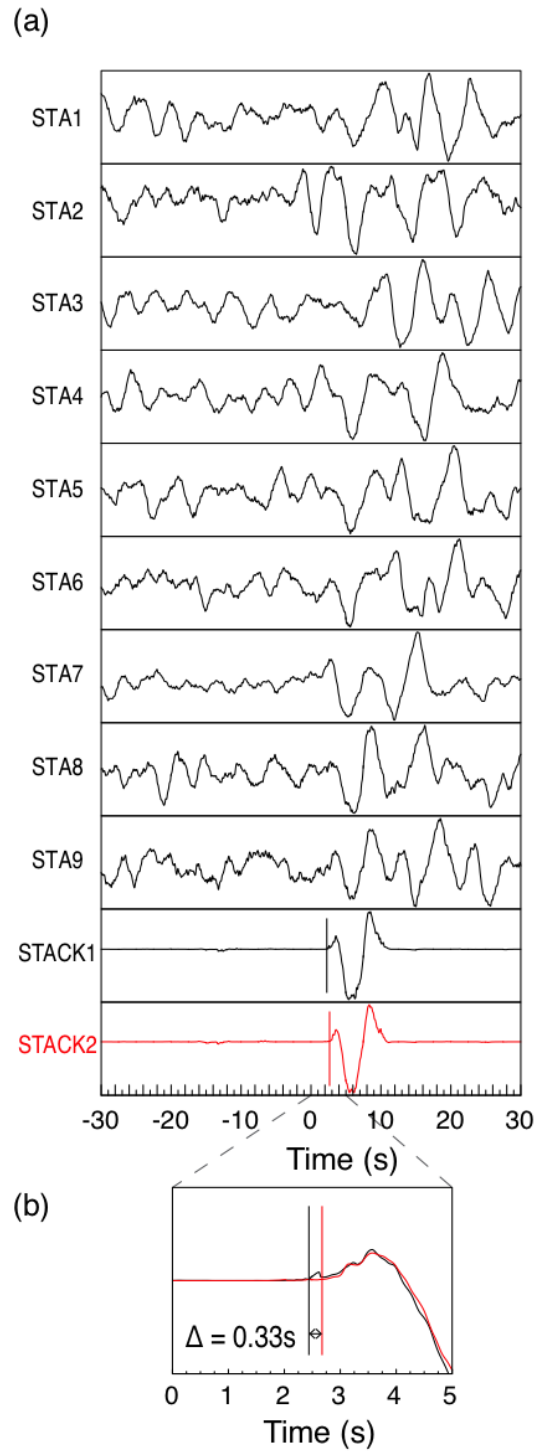


Figure 4.7: **Figure to show nine variably noisy waveforms (STA 1 – 9) that contribute to the initial (stack1 - black) and final stack (stack2 - red) - (a).** Vertical lines show manually picked arrival-times. An expanded view of the stacks (b) shows the difference between the picks (Δ).

First arrival picking

The linear weighting process leads to an improved second phaseweighted stack (Equation 4.6), on which the first arrival or onset time is picked manually by an analyst (Figure 4.2e). We use a consistent window size and axes scaling for the entire dataset whilst picking the first arrival (e.g., *Diehl et al.*, 2009; *Douglas et al.*, 1997). Manual picking does introduce some human bias into the data (e.g., *Leonard*, 2000) but is more randomly distributed than automatic picking methods (e.g., *Aldersons*, 2004; *Amaru et al.*, 2008). Furthermore, the variably emergent nature of unfiltered, stacked signals, necessary training of algorithms (for pick accuracy) and generally low numbers of earthquakes within a relative arrival-time dataset (often ~ 200) dictate that automatic picking methods would provide no logical benefit here.

For each earthquake, enough stations across the network must record at high enough SNR to form a high SNR stack (Figure 4.2e) where random pre-arrival noise is sufficiently suppressed. This is required to accurately pick the stack manually.

Computation of absolute arrival-times and residuals

Following manual picking of the final stack (Figure 4.2e), similarly to Equation 4.2, absolute arrival-time residuals ($RES_{abs}[i]$) can be calculated with the following expression:

$$\mathbf{RES}_{abs}[\mathbf{i}] = \left(\mathbf{T}_{align}[\mathbf{i}] + \mathbf{T}_{adj}[\mathbf{i}] + \mathbf{T}_{corr} \right) - \mathbf{T}_{exp}[\mathbf{i}]. \quad (4.7)$$

The difference between the final alignment point in the stack and the user picked onset time (negative when the onset occurs before the final alignment point, see Figure 4.1) gives the time correction ($\mathbf{T}_{corr}$) to be applied to each trace. This value is consistent across all traces for a teleseismic earthquake where the waveform does not change shape over a network (an assumption of prior relative arrival-time analysis). The absolute arrival-times $\mathbf{T}_{abs}$ (i.e. the term in parentheses in Equation 4.7) are easily calculated by adding the correction to each of the alignment times ($\mathbf{T}_{align}[\mathbf{i}]$). This may also be adjusted using the cross-correlation correction ($\mathbf{T}_{adj}[\mathbf{i}]$). These times are then compared with a predicted time for each ray ($\mathbf{T}_{exp}[\mathbf{i}]$), ak135 in this case, producing absolute arrival-time residuals, $\mathbf{RES}_{abs}[\mathbf{i}]$. Upon processing of each event, visual inspection of residual distribution enables fast

identification of outliers. We also use our pick quality assessment parameters discussed below to remove further picking inconsistencies.

4.3.3 Methodological assessment of pick quality

A visual first-order indication of picking error is given by the broadness of the final stacked trace (*Rawlinson and Kennett, 2004*) controlled by the impulsivity of the unknown source wavelet. Unlike controlled source experiments where the wavelet can be tuned for the purpose, earthquakes exhibit variability of signal onsets, from impulsive to emergent (Figures 4.5, 4.6). In extreme cases less impulsive first-arrivals from long rise-time earthquakes can give onset timing errors of up to 0.5 s (*Douglas et al., 1997*).

Chevrot (2002) proposed a quantitative method for evaluating this relative picking error. First, the autocorrelation of the final stack is calculated. Then each trace is individually cross-correlated with the final stack to give a cross-correlation function. As the autocorrelation of the stack decreases from 1 from its centre (in a normalised case) it will pass through the absolute maximum value observed on the cross correlation function of the stack and the individual trace. This point will be offset from zero in time and thus gives an approximate measure of relative picking error. We set an upper bound on this autocorrelation derived relative error estimate of 0.25 s above which we exclude the picks from the output absolute arrival-times. This ensures that the individual relative pick uncertainty is lower than the absolute errors of global pick databases.

The cross-correlation of the primary stack with each trace can be used to adjust the alignment of traces within the final stack if the maximum ($\mathbf{T}_{\text{adj}}[\mathbf{i}]$) is offset from zero (e.g., *Amaru et al., 2008; Pavlis and Vernon, 2010*). This indicates that the trace is better aligned with the stack by a small time shift. When using the cross-correlation derived weighting scheme it is possible to correct for this, therefore individual traces are better aligned for the second stack, giving more accurate arrival-time picks. However, in some cases after the second stack, the value of the maximum offset can be non-zero (more often using the SNR weighting scheme which does not make this correction). Because our stacked traces are well aligned initially by the relative arrival-time analysis, we find this measure to be on the order of the sample interval for >95 % of traces in our test datasets. Thus we use this as a tool to remove poorly aligned traces from our absolute arrival-time dataset (e.g., *Pavlis and Vernon, 2010*) rather than to directly assess pick quality.

As described above, the cross-correlation function maximum between the primary stack and each trace can be used to weight a second stack. The distribution of these weights is a useful indicator of sources of error although is not a direct measure in the time domain. For earthquakes where a high number of traces have weights > 0.6 the stack will be a reliable approximation to the ideal trace for the network (e.g., Figures 4.5, 4.6) or array beam. Conversely, in cases where $< 10\%$ of traces have weights > 0.6 the majority of traces are downweighted heavily in the final phaseweight stack. The final stack is not necessarily a reliable estimate for the array beam and thus the first arrival on the stack may not be well constrained. This does provide a strategy for weighting of individual earthquakes within a subsequent tomographic inversion however. We note that traces with low weightings in the second stack do not contribute significantly but arrival-time picks can still be produced or removed as necessary.

4.4 Testing

4.4.1 Testing of observed and synthetic datasets

We test AARM using both synthetic and observed datasets. All synthetic and observed input data are normalised to the broadband velocity response of a standard Blacknest type seismometer (available in Seismic Analysis Code, *Goldstein and Snoke*, 2005; *Helffrich et al.*, 2013). Deconvolution with the individual seismometer response directly to displacement is ill-advised due to the imposed filtering and therefore possible distortion of the onset time (e.g., *Di Stefano et al.*, 2006; *Küperkoch et al.*, 2010; *Leonard*, 2000). We have also tested the picking regime on both displacement and velocity seismograms. We find that high-frequency noise suppression during stacking is strong enough to use velocity seismograms to pick absolute arrival-times for our test datasets. Despite the lack of direct absolute error estimates derived from manual picking, we nevertheless explore quantitatively, how well absolute arrival-times are recovered by AARM.

Absolute arrival-time recovery for observed datasets

AARM is first tested on a published observed relative arrival-time dataset (Figure 4.8) of $\sim 10,000$ picks from southeast Canada (*Boyce et al.*, 2016, see Chapter 3) and compared to the absolute arrival-time database of the International Seismological Centre (ISC) (*Di Giacomo et al.*, 2014; *International Seismological Centre*, 2016). Due to strict quality control measures imposed (e.g. $\text{SNR} < 1$ and

relative pick accuracy < 0.25 s), we recover 9053 ($> 90\%$) absolute arrival-time measurements (Figure 4.8a) for subsequent use in a global tomographic inversion (such as *Li et al.*, 2008a). Almost all (99.7%) residuals are distributed between ± 3 s with a mean of -0.44 s. This is expected for a region of generally elevated wavespeed relative to the global average (e.g., *Li et al.*, 2008a).

The International Seismological Centre (ISC) (*Di Giacomo et al.*, 2014; *International Seismological Centre*, 2016) collates travel-time data from a number of sources and contributing agencies, where picks can be manual or automatic. However, it does not routinely provide associated pick errors, let alone any method of calculating them. The oldest records available on the ISC (1964 onwards) are thought to be accurate to only ± 2 s (*Kennett and Engdahl*, 1991), whilst present average origin-time uncertainties are ~ 1 s for the ISC catalog (*Kagan*, 2003). Random picking errors are also thought to be high (~ 0.5 s e.g., *Gudmundsson et al.*, 1990). The regional hypocentre determination work of *Husen et al.* (1999) indicates that modern techniques are accurate to ~ 2 km in source depth. This translates to a timing uncertainty of ~ 0.3 s for a typical P-wave arrival, and ~ 0.55 s for S-waves.

Within our observed dataset, 535 equivalent earthquake-station pairs are available on the ISC pick database (Figure 4.8b). AARM recovers 83% of absolute arrival-time picks within ± 0.5 s of the ISC picks (approximately normally distributed). ISC picks are on average slower by 0.05 s in agreement with *Amaru et al.* (2008). The autocorrelation-derived measure of relative error (Figure 4.8c) indicates that picks are accurate to 0.1 s on average with 94.8% of picks ≤ 0.15 s. As expected, as the mean trace SNR increases, mean relative pick error decreases (consistently low above a mean SNR of 2, Figure 4.8d) and the SNR of the stack increases (Figure 4.8e). However, high SNR stacks can result from generally lower SNR data. This is likely due to a small proportion of noisy traces that lower the average SNR but are down-weighted heavily in the final stack. Additionally, higher SNR results in marginally improved cross-correlation coefficients with the final stack (Figure 4.8f).

The comparison to the ISC database and relative assessments of error (that provide a weighting for arrivals during tomographic inversion) show that AARM works reliably for our test dataset. However, using waveforms of high SNR, it is possible to go one step further and compare manual reference picks with the picks derived from AARM (e.g. *Amaru et al.*, 2008; *Di Stefano et al.*, 2006; *Küperkoch et al.*, 2010) to obtain an estimate for absolute error. The temporary network deployments, on which AARM is designed to be used, will not allow for manual reference picking in most cases due to high background noise levels. Instead we are able to directly compare picks produced by AARM and manual picks on noise-free synthetic seismograms described below.

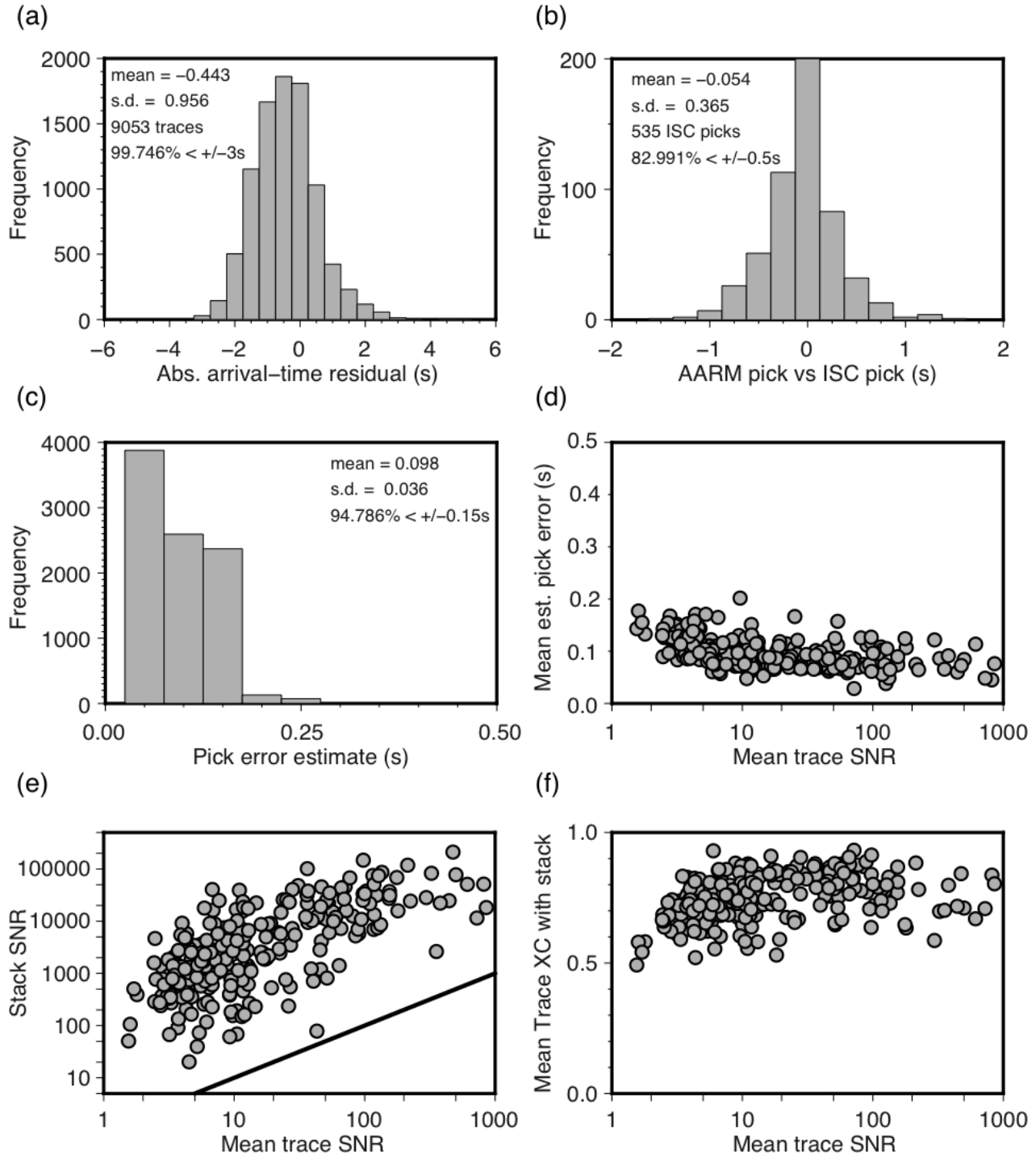


Figure 4.8: **Dataset overview** for the methodology (AARM) applied to the relative arrival-time picks of Boyce et al. (2016, Chapter 3). (a): The distribution of absolute arrival-time residuals for the dataset. (b): The difference between the calculated absolute arrival-times (from AARM) and ISC picks where available. (c): The autocorrelation pick error estimate for the entire dataset. (d): The variation of mean autocorrelation pick errors with event averaged SNR. (e): The variation in SNR of the final stacked trace against event averaged SNR. The black line shows the line $y = x$. (f): Shows the mean cross-correlation coefficient of each trace with the stack as a function of event-averaged SNR.

Absolute arrival-time recovery for synthetic datasets

For our synthetic testing we model a linear network of nine stations evenly spaced between epicentral distances of 3400 km and 4200 km (Figure 4.9) simulating a teleseismic earthquake from the western U.S. recorded in eastern Canada (e.g., *Boyce et al.*, 2016). The velocity model (Figure 4.9) is perturbed systematically by $\pm 2\%$ to replicate a fast wavespeed Superior province (SUP), average velocity Grenville (GRN) and slow wavespeed Appalachians (APP) (see *Boyce et al.*, 2016) through which synthetic P-wave seismograms (dip-slip source-time function: Figures 4.10a, 4.12a) are generated using the Computer Programs for Seismology generalised ray package (*Herrmann*, 2013). The equivalent procedure is followed to generate the S-wave synthetics in Figures 4.11a, 4.13a although a strike-slip source-time function is used to increase testing heterogeneity. These waveforms are initially aligned (Figure 4.14) using the method of *VanDecar and Crosson* (1990) to mirror the approach used for calculation of relative arrival-times for observed datasets (See Chapter 3). We note that the large maximum interstation distance results in seismograms on the limit of necessary waveform coherence (Figures 4.12a, 4.13a) for a relative arrival-time study, thus in practice errors are likely to be lower than those observed for our synthetic datasets. Using the teleseismic noise spectra from the study of *Peterson* (1993), the synthetic seismograms are subject to increasing levels of random noise distributed between the low (NLNM) and high (NHNM) power spectra (coloured curves of Figure 4.15). Example outputs for a synthetic station (SUP2) are given in Figures 4.10a, 4.11a for P- and S-waves respectively. The increasing uncorrelated teleseismic noise spectra (0.0 - 1.0: Figure 4.15) give rise to decreased SNR of each trace (Figures 4.10a, 4.11a).

AARM picks produced for all nine zero-noise synthetic stations are within ± 0.25 s of the manual reference pick (Figure 4.12f). Thus we consider AARM to be able to produce P-wave picks to accuracies of < 0.25 s. This is consistent with the ISC database and other accurate databases worldwide (*Leonard*, 2000) where 90 % of picks made by an experienced analyst are typically accurate to within ± 0.2 s. AARM picks preferentially occur slightly after the manual reference picks (on average less than the sampling interval) in agreement with *Amaru et al.* (2008) and *Küperkoch et al.* (2010).

When compared to the zero-noise case, synthetic seismograms (with average SNR > 8) produce very similar results to the zero-noise case in terms of absolute arrival-time residuals (Figure 4.12c,d), estimated errors (Figure 4.12b,e) and the manual pick time on the stack (Figure 4.10b). Figure 4.10a shows that between a trace SNR of 8 and 2, the resulting stack undergoes mild distortion of the

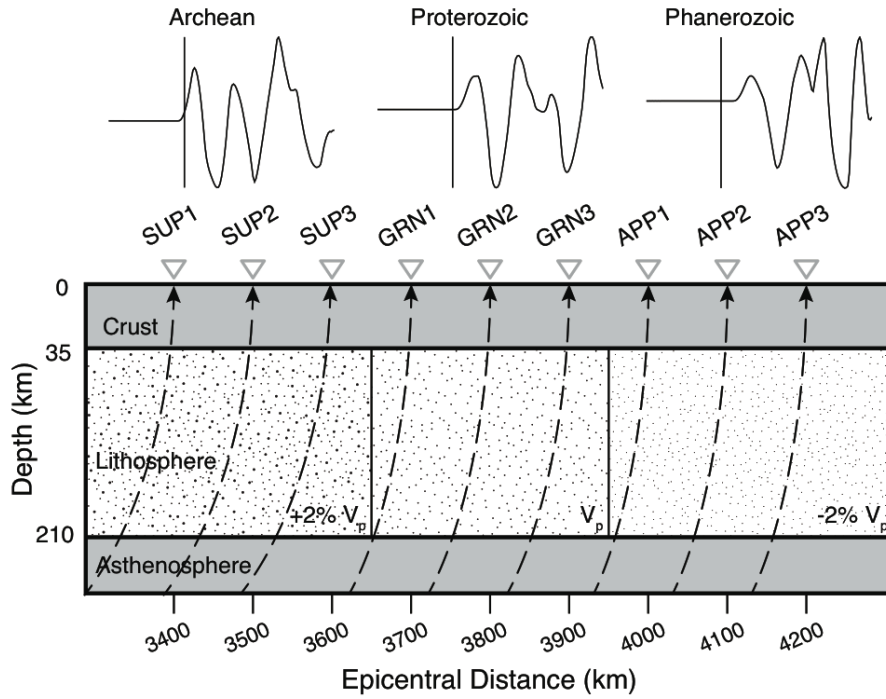


Figure 4.9: **Schematic showing perturbed 2D ak135 velocity model** ($+2\%$, 0% , $-2\% dV_p$), between 35 and 210 km depth, used to produce symmetrically distributed synthetic waveforms, modelling early, average and late arrivals (example delayed waveforms are shown above) at increasing epicentral distance (3400–4200 km). This corresponds to the Archean Superior Province (SUP - dense fill), Proterozoic Grenville Province (GRN - average fill) and Phanerozoic Appalachians (APP - light fill) in southeastern Canada, akin to the study of *Boyce et al.* (2016, Chapter 3). Ray tracing is used to produce Green’s Functions that are convolved with a parabolic source-time function representative of a 90° dip-slip earthquake placed at 210 km depth (after *Herrmann*, 2013). Output seismograms are convolved with a standardised broadband velocity response of a standard Blacknest type seismometer (available in Seismic Analysis Code, *Goldstein and Snoke*, 2005; *Helffrich et al.*, 2013).

first arriving peak (Figure 4.10b) and thus manual pick time errors result. Below this average trace SNR level, the stack completely breaks down, containing significant high frequency noise and does not show any resemblance to the zero-noise stack (Figure 4.10b). However, the previously described quality control steps ensure that such low SNR stacks, with little representation of the average trace for the network, rarely occur.

For the S-wave synthetic dataset the change in source-time function does not affect the accuracy of the picks. The stack breaks down at a lower noise level (Figure 4.11b) because the noise spectrum has greater amplitude at dominant S-wave periods ($\sim 5 - 25$ s) when compared to P-wave data ($\sim 0.5 - 2.5$ s), so this is expected (see Figure 4.15). The absolute errors, compared to individual manual picks, are in agreement to within ± 0.5 s (Figure 4.13f), again a slight increase over the P-wave tests.

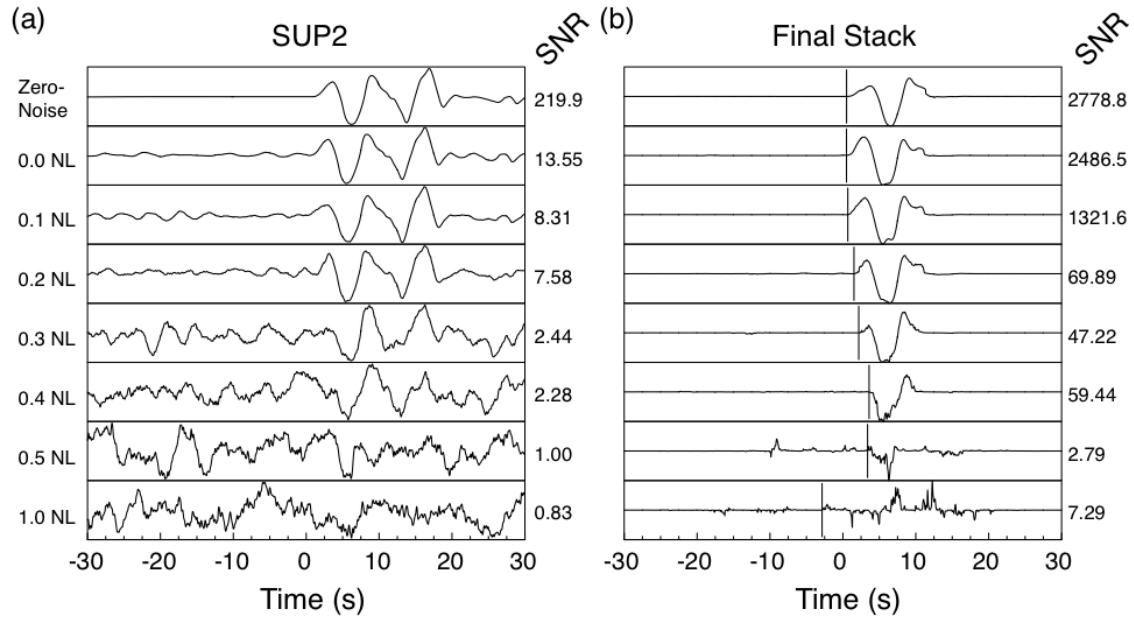


Figure 4.10: **Plot to show how increasing levels of teleseismic background noise** (*Peterson, 1993*) affects the input P-wave velocity synthetic seismograms (a) and the output stacked traces (b). (a): Increasing levels of teleseismic noise are added to the synthetic velocity seismogram SUP2 resulting in decreased SNR as shown. (b): The increasing levels of noise on the input traces produce velocity seismogram stacks of decreasing SNR and thus inaccurate picks for the onset time. Vertical lines are the manual picks used for absolute arrival-time calculation in AARM.

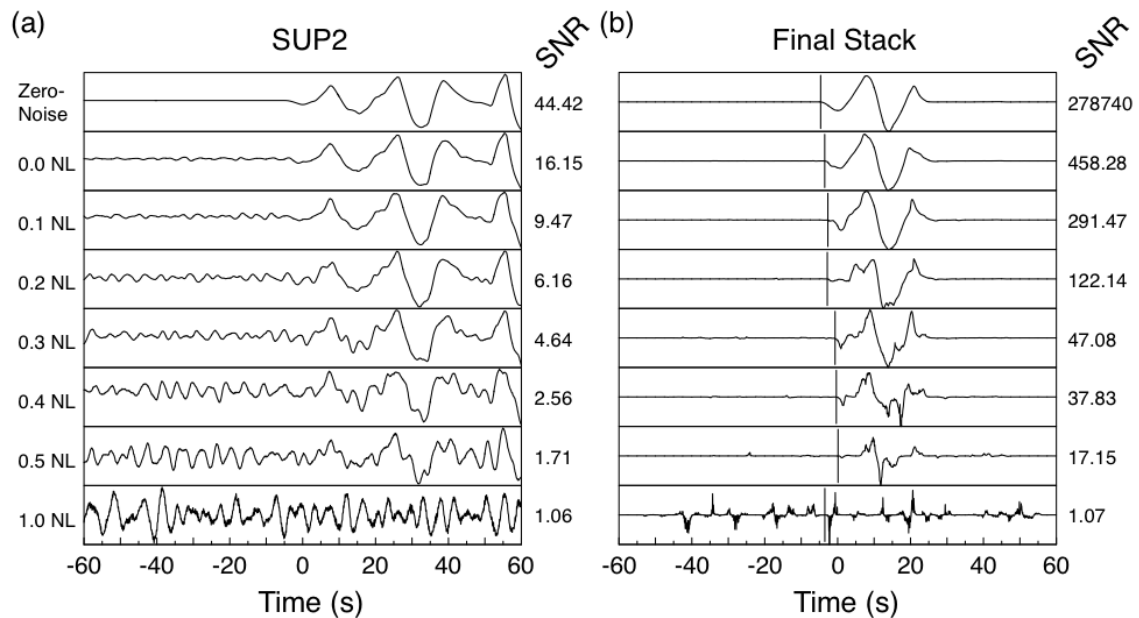
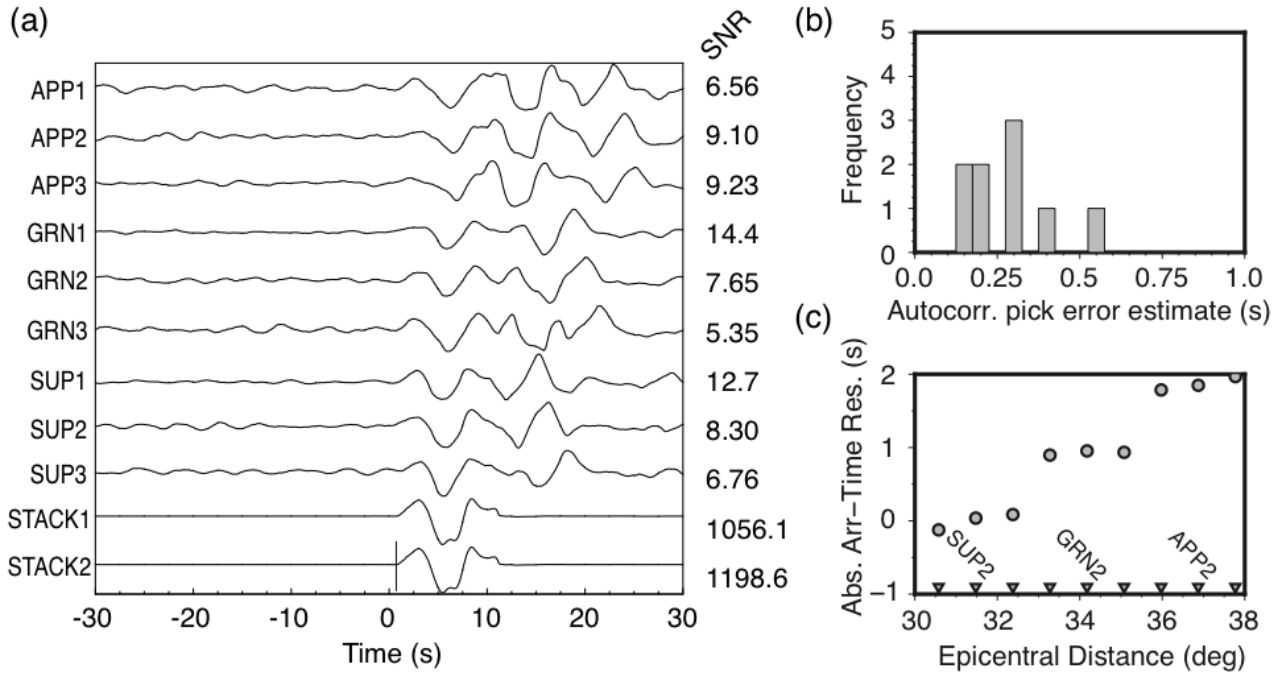
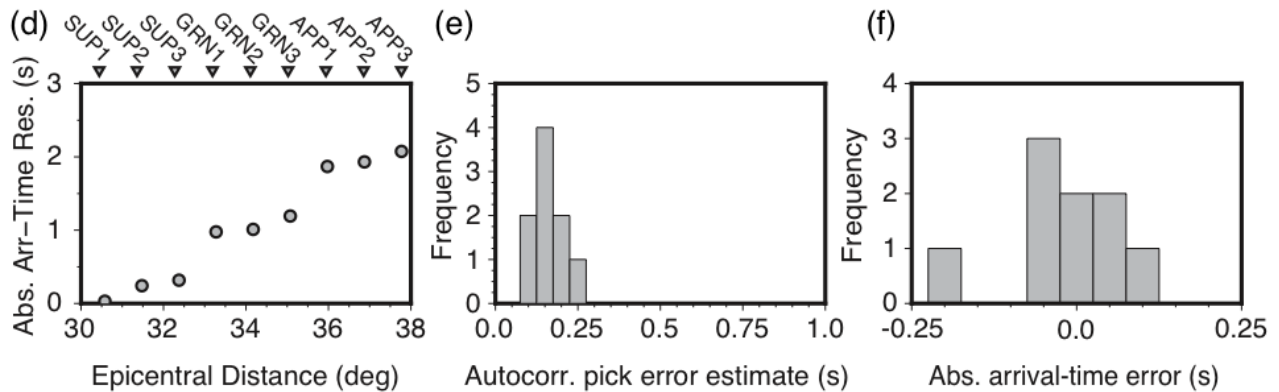


Figure 4.11: **Plot to show how increasing levels of teleseismic background noise** (*Peterson, 1993*) affect individual traces at station SUP2 (a) and the final stack (b) for the S-wave synthetics (labels are the same as Figure 4.10 P-wave data).



Synthetic seismograms at 0.1 noise-level of Peterson (1993).



Idealised zero-noise synthetic data.

Figure 4.12: **Figure to show results from alignment of noisy synthetic traces compared to key results from zero-noise case.** (a): Nine displacement synthetic seismograms convolved with the 0.1 level power spectra of teleseismic noise (*Peterson, 1993*). The approximate SNR is shown beside each trace. Traces are aligned and stacked using phaseweighting (STACK1), then restacked using cross-correlation (used here) or SNR derived weights to form the higher SNR final stack (STACK2). The manually picked onset or first arrival is shown as a vertical line (a). (b): The autocorrelation pick error estimate for the 0.1 noise-level synthetic traces in (a). (c): Distribution of absolute arrival-time residuals for the 0.1 noise-level synthetic traces in (a). (d): Plot showing absolute arrival-time residuals for the idealised zero-noise case for comparison with 0.1 noise-level in (a), (b) and (c). (e): Autocorrelation derived pick error estimates for the zero-noise case. (f): The zero-noise case allows calculated absolute arrival-time residuals (from AARM) to be compared with manually picked arrivals. The plot shows a histogram of difference between two measurements where negative values indicate the AARM pick to be after the manual hand pick.

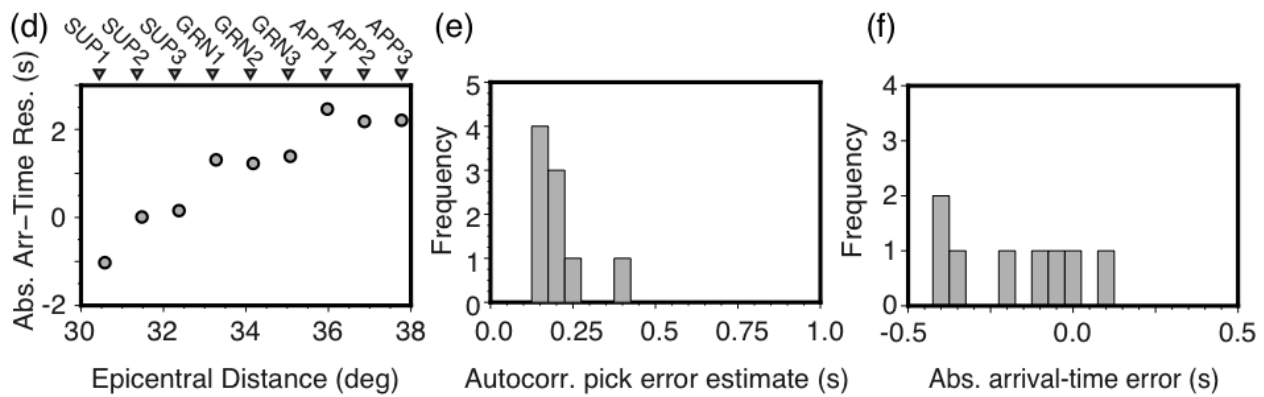
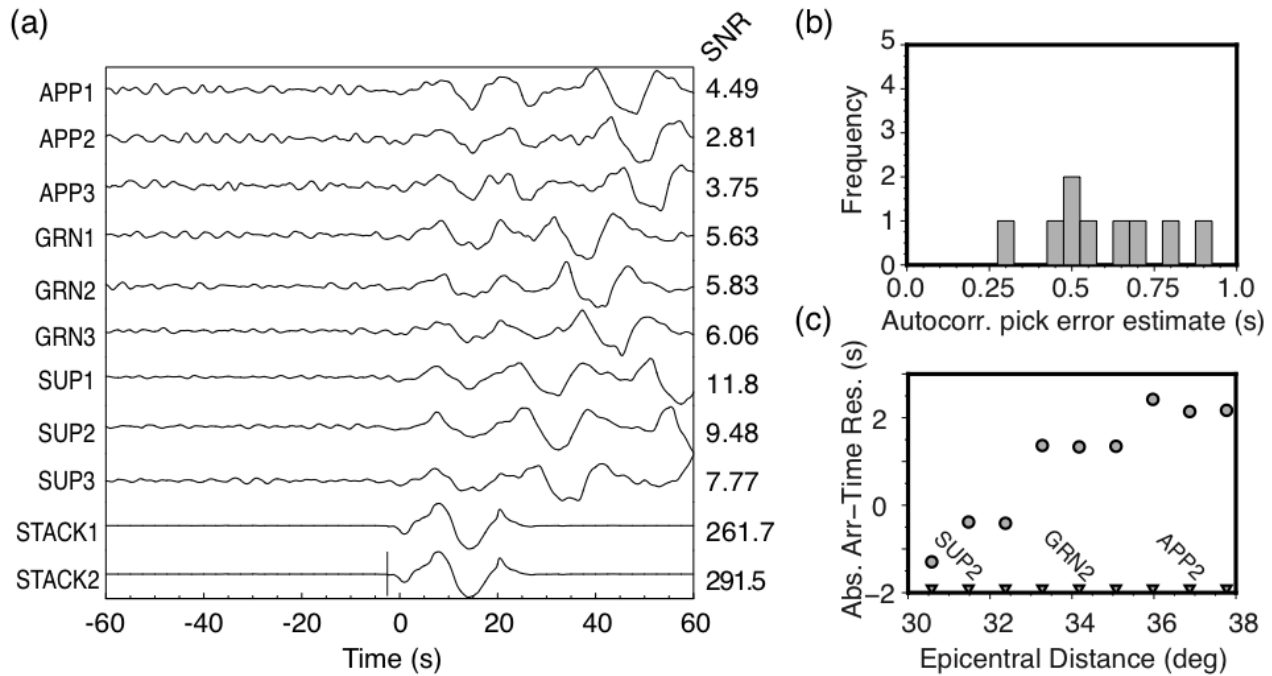


Figure 4.13: **Figure to show results from S-wave synthetic testing**, subplots are identical to Figure 4.12.

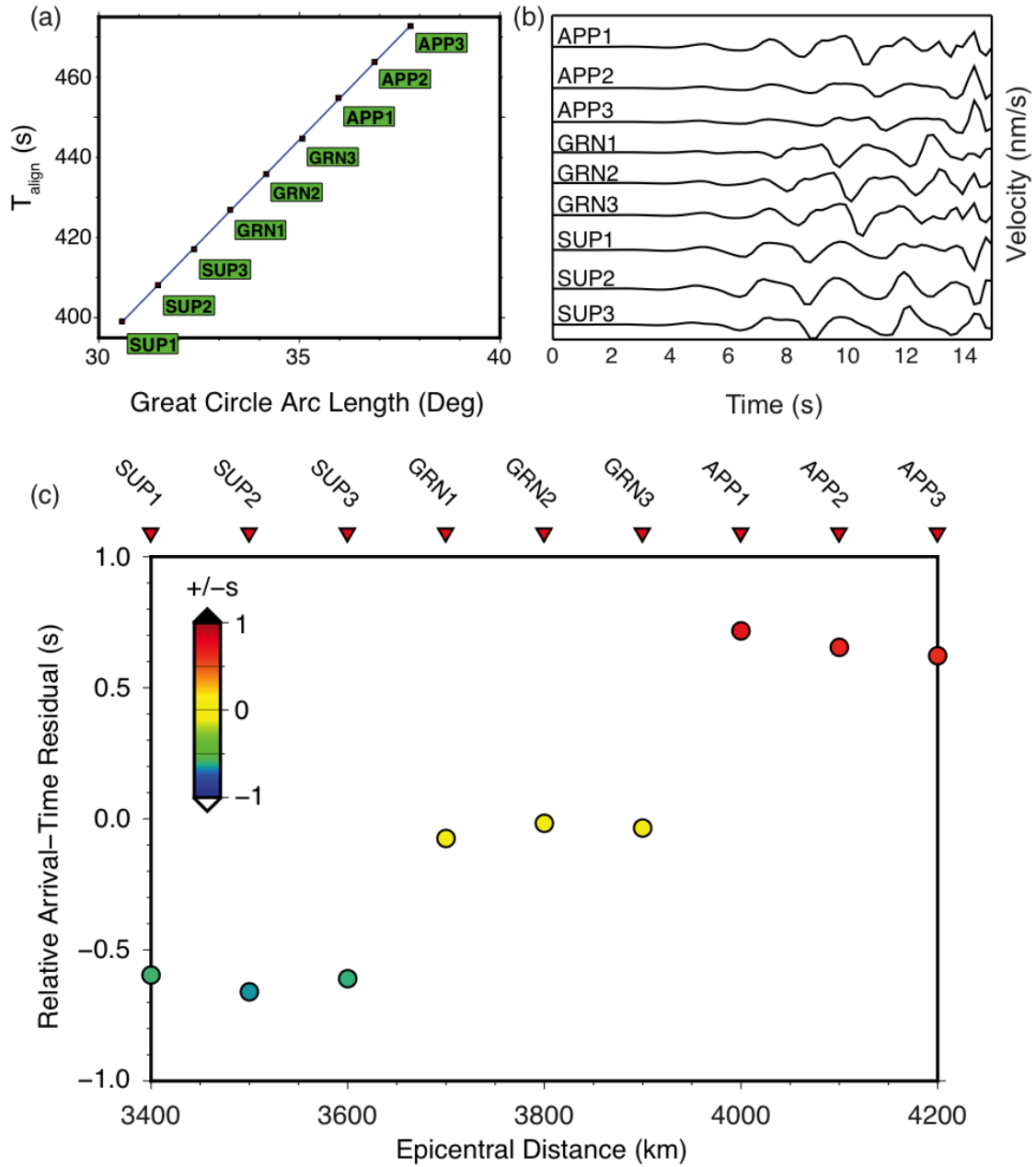


Figure 4.14: **(a): Distribution of MCCC derived arrival-times T_{align}** (VanDecar and Crosson, 1990) with great circle arc length for the synthetic P-wave arrivals. **(b):** Filtered velocity seismograms aligned at 5 seconds on MCCC derived arrival-times. **(c):** The distribution of relative arrival-time residuals for the array of synthetic seismometers (Figure 4.9) at increasing epicentral distance. The removal of the mean arrival-time for the array results in approximately zero residuals for central stations in the symmetric synthetic array.

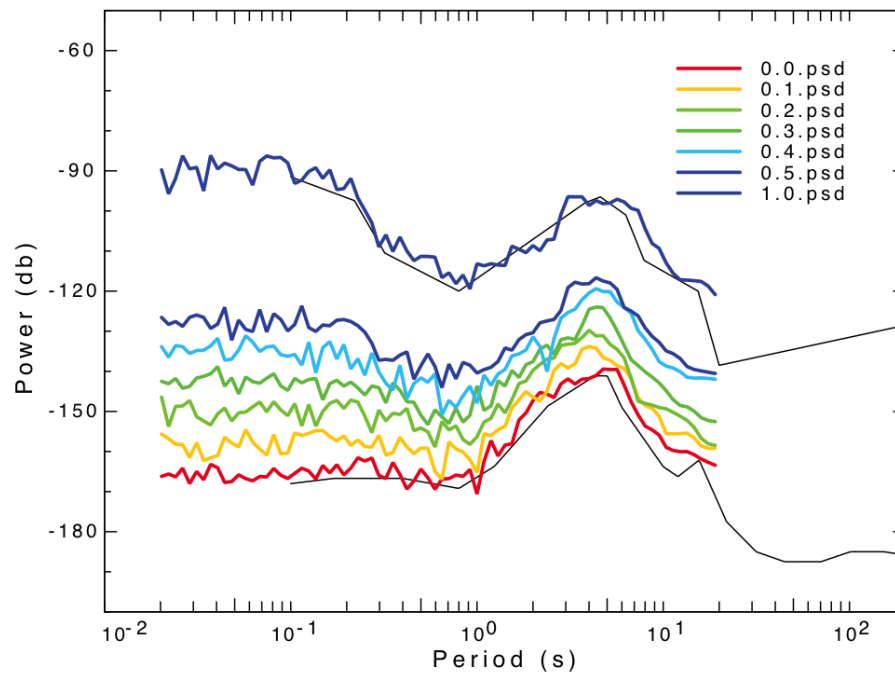


Figure 4.15: **Modelled power spectra for teleseismic background noise** models (after *Peterson*, 1993) varying between high (1.0) and low (0.0) extremes. These are used to add progressive levels of noise to synthetic seismograms. Black curves represent the upper and lower bound on observed noise.

4.5 Conclusions

We have developed a strategy (AARM) to determine absolute arrival-times using routinely processed, teleseismic, relative arrival-time datasets from regional networks. A manual pick of the first arrival of stacked, pre-aligned, unfiltered traces allows absolute arrival-times to be calculated for each station record.

A database of $\sim 10,000$ picks from networks in southeast Canada is used to test AARM. When checked against the ISC archive, the difference in absolute arrival-times shows 83 % are in agreement to within ± 0.5 s. The tests also indicate that AARM is effective for a low average trace SNR (> 2). Further testing of synthetic data indicates that AARM is accurate to better than 0.25 s in an absolute sense, on par with modern worldwide databases. AARM can thus be used to incorporate data from dense, but short-duration temporary seismograph networks into global pick databases and absolute wavespeed tomographic inversions.

Chapter 5

Absolute arrival-time tomography of North America

5.1 Chapter overview

Cratons, the ancient cores of the continents, have survived thermal and mechanical erosion over multiple Wilson cycles, but their ability to withstand modification during orogenesis is debated. The $\sim 1.1\text{ Ga}$ Grenville Orogen of North America documents protracted orogenesis ($\geq 300\text{ Ma}$) at the eastern flank of the Proterozoic Laurentian continent. It is thus an ideal locale to study the impact of orogeny on Precambrian lithosphere of varying age. Critically lacking from the geophysics community is a continental-scale seismic image capable of sampling the mantle lithosphere of the entire Grenville Orogen at adequate resolution.

Building on previous work on the Grenville Orogen in southeast Canada (Chapter 3), coupled with the ability to recover absolute delay times for temporary seismograph deployments in this region (Chapter 4), in this chapter I present the first P-wave tomographic image of the North American continent capable of imaging the entire Grenville Orogen. Again, the seismic images provoke discussion on modification of the lithosphere during the Grenville Orogeny, when considered in the context of the Proterozoic geological history of North America. Specifically, I explore whether these seismic images can be best explained by variations in the extent of lithospheric delamination during orogenesis. This debate furnishes the opportunity to speculate on the initiation of modern-day orogenic processes and the timescales of cratonisation.

5.2 Methods and data

5.2.1 Global and continental datasets

The global database of absolute arrival-time picks utilises the P, PP, Pn, Pg and pP phases (Figure 5.1) from the “EHB” dataset (*Engdahl et al.*, 1998) comprising earthquakes recorded during the period 1964 – 2007 (*Li et al.*, 2008a). The global database is supplemented with two primary sources of continental-scale datasets. The USArray Transportable Array study of *Burdick et al.* (2017) provides absolute arrival-time residuals for time period January 2013 to May 2016. The dataset incorporates absolute arrival-time picks for the phases P, Pn and PP from 6130 earthquakes. Typical interstation spacing is $\sim 70\text{ km}$ which provides unprecedented spatial sampling for a continental seismic network. Further constraints on the background model are provided by the Annual Bulletin of Chinese Earthquakes (CSN) (*Li et al.*, 2008a). All phase ray paths, earthquakes and stations used in the inversion presented here, are shown in Figures 5.1, 5.2 and 5.3.

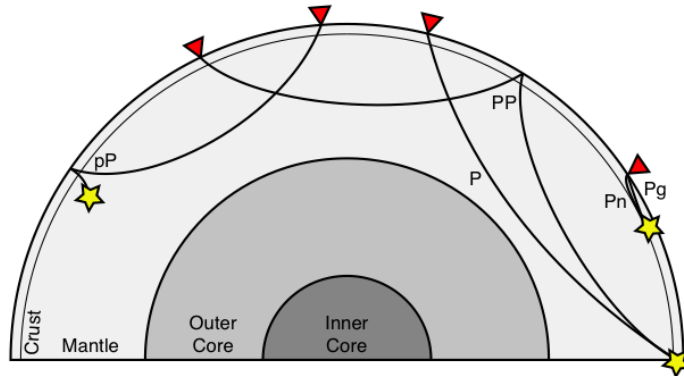


Figure 5.1: **Schematic of P-wave phases** derived from the datasets published by *Li et al.* (2008a) and *Burdick et al.* (2017) and P-wave measurements from southeast Canada, used here for the first time in a continental absolute arrival-time inversion of North America.

5.2.2 Uniquely available datasets

To improve previous continental inversions for the U.S. (e.g., *Burdick et al.*, 2017), resolution is extended northwards into the Canadian Shield, by incorporating numerous smaller regional seismic networks (e.g., *Boyce et al.*, 2016, see Chapter 3). These include 9032 P arrivals from 238 earthquakes recorded in SE-Canada at 160 stations over the period 2007 – 2016 (*Boyce et al.*, 2016), and arrivals from temporary seismic networks in the western Superior Province and the QM-III seismic

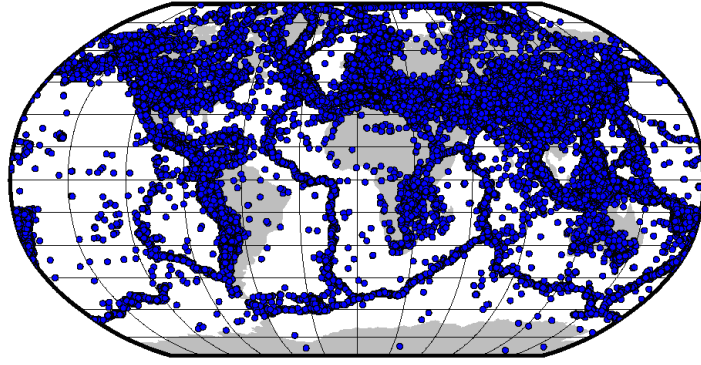


Figure 5.2: **Global earthquake map** derived from the published datasets of *Li et al.* (2008a) and *Burdick et al.* (2017) combined with teleseismic earthquakes recorded in southeast Canada. In total 476,783 earthquakes are included in the inversion.

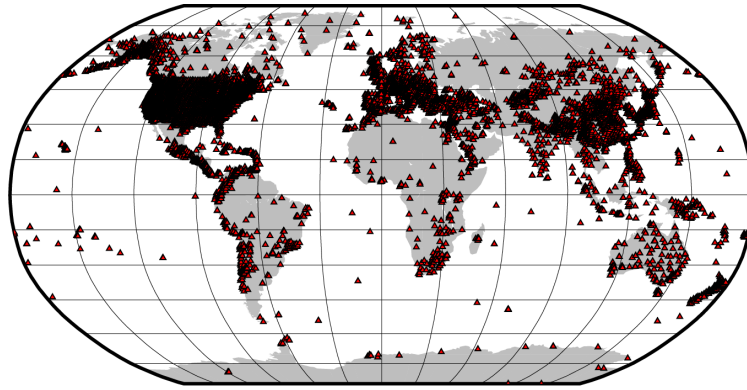


Figure 5.3: **Global map of stations** derived from the datasets published by *Li et al.* (2008a) and *Burdick et al.* (2017) and P-wave datasets southeast Canada.

network which traverses the interior of Quebec to the Canadian Maritimes (*Petrescu et al.*, 2016). Details of these datasets are included below.

Identical data processing steps were followed for these datasets to ensure compatibility (see sections 3.3, 4.3). Such data were not included in any further relative arrival-time inversions but are used to supplement datasets in the P-wave absolute arrival-time inversion after subsequent conversion to absolute arrival-times.

Western Superior networks

Data were sourced from the Canadian National Data centre, ‘Superior Province Rifting Earthscope Experiment’ and ‘Teleseismic Studies in the Western Superior Transect’ networks from the western Superior province. From the period 1997 – 2016 this yielded 446 $m_b > 5.5$ earthquakes with high SNR

resulting in 4042 relative arrival-time picks from 34 stations. Results from this analysis are shown in Figure 5.4. Once converted to absolute arrival-times using the AARM method (Boyce *et al.*, 2017, see Section 4.3) the western Superior dataset comprised 3766 picks from 408 earthquakes (Figure 5.5).

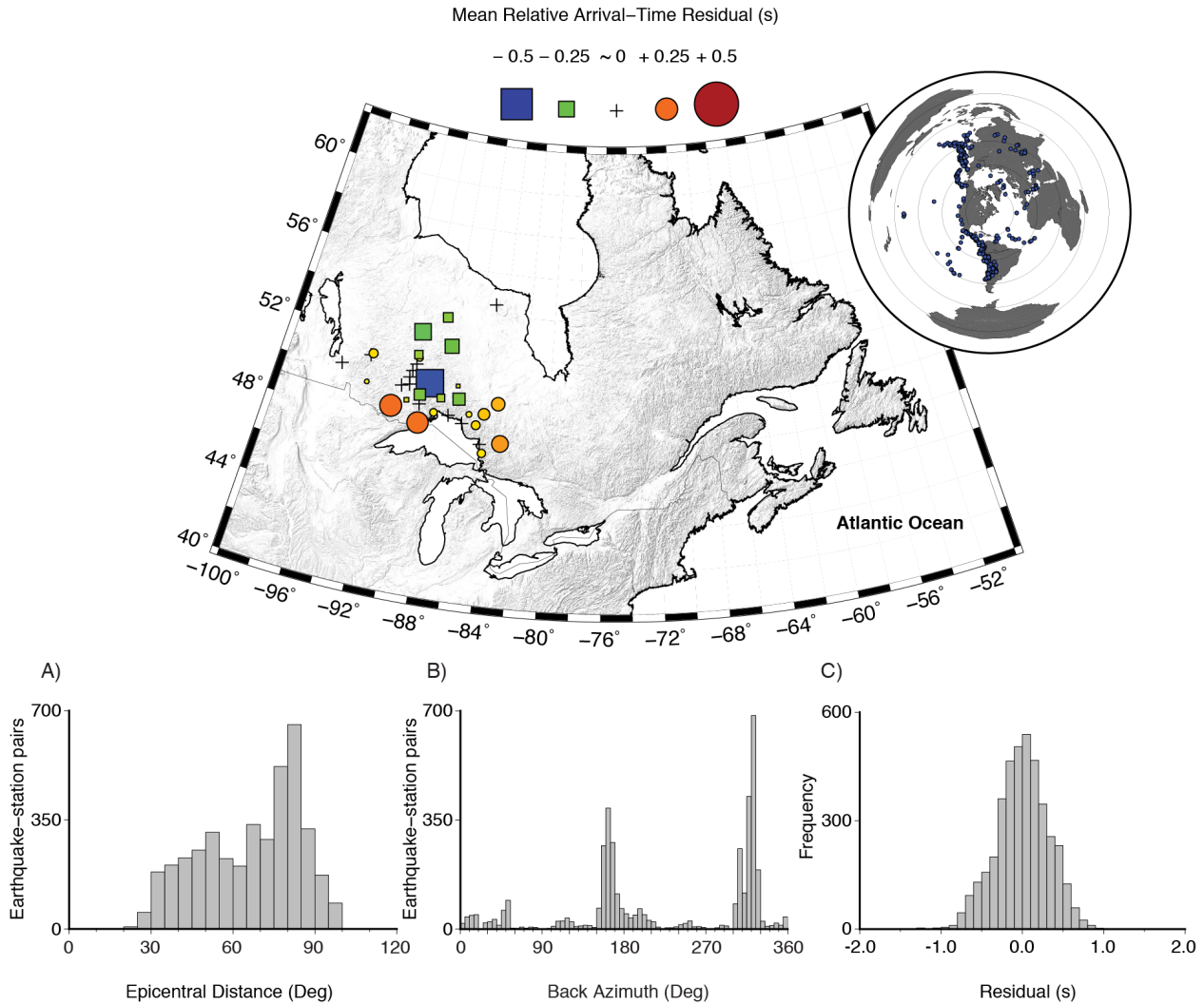


Figure 5.4: **Relative arrival-time dataset for the western Superior.** **ABOVE** - Mean relative arrival-time residual for 34 stations across the Western Superior Network. Early arrivals (faster) are represented by blue squares, later arrivals (slower) are shown as red circles. Inset globe shows the distribution of magnitude $m_b > 5.5$ earthquakes recorded with adequate SNR at ≥ 4 stations across the network between 1997 – 2016. Concentric rings mark 30° increments in epicentral distance from the Superior province in southeast Canada. **BELOW** - Plot of earthquake distribution with epicentral distance (A), earthquake frequency with backazimuth (B) and frequency distribution of residuals (C) for the P-wave dataset.

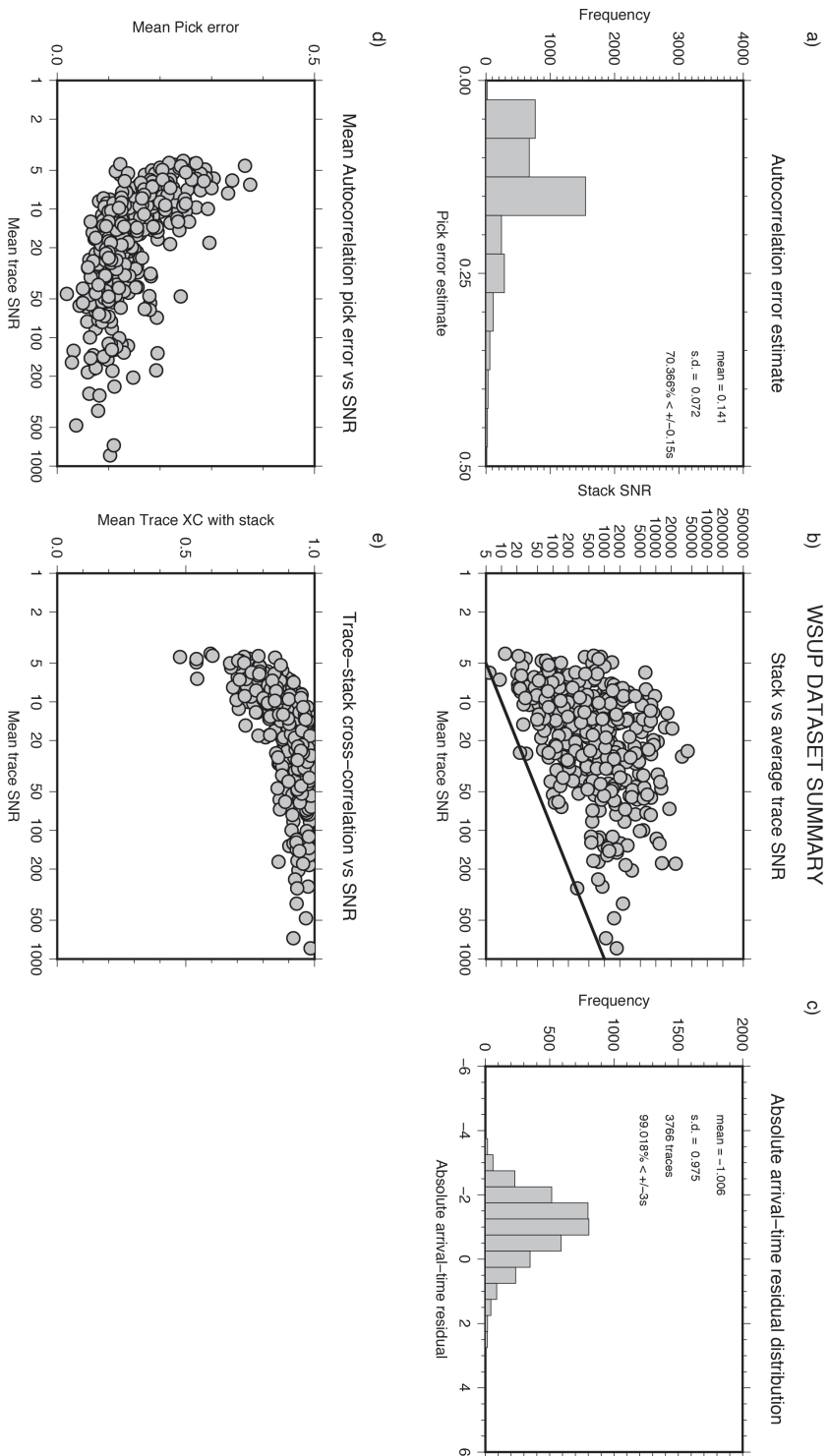


Figure 5.5: **AAFM overview (using cross-correlation scheme) for the western Superior network.** (a): The autocorrelation pick error estimate for the entire dataset. (b): The variation in SNR of the final stacked trace against earthquake averaged SNR. The black line shows the line $y = x$. (c): The distribution of absolute arrival-time residuals for the dataset. (d): The variation of mean autocorrelation pick errors with earthquake averaged SNR. (e): Shows the mean cross-correlation coefficient of each trace with the stack as a function of earthquake-averaged SNR. For further detail on this output see Chapter 4.

QM-III network

Additional data for southeast Canada were also sourced from the QM-III network comprising 34 stations which became publicly available through the IRIS Data Management Center in August 2017. Between 2012 and 2015, 297 $m_b > 5.5$ earthquakes yielded 4142 relative arrival-time residuals (Figure 5.6). Following the AARM method (Boyce *et al.*, 2017, see Section 4.3) the QM-III network yielded 3599 absolute P-wave arrival-time picks from 250 earthquakes (Figure 5.7).

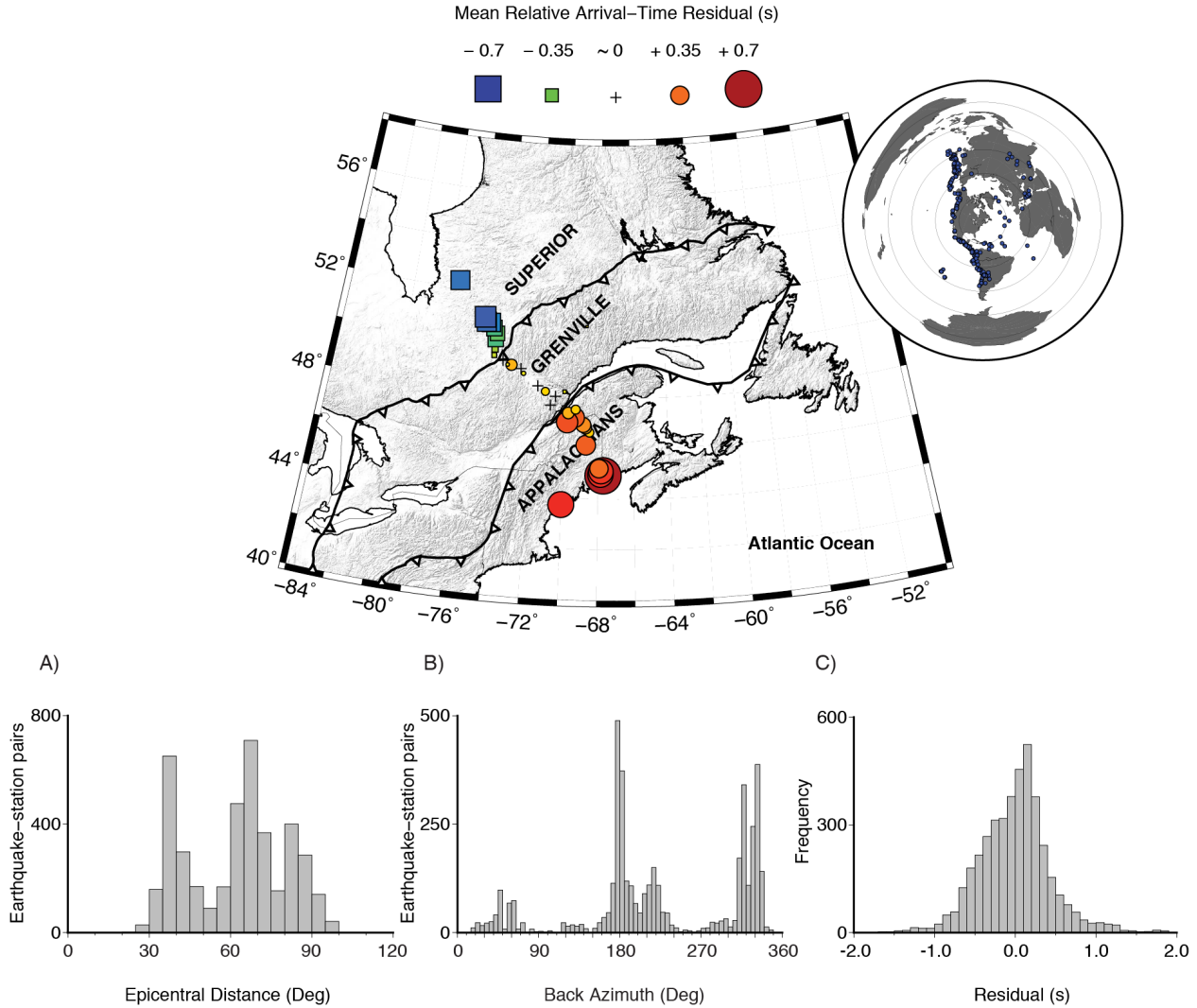


Figure 5.6: **Relative arrival-time dataset for the QM-III network.** **ABOVE** - Mean relative arrival-time residual for 34 stations across the QM-III Network. Early arrivals (faster) are represented by blue squares, later arrivals (slower) are shown as red circles. Inset globe shows the distribution of magnitude $m_b > 5.5$ earthquakes recorded with adequate SNR at ≥ 4 stations across the network between 2012 – 2015. Concentric rings mark 30° increments in epicentral distance from the Superior province in southeast Canada. **BELOW** - Plot of earthquake distribution with epicentral distance (A), earthquake frequency with backazimuth (B) and frequency distribution of residuals (C) for the P-wave dataset.

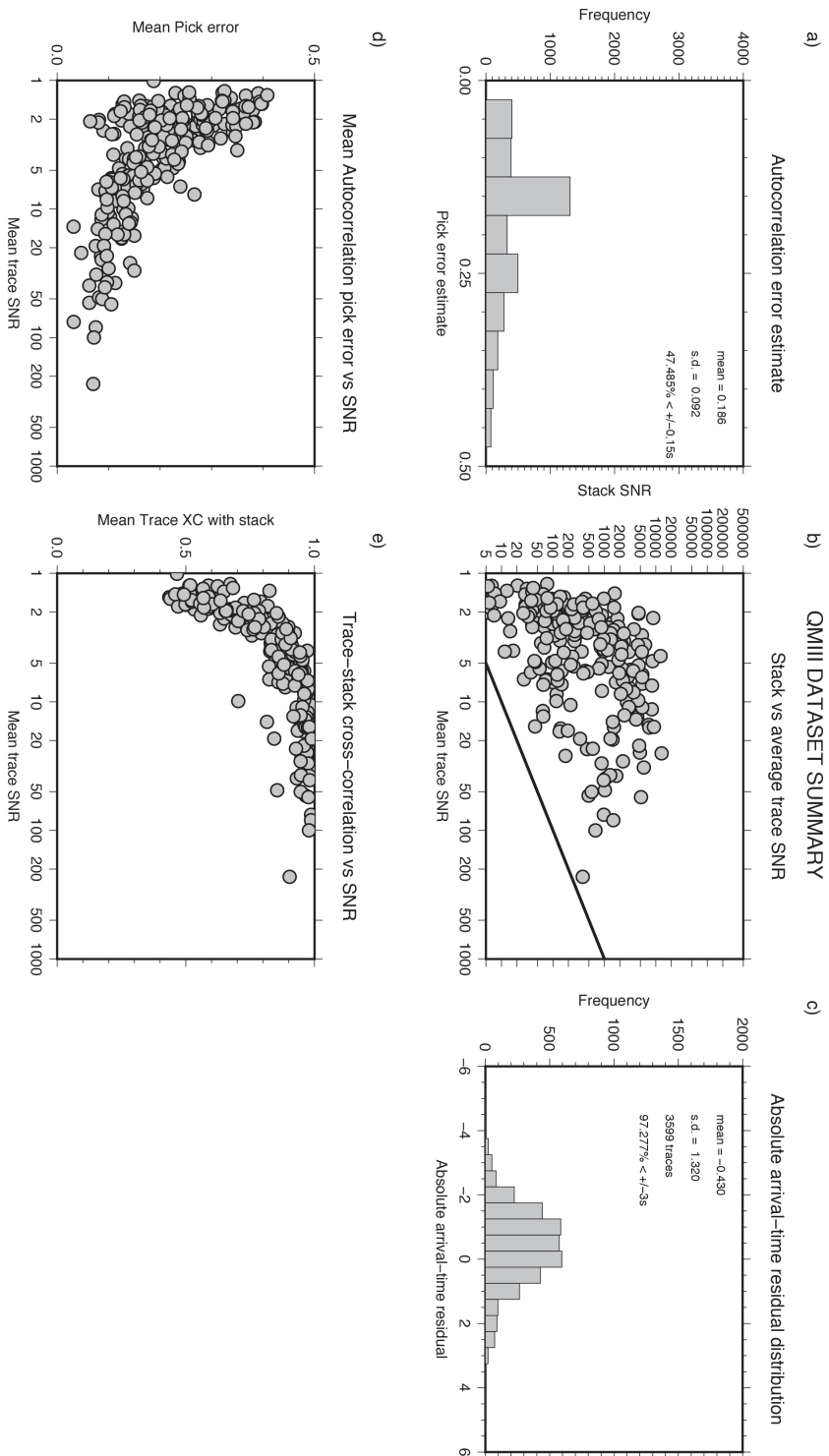


Figure 5.7: **AARM overview (using cross-correlation scheme) for the QM-III network.** (a): The autocorrelation pick error estimate for the entire dataset. (b): The variation in SNR of the final stacked trace against earthquake averaged SNR. The black line shows the line $y = x$. (c): The distribution of absolute arrival-time residuals for the dataset. (d): The variation of mean autocorrelation pick errors with earthquake averaged SNR. (e): Shows the mean cross-correlation coefficient of each trace with the stack as a function of earthquake-averaged SNR. For further detail on this output see Chapter 4.

New absolute arrival-time measurements

These additional networks have, until now, only been used to report relative arrival-time residuals for use in regional inversion models to study the local mantle velocity structure (e.g., *Boyce et al.*, 2016; *Frederiksen et al.*, 2013). By utilising the AARM method (*Boyce et al.*, 2017, Chapter 4), absolute arrival-times for the regional seismic networks can be recovered. Consequently, these temporary networks deployed in Canada (Figure 5.8) can be directly included in the absolute arrival-time tomographic inversions.

Figure 5.8 shows mean absolute arrival-time residuals plotted for stations in southeast Canada derived from the AARM method (*Boyce et al.*, 2017, Chapter 4). Variations in relative arrival-time residuals result from velocity structure beneath the local network. However, absolute arrival-time residuals shown here, record velocity variation along the entire ray path from source to receiver. Residuals may therefore not represent upper mantle or lithospheric velocity structure beneath the station network in which first order variations may be easily seen (*Boyce et al.*, 2016). However, it is reasonable to assume that a good azimuthal coverage of earthquakes for each station (Figure 5.8) will result in an observable signature of velocity structure from beneath the network. Prior to inversion, these data are corrected for station elevation and the Earth’s ellipticity (e.g., *Li et al.*, 2008a).

5.2.3 Inversion technique

The inversion approach closely follows that of *Burdick et al.* (2008, 2017); *Li et al.* (2008a). Rays are traced (e.g., *Červený*, 1972) in the 1D reference model ak135 (*Kennett et al.*, 1995), and are clustered to form composite rays in regions of dense sampling (e.g., *Li et al.*, 2008a; *Zhang and Thurber*, 2005). Initially, the model is parameterised globally on a regular non-overlapping constant velocity cells with spacing of 45 km in depth, 0.35° in latitude and longitude. Where ray sampling is poor, adjacent grid cells are combined up to a minimum ray count per cell. As spatial sampling provided by the USArray network dominates the grid, a lower minimum hit count of 900 is used to allow finer model resolution north of the US-Canadian border. A comparison of the adaptive parametrisation result for increasing minimum ray hit count is shown in Figure 5.9. An iterative, linearised, least squares inversion (LSQR) (*Paige and Saunders*, 1982) is used to solve for slowness and hypocentre mislocation parameters (e.g., *Spakman and Nolet*, 1988) with respect to a 1D reference model ak135 (*Kennett et al.*, 1995) to

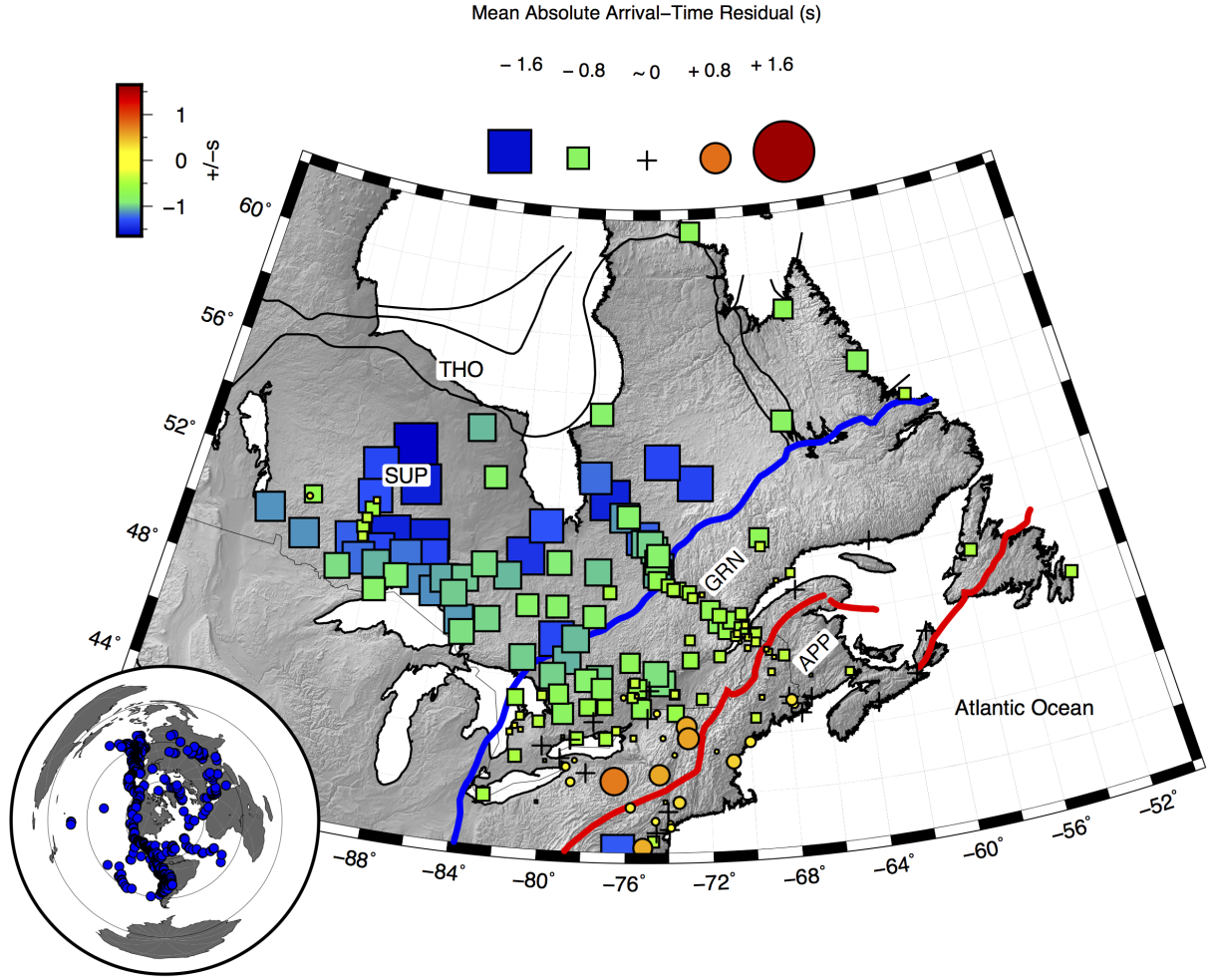


Figure 5.8: **Plot of mean absolute arrival-time residuals** plotted for stations in southeast Canada derived from the AARM method (Boyce *et al.*, 2017, see Chapter 4). Negative values represent early arrivals (fast) and positive values represent later arrivals (slow). The Superior Province is dominated by the negative residuals of ≥ 1.0 s. In the Grenville Province there is a marked decrease to negative residuals predominately ≤ 0.8 s. The southeast Grenville and Appalachian province are dominated by slower residuals of ~ 0 s and low amplitude positive residuals. The tectonically stable terranes of eastern Canada are typically dominated by fast seismic wavespeeds in global tomographic models (e.g., Li *et al.*, 2008a; Ritsema *et al.*, 2011) consistent with these observations. Inset is a global plot of earthquakes contributing to this dataset.

minimise the following function:

$$\epsilon = \| \mathbf{Gm} - \mathbf{d} \|^2 + \lambda_1 \| \mathbf{Lm} \|^2 + \lambda_2 \| \mathbf{m} \|^2. \quad (5.1)$$

$\mathbf{G}$ is the sensitivity matrix, $\mathbf{m}$ is the model, $\mathbf{d}$ is the data, $\mathbf{L}$ is a smoothing operator. Regularisation

is required to overcome incomplete sampling (e.g., *VanDecar and Snieder, 1994*). In the least squares minimisation of Equation 5.1, model roughness is controlled by the smoothing operator, $\mathbf{L}$ and the value of λ_1 . Model damping is controlled by the model variance from the starting model weighted by the term λ_2 . Although the majority of LSQR convergence is reached quickly, 400 iterations of the inversion scheme are conducted for all models. P-wave velocity anomalies are plotted as a percentage deviation from ak135 (*Kennett et al., 1995*).

Although it is possible to include a crustal correction in the cost function (Equation 5.1: $\lambda_3 \|\mathbf{C} - \mathbf{M}_C\|^2$, where $\mathbf{C}$ is a crustal model and $\mathbf{M}_C$ is the crustal component of the model), without a detailed crustal model for the eastern U.S. and Canada, it is uncertain whether a distinct improvement in the model is gained (*Burdick et al., 2008, 2014, 2017; Li et al., 2006, 2008a*). This issue is partly addressed by the inclusion of the Pn and Pg crustal phases which help to reduce downward smearing of anomalies in the lithosphere when compared to the recent model of *Burdick et al. (2017)*.

Figure 5.10 shows a comparison of the inversion output at 100 km depth and a cross section at 28°N including only mantle P-phases (M), Mantle and crustal phases (MC) and mantle, crust and free surface reflected phases (MCF). The inclusion of crustal phases decreases downwards smearing of fast velocity anomalies beneath eastern North America. Transition zone depths appear to be better resolved with the inclusion of free-surface reflected phases. Mantle, crustal and free-surface reflected phases are therefore all included in the inversion.

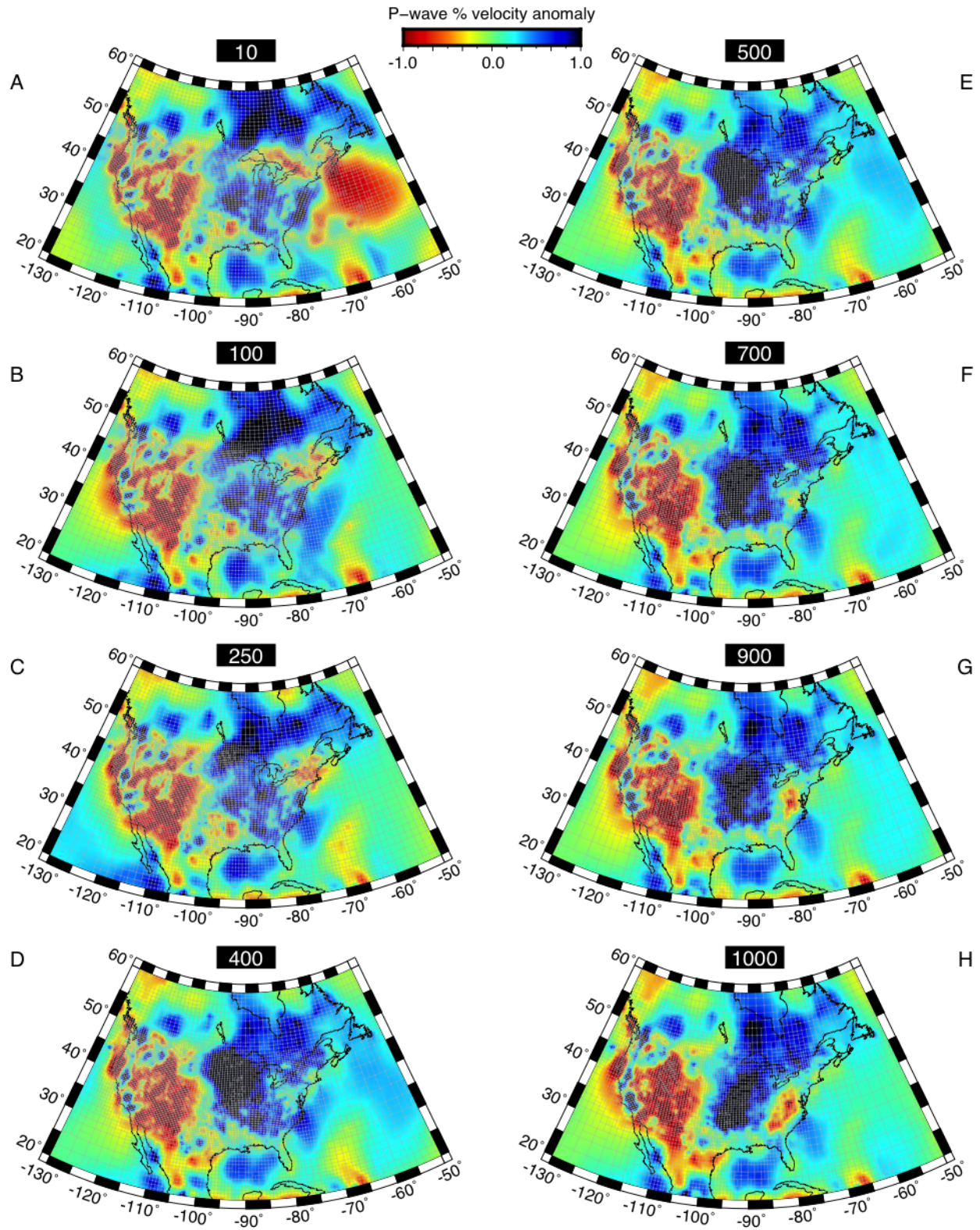


Figure 5.9: **Comparison of grid cell density** resulting from adaptive parameterisation routine for inversion outputs at 200 km depth. A through H are the result of a minimum cell ray hit count of 10, 100, 250, 400, 500, 700, 900 and 1000 respectively, indicated above each plot. The recovered model is more stable to perturbation in grid cell threshold at higher values.

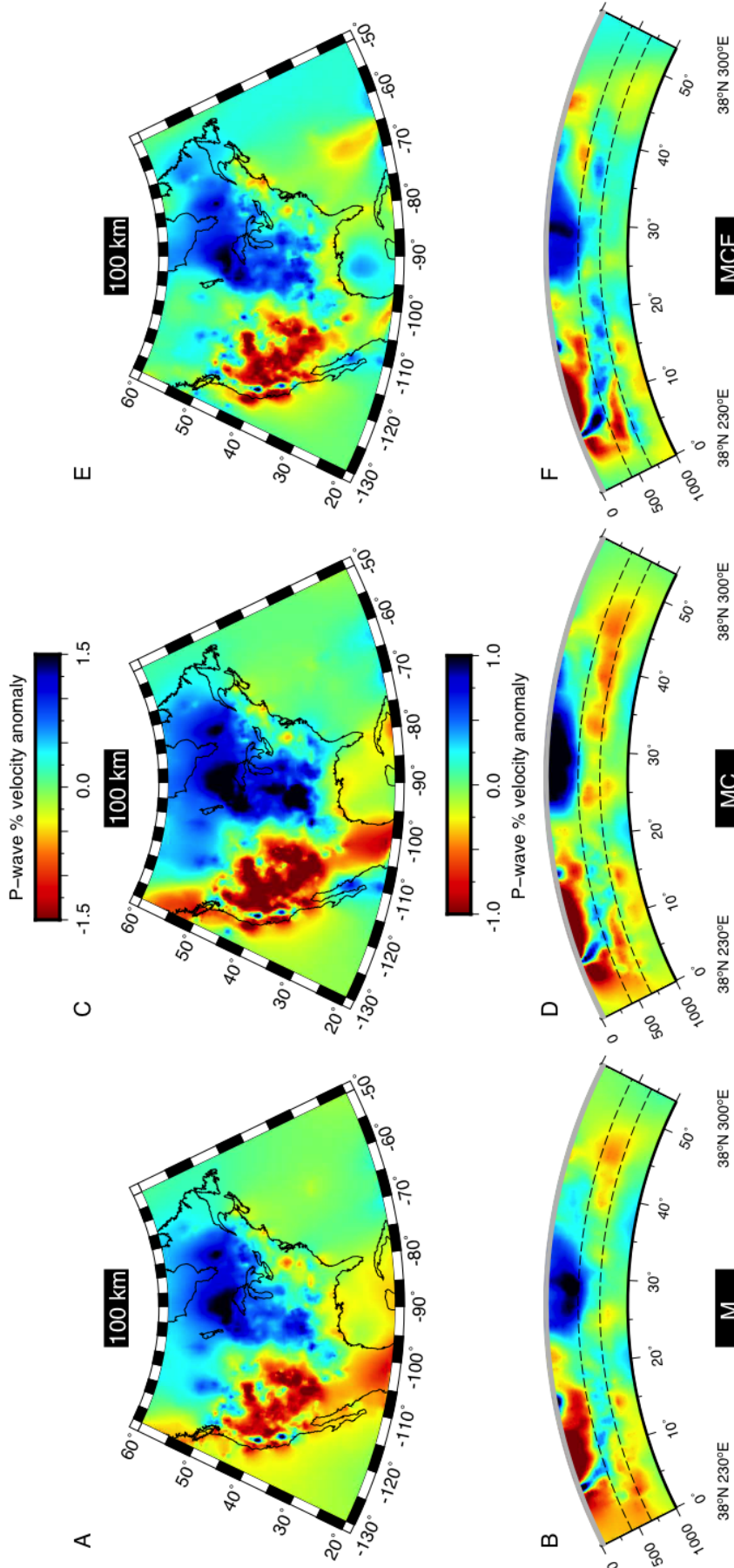


Figure 5.10: **P-wave model comparison including differing phases in the inversion.** The left-hand model (A and B) uses solely direct P-wave arrivals from phases passing through the mantle (M). The central model (C and D), also includes crustal phases (MC) such as Pn and Pg. The right-hand model (E and F) shows the results of the inversion adding free surface reflections (MCF) such as pP and PP. These additional phases are sourced from USArray and EHB datasets.

5.3 Resolution testing

5.3.1 Resolution assessment and dataset weighting

Synthetic resolution testing of a Canadian Shield structure is used to balance the contributions of the dense USArray network and the temporary networks added in Canada using the AARM technique described previously (Boyce *et al.*, 2017, see Chapter 4). It is reasonable to expect that Precambrian unmodified lithospheric structures will have similar seismic velocities, both north and south of the U.S. - Canadian border. A synthetic anomaly model, centred at 100 km depth, is placed beneath both the USArray and temporary networks sampled at a minimum of 900 ray-hits per cell (Figure 5.11). Accurate recovery of this model can be obtained by balancing individual dataset weights in the forward problem. Results show that smaller Canadian networks can be weighted twice that of the USArray dataset in order to achieve this (Figure 5.11). Whilst this appears to be a some-what subjective approach, equal dataset weighting would not present a viable option due to spatial and temporal sampling differences between networks that should not be ignored. Notably, the mean minimum interstation spacing for the USArray is 59.7 km, compared to 68.3 km for the Canadian datasets to the north. Downsampling of other datasets is not a viable option as this may introduce statistical bias and would throw away large quantities of good data. Further, spatially variant regularisation and imposed parameterisation threshold is not a current feature of this code and again require further ad-hoc assumptions.

5.3.2 Synthetic and checkerboard resolution tests

To address the lack of quantitative assessment of model resolution (e.g., Spakman *et al.*, 1989), using the dataset weighting criteria above, synthetic structure (Figure 5.12) and checkerboard resolution tests are performed (Figures 5.13 and 5.14). Using identical rays to the observed data, travel-time residuals are calculated through a specified (synthetic) model of velocity perturbations, onto which noise is added to simulate real-world picking error. Upon inversion of the synthetic data, the recovery of the model can be inspected, observing relative differences between parts of the model (e.g., Bastow *et al.*, 2008; Boyce *et al.*, 2016; Li *et al.*, 2008a; Villemaire *et al.*, 2012).

These synthetic anomaly tests are motivated by structure recovered in the observed data inversion (Section 5.4). In both synthetic models, a slow wavespeed anomaly ($-2.0\% \delta V_p$) is placed in the

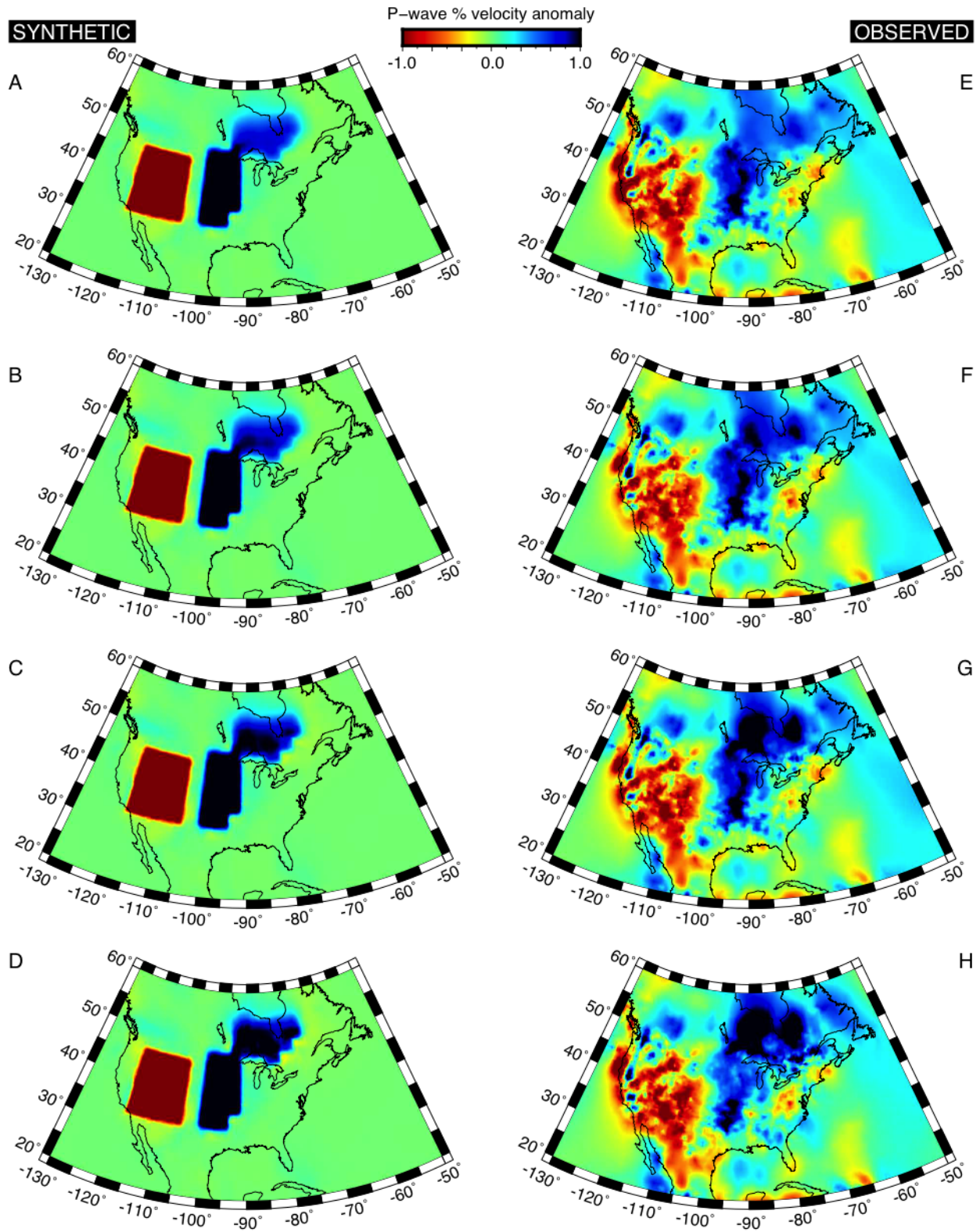


Figure 5.11: **Dataset weighting tests** showing recovery of $\delta V_P = \pm 2.0\%$ synthetic velocity model from Figure 5.12 (LHS - A-D) that crosses the U.S. - Canadian border. Dataset weightings for Canadian datasets of 1,2,5,10 are used relative to USArray and EHB dataset weighted at 1. The corresponding inversions of observed data are shown on the RHS (E-H). All models are plotted at 200 km depth.

western U.S. however in the east the models differ. The first model shows a more pervasive fast velocity structure centred in the upper lithosphere (100 km) designed to represent the fast wavespeeds of the Archean and flanking Canadian Proterozoic domains (Grenville Province), whilst the second shows a fast velocity anomaly centred in the lower lithosphere (200 km), representing the unmodified Mesoproterozoic and Archean domains further west (Figure 5.12).

Across both models the slow velocity anomaly in the western U.S. is well recovered (amplitude recovery $\geq 50\%$), for which the USArray is largely responsible. The fast velocity anomaly in the upper lithosphere in the east is well recovered ($\geq 50\%$ amplitude), however in the northeast Grenville and the northern Superior Province amplitude recovery does reduce due to sparser station coverage (Figure 5.12). In the lower lithosphere, amplitudes are again well recovered in the U.S. but farther north in Canada, recovery amplitude drops to $\sim 25\%$. Lateral smearing of anomalies appears very low across both models however. These tests indicate that an east-west striking velocity boundary separating a faster northern Grenville from a slower south, is a recoverable feature.

Next, standard checkerboard tests are conducted, where alternating anomalies of $\pm 2.0\% \delta V_p$ at 5° and 10° widths, are input to estimate variations in short wavelength spatial recovery (Figures 5.13 and 5.14). Beneath the USArray, coverage is near-ideal. For the shortest wavelength model (5°) (Figure 5.13), amplitude recovery in the western U.S. is $\geq 50\%$, but slightly lower in the east at lithospheric depths (200 km). Deeper in the mantle (500 – 800 km) amplitude is pervasively $\geq 50\%$. In the lithosphere (200 km), beneath the newly added stations in southeast Canada, short wavelength resolution is improved over *Burdick et al.* (2017), amplitude recovery of up to 35% is seen, comparable to some regions beneath the USArray. In the western Canadian Cordilleran regions that suffer sparse station density, and the mid-Superior recovery of the checkerboard anomalies is poor ($< 10\%$ of input amplitudes). Using the 10° model (Figure 5.14), anomalies are well retrieved in the Grenville and mid-Superior, however the western Superior, southwest of James Bay shows weaker recovery. Amplitude recovery for this model is $\geq 50\%$ everywhere else in this model (Figure 5.14).

5.3.3 Trade-off analysis

By varying the regularisation of the model via parameters λ_i in Equation 5.1 a suite of models, ranging from smooth models with lower data fit, through to rough models with high data fit can be generated and plotted on a trade-off curve in Figure 5.15. Errors within the input dataset mean that the low

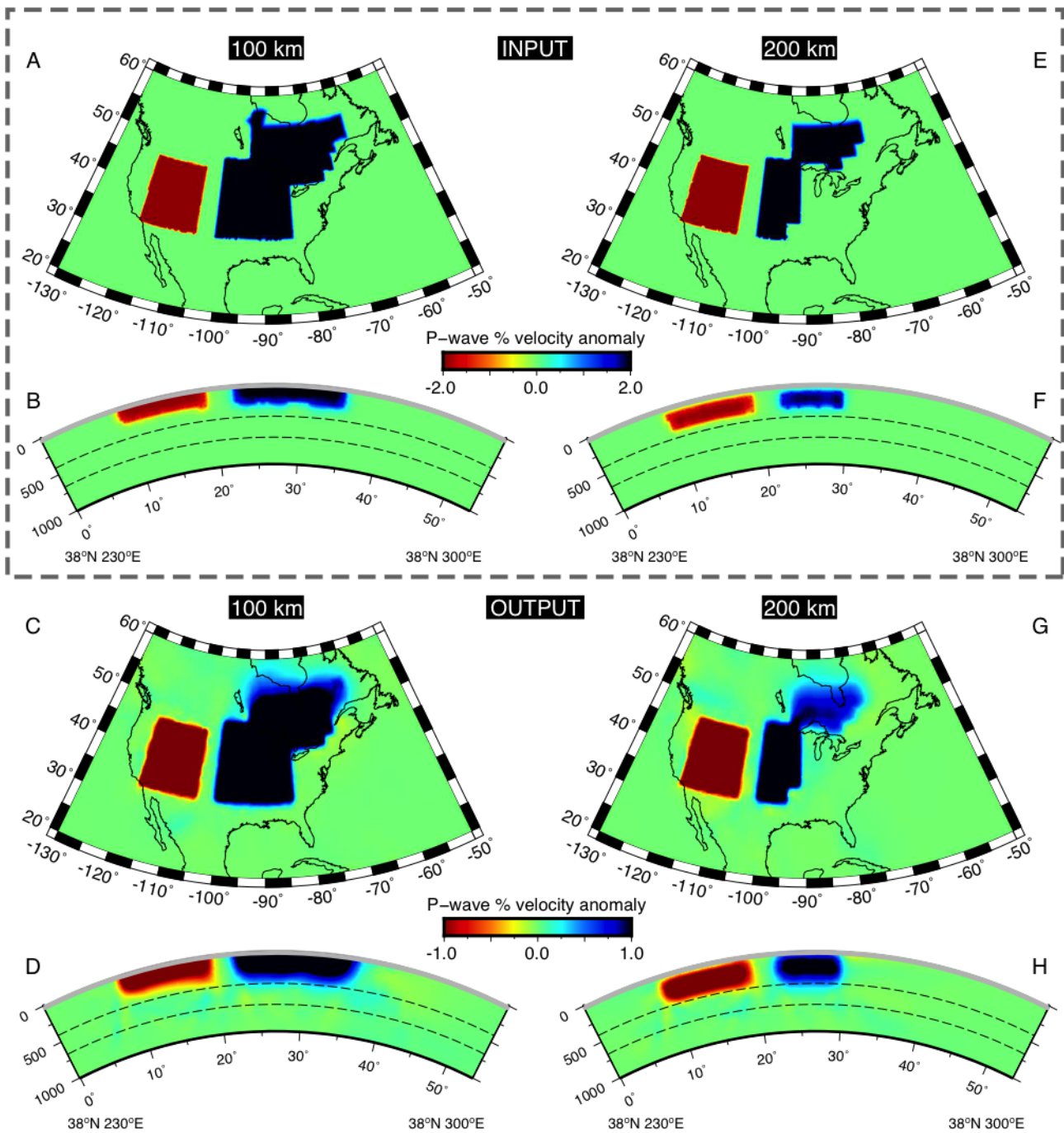


Figure 5.12: **Synthetic structure resolution tests** showing recovery of $\delta V_P = 2.0\%$ positive velocity synthetic anomaly structure from upper lithospheric depths (A-D - 100 km) and lower lithosphere (E-H - 200 km) in the P-wave model. A,B and E,F are the input models, C,D and G,H, are the recovered models, note the change in colour scale between input and output. These synthetic models are motivated by structure found in the inversion of observed data (Section 5.4).

misfit models are not appropriate whilst smooth models are unlikely to represent the true Earth that is not smooth. The most appropriate model is chosen on a somewhat ad-hoc basis. Commonly we seek the minimum structure (simplest) model that fits the data most well. This point typically exists at the knee of the trade-off curve in an ‘Occam’s razor’ approach (Constable *et al.*, 1987).

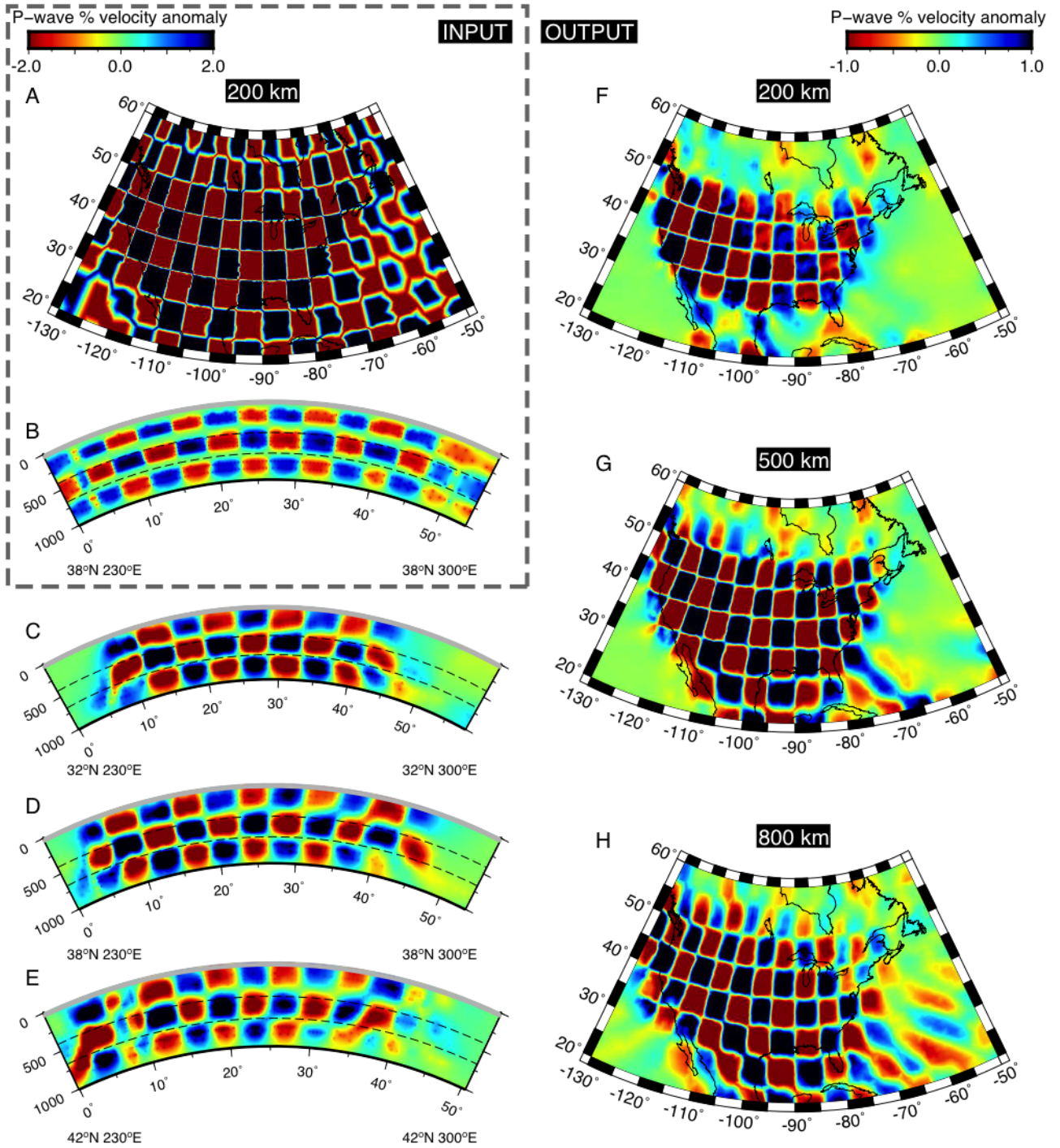


Figure 5.13: **Checkerboard resolution tests** showing results for 5° checkerboard velocity anomalies. A and B show the input velocity $\delta V_P = \pm 2.0\%$ anomalies, C, D and E show cross sections at 32°N , 38°N and 42°N through the checkerboard model. F, G and H show recovery of the model at 200 km , 500 km and 800 km depth.

The results of plotting RMS misfit against the model norm (i.e. a measure of data fit versus model complexity) does not produce a conventional exponential decay as seen in most trade-off curves (see Section 3.3). Models I-M produce overly smoothed results that do not fit the data well. At the knee of the curve, F-D, the models show too much velocity structure to be considered geologically feasible

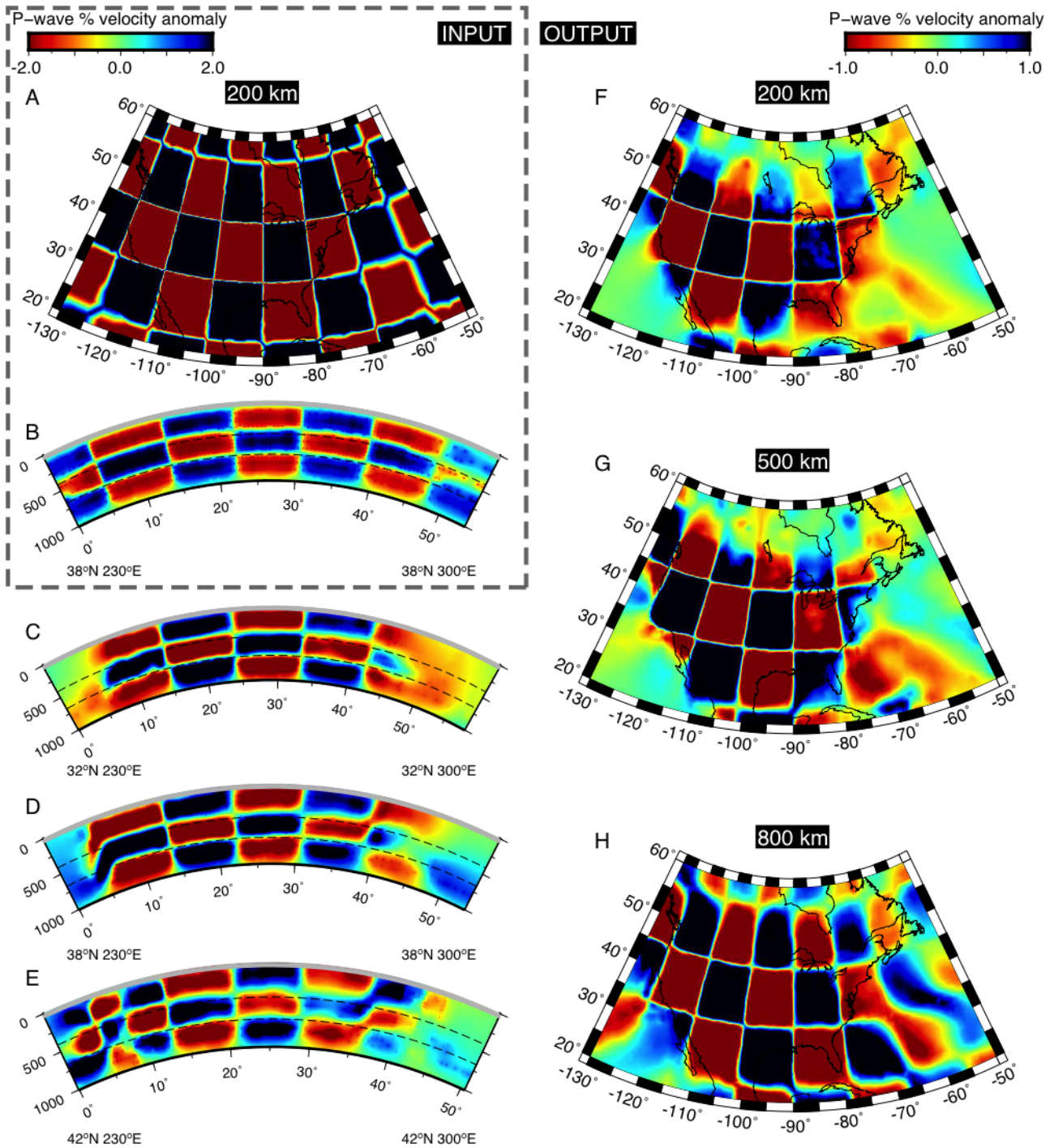


Figure 5.14: **Checkerboard resolution tests** showing results for 10° checkerboard velocity anomalies. A and B show the input velocity $\delta V_P = \pm 2.0\%$ anomalies, C, D and E show cross sections at 32°N , 38°N and 42°N through the checkerboard model. F, G and H show recovery of the model at 200 km , 500 km and 800 km depth.

or resolvable given the datasets used. Unusually, models C-A show an increase in data misfit despite even greater model complexity.

Typically for lithospheric scale models, continuing to increase model complexity results in reduced

data misfit likely due to the ‘white’ spectrum of heterogeneity within the lithosphere. The model is parameterised into the lower mantle and thus increasing model complexity will increase structure in the deep mantle as well. In the Earth, deep mantle heterogeneity is dominated by long wavelengths (*Su and Dziewonski, 1991*), so increased short-wavelength structure may result in the increased global measure of misfit for models C-A. Due to the above reasoning, model H is selected, above the knee of the curve to represent the most appropriate inversion model.

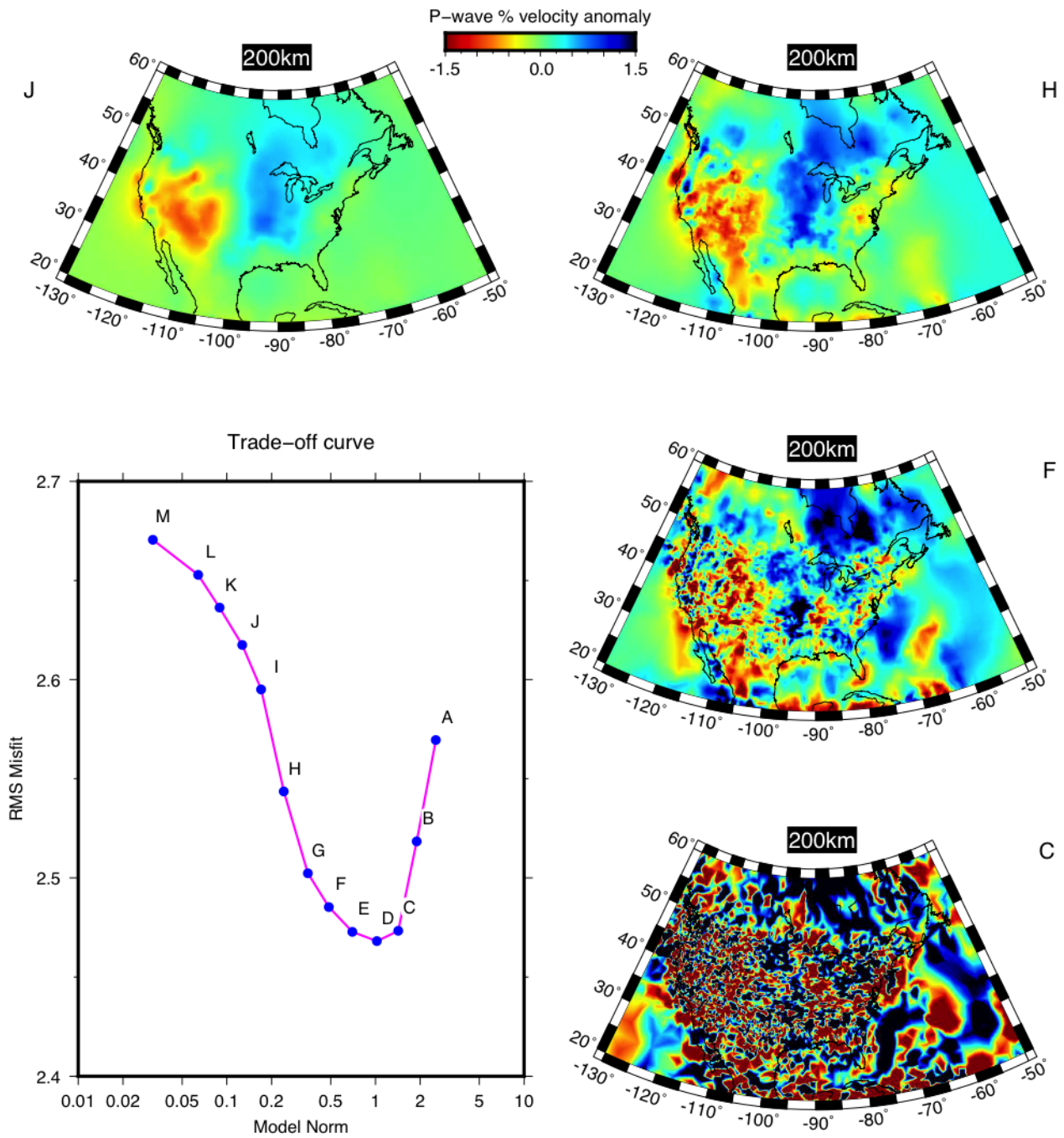


Figure 5.15: **Trade-off curve for the P-wave tomographic model** plotting the RMS data-misfit against the norm of the model. The chosen model for the inversion focusing on North America is H.

5.4 Tomographic model

5.4.1 Features of the new P-wave tomography model for North America

The following section describes the main features of the BBNAP18 tomographic model (Boyce-Bastow North America P-wave model 2018). At the continent scale, the shallow mantle velocity structure of North America (100 – 300 *km* depth) is dominated by pervasive slow velocities ($\delta V_P \leq -0.5\%$) in the west transitioning to fast velocities ($\delta V_P \geq 0.5\%$) in the central-eastern regions that about the average velocity ($-0.3\% \leq \delta V_P \leq 0.3\%$) coastal Appalachian terranes (Figures 5.16, 5.17). Within the well resolved continental U.S. and southeast Canada (Figures 5.12, 5.13 and 5.14), numerous smaller wavelength anomalies exist. At 36°N, 120°W a vertical fast velocity anomaly (continuous from 100 – 250 *km* depth) disrupts the pervasive slow velocities (Figure 5.16). To the northwest, a north-south striking fast velocity anomaly also extends from $\sim 40 - 50^\circ\text{N}$, in cross section this anomaly dips to the east (Figure 5.17). Beneath the Wyoming craton (Figure 5.16), seismic velocities are only slightly elevated with respect to the global mean ($\delta V_P \leq 0.4\%$). In the central Proterozoic regions of North America, fast velocities dominate at lithospheric depths.

In the Appalachians, the average seismic velocities are only disturbed by two slow velocity anomalies, the first at 41°N, 72°W, and a second at 37°N, 80°W. The first is constrained to the Appalachians in the upper lithosphere but may show an extension to the northwest in the lower lithosphere. The second anomaly is less spatially extensive but it widens with depth maintaining an approximately circular shape. Both slow velocity anomalies become less distinct below ~ 300 *km* depth. These anomalies appear most clearly along great circle path cross sections at 36°N and 32°N respectively (Figure 5.17).

Below the lithosphere, in the mid mantle (500 – 700 *km* depth) a fast velocity anomaly is seen at 35°N, 85°W (Figures 5.17, 5.18). In cross section at 32°N and 34°N it is unclear whether a fast velocity anomaly, exists between that at depth, to the lithosphere above. Deeper still at 1150 – 1350 *km* depth, a broad fast velocity anomaly exists, centred around 42°N, 100°W. To the east, a north-south striking fast velocity anomaly ($\delta V_P \approx 0.4\%$) extends from 1150 – 1650 *km*, turning to the north west below James Bay. These lower mantle anomalies have been a consistent feature of the model throughout testing, but may be on the limit of resolution of the dataset presented here. Due to the marked improvement in resolution over previous studies, the next section is dedicated entirely to exploring further the lithospheric velocity structure of eastern North America and southeast Canada within BBNAP18.

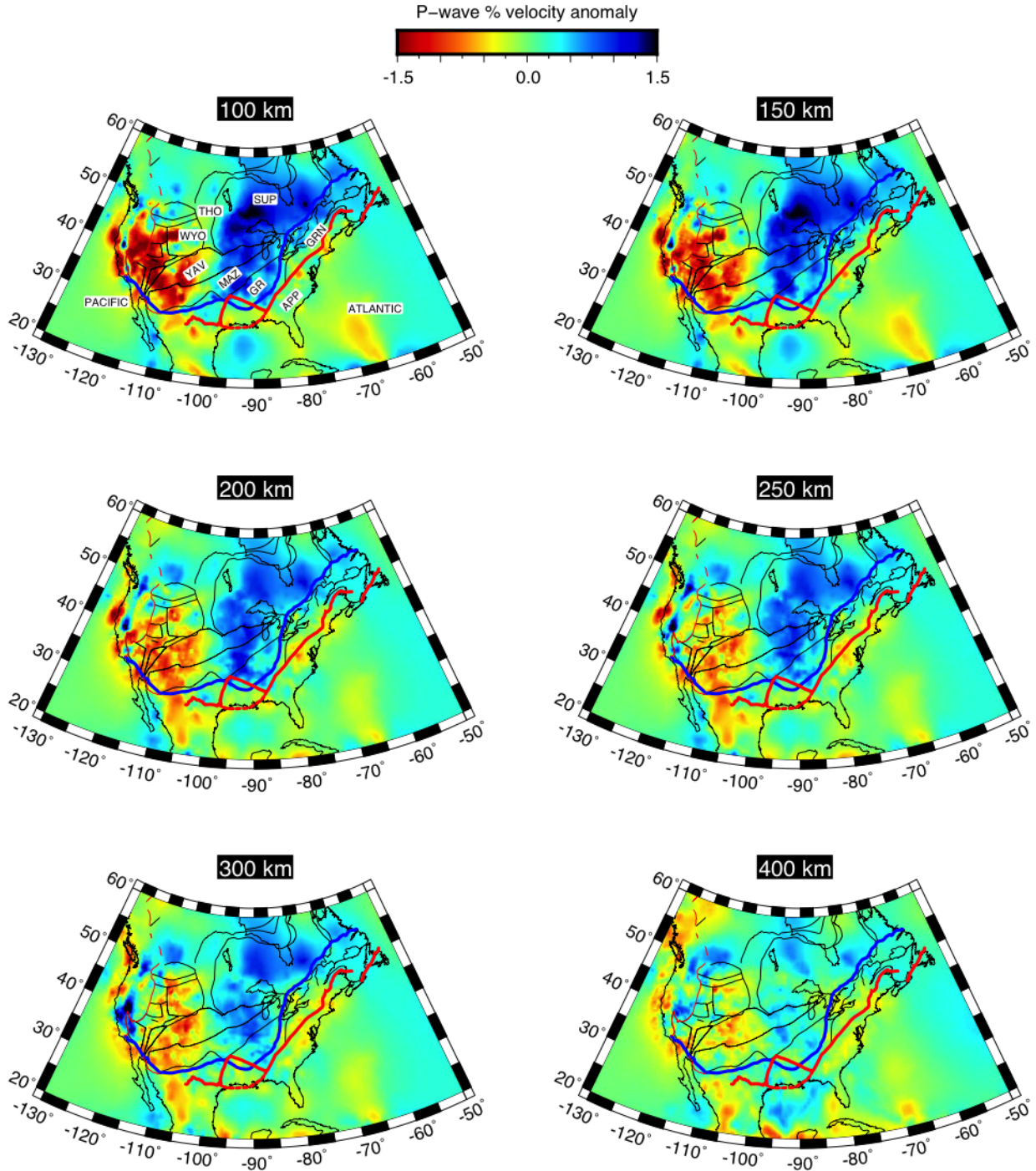


Figure 5.16: **Figure to show upper mantle results from the final P-wave velocity model for North America (BBNAP18) plotted as a percentage deviation from ak135. For reference, tectonic boundaries are plotted, see Figure 2.1.**

5.4.2 Eastern North American lithosphere

At depths of $75 - 150 \text{ km}$ in the north-eastern continental region ($\geq 40^\circ\text{N}$), fast velocities ($\delta V_P \geq 1.0\%$) exist throughout Ontario and Québec within the Superior Province but also extend eastward into the

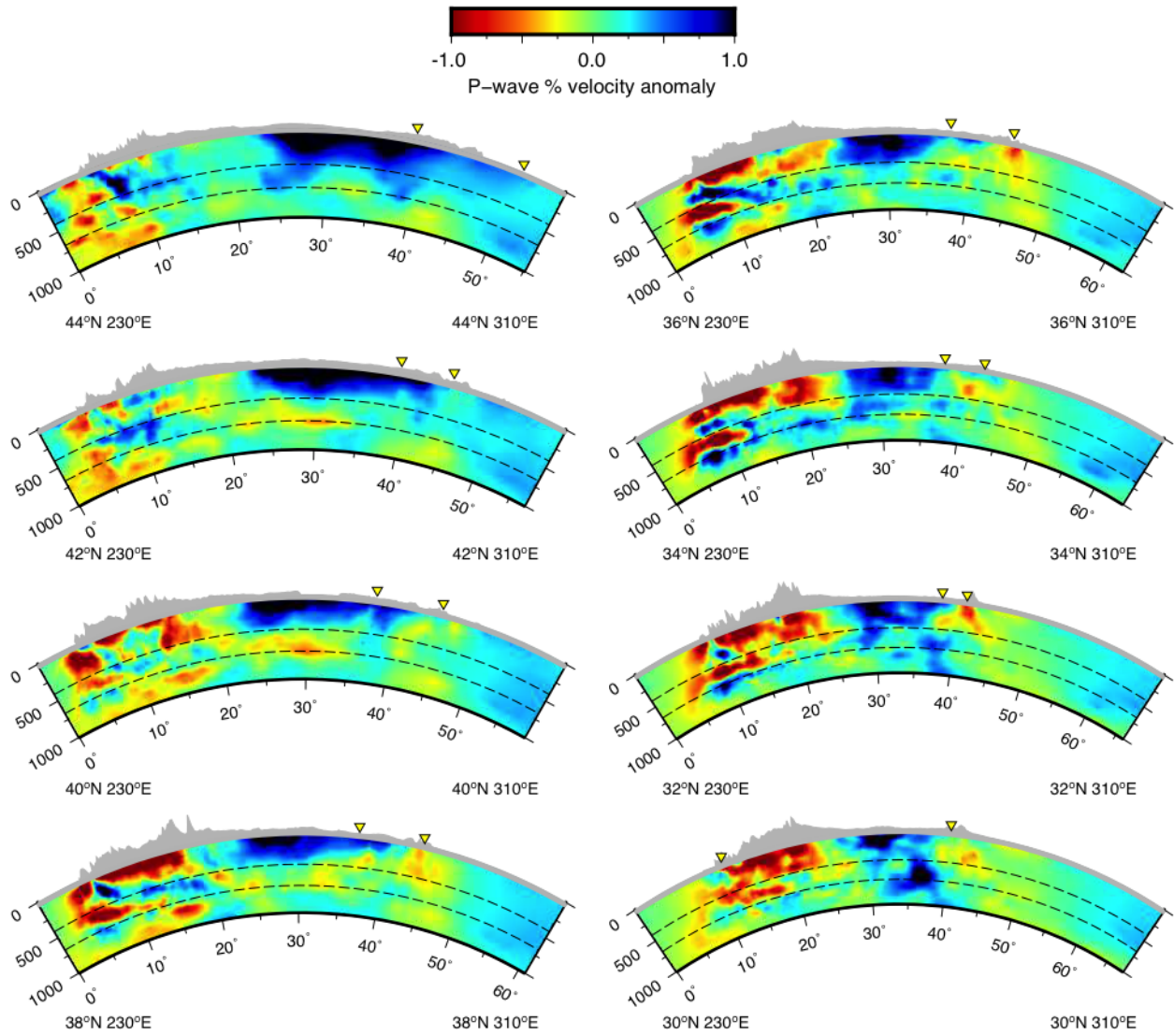


Figure 5.17: **Figure to show cross sections through the final P-wave velocity model for North America (BBNAP18)**, plotted as a percentage deviation from ak135. Note the slight difference in velocity scale between maps and cross sections. Cross sections show exaggerated topography in grey and location of GF and AF tectonic boundaries at the surface as yellow triangles (Figure 2.1).

Grenville Province (Figures 5.16, 5.17), albeit slightly slower. In the southern U.S. mid-continent, the fast velocities of the North American shield stretch from the N-S trending boundary at $\sim 105^\circ\text{W}$ but terminate abruptly, coincident with the inferred location of the GF (Figures 5.16, 5.17) beneath the sedimentary cover. This signature is only interrupted by the Mississippi Embayment at 35°N , 90°W . The Grenville Province in the U.S. is dominated by low amplitude positive velocity perturbations ($\delta V_P \leq 0.4\%$) in contrast to the continental interior. These lithospheric seismic velocity boundaries are well resolved in BBNAP18 (Figures 5.12, 5.13 and 5.14).

In the lower lithosphere ($150 - 250\text{ km}$ depth) of the continental north-east, the fast velocities are

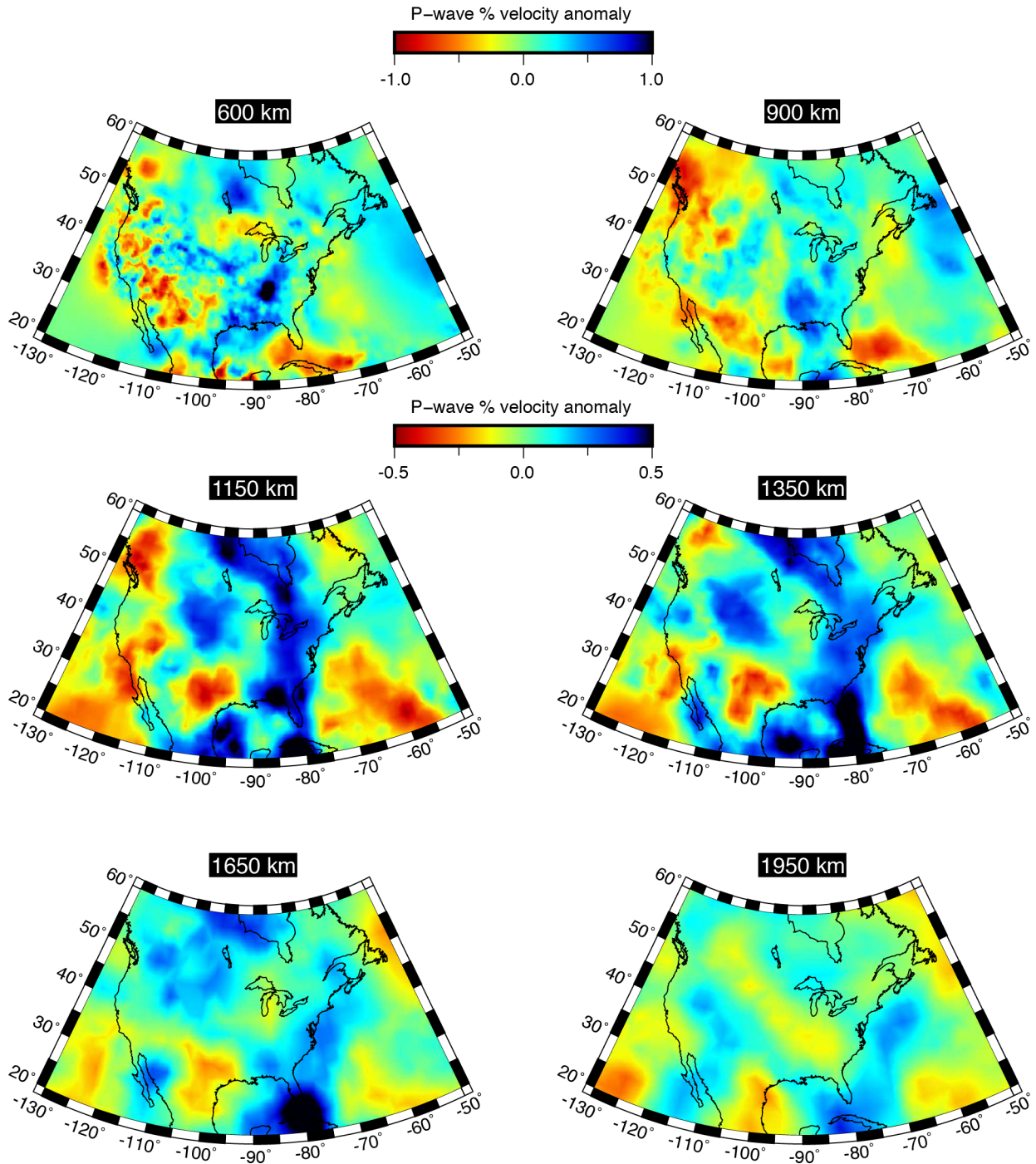


Figure 5.18: Figure to show deep mantle results from the final P-wave velocity model for North America plotted as a percentage deviation from ak135.

confined to the Superior Province, west of the GF (Figures 5.16, 5.17). Low amplitude anomalies are pervasive in the Grenville Province ($-0.3\% \leq \delta V_P \leq 0.3\%$). In the U.S., fast velocities are restricted to $\sim 105 - 95^\circ\text{W}$. The low amplitude anomalies of the Grenville Province also extend westward beyond the GF into the Granite-Rhyolite and northern Mazatzal Provinces. This is also a well resolved observation.

East-west striking cross-sections at $38 - 42^\circ\text{N}$ across the continent also show the fast velocities to extend from the Superior into the Grenville Province in Canada at upper lithospheric depths (Figure 5.17). However the cross-sections to the south, at $32 - 36^\circ\text{N}$ in the U.S., show the abrupt velocity transition coincident with the GF in the upper lithosphere (Figure 5.17).

5.5 Discussion

5.5.1 Comparisons to previous studies of North American mantle velocity structure

The final tomographic inversion model (BBNAP18) presented in Section 5.4 corroborates previous body-wave tomographic images of the U.S. continent (e.g., *Burdick et al.*, 2017; *Schmandt and Lin*, 2014), with the transition from slow to fast velocities in the lithosphere (west to east) occurring at $105 - 100^\circ\text{W}$ (Figures 5.16, 5.17). In western Canada, BBNAP18 does not recover significant lithospheric velocity structure due to lack of station coverage (Figures 1.2 and 5.3). Broad scale tectonic investigations of western Canada have mostly been derived from surface wave studies to date (e.g., *Schaeffer and Lebedev*, 2014; *Yuan et al.*, 2014). In the U.S., BBNAP18 (Figures 5.16 and 5.17) agrees well with the long wavelength variations in the surface wave models (e.g., *Schaeffer and Lebedev*, 2014; *Yuan et al.*, 2014). The model does not recover a low velocity anomaly beneath the Great Lakes as imaged by *Schaeffer and Lebedev* (2014), but this feature is not present in other models either (e.g., *Clouzet et al.*, 2018; *Yuan et al.*, 2011, 2014).

Features of BBNAP18 (Figures 5.16 and 5.17) in the western U.S., consistent with previous work, are the fast velocity anomaly of the Isabella anomaly (e.g., *Jones et al.*, 2004) at 36°N , 120°W , the subducting Juan de Fuca plate that strikes north-south as a fast velocity anomaly (e.g., *Hawley et al.*, 2016) and the reduced velocities (comparing to other cratonic regions worldwide) of the Wyoming craton (*Dave and Li*, 2016; *Schmandt and Lin*, 2014). In the east, the northern and central Appalachian anomalies are well imaged in agreement with *Schmandt and Lin* (2014) and *Burdick et al.* (2017). Broadening of the central Appalachian anomaly in Virginia agrees with results of *Schmandt and Lin* (2014).

In the upper asthenosphere, BBNAP18 (Figures 5.16 and 5.17) does not show a region of suppressed velocities with respect to the global average parallel to the Appalachians. This contrasts with the deep

slab de-watering hypothesis of *van der Lee et al.* (2008) or smaller anomalies associated with edge driven convection (*Savage et al.*, 2017). In the transition zone at 500 – 700 km depth, a fast velocity anomaly is seen at 35°N, 85°W in agreement with previous studies (*Biryol et al.*, 2016; *Schmandt and Lin*, 2014; *Sigloch and Mihalynuk*, 2013). Corroborating the results of *Schmandt and Lin* (2014) and *Sigloch and Mihalynuk* (2013), BBNAP18 does not show a clear connection to the lithosphere unlike the model of *Biryol et al.* (2016). This anomaly has been interpreted as Farallon-related (*Schmandt and Lin*, 2014), vertical sinking slab piles (*Sigloch and Mihalynuk*, 2013) or a lithospheric drip (*Biryol et al.*, 2016). Deeper still, fast velocity anomalies beneath the western U.S., eastern U.S. and Canada are consistent with the observations of remnant subduction systems (*Schmandt and Lin*, 2014; *Sigloch*, 2011) suggesting the model does have resolution in the deep mantle (Figure 5.18).

5.5.2 Comparisons to previous studies of Grenville Orogen lithospheric velocity structure

By incorporating numerous seismic networks from eastern Canada into a continental model for North America, BBNAP18 developed here is the first body wave model capable of resolving seismic heterogeneity along the entire Grenville Province (significantly faster velocities in the upper lithosphere of the northern Grenville than the south - Figures 5.16, 5.17, 5.19). Previous continental body wave tomographic models use data only from the U.S. to the south (e.g., *Burdick et al.*, 2017; *Schmandt and Lin*, 2014) or a low number of permanent stations from Canada (e.g., *Grand*, 1987; *Li et al.*, 2008a; *Sigloch*, 2011). BBNAP18 shows less heterogeneity in the southern Grenville Province than that of *Schmandt and Lin* (2014), possibly due to two years more data, and a much stronger lithospheric seismic velocity boundary at the GF than the model of *Burdick et al.* (2017). At upper lithospheric depths, the fastest seismic velocities in the Archean Superior are adjacent to slightly slower velocities in the Canadian Grenville Province (Figures 5.16), matching regional body wave imaging (*Boyce et al.*, 2016, see Chapter 3). *Biryol et al.* (2016) performed regional body wave imaging of the south eastern U.S. High velocities extend from the Proterozoic core of North America into the Grenville Province in the upper lithosphere, in contrast to BBNAP18, but truncate approximately coincident with the GF in the lower lithosphere (*Biryol et al.*, 2016).

Surface wave studies tend to show laterally smoothed velocity structure (*Lebedev and Van der Hilst*, 2008; *Petrescu et al.*, 2017; *Schaeffer and Lebedev*, 2014) compared to body wave images (*Boyce et al.*, 2016; *Li et al.*, 2008b; *Schmandt and Lin*, 2014). Despite this, the continental surface wave models of

Clouzet et al. (2018); *Schaeffer and Lebedev* (2014); *Yuan et al.* (2014) show elevated velocities from the cratonic core of North America to extend across the GF into the upper lithosphere of the Canadian Grenville in the north but are generally restricted to west of the GF in the south throughout the lithosphere. This observation is not mentioned as a significant feature in these studies perhaps due to lower lateral resolution in the Grenville, but agrees with BBNAP18 and regional surface wave imaging in Canada (*Petrescu et al.*, 2017) and the U.S. (*Pollitz and Mooney*, 2016). These observations are thus consistent with the seismic root of the North American craton extending beneath the Canadian Grenville in the north but not the U.S. Grenville in the south (*Griffin et al.*, 2004) and underthrusting of Archean Superior rocks in the Canadian Grenville (*Ludden and Hynes*, 2000). Beneath the western Superior, no evidence for a lateral shift in the cratonic keel (*Kaban et al.*, 2015) is shown by BBNAP18.

5.5.3 Causes of seismic heterogeneity

Eastern continental North America is located far from active plate boundaries, remaining tectonically stable for ≥ 250 Ma, since the closure of the Iapetus Ocean. Thus partial melt can be eliminated as a potential source of seismic anomalies (e.g., *Sobolev et al.*, 1996). The Great Meteor hotspot ca. 190–110 Ma and the ~ 48 Ma Virginia volcanics are the most recent magmatic episodes but these would not manifest as significant, abrupt, present day thermal anomalies (e.g., *Eaton and Frederiksen*, 2007). Thus seismic anomalies in eastern North America are not likely to be temperature driven (*Boyce et al.*, 2016). Seismic anisotropy is neglected in the isotropic velocity model presented here. This is justified because of good backazimuthal coverage of earthquakes (Figure 5.2) and because both shear-wave splitting (*Darbyshire et al.*, 2015; *Frederiksen et al.*, 2006; *Long et al.*, 2016; *White-Gaynor and Nyblade*, 2017) and anisotropic shear-wave inversions (e.g., *Clouzet et al.*, 2018) show the region can only be weakly anisotropic compared to the more active western side of the continent (e.g., *Martin-Short et al.*, 2015).

High heat flow in the south-eastern, sediment-covered, portion of the continent contrasts low heat flow in the exposed, eroded, shield region to the north (*Artemieva and Mooney*, 2001). This is best explained by variation in crustal radiogenic heat production (e.g., *Jaupart and Mareschal*, 2010). Heat flow variation across tectonic boundaries at the base of the lithosphere in the Canadian shield is considered low (*Mareschal and Jaupart*, 2004), so heat flow cannot explain the short-wavelength velocity variations in the lithosphere (Figures 5.16, 5.17). Abrupt changes in seismic velocity in stable eastern North America are best explained by chemical heterogeneity or variations in iron depletion

laterally and with depth (*Goes and van der Lee, 2002; Hieronymus and Goes, 2010*), in agreement with combined gravity, seismic and mineral physics analysis (e.g., *Kaban et al., 2014*).

Studies of LAB depth (*Artemieva and Mooney, 2001; Bechtel et al., 1990; Fischer et al., 2010; P. and Carolina, 2006; Pasyanos et al., 2014; Priestley and McKenzie, 2013; Rychert and Shearer, 2009*) have not identified a clear step in lithospheric thickness across the Grenville Front that could produce the observed vertical boundaries in seismic velocity in both the north (see Chapter 3) and south (Figures 5.16, 5.17). Because the western Grenville has remained largely tectonically quiescent since $\sim 0.9\text{ Ga}$, over which time abrupt lithospheric thickness variations would not be maintained (*Wang et al., 2014*), a wavespeed boundary within the lithosphere of constant or gradually changing thickness across the GF is the most likely scenario.

5.5.4 Evidence for a north-south divide in the Grenville Orogen

BBNAP18 indicates that in the upper 150 km , there is a significant difference in seismic velocities along strike in the Grenville Orogen (Figures 5.16, 5.17). The difference between the north and south Grenville is shown even more strikingly in Figure 5.19 in which the average velocity of the eastern U.S. ($+0.30\% \delta V_p$) is subtracted before plotting. The northern Grenville exhibits faster velocities than the regional average; the south is slower. This first order observation suggests the Grenville Orogeny was heterogeneous along strike. The following observations also indicate a fundamental difference between the north and south Grenville.

In the crust, receiver function profiles in the northern Grenville show a thickened crust across the GF (*Petrescu et al., 2016*) with respect to the Superior Province. In the south, a transition to thinner crust across the GF is observed (*Abt et al., 2010*). A possible difference in tectonic regime is also indicated by active-source seismic profiling across the GF. In Canada, overthrusting involves only basement rocks (e.g., *Ludden and Hynes, 2000*); Grenville-age crustal material overlies Archean Superior lithosphere for $\geq 300\text{ km}$ southeast of the GF. In the U.S. however, thrust faulting involves post-Grenvillian sedimentary formations and the reworked rocks of the Grenville Orogen (*Santos et al., 2002*).

Proterozoic reconstructions of Laurentia (Figure 2.1) suggest along strike heterogeneity in the basement rocks and amalgamated terranes existed prior to the Grenville collision (*Hoffman, 1988; Whitmeyer and Karlstrom, 2007*). In the north, Archean Superior rocks underthrust much of the Grenville province along with Palaeoproterozoic Mazatzal and Granite-Rhyolite material. In the south, the

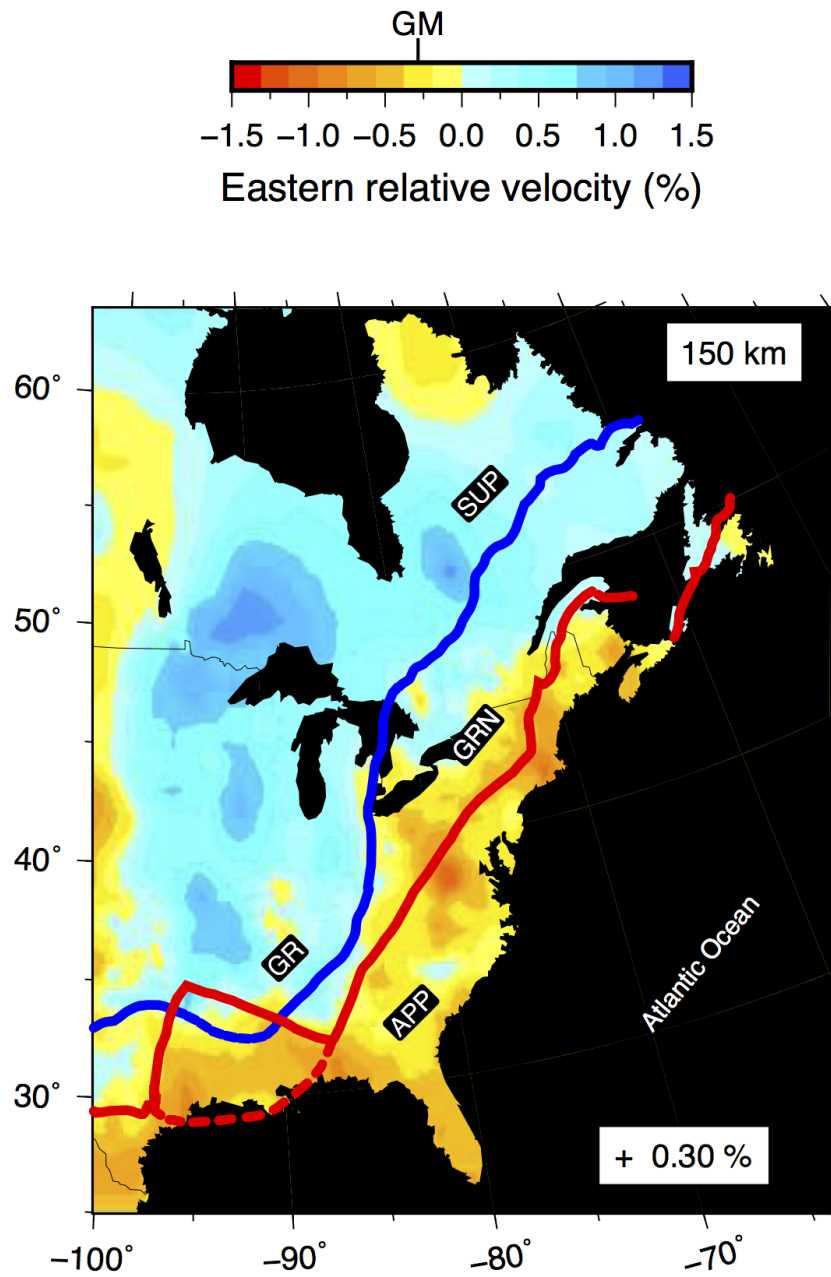


Figure 5.19: **Plot of tomographic model (BBNAP18) from eastern North America.** Velocity is plotted as a percentage deviation from the regional average ($+0.30\% \delta V_p$) from the final inversion within the plotted area ($105 - 50^\circ\text{W}$, $25 - 65^\circ\text{N}$). The colour scale is used to highlight the change from fast to slow velocities, with the position of the global mean indicated (GM). Of particular note is the change from relatively fast velocities in the northern Grenville to slow velocities in the southern Grenville, compared to the regional average.

Grenville Province comprises younger Granite-Rhyolite material and some accreted terranes adjacent to the Appalachian Front (*Hoffman, 1988; Whitmeyer and Karlstrom, 2007*). Plate reconstructions

(e.g., *Hoffman*, 1988; *Hoffman*, 1991) also tentatively suggest that up to four incoming cratonic blocks interacted with present day eastern Laurentia during Rodinia formation.

Figure 2.1 shows that basement rocks outcrop in the Canadian Grenville, but are overlain by sediments south of the Adirondack mountains (e.g., *Laske and Masters*, 1997). Further, the distribution of A-type granitoid suites, emplaced concomitant with the Grenville, show a pervasive network of intrusions that likely subcrop beneath the Appalachians (*McLelland et al.*, 2013; *Whitmeyer and Karlstrom*, 2007); these are far less extensive in the Canadian Grenville terranes (e.g., *Corrigan and Hanmer*, 1997, see Figure 2.1). Such rocks are characteristic of high temperatures at shallow depths causing remelting of a crustal source (e.g., *Whalen et al.*, 1987). Finally, the macro-scale geometry shows the northern Grenville province of Canada to be much wider than that in the south (Figure 2.1). The sutures between the colliding continents of Laurentia and those to the east (including Amazonia *Hoffman*, 1988) have been lost to Appalachian orogenic overprinting (*Thomas*, 2006), and therefore the original width of the Grenville Orogenic belt is difficult to ascertain. However, the north-south preserved widths in the Appalachians is in anti correlation with the Grenville orogen, implications of which are discussed later in this chapter.

5.5.5 Modification derived velocity structure

Earlier relative arrival-time tomography from southeast Canada (*Boyce et al.*, 2016, Chapter 3) suggests that the age progressive lithospheric subdivisions identified by increasing seismic velocities are best explained by modification derived processes rather than by a secular change in plate forming processes. Additionally the formation of a vertical boundary in lithospheric seismic velocities is more easily reconciled with post formation processes (*Boyce et al.*, 2016, Chapter 3). Unlike the north, there is no evidence for overthrusting of basement rocks (*Santos et al.*, 2002) beneath the GF in the south, so the same lithologies are expected either side of the GF (i.e. Granite-Rhyolite rocks *Whitmeyer and Karlstrom*, 2007). A vertical boundary in seismic velocity (Figure 5.17) at the GF cross-cutting the continuous Granite-Rhyolite terrane (*Whitmeyer and Karlstrom*, 2007) implies that the Grenville imparted a preserved deformation signature on the lithosphere, thus processes during collision modified the lithosphere in the south of the Grenville Orogen as well.

Subduction, hotspot tectonism, or gravitational destabilisation can all cause modification of continental lithosphere that would manifest as observable vertical boundaries in seismic velocity (e.g.,

Frederiksen et al., 2013; *Kusky et al.*, 2007; *Menzies et al.*, 2007; *Nolet and Zielhuis*, 1994; *Pedersen et al.*, 2013; *Villemaire et al.*, 2012; *Youssof et al.*, 2015). These mechanisms have been invoked by numerous authors to describe localised modification of North America/Laurentia over the past 1.3 Ga as well (e.g., *Bao et al.*, 2014; *Biryol et al.*, 2016; *Boyce et al.*, 2016; *Dave and Li*, 2016; *Jones et al.*, 2004; *Levander et al.*, 2011; *van der Lee et al.*, 2008; *Zandt et al.*, 2004).

Deep slab dewatering processes have been used to explain large low velocity zones in the mid-mantle beneath the east coast of the U.S. (*van der Lee et al.*, 2008), but lithospheric effects of this deep dewatering are likely to be diffuse. As such, it is difficult to reconcile the abrupt boundary in velocity coincident with the GF in the U.S. and less dramatic decrease in velocity across the boundary in Canada, beneath which no slab remnants exist (*Sigloch*, 2011, also see Figure 5.18). Removal of younger lithosphere by edge driven convection at the margin of a craton (*Bao et al.*, 2014) can produce an abrupt boundary in seismic velocity. This mechanism does require a point of initiation, either a step in lithospheric thickness or an exploitable lithological boundary across which susceptibility to modification varies. However, as Superior material was thought to underlie the Grenville Orogen at the time of collision (*Ludden and Hynes*, 2000) in the north, and no exploitable lithological variation across the GF existed in the south (e.g., *Whitmeyer and Karlstrom*, 2007), neither a lithospheric proto-step or exploitable lithological variation is likely to be present (e.g., *Boyce et al.*, 2016).

Lithospheric modification by subduction and hotspot interaction has also been proposed (*Boyce et al.*, 2016; *Dave and Li*, 2016). Subduction can lead to abrupt velocity boundaries in the lithosphere (*Bao et al.*, 2014; *Boyce et al.*, 2016), and the possibility that subduction processes varied along-strike cannot be precluded. Subduction is not best-placed to explain the observed lithospheric heterogeneity since this would require speculation on incoming plates which are poorly constrained in the Grenville Orogeny (*Hoffman*, 1991) and is less easily reconciled with the formation of A-type granitoid intrusives (*Corrigan and Hanmer*, 1997; *McLelland et al.*, 2013). Hotspot activity is also less likely to produce the broad, abruptly bounded, features observed (*Boyce et al.*, 2016). It appears unlikely that present day processes (*Biryol et al.*, 2016), away from an active continental margin, are capable of affecting “stable” Proterozoic lithosphere (*Dave and Li*, 2016; *Poudjom-Djomani et al.*, 2001). Seismic velocity anomalies in BBNAP18 and magnetotelluric observations (*Murphy and Egbert*, 2017) are inconsistent with present-day modification of the mantle lithosphere (*Biryol et al.*, 2016; *van der Lee et al.*, 2008).

5.5.6 A preferred interpretation for Grenville orogenic development

The remaining candidate to explain the observed seismic velocities within the Grenville Orogen is gravitational instability leading to delamination of the lithosphere (e.g., *Jones et al.*, 2004; *Levander et al.*, 2011; *Zandt et al.*, 2004). I discuss this possibility in the following section. A critical observation that has to be reconciled in any model for Grenville development is the distribution of A-type granitoid intrusions that are pervasive beneath the southern Grenville and Appalachians (*McLelland et al.*, 2013; *Whitmeyer and Karlstrom*, 2007), but are far less extensive in the Canadian Grenville terranes (e.g., *Corrigan and Hanmer*, 1997, see Figure 2.1).

During orogenesis, the crust and mantle lithosphere of both plates thicken (Figure 5.20), due to horizontal shortening (e.g., *Houseman et al.*, 1981; *Nelson*, 1992). The over thickened lithospheric column becomes gravitationally unstable and the lower part of the lithosphere delaminates (e.g., *Houseman et al.*, 1981; *Nelson*, 1992) and sinks into the deep mantle (e.g., *Harig et al.*, 2010). This instability can be enhanced by water released from the down going subducting oceanic plate prior to collision (*Kusky et al.*, 2007; *Leech*, 2001). The lithospheric removal may lead to further surface uplift during this phase (*England*, 1993). Thickened crust is exposed to an increased geotherm leading to sparse emplacement of A-type granitoid intrusions (e.g., *England*, 1993; *Kay and Kay*, 1993; *Whalen et al.*, 1987). Unable to maintain elevation isostatically after orogenic forces subside, the orogen next undergoes collapse (e.g., *Dewey*, 1988). Both the crust and mantle lithosphere are thinned to less than initial thicknesses (e.g., *Dewey*, 1988; *Nelson*, 1992, see Figure 5.20). Thermal subsidence promotes sag basin formation (e.g., *Nelson*, 1992) and asthenosphere accretes to the base of the lithosphere via cooling (e.g., *Zheng et al.*, 2007, see Figure 5.20). Due to erosion of topography through time and loss of the delaminated lithosphere to the deep mantle (*Harig et al.*, 2010), the enduring signatures of delamination within an ancient orogen are (i) a thinned crust and associated sag basin formation, (ii) emplacement of A-type granitoid intrusions and (iii) preserved lithospheric mantle seismic velocity variation (*Dewey*, 1988; *England*, 1993; *Houseman et al.*, 1981; *Kay and Kay*, 1993; *Nelson*, 1992), each of which is observed.

Consider initially the northern Grenville in Canada at $\geq 1.3\text{ Ga}$, thought to be underlain by depleted (*Jordan*, 1988), buoyant and viscous (*Sleep*, 2003), Archean Superior lithosphere at time of collision (*Ludden and Hynes*, 2000; *Whitmeyer and Karlstrom*, 2007). Under continental collision and subsequent thickening of the crust and mantle lithosphere, buoyant Archean lithosphere could have become

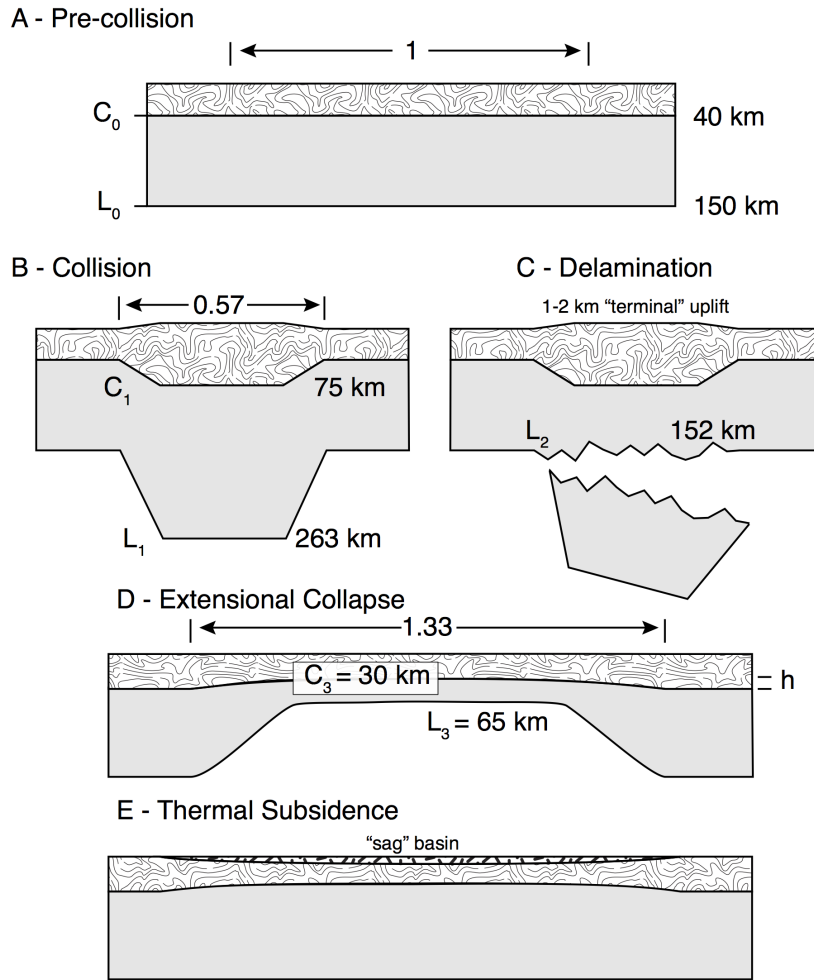


Figure 5.20: **Schematic diagram of collisional orogeny re-drawn after Nelson (1992)** that is followed by delamination, extensional collapse and thermal subsidence. Initially, the example crust (patterned) and lithosphere (grey) ($C_0 = 40$ km and $L_0 = 150$ km thickness) are shortened to produce a thicker crust and lithosphere ($C_1 = 75$ km and $L_1 = 263$ km thickness). Subsequently the thick lithospheric root delaminates and sinks into the underlying asthenosphere. Compressional forces subside and the orogen is allowed to collapse. Thermal subsidence results in a thinner crust and lithosphere ($C_3 = 30$ km and $L_3 = 65$ km) within the orogen, and younger asthenosphere is accreted by cooling to the base of the lithosphere.

partially unstable leading to delamination (e.g., Griffin *et al.*, 2003; Houseman *et al.*, 1981). Incomplete/partial delamination would be expected to result in sparsely-developed A-type granitoids, and an only partially modified (i.e. still seismically fast) mantle lithosphere. These observations satisfy the geological record (Corrigan and Hanmer, 1997; Whitmeyer and Karlstrom, 2007) and my seismic tomographic observations (Figures 5.16, 5.17, 5.19) for the Canadian Grenville province. If this old, strong lithosphere was more resistant to orogenic collapse than younger terranes (e.g., Leech, 2001), then a thickened crust relative to the Superior Province would have been preserved below the Grenville

Orogen, as is observed (*Petrescu et al.*, 2016). The Canadian Grenville, is presently supported by the North American cratonic root (*Griffin et al.*, 2004); it is thus exposed and heavily eroded at the surface but resistant to basin formation.

To the south in the U.S., the Grenville Province comprised young (~ 200 Ma), less buoyant (*Hawkesworth et al.*, 1990; *Poudjom-Djomani et al.*, 2001) lithosphere of the Granite-Rhyolite terrane (*Whitmeyer and Karlstrom*, 2007) at the time of orogeny (Figure 2.1). During continental collision, tectonic shortening of weaker lithosphere would thus have promoted the development of a wide orogenic plateau (e.g., *Li et al.*, 2016). The greater shortening (and thus increased crustal and lithospheric mantle thickening) of this less buoyant lithosphere, would have rendered it susceptible to delamination (*Hawkesworth et al.*, 1990; *Houseman et al.*, 1981; *Poudjom-Djomani et al.*, 2001). Prior to Grenvillian continental collision, water released from a subducting slab (e.g., *Boyce et al.*, 2016, see Chapter 3) would have metasomatised this younger lithosphere to a greater extent, leading to increased gravitational instability (*Kusky et al.*, 2007; *Leech*, 2001).

Pervasive delamination below the southern Grenville terranes offers an elegant explanation for the large volumes of A-type granitoids, observed there (*McLelland et al.*, 2013; *Whitmeyer and Karlstrom*, 2007, Figure 2.1). The vertical boundary in seismic velocity seen beneath the GF in the U.S. (Figure 5.17) can be explained by hot, less depleted asthenospheric material rising to shallower depths in place of the unstable lithosphere (e.g., *Nelson*, 1992; *Zheng et al.*, 2007, see Figure 5.20). Thus faster velocities of the unmodified Proterozoic Granite-Rhyolite province can be maintained west of the GF, whilst to the east, the younger lithosphere in the Grenville Province manifests as average velocity (Figure 5.16, 5.19). Whilst it may be possible to produce post collisional, A-type granites through preservation of very large crustal thicknesses over protracted periods (e.g., *Jackson et al.*, 2008), this would not explain the development of an abrupt lithospheric velocity boundary. Because the present crustal thickness in the southern Grenville is reduced with respect to the Granite-Rhyolite to the west (*Abt et al.*, 2010), the Grenville Orogen in the south may have undergone widespread orogenic collapse (*Dewey*, 1988; *Nelson*, 1992). The potential for sedimentary sag basin formation in the south by thermal subsidence (Figure 5.20 *Nelson*, 1992) in this region is consistent with the present day sedimentary cover of the Grenville Province in the U.S. and the Middle Run formation (*Baranoski et al.*, 2009; *Santos et al.*, 2002), a depositional environment active between 1.05 – 0.7 Ga.

Vertical tectonic processes may also offer a suitable explanation for the lack of strong, orogen-parallel lithospheric deformation fabrics preserved in the Grenville (*Long et al.*, 2016; *White-Gaynor and*

Nyblade, 2017) when compared to other orogens (*Bostock and Cassidy*, 1995). I speculate that my study provides seismic evidence for the oldest record of delamination of subcontinental lithospheric mantle in a Proterozoic large, hot, orogen, which is supported by emerging geological evidence (e.g., *Rivers*, 2009, 2012, 2015).

5.5.7 Speculations on the extent of Grenville deformation

I now review the N-S divide in the Grenville orogeny in light of recent geodynamic modelling studies. Delamination in a down-going-plate is favoured in a scenario of slow convergence with a relatively strong overriding plate (*Li et al.*, 2016), whilst delamination of the overriding plate is favoured when it is rheologically weak and convergence rate is faster. Wide, flat orogenic plateaux are predicted where delamination is widespread in an orogen, however where delamination is less pervasive mountain belts are narrow (*Li et al.*, 2016). In the Himalaya, a wide plateau is favoured in the east where the overriding Asian crust is weaker (*Chen et al.*, 2017) whereas a stiff Asian crust in the west promotes a narrow orogeny to accommodate shortening and leads to reversed polarity underthrusting.

The preserved northern Grenville province of Canada is wider than that to the south (Figure 2.1). The sutures between the colliding continents of Laurentia and those to the east (including Amazonia, *Hoffman*, 1988) have been reworked during Appalachian orogenic overprinting (*Thomas*, 2006), and thus the original width of the Grenville Orogenic belt is difficult to ascertain. However, the present width of the Appalachian terranes may indicate lithospheric strength heterogeneity in the Grenville by inference (*Chen et al.*, 2017; *Li et al.*, 2016).

In the northern Grenville, underlain by stiff Superior material (*Griffin et al.*, 2004; *Ludden and Hynes*, 2000), less delamination is consistent with a strong overriding-plate (*Li et al.*, 2016), and a relatively narrow preserved orogeny leading to reversed polarity underthrusting, as is observed (*Hynes and Rivers*, 2010). In the south, the hypothesis of pervasive delamination suggests the Granite-Rhyolite province was therefore weak (*Chen et al.*, 2017; *Li et al.*, 2016) at the time of collision. Speculatively, increased Appalachian overprinting in the south (*Thomas*, 2006), evidenced by a narrow preserved Grenville orogen, implies a weaker preserved Grenville lithosphere, consistent with this view. This hypothesis suggests that the southern Grenville Orogen displays modern-day lithospheric rheological characteristics and resulting structural features - large plateau and subsequent collapse. Therefore a secular change in rheological properties of the lithosphere may have occurred before the beginning of

the Grenville Orogen (1.3 Ga) prior to which, large orogenic plateaux and orogenic collapse were not seen in large, hot, orogens (Leech, 2001; Rivers, 2012).

5.5.8 Speculations on the timescales of cratonisation

The North American craton comprises the Canadian shield region to the north and the platformal regions of the U.S. to the south, atop which significant sedimentary thickness has developed (e.g., Laske and Masters, 1997, see Figure 2.1). A buoyant lithospheric keel is thought to extend from the Archean Superior core to beneath the northern Canadian Grenville (Griffin *et al.*, 2004). This is consistent with my tomographic results (Figures 5.16 and 5.17), the delamination hypothesis for Grenville development and widespread underthrusting of reworked Archean material (Ludden and Hynes, 2000) that may not have been limited to the crust (Boyce *et al.*, 2016). Conceivably, juvenile Proterozoic lithosphere ($\leq 0.25\text{ Gyr}$) of the Granite-Rhyolite province was not buoyant and refractory (Hawkesworth *et al.*, 1990) at the time of the Grenville collision, thus rendering it more susceptible to replacement by delamination. The remaining Granite-Rhyolite, Mazatzal and Yavapai provinces (west of the GF) did not encounter any significant tectonic processes throughout the Proterozoic so are likely to have thermally stabilised over time (Hawkesworth *et al.*, 1990). This indicates that Proterozoic lithosphere may only be long-lived when it is isolated from tectonic processes or is formed from reworked (and underthrusting) Archean constituents. I speculate that there may have been only two periods of lithospheric generation as a result, an Archean highly depleted regime and a modern fertile regime (Griffin *et al.*, 2003). Variations between these regimes, producing an apparent continuous secular change in depletion levels (Poudjom-Djomani *et al.*, 2001), could therefore be a modification-derived affect (Boyce *et al.*, 2016; Griffin *et al.*, 2003), unrelated to formation. Juvenile lithosphere formed in the Proterozoic could thus be more similar to present-day continental lithosphere (Hawkesworth *et al.*, 1990; Jordan, 1988) than traditionally thought. Thus, under modern plate recycling conditions it may not be the case that Proterozoic lithosphere would survive significantly longer than Phanerozoic continental lithosphere.

Eaton and Perry (2013) showed that from a hot or cold starting model, Proterozoic regions of cratons obtain their long-lived buoyancy after $\sim 1\text{ Ga}$. By calculating the time difference between the final stage of Mazatzal accretion, a stable Proterozoic province, and the start of the Grenville Orogeny, or potential lithospheric modification, the results of this chapter place a lower bound on this process, implying at least $\sim 0.3\text{ Ga}$ is required to prevent modification. This equates to a loss of $\geq 75\%$ of

initial thermal state (*Eaton and Perry, 2013*) consistent with the approximate thermal time constant of the lithosphere (~ 250 Ma; *McKenzie and Priestley, 2016*). The location of the Illinois and Michigan cratonic basins west of the GF, a region where thick crust has been maintained for at least this time (*McKenzie and Priestley, 2016*), is consistent with this view.

5.6 Conclusion

This chapter has presented the highest resolution P-wave model of North American mantle structure to-date using, for the first time, data from temporary seismograph networks from the Precambrian core of Laurentia in Canada. BBNAP18 shows, along strike variation in the upper lithosphere (≤ 150 km depth) of the Grenville Orogen, with fast velocities in Canada to the north and average velocities in the U.S. to the south. When discussed in light of other geological and geophysical observations, I suggest these wavespeed variations are best explained by along strike differences in the extent of lithospheric delamination during the Grenville Orogen. These results may therefore present the first mantle seismic evidence for delamination of subcontinental lithospheric mantle in a Proterozoic large, hot, orogen.

Chapter 6

Concluding remarks

I have presented the highest resolution P- and S-wave tomographic models of upper-mantle wavespeeds beneath southeast Canada. Both models show increasing seismic wavespeed across the three tectonic domains: slow Phanerozoic Appalachian terranes, an average wavespeed Proterozoic Grenville Province and a fast wavespeed Archean Superior Province in central Québec. Consistent with predictions of two-stage formation of lithospheric keels (chemically depleted Archean core and thermally accreted, less depleted Proterozoic boundary layer), wavespeeds decrease dramatically across the Grenville Front. This suggests keel growth may not have been limited to the Archean.

Crustal studies suggest that Archean Superior material underlies the Grenville province for up to ~ 300 km southeast of the Grenville Front however. To reconcile these results with a vertical wavespeed boundary seen in tomographic imaging, I propose that subduction beneath the margin of Laurentia modified the edge of the continent during the Grenville orogeny. Metasomatic alteration of the Laurentian margin in the Proterozoic prior to keel stabilisation would result in the imaged vertical wavespeed boundary beneath the Grenville Front. This work confirms that cratons, particularly their margins may not be stagnant features throughout geological time. Furthermore, I suggest that post-formation metasomatic enrichment of Proterozoic domains, rather than inherently less depleted compositions at the time of formation, may better explain variable seismic wavespeeds and thus chemical depletion levels across Archean-Proterozoic boundaries worldwide.

To combine the P-wave relative arrival-time dataset with global absolute arrival-time datasets, I developed the Absolute Arrival-time Recovery Method (AARM). A relative arrival-time technique is first used to align waveforms. These timeshifts are applied to unfiltered data which are stacked using

phaseweighting. A weighting calculation results in a higher SNR stack on which the arrival-time is picked, allowing absolute arrival-times to be retrieved. This method also produces a relative estimate of uncertainty in each pick.

AARM is tested initially using a database of $\sim 10,000$ picks from networks in southeast Canada. $\sim 83\%$ of AARM picks agree to within $\pm 0.5\text{ s}$ of ISC database picks for equivalent earthquake-station pairs. Reliable picks are derived from aligned waveforms of average $\text{SNR} \geq 2$. Testing using synthetic datasets shows AARM produces arrival-time picks to $\geq 0.25\text{ s}$ accuracy, comparable to the ISC databases. Data from short-duration temporary networks, when processed using AARM, can therefore be used to fill gaps in spatial coverage in global wavespeed models.

By utilising AARM, to combine data from southeast Canada with absolute arrival-time datasets from the U.S. to the south, I have imaged the entire eastern margin of the North American continent, for the first time using body-waves. This provides the opportunity to address the along-strike heterogeneity in the rock record, observed in the $\sim 1.1\text{ Ga}$ Grenville Orogen. The model BBNAP18 shows along-strike variation in the upper lithosphere ($\leq 150\text{ km}$ depth) of the Grenville Orogen, with fast velocities in Canada to the north and average velocities in the U.S. to the south. The abrupt vertical boundaries in seismic velocity beneath the Grenville Front are best explained by chemical heterogeneity within the lithosphere.

When interpreted in light of available geological evidence, these seismic images suggest that Canadian Grenville Province in the north retains faster wavespeeds having undergone sparse lithospheric delamination and resisted orogenic collapse due to inherent strength. By contrast, in the Grenville orogen in the U.S. to the south, pervasive lithospheric delamination and orogenic collapse provides the most suitable explanation for the observed lowered seismic velocities in the upper lithosphere. These results may thus present the first seismic evidence for delamination of lithospheric mantle in a large, hot, orogen in the Proterozoic. This hypothesis can be reconciled with metasomatic alteration of preserved Archean lithosphere beneath the Grenville in Canada indicated by the P- and S-wave relative arrival-time study earlier in this thesis. Cratons, particularly their margins, are susceptible to modification.

Further work

Motivation for further work on North America, following this thesis can be divided in two: data-driven and tectonics-driven. After publication of the manuscript (*Boyce et al.*, 2016), no further relative arrival-time inversions were performed. However sections 5.2.2 and 5.2.2 detail two extra datasets that were processed using multichannel cross-correlation (*VanDecar and Crosson*, 1990) prior to conversion to absolute time using AARM (*Boyce et al.*, 2017). The western Superior (Figure 5.4) and QM-III (Figure 5.6) could be added to the relative arrival-time inversion of presented in Chapter 3 (*Boyce et al.*, 2016) as these data overlap spatially and temporally. These datasets would almost certainly strengthen the arguments for modification of the edge of the North American craton from visual inspection of residual distributions (Figure 5.4, 5.6).

To more thoroughly test AARM presented in Chapter 4 (*Boyce et al.*, 2017), picking of the entire S-wave dataset presented in Chapter 3 (*Boyce et al.*, 2016) would highlight error variation in an “observed” S-wave dataset (see Figure 4.8 for P-wave example). This would likely corroborate the findings presented in section 4.4 that indicate reliable results albeit with slightly larger errors than an equivalent P-wave dataset. Such an S-wave dataset may then be incorporated in a continental S-wave model for North America.

Furthermore the opportunity to incorporate data from other available datasets (e.g., *Liddell et al.*, 2018) into the absolute arrival-time tomographic model of North America (BBNAP18) would vastly improve data coverage in the far north and northwest of the continent (Figure 6.1), including the remains of the THO. Further data could also be added from the ongoing USArray. This dataset of new picks when compiled would be suitable for submission to the ISC (*Di Giacomo et al.*, 2014; *International Seismological Centre*, 2016) for further distribution to other authors.

The North American continental model, presented in Chapter 5, may also be used to investigate the formation of the Alaskan peninsular. Available geological evidence fails to explain how Alaska formed ≥ 50 Ma, however, subducted slabs which are still present beneath the continent (Figure 6.2) can provide insight into how Phanerozoic subduction shaped the evolution of western North America (e.g. *Sigloch and Mihalynuk*, 2013). Critically lacking is a high-resolution image of the northern Alaskan slab pile (Figure 6.2) and a lithospheric image capable of providing constraints on subduction processes at trenches lost to subduction. By incorporating newly available data (Figure 6.1) further insights into the Farallon-Mezcalera-Angayucham subduction system can be gleaned from the lithosphere and

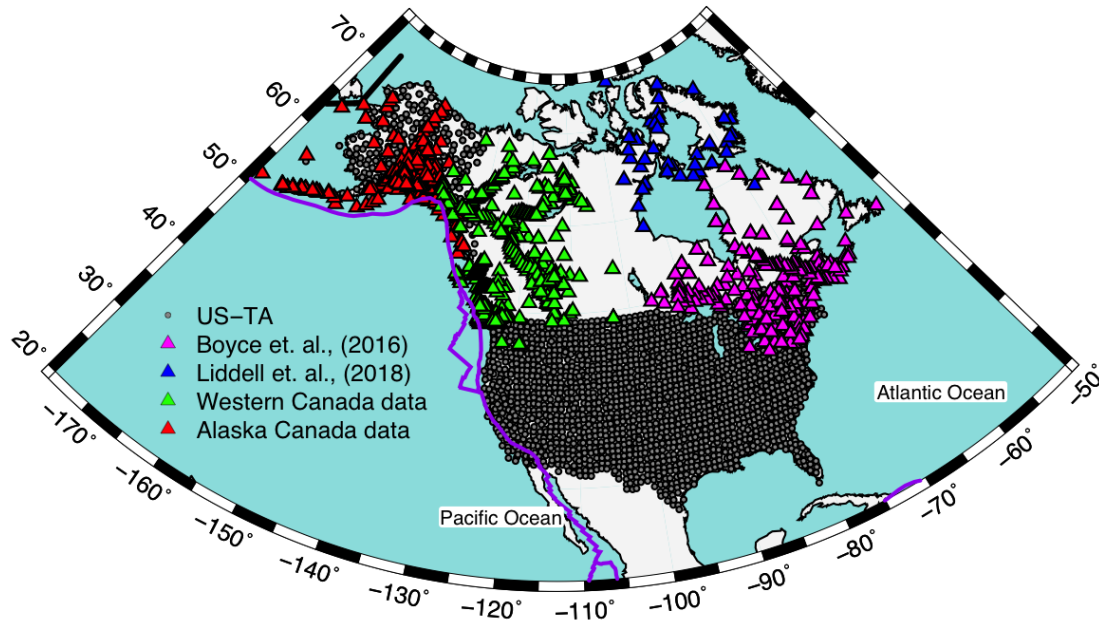


Figure 6.1: **Map of available datasets** that may be included into the North American continental tomographic inversion. Stations already incorporated (magenta triangles) with USArray stations (grey circles) up to May 2016 are shown. Stations to be added include those from Alaskan networks (red triangles), western Canadian networks (green triangles), and northern Canada (blue triangles e.g., *Liddell et al.*, 2018), available through the IRIS Data Management Center and Canadian National Data Center. Purple lines indicate active plate boundaries.

deep mantle (Figure 6.2). This would likely include incorporation of a variety of deeper mantle/core phases taking account of the appropriate sensitivity kernels (*Li et al.*, 2008a).

Numerous studies have converted seismic P- and S- wave velocities to the thermochemical parameters of the mantle (e.g. *Goes and van der Lee*, 2002; *Goes et al.*, 2000). Using the North American continental model for this must be done with caution due to the large difference in spatial sampling between the USArray and surrounding regions and the inherent non-uniqueness of amplitudes in tomographic inversions.

The analysis undertaken in Chapter 3 followed by calculation of absolute arrival-times (Chapter 4) can also be used to investigate Cenozoic volcanism and hotspot tectonism in Africa. Using 64,593 absolute arrival-time residuals (Figure 6.3) derived from all publicly available temporary networks from 1990 – 2017, I have developed a new tomographic image for Africa (Figure 6.4). Identical inversion procedure to Chapter 5 is followed.

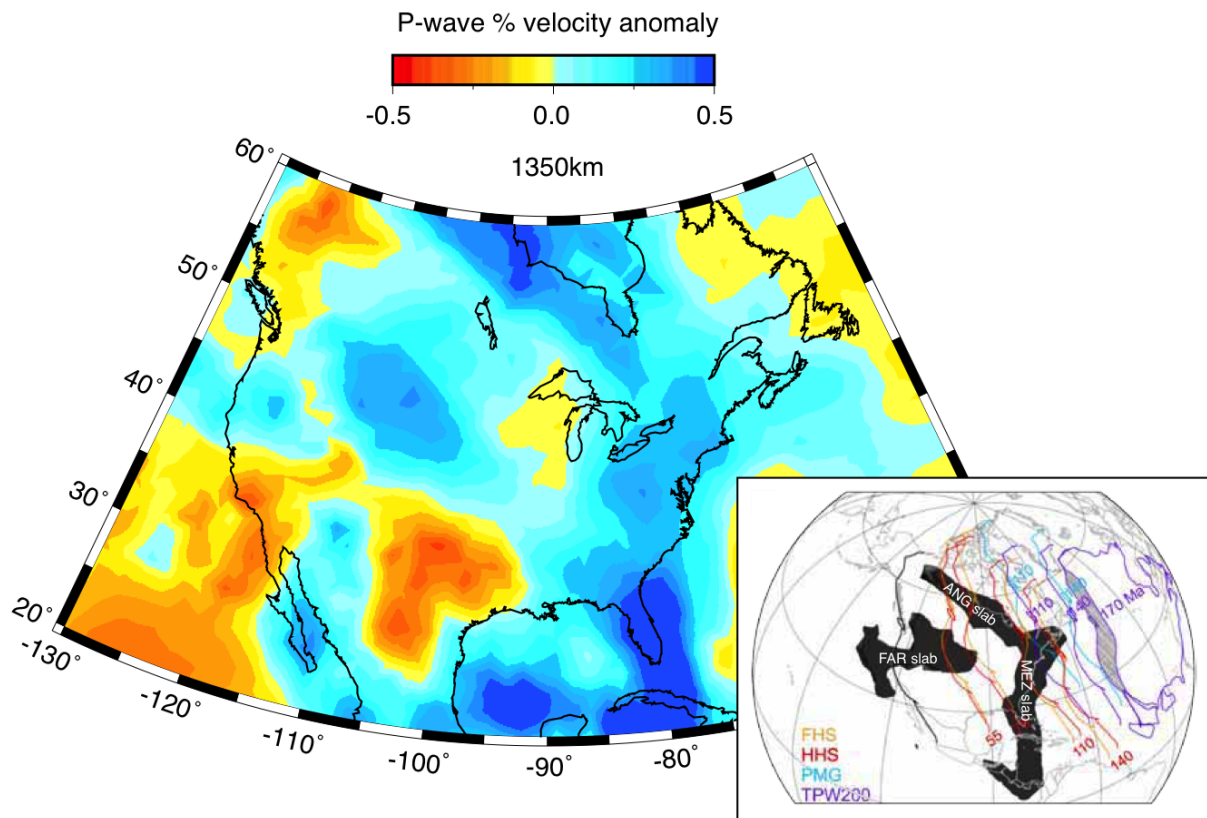


Figure 6.2: **Tomographic image of Mezcalera, Angayucham, Farallon Subduction system** at 1350 km in the North American continental inversion presented in Chapter 5. The colour scale is selected to highlight transition from fast (blue) to slow (red) velocity anomalies. Inset, (adapted from *Sigloch and Mihalynuk, 2013*) shows deep mantle “slab piles” resulting from ≥ 200 Ma of subduction history as the North American continent moves westwards. Mezcalera (MEZ slab), Angayucham (ANG slab) and Farallon (FAR slab) are labelled.

Initial results show the African super plume to extend from the lower mantle beneath South-Africa, crosses the transition zone beneath the East African Rift terminating beneath Ethiopia. Here, some of the slowest wavespeeds in the model are seen. The Archean cratonic cores of southern Africa separated by the Proterozoic Limpopo belt is also well imaged.

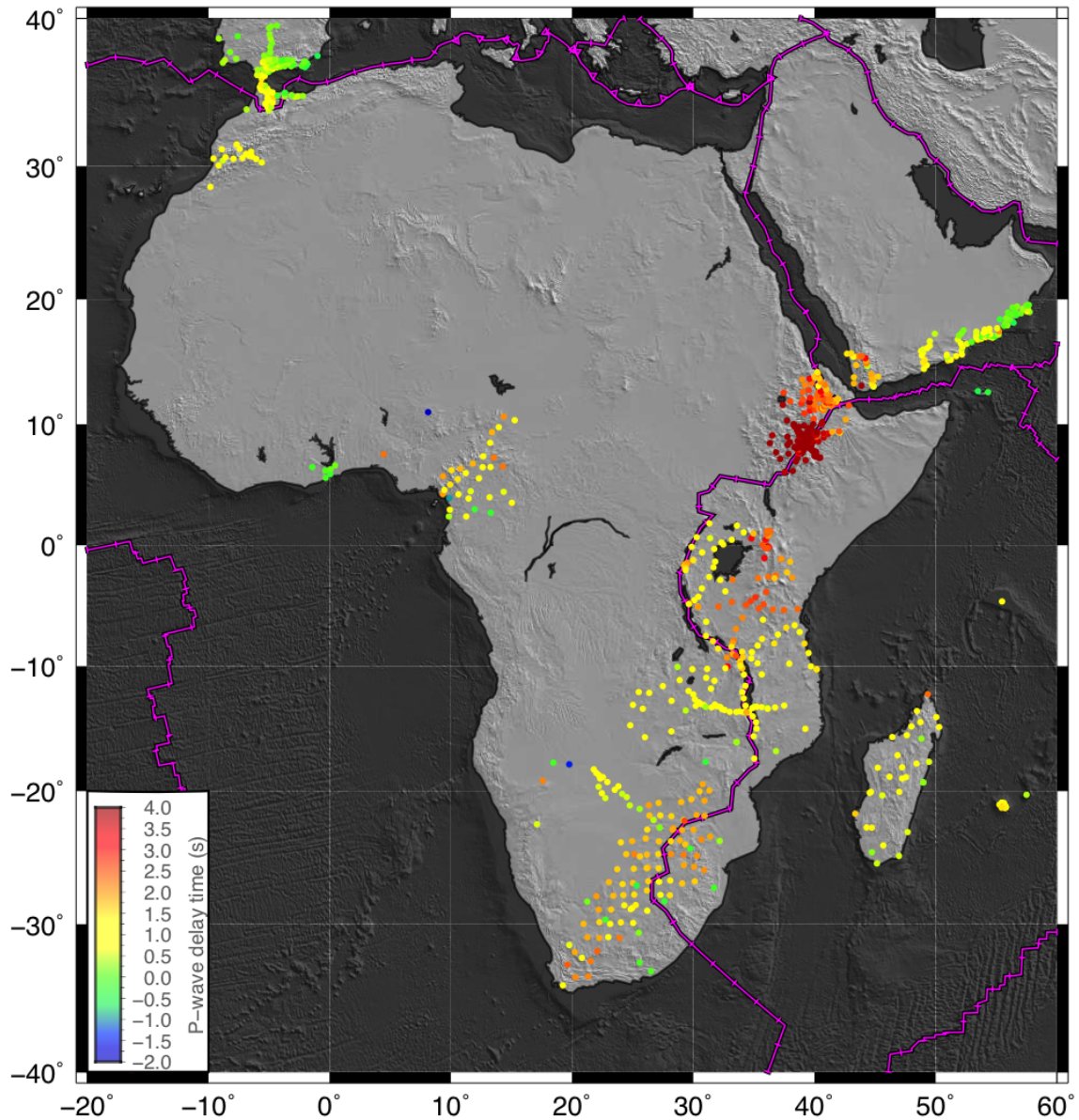


Figure 6.3: **Absolute arrival-time residuals** calculated for all temporary networks in Africa since 1990. Data were processed using identical relative arrival-time analysis as in Chapter 3 and processed to obtain absolute arrival-times using AARM (Chapter 4). Fast residuals are blue, slow are red, plate boundaries are marked in magenta.

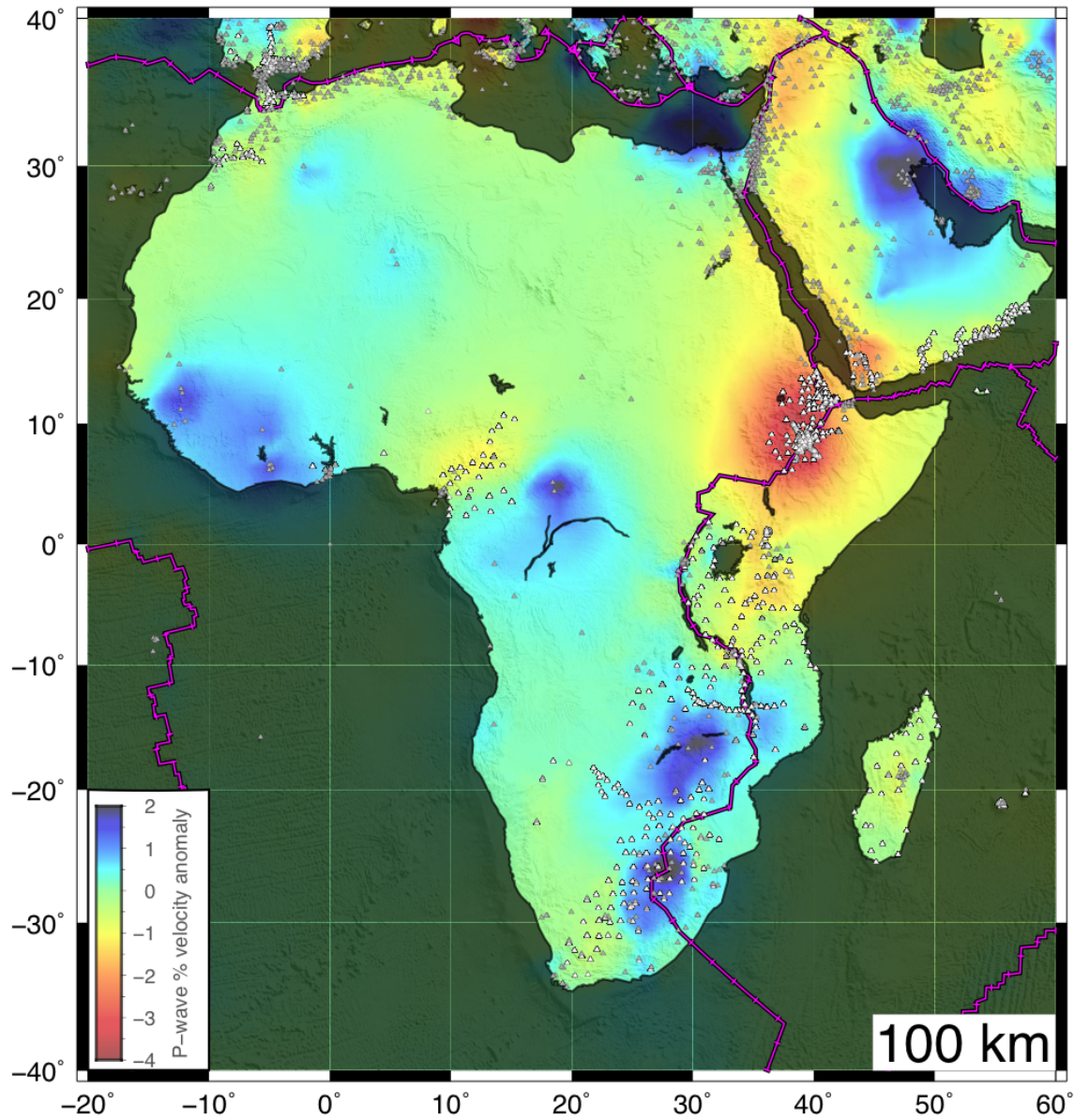


Figure 6.4: **Tomographic image of Africa** at 100 *km* depth. Inversion methodology is identical to that presented in Chapter 5. Plate boundaries are plotted in magenta, station locations of both the global dataset and new additional networks are shown as grey triangles.

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Appendix A

Fieldwork

A.1 Seismic data collection

A.1.1 Canadian Maritimes

During the period May 2015 - August 2015 I took part in the servicing and pull in of 10 Seis-Uk Instruments deployed in the Maritimes region of Eastern Canada (e.g., *Gilligan et al.*, 2016a,b). The stations were situated in gardens of local residents, a blueberry farm, a Geological Survey of Canada rock store and a geological museum throughout Nova Scotia and New Brunswick (Figure A.1). These data were used to improve coverage in the eastern portions of tomographic images (e.g., *Boyce et al.*, 2016). Unfortunately however, proximity to nearby noise sources (the ocean) limited the use of this dataset from broadband arrival-time picking.

A.1.2 The QM-III network

The QM-III network (QuébecMaine across three sutures) was formed from a collaboration between researchers at the Université du Québec à Montréal, Lamont Doherty Earth Observatory and Rutgers University. In the summer of 2015, I joined a field team to pull in data and stations deployed from the US-Canadian border into the centre of Québec (Figure A.1). The data became publicly available through IRIS (Incorporated Research Institutions for Seismology) in Autumn 2017 and has thus been incorporated in tomographic inversions later in this thesis (Chapter 5). Arrival-times are calculated

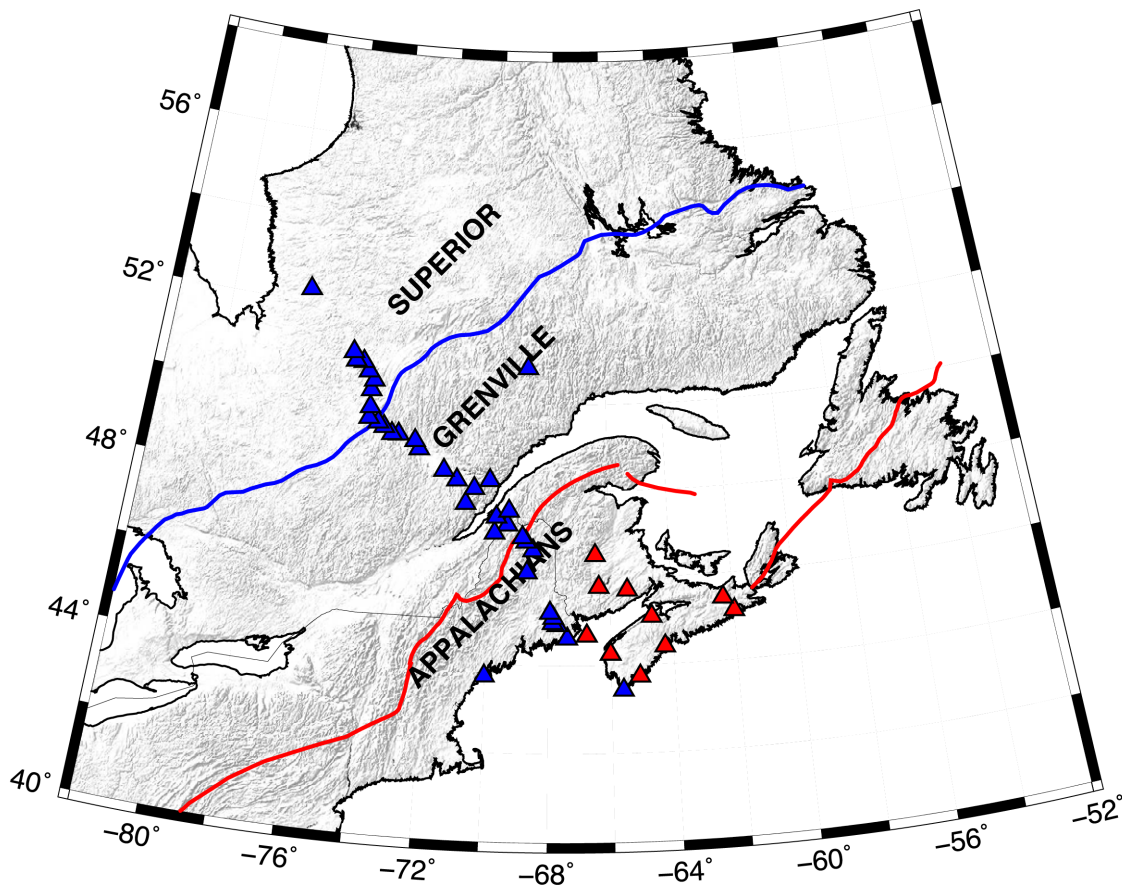


Figure A.1: **Figure to show location of field sites** for networks serviced during field campaigns of 2015. Blue Triangles refer to the QM-III network, red are the Imperial College Maritimes Network (e.g., *Gilligan et al.*, 2016a,b).

using the methods of *VanDecar and Crosson* (1990) and subsequently that developed in Chapter 4. See Chapter 5 for more detail on this dataset.

Appendix B

Data processing - further details

B.0.1 Instrument response

Broadband (BB) and short period (SP) seismometers display a large discrepancy in frequency sensitivity at long periods, below the 1 Hz corner frequency of SP instruments. The instrument response is defined mathematically by poles and zeros which amplify and attenuate frequencies respectively (*Scherbaum, 2001*). These parameters control the shape of the response spectrum for a given instrument, and thus determine the bandwidth (amplitude response is constant in this part of the spectrum) and gradient of “drop-off” outside the corner frequencies (e.g., *Kim and Ekström, 1996*). In the Fourier domain the impulse response $\mathbf{T}(\mathbf{i}\omega)$ is described by $\mathbf{L}$ complex zeros ($\mathbf{z}_i$) and $\mathbf{N}$ complex poles ($\mathbf{p}_i$) (Equation B.1 *Kim and Ekström, 1996*). The number and location of zeros controls where the amplitude response is zero, and poles, where the instrument response becomes infinity (denominator is zero in Equation B.1). Increasing the number of poles, increases the gradient of drop-off in amplitude outside the corner frequencies (*Scherbaum, 2001*).

$$\mathbf{T}(\mathbf{i}\omega) = \frac{\beta \prod_{j=1}^{\mathbf{L}} (\mathbf{i}\omega - \mathbf{z}_j)}{\alpha \prod_{k=1}^{\mathbf{N}} (\mathbf{i}\omega - \mathbf{p}_k)} = \frac{\mathbf{Output}(\mathbf{i}\omega)}{\mathbf{Input}(\mathbf{i}\omega)} \quad (\text{B.1})$$

Preprocessing steps must be followed to make use of both BB and SP instruments in a relative arrival-time study because BB seismic recordings will contain a much greater range of frequencies. This ensures the first arriving peak or trough of energy is picked consistently across the network. Therefore for networks including SP alongside BB instruments, such as that presented here, the instrument response of the BB instrument must be removed and a SP response reconvolved with that signal. This

accounts for the differing sensitivities below the corner frequency of the SP instruments of 1 Hz , but retains necessary frequency content required for picking of P-wave arrivals.

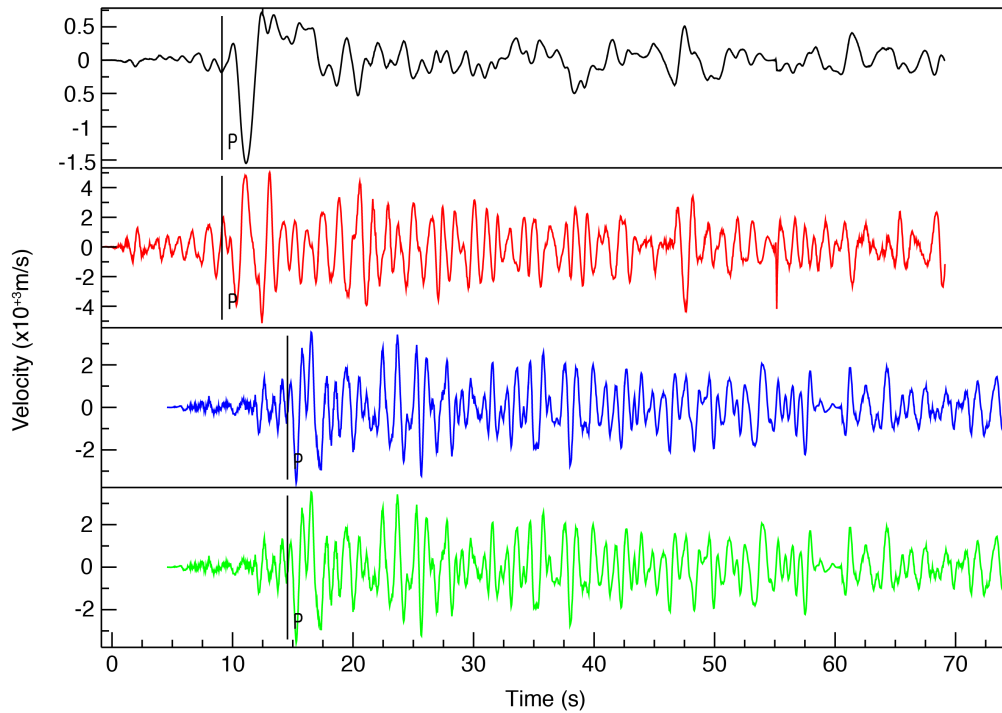


Figure B.1: **Removal of instrument response in the time domain.** BB seismic signal (A11 - black) has instrument response removed and is reconvolved with an SP response (A11 normalised - red) from station CNQ. SP seismic signal from station CNQ (blue) and normalised signal (green) show that normalisation to SP response has no effect for this SP station as expected.

Figures B.1 and B.2 show SP seismograms are unaltered by this process but BB traces are normalised to reveal the higher frequency part of the signal. This process is adequate to account for differences in the frequency content of recorded signals across a network. For some networks where only one instrument type is utilised, this will not be a necessary step but can be performed for consistency without detriment.

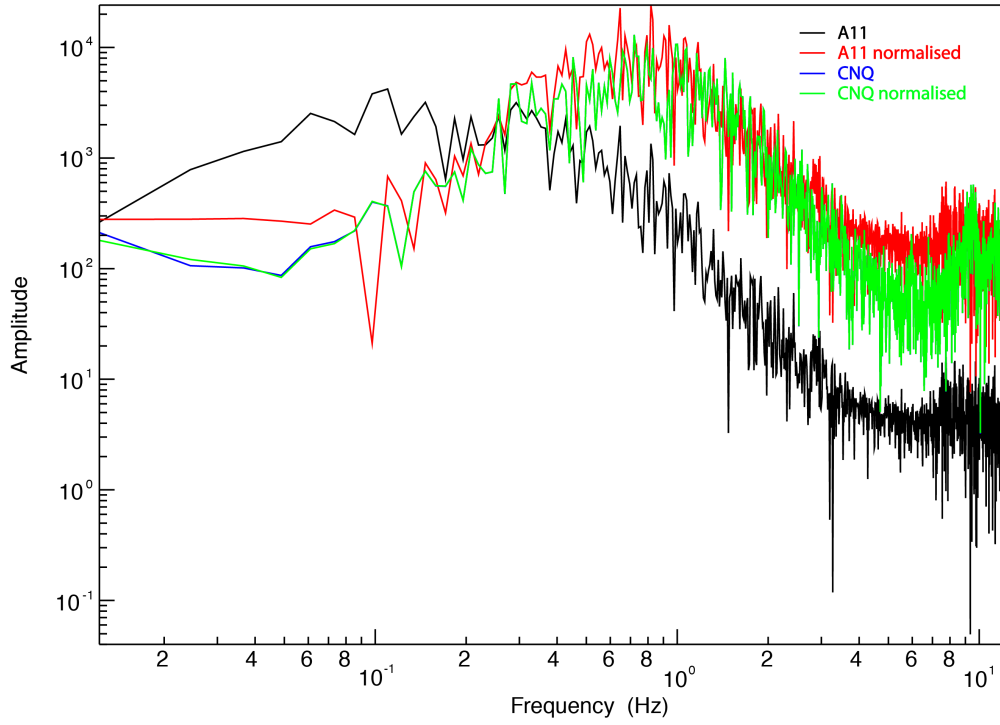


Figure B.2: **Removal of instrument response in the frequency domain** shows suppression of low frequencies and amplification of high frequencies of BB input signal (A11 - black) when reconvolved with a SP response (red). SP input seismogram shows no change after normalisation from input (CNQ - blue) to output (green) as expected.

B.1 Event selection - details

Of all the magnitude $m_b > 5.5$ earthquakes recorded across networks requested from data centres such as the IRIS Data Management Center and Canadian National Data centre *Boyce et al.* (in the case of 2016), fewer than half may contain data of sufficient quality to use. A first pass inspection and sort is performed to rapidly cut each dataset down to useable earthquakes only, with greater than four traces with adequate signal-to-noise around the predicted first arrival set using the *TauP* toolkit (*Crotwell and Ritsema, 1999*). This is a quality control measure imposed to ensure a more accurate cross-correlation using MCCC (*VanDecar and Crosson, 1990*) within the chosen frequency band of 0.4-2.0 Hz for P-waves, 0.04-0.2 Hz for S-waves.

B.2 Phase picking - details

Within each event, previously determined to have above average SNR at ≥ 4 stations, traces are then picked to identify the first peak or trough of energy. An automated script then runs to determine the optimised cross-correlation derived alignments using MCCC (*VanDecar and Crosson, 1990*) from which relative arrival-time residuals can be calculated. This code also generates input for the inversion method of *VanDecar et al. (1995)*. If the average cross-correlation coefficient r_{ij} , for a station is ≤ 0.85 , it is removed and the MCCC code is re-run to ensure improved alignment of traces (Figure B.3). This is repeated until this quality control criterium is satisfied. Visual inspection of alignment (e.g. Figure B.3) and alignment-time versus great circle arc distance (Figure B.4) enables quick identification of cycle skips and timing errors in the picked data. Figure B.4 shows arrival-time increases approximately linearly with great circle arc distance, slight deviations result from local heterogeneity in the subsurface. Cycle skips are easily identified as outliers away from the best fit line (blue).

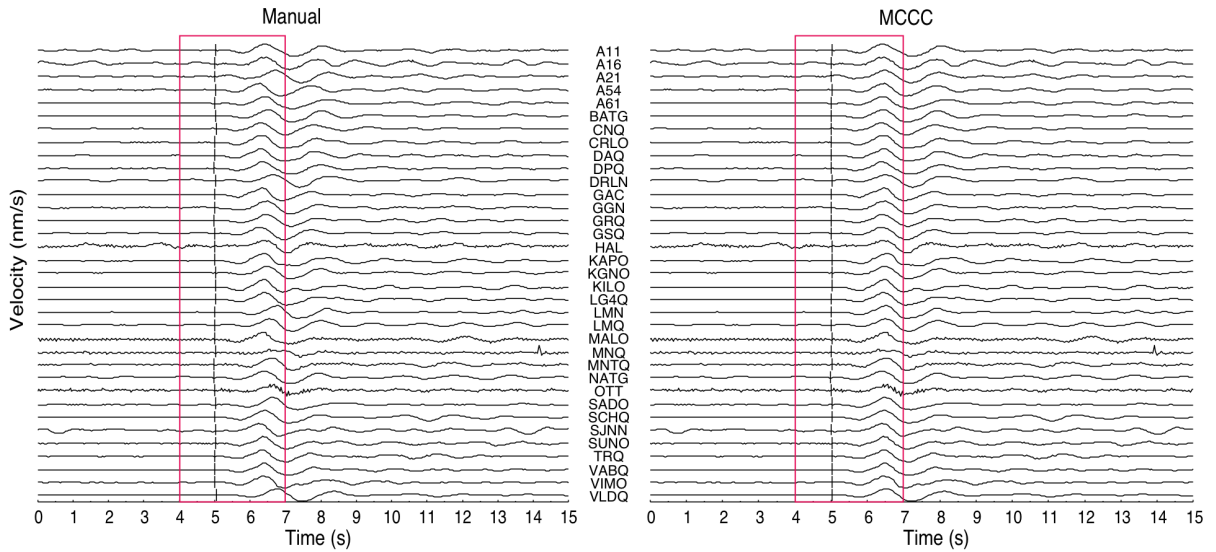


Figure B.3: **Improvement in alignment of traces** using optimised arrival-times derived from MCCC (*VanDecar and Crosson, 1990*) (right) over manual picks (left). The red region highlights the three second cross-correlation window on each trace used in the MCCC for P-waves.

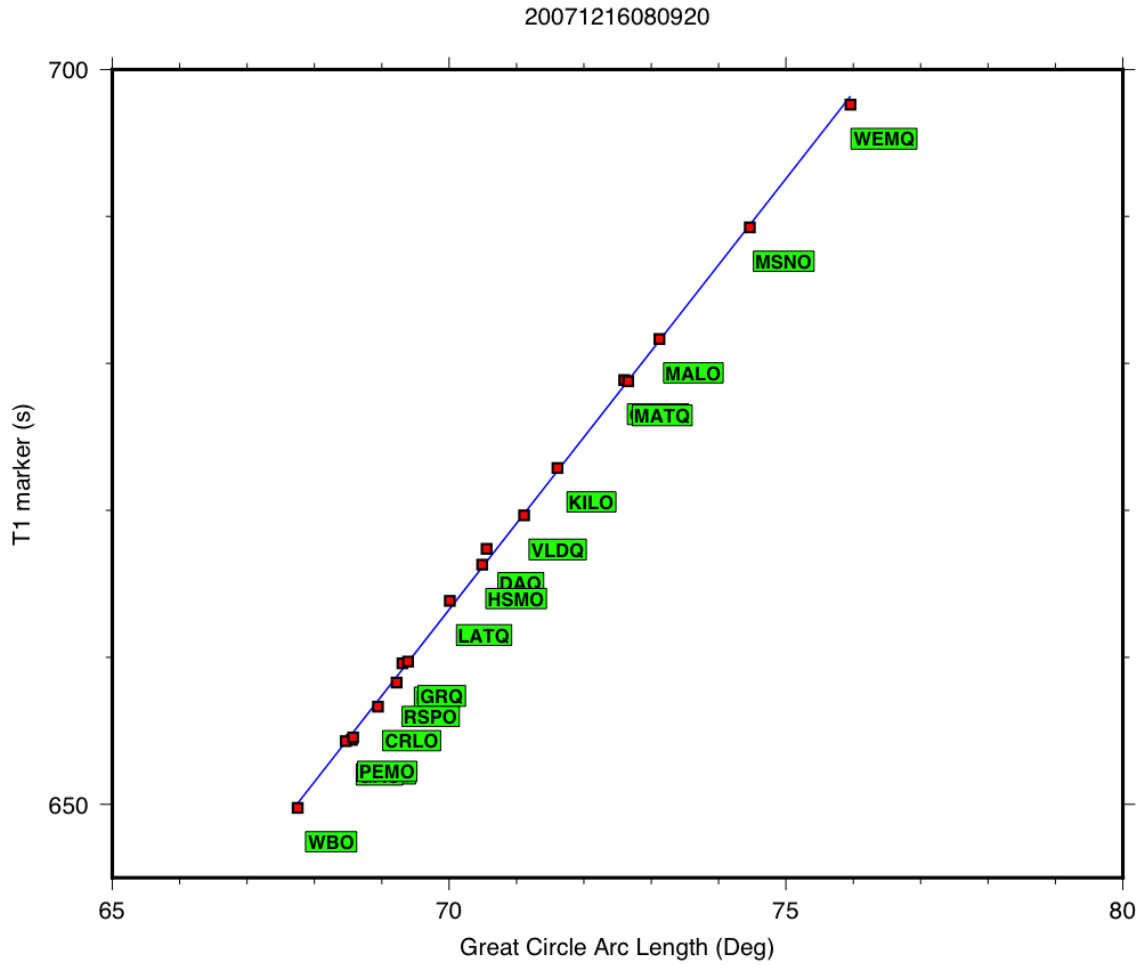


Figure B.4: **Plot of great circle arc length against arrival-time** for stations recording for earthquake 2007 – 12 – 16 08 : 09 : 20 in southeast Canada. A first order line of best fit is plotted in blue. Any optimised picks that deviate by ± 3 s are subject to further analysis for errors.

Whole dataset quality control measures are also undertaken to ensure residuals fall in an appropriate range (most residuals fall within ± 3 s for P-waves, ± 6 s for S-waves). See Figure 3.2 of relative arrival-time plotted against back azimuth for stations LG4Q, FRNY and FFD alongside the distribution of arrivals plotted against epicentral distance, back azimuth and residual for the P-wave dataset. The corresponding plot for the S-wave dataset is shown in Figure 3.3. All individual station plots for all analysed datasets are not included here in the interest of the environment.

B.3 Validation of short period data and filtering

Figure B.5 shows how the cross-correlation coefficient derived from MCCC (*VanDecar and Crosson, 1990*), for a given event, for broadband stations varies with (red) and without (blue) the inclusion of

SP stations in the dataset. Because there is very little effect on cross-correlation coefficients when SP stations are removed, these seismograms are included in the analysis to increase spatial coverage (see Figure 3.1). This analysis was performed on traces that have undergone instrument response normalisation described above.

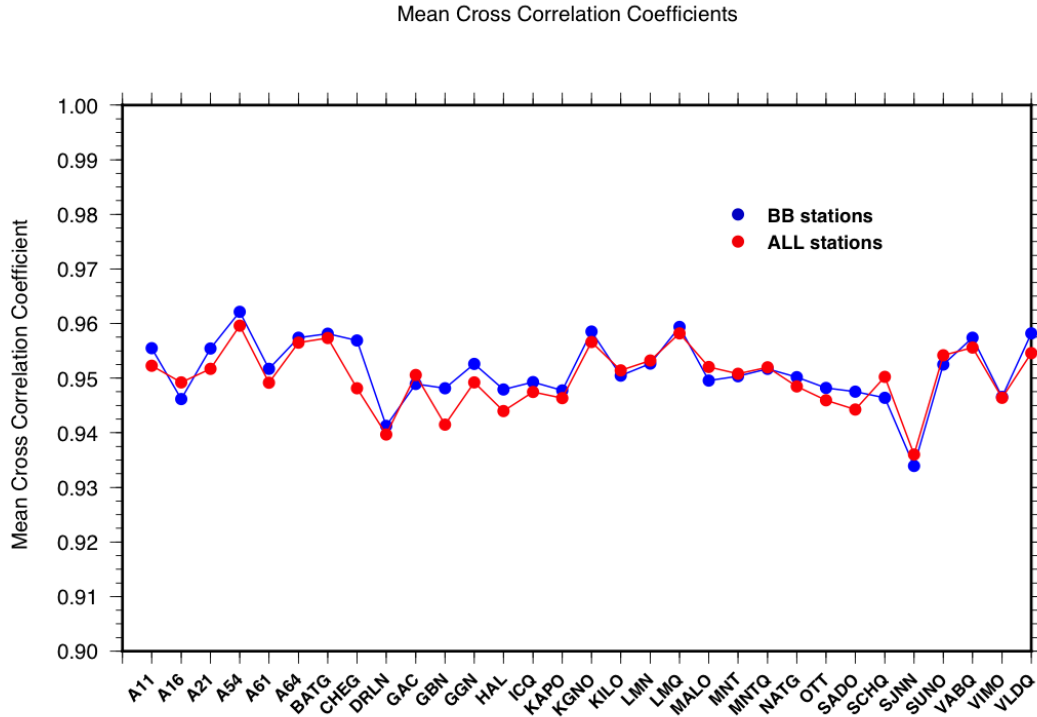


Figure B.5: **Effect on the average MCCC coefficient** of the broadband stations with (red) and without (blue) the inclusion of short period data.

Consistent filtering for the P- and S-wave datasets was used throughout the analysis (0.4-2.0 Hz for P-waves, 0.04-0.2 Hz for S-waves) and are in line with other recent tomographic studies (e.g., *Bastow et al.*, 2008; *Sol et al.*, 2002; *Villemaire et al.*, 2012). Filtering is required to suppress high frequency noise that is unique to each station whilst maintaining signal that is coherent across the network (*VanDecar and Crosson*, 1990). Harshly filtered waveforms (blue and green in Figure B.6) become “artificially” more alike due to a “ringing” effect that results from the convolution of the original waveform with a “boxcar” function in the frequency domain and the associated side lobes.

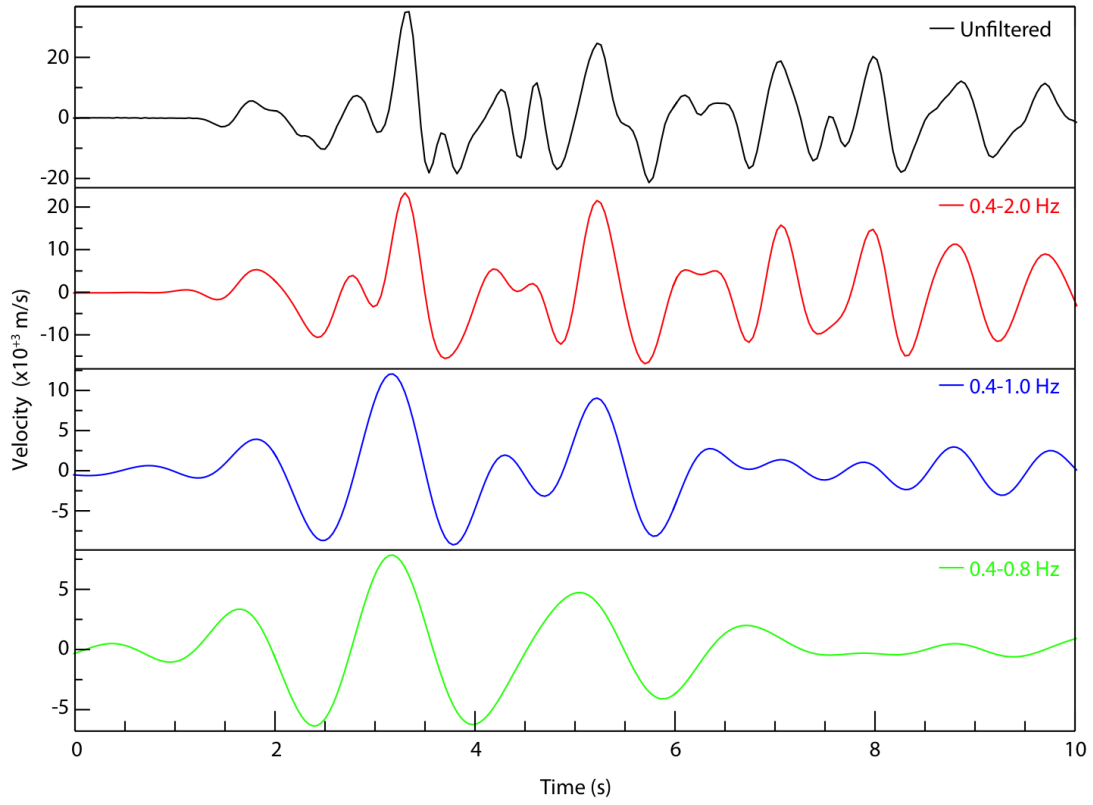


Figure B.6: **The effect of applying decreasing bandwidth filters** to a waveform recorded at station BATG from 02/09/2011. The majority of dominant frequencies contained within the unfiltered waveform (black) appear to be contained in the waveform filtered at $0.4 - 2.0\text{ Hz}$ (red). Decreasing bandwidths ($0.4 - 1.0\text{ Hz}$ - blue, $0.4 - 0.8\text{ Hz}$ - green) progressively suppress higher frequencies.

Various filters are tested to show that decreasing the bandwidth of filter imposed also artificially increases the cross-correlation coefficient as frequency content of traces are limited (Figures B.7). The increase in cross correlation coefficient is 2.6% and 2.4% for $0.4 - 0.8\text{ Hz}$ and $0.4 - 1.0\text{ Hz}$ respectively over the frequency band used in the cross-correlation for P-waves ($0.4 - 2.0\text{ Hz}$). Including higher frequencies results in a large drop in cross-correlation coefficient due to the incoherent noise present in each of the original records.

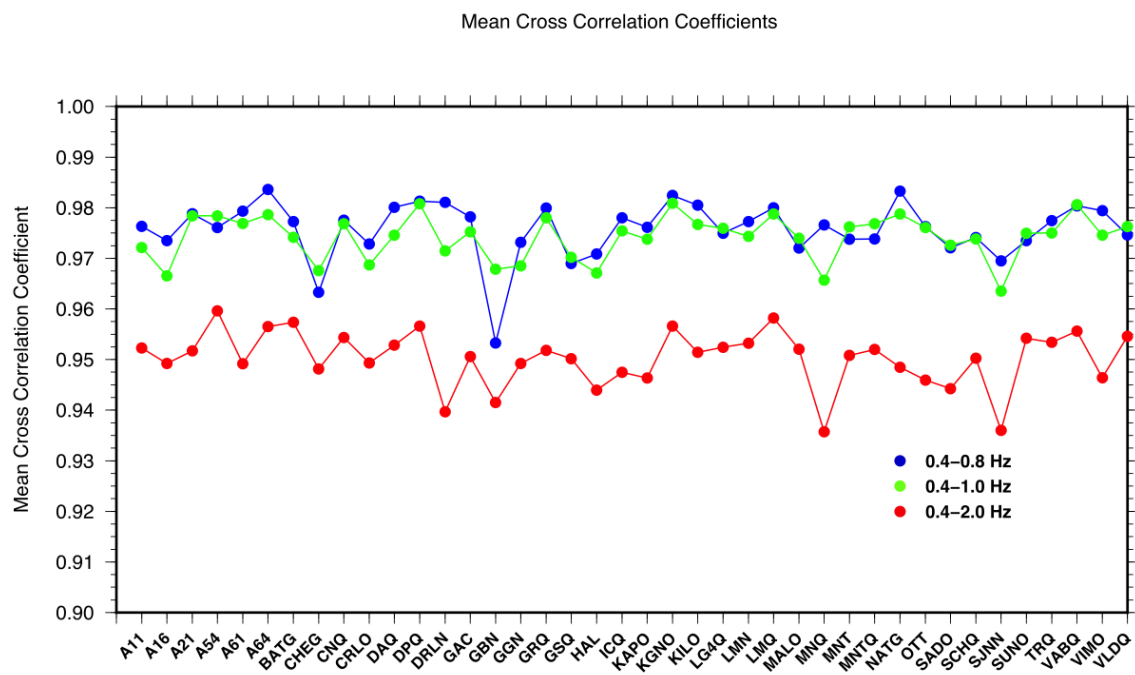


Figure B.7: Increase in average Cross-Correlation coefficient using harsher filters in the MCCC.

Appendix C

Supplemental information - Relative arrival-time tomography of SE Canada

C.1 Details of supplementary tables C.1, C.2

Table C.1

This table gives details of all earthquakes used in the analysis of data from southeast Canada (Chapter 3), excluding the QM-III network and western Superior stations (see sections 5.2.2, 5.2.2).

Table C.1: Earthquakes used for the relative arrival-time study of Chapter 3 - details and phases. Times are in ‘YYYY:MM:DD HH:MM:SS’ format.

Date.	Time.	Lat.	Long.	Depth (Km)	Back. Azi.	Great Circle Arc.	Phases.
2007:10:08	17:10:37	43.5170	146.708	55.30	30.1144	83.3061	P
2007:10:26	15:23:30	54.520	-161.710	43.20	62.882	51.1699	P
2007:11:14	17:44:05	-23.2150	-70.5260	38.0	357.5	66.3312	P
2007:11:15	15:03:09	-22.8670	-70.4070	27.0	357.312	72.5362	P
2007:11:15	15:06:00	-22.980	-70.940	20.70	353.611	68.5271	P
2007:11:16	03:13:00	-2.50	-78.0	114.30	359.952	48.2454	P
2007:11:18	05:40:09	-22.670	-66.480	262.40	351.385	69.1009	P
2007:11:20	17:55:51	-22.220	-70.540	33.0	354.325	67.2524	P

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Table C.1: *continued*

Date.	Time.	Lat.	Long.	Depth (Km)	Back. Azi.	Great Circle Arc.	Phases.
2007:11:21	12:55:06	-25.020	-68.810	127.30	356.188	68.2117	P
2007:11:24	05:02:08	-23.7720	-69.0060	94.40	353.231	68.9412	P
2007:11:26	13:51:39	37.3340	141.637	38.0	24.969	93.4923	P
2007:11:26	21:56:10	18.80	-101.580	70.10	37.6726	33.9476	P
2007:11:29	19:00:19	15.060	-61.410	147.90	338.252	32.9721	P
2007:12:16	08:09:20	-22.9140	-70.060	58.0	352.95	68.5567	P
2007:12:19	10:24:35	51.3450	-179.493	30.0	50.2135	65.8957	P
2007:12:21	07:23:37	51.2440	-179.009	30.0	50.2482	60.3822	P
2007:12:21	07:24:35	51.3350	-178.963	35.0	50.3134	60.3017	P
2007:12:26	23:40:54	-22.450	-68.6270	106.60	352.953	68.6135	P
2007:12:29	22:58:07	52.210	-168.149	15.80	57.6548	55.6661	P
2008:01:01	06:32:28	40.2880	72.9850	6.0	336.366	91.4334	P
2008:01:05	01:56:46	14.190	-91.410	108.70	17.7765	32.792	P
2008:01:05	07:29:33	-22.820	-68.4840	103.0	352.665	70.4926	P
2008:01:05	11:01:05	51.2540	-130.746	15.0	77.6653	34.6129	P
2008:01:05	11:44:48	51.1630	-130.542	10.0	77.6548	34.507	P
2008:01:06	05:14:21	37.2160	22.6930	75.0	309.394	69.6519	P
2008:02:04	17:01:31	-20.1660	-70.0370	35.0	357.038	63.3222	P
2008:02:21	02:46:18	77.0790	18.5710	12.0	289.074	47.1589	P
2008:02:21	14:16:04	41.1530	-114.867	6.70	67.2993	26.9469	P
2008:03:01	18:51:27	53.8860	159.295	112.30	37.4757	73.0418	P
2008:03:03	09:31:06	46.4060	153.175	10.0	33.8421	77.5901	P
2008:03:06	09:38:43	51.5220	156.546	134.40	35.2777	75.9401	P
2008:03:10	09:43:01	-28.3370	-69.2940	89.50	353.668	74.3655	P
2008:03:13	08:41:07	43.0660	146.528	52.90	28.12	86.6994	P
2008:03:14	10:45:34	47.5660	154.161	26.40	33.266	80.0871	P
2008:03:22	21:24:12	52.1760	-178.716	132.0	52.2871	61.7254	P
2008:03:24	20:39:07	-20.0430	-68.9630	120.0	353.083	66.1981	P
2008:04:12	00:46:28	43.8790	147.501	64.40	28.7713	85.6474	P

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Table C.1: *continued*

Date.	Time.	Lat.	Long.	Depth (Km)	Back. Azi.	Great Circle Arc.	Phases.
2008:04:15	22:59:54	51.8560	-179.361	11.0	50.4708	65.5057	P
2008:04:16	05:54:21	51.8890	-179.188	13.0	51.2294	61.396	P
2008:04:28	00:06:30	17.8460	-100.171	56.90	35.6743	33.9092	P
2008:05:02	01:33:35	51.8640	-177.528	14.0	51.5792	64.6169	P
2008:05:19	10:08:36	42.5030	131.872	513.50	18.3264	91.4736	P
2008:05:23	19:35:35	7.3130	-34.8970	8.0	323.564	53.4552	P
2008:05:24	04:58:18	42.3860	-30.5150	10.0	292.621	33.886	P
2008:05:24	19:20:43	4.330	-73.7640	8.90	355.498	41.6291	P
2008:05:29	15:46:00	64.0050	-21.0130	9.0	267.553	35.8648	P
2008:06:04	17:03:07	41.5380	139.058	205.0	24.9332	86.8111	P
2008:06:13	23:43:46	39.030	140.881	7.80	24.4414	92.203	P
2008:06:14	00:20:12	38.8820	140.678	6.40	24.3109	92.4029	P
2008:06:15	08:37:17	-36.6230	-107.450	10.0	20.0692	86.3637	P
2008:06:17	17:42:14	5.0080	-82.6520	10.0	5.58444	41.1269	P
2008:06:25	23:37:12	41.9370	142.486	55.60	25.4785	89.0539	P
2008:06:29	20:53:03	45.1560	137.446	326.0	24.0364	84.0071	P
2008:07:05	02:12:06	53.8820	152.886	632.80	35.2267	72.3715	P
2008:07:06	01:00:09	45.3540	151.044	22.0	31.1084	83.107	P
2008:07:06	09:08:24	45.3870	150.965	17.0	31.0602	83.1075	P
2008:07:08	09:13:07	-15.9860	-71.7480	123.0	358.369	59.1029	P
2008:07:15	03:26:36	35.80	27.860	52.0	314.102	69.3679	P
2008:07:17	22:51:55	44.3710	-129.417	10.0	68.9011	35.7409	P
2008:07:19	02:39:29	37.5520	142.214	22.0	21.2073	90.5896	P
2008:07:21	11:30:30	37.1860	142.051	22.0	25.2406	93.4859	P
2007:10:08	17:10:37	43.5170	146.708	55.30	30.1144	83.3061	P
2007:10:26	15:23:30	54.520	-161.710	43.20	62.882	51.1699	P
2007:11:14	17:44:05	-23.2150	-70.5260	38.0	357.5	66.3312	P
2007:11:15	15:03:09	-22.8670	-70.4070	27.0	357.312	72.5362	P,S
2007:11:15	15:06:00	-22.980	-70.940	20.70	353.611	68.5271	P

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Table C.1: *continued*

Date.	Time.	Lat.	Long.	Depth (Km)	Back. Azi.	Great Circle Arc.	Phases.
2007:11:16	03:13:00	-2.50	-78.0	114.30	359.952	48.2454	P,S
2007:11:18	05:40:09	-22.670	-66.480	262.40	351.385	69.1009	P,S
2007:11:20	17:55:51	-22.220	-70.540	33.0	354.325	67.2524	P,S
2007:11:21	12:55:06	-25.020	-68.810	127.30	356.188	68.2117	P,S
2007:11:24	05:02:08	-23.7720	-69.0060	94.40	353.231	68.9412	P
2007:11:26	13:51:39	37.3340	141.637	38.0	24.969	93.4923	P
2007:11:26	21:56:10	18.80	-101.580	70.10	37.6726	33.9476	P
2007:11:29	19:00:19	15.060	-61.410	147.90	338.252	32.9721	P,S
2007:12:16	08:09:20	-22.9140	-70.060	58.0	352.95	68.5567	P,S
2007:12:19	10:24:35	51.3450	-179.493	30.0	50.2135	65.8957	P
2007:12:21	07:23:37	51.2440	-179.009	30.0	50.2482	60.3822	P
2007:12:21	07:24:35	51.3350	-178.963	35.0	50.3134	60.3017	P,S
2007:12:26	23:40:54	-22.450	-68.6270	106.60	352.953	68.6135	P
2007:12:29	22:58:07	52.210	-168.149	15.80	57.6548	55.6661	P,S
2008:01:01	06:32:28	40.2880	72.9850	6.0	336.366	91.4334	P
2008:01:05	01:56:46	14.190	-91.410	108.70	17.7765	32.792	P
2008:01:05	07:29:33	-22.820	-68.4840	103.0	352.665	70.4926	P
2008:01:05	11:01:05	51.2540	-130.746	15.0	77.6653	34.6129	P,S
2008:01:05	11:44:48	51.1630	-130.542	10.0	77.6548	34.507	P
2008:01:06	05:14:21	37.2160	22.6930	75.0	309.394	69.6519	P,S
2008:02:04	17:01:31	-20.1660	-70.0370	35.0	357.038	63.3222	P,S
2008:02:21	02:46:18	77.0790	18.5710	12.0	289.074	47.1589	P
2008:02:21	14:16:04	41.1530	-114.867	6.70	67.2993	26.9469	P
2008:03:01	18:51:27	53.8860	159.295	112.30	37.4757	73.0418	P
2008:03:03	09:31:06	46.4060	153.175	10.0	33.8421	77.5901	P,S
2008:03:06	09:38:43	51.5220	156.546	134.40	35.2777	75.9401	P
2008:03:10	09:43:01	-28.3370	-69.2940	89.50	353.668	74.3655	P
2008:03:13	08:41:07	43.0660	146.528	52.90	28.12	86.6994	P
2008:03:14	10:45:34	47.5660	154.161	26.40	33.266	80.0871	P

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Table C.1: *continued*

Date.	Time.	Lat.	Long.	Depth (Km)	Back. Azi.	Great Circle Arc.	Phases.
2008:03:22	21:24:12	52.1760	-178.716	132.0	52.2871	61.7254	P,S
2008:03:24	20:39:07	-20.0430	-68.9630	120.0	353.083	66.1981	P,S
2008:04:12	00:46:28	43.8790	147.501	64.40	28.7713	85.6474	P
2008:04:15	22:59:54	51.8560	-179.361	11.0	50.4708	65.5057	P,S
2008:04:16	05:54:21	51.8890	-179.188	13.0	51.2294	61.396	P,S
2008:04:28	00:06:30	17.8460	-100.171	56.90	35.6743	33.9092	P
2008:05:02	01:33:35	51.8640	-177.528	14.0	51.5792	64.6169	P,S
2008:05:19	10:08:36	42.5030	131.872	513.50	18.3264	91.4736	P
2008:05:23	19:35:35	7.3130	-34.8970	8.0	323.564	53.4552	P,S
2008:05:24	04:58:18	42.3860	-30.5150	10.0	292.621	33.886	P
2008:05:24	19:20:43	4.330	-73.7640	8.90	355.498	41.6291	P
2008:05:29	15:46:00	64.0050	-21.0130	9.0	267.553	35.8648	P,S
2008:06:04	17:03:07	41.5380	139.058	205.0	24.9332	86.8111	P,S
2008:06:13	23:43:46	39.030	140.881	7.80	24.4414	92.203	P
2008:06:14	00:20:12	38.8820	140.678	6.40	24.3109	92.4029	P
2008:06:15	08:37:17	-36.6230	-107.450	10.0	20.0692	86.3637	P
2008:06:17	17:42:14	5.0080	-82.6520	10.0	5.58444	41.1269	P,S
2008:06:25	23:37:12	41.9370	142.486	55.60	25.4785	89.0539	P
2008:06:29	20:53:03	45.1560	137.446	326.0	24.0364	84.0071	P,SKS
2008:07:05	02:12:06	53.8820	152.886	632.80	35.2267	72.3715	P,S
2008:07:06	01:00:09	45.3540	151.044	22.0	31.1084	83.107	P
2008:07:06	09:08:24	45.3870	150.965	17.0	31.0602	83.1075	P
2008:07:08	09:13:07	-15.9860	-71.7480	123.0	358.369	59.1029	P,S
2008:07:15	03:26:36	35.80	27.860	52.0	314.102	69.3679	P,S
2008:07:17	22:51:55	44.3710	-129.417	10.0	68.9011	35.7409	P
2008:07:19	02:39:29	37.5520	142.214	22.0	21.2073	90.5896	P,SKS
2008:07:21	11:30:30	37.1860	142.051	22.0	25.2406	93.4859	P
2008:07:23	15:26:20	39.8020	141.464	108.0	20.7488	88.702	P,SKS
2008:07:24	01:43:18	50.9670	157.584	27.0	31.4503	73.7498	P,S

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Table C.1: *continued*

Date.	Time.	Lat.	Long.	Depth (Km)	Back. Azi.	Great Circle Arc.	Phases.
2008:07:25	00:29:25	51.2790	178.982	21.0	49.2758	66.6688	P
2008:07:27	21:15:42	-0.2530	-18.2870	17.0	319.102	69.1843	P,S
2008:08:04	04:42:15	49.8520	156.380	76.90	36.2803	73.5348	P,S
2008:08:13	18:30:57	83.550	114.516	10.0	11.3301	50.591	P,S
2008:08:14	22:55:52	7.2920	-82.7260	6.0	5.21443	38.7297	P
2008:08:16	04:01:07	52.2720	98.2070	12.10	357.365	82.0995	P
2008:08:26	21:00:36	-7.6410	-74.3770	154.0	355.576	53.0327	P,S
2008:08:27	01:35:31	51.6070	104.158	16.0	1.55309	82.7997	P,S
2008:08:28	15:22:23	-0.2520	-17.3580	12.0	318.523	66.3331	P,S
2008:09:03	11:25:14	-26.7360	-63.2250	569.60	349.275	73.5885	P,S
2008:09:10	16:11:59	-20.2620	-69.2240	37.70	353.29	66.3854	P
2008:09:13	09:32:02	4.790	-75.5170	132.90	2.1673	38.469	P
2008:09:16	07:28:25	0.9020	-28.9980	10.0	323.388	62.0725	P
2008:09:18	11:58:51	51.940	158.363	35.0	36.5266	74.9384	P
2008:09:22	07:31:59	41.5780	140.450	149.0	26.1584	87.1943	P
2008:09:24	02:33:05	17.6120	-105.497	12.0	39.4969	37.1945	P,S
2008:10:05	15:52:50	39.5330	73.8240	27.40	336.912	92.3815	P,SKS
2008:10:05	16:11:11	39.5080	73.8880	35.0	336.954	92.4239	P
2008:10:07	10:00:48	79.8170	-115.448	10.0	134.605	36.521	P
2008:10:11	09:06:10	43.3720	46.2540	16.0	323.942	78.2557	P
2008:10:12	20:55:41	-20.1230	-64.9710	352.70	356.303	67.4306	P,S
2008:10:16	19:41:33	14.4230	-92.3640	24.0	19.0158	32.8641	P
2008:11:24	09:03:00	54.2030	154.322	492.30	36.224	71.6159	P
2008:12:25	08:11:59	49.1280	-128.621	10.0	75.782	33.7718	P
2009:01:03	20:23:20	36.4190	70.7430	204.80	338.773	93.3135	P
2009:01:04	23:12:59	36.4420	70.8830	186.70	338.862	93.3329	P
2009:01:08	19:21:34	10.1650	-84.1970	14.0	13.6603	34.3413	P,S
2009:01:15	17:49:38	46.8570	155.154	36.0	29.6782	78.2412	P,S
2009:01:26	19:11:48	51.9550	-171.159	25.30	55.4227	61.3916	P

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Table C.1: *continued*

Date.	Time.	Lat.	Long.	Depth (Km)	Back. Azi.	Great Circle Arc.	Phases.
2009:01:29	22:28:06	-8.1080	-71.4020	602.40	354.01	55.6384	P
2009:01:31	21:51:52	36.7190	141.148	34.0	24.6692	94.2137	P
2009:02:02	17:53:23	-13.5820	-76.5580	21.0	358.788	59.2718	P
2009:02:06	02:58:35	-27.9870	-71.1040	52.20	354.942	72.9217	P
2009:02:09	14:09:03	-6.5670	-81.150	15.0	2.92238	51.4325	P,S
2009:02:15	09:24:33	40.2390	142.236	33.0	26.8563	86.926	P
2009:02:15	10:04:50	-5.8540	-80.8910	21.0	2.52662	51.6395	P
2009:02:16	02:22:48	-29.1340	-69.8070	75.0	354.022	74.1689	P
2009:02:16	23:16:39	37.130	20.780	15.0	308.636	68.5186	P
2009:02:28	00:35:56	42.610	142.104	105.0	25.2338	88.5675	P
2009:03:05	19:41:41	80.2840	-1.8320	10.0	266.447	42.7547	P
2009:03:06	10:50:30	80.3240	-1.8530	9.0	258.923	40.1871	P,S
2009:03:07	14:33:04	41.7960	143.693	32.70	26.2646	88.7873	P
2009:03:11	13:06:53	-31.8850	-69.1890	28.0	353.649	76.9523	P
2009:03:11	17:24:37	8.5040	-83.2190	14.0	11.9537	35.737	P
2009:03:11	21:03:58	8.4930	-83.2060	17.0	11.935	35.745	P,S
2009:03:15	03:14:31	2.4290	-95.1240	10.0	22.1453	45.029	P,S
2009:03:15	08:19:07	-14.4520	-70.3650	189.70	353.786	59.5846	P
2009:03:30	07:13:06	56.5460	-152.743	21.0	70.9937	46.3806	P,S
2009:03:30	12:07:27	51.5350	-178.258	31.0	46.2087	62.9065	P
2009:04:03	17:54:47	-27.8420	-66.5930	174.20	351.629	73.2452	P
2009:04:06	01:32:42	42.3340	13.3340	8.80	303.501	60.8221	P,S
2009:04:07	04:23:34	46.0490	151.548	31.0	27.4168	80.1424	P,S
2009:04:07	17:47:38	42.2750	13.4640	15.10	303.586	60.935	P
2009:04:14	03:04:26	48.8370	154.883	43.50	33.8589	78.7602	P
2009:04:17	02:08:09	-19.5840	-70.4830	25.0	354.163	64.6481	P,S
2009:04:18	19:18:01	46.0150	151.427	35.0	27.3403	80.2113	P
2009:04:21	05:26:11	50.8330	155.009	152.0	36.0491	74.095	P,S
2009:04:27	16:46:27	17.0320	-99.4460	35.0	34.3636	34.175	P,S

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Table C.1: *continued*

Date.	Time.	Lat.	Long.	Depth (Km)	Back. Azi.	Great Circle Arc.	Phases.
2009:04:28	11:21:19	42.6530	144.950	14.70	27.0921	87.6029	P
2009:05:02	02:19:09	54.5310	-161.520	9.40	63.8154	51.758	P,S
2009:05:03	16:21:46	14.5710	-91.1690	123.80	24.1546	32.3979	P,S
2009:05:08	13:44:53	-32.0570	-69.6780	111.90	354.039	78.0163	P
2009:05:08	21:22:30	58.1280	164.258	8.70	41.785	67.9912	P
2009:05:10	01:16:06	1.3930	-85.1690	6.0	5.83112	44.1602	P,S
2009:05:17	19:24:21	82.5160	-8.6220	10.0	256.766	42.9669	P
2009:05:19	17:35:02	25.2920	37.7440	2.0	320.255	88.774	P
2009:05:21	12:33:54	36.2670	77.5380	99.30	343.17	95.2463	P
2009:05:22	00:24:20	13.8750	-90.7420	69.10	16.0249	33.7249	P
2009:05:22	19:24:17	18.1140	-98.4570	62.30	26.907	32.5144	P
2009:05:28	08:24:48	16.7310	-86.2170	19.0	20.2321	33.7527	P
2009:06:03	21:37:38	19.5530	-109.165	6.0	43.3795	38.0515	P,S
2009:06:05	03:30:33	41.8240	143.445	29.0	26.1036	88.844	P,SKS
2009:06:06	05:52:43	35.4830	140.914	34.0	26.0329	91.6433	P
2009:06:06	20:33:29	23.8640	-46.1050	14.0	317.137	33.3513	P,S
2009:06:08	05:13:15	15.7920	-86.8650	10.0	19.6183	29.7189	P
2009:06:10	23:13:21	-41.7160	-83.9950	10.0	7.53625	85.2257	P
2009:06:13	17:17:39	44.7240	78.8640	14.50	340.346	88.9186	P
2009:06:17	07:08:18	51.6060	-175.234	31.0	53.3498	59.6246	P
2009:06:19	14:05:00	35.3580	28.4530	28.0	316.293	75.478	P
2009:06:22	18:15:40	76.2240	6.7870	10.0	279.057	44.715	P
2009:06:22	19:28:03	61.9390	-150.704	64.60	76.8552	43.1433	P
2009:06:23	07:37:18	38.850	142.389	37.0	25.426	91.8702	P
2009:07:01	09:30:10	34.1640	25.4710	19.0	313.55	69.308	P,S
2009:07:04	06:49:36	9.590	-78.9660	38.0	9.65665	38.2563	P,S
2009:07:06	14:53:12	50.4350	176.992	22.0	47.8203	68.1736	P,S
2009:07:08	19:23:38	-36.0040	-102.686	12.0	16.9849	84.5553	P
2009:07:14	18:38:08	-21.8220	-67.0870	175.60	351.783	68.1837	P,S

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Table C.1: *continued*

Date.	Time.	Lat.	Long.	Depth (Km)	Back. Azi.	Great Circle Arc.	Phases.
2009:08:05	09:13:12	29.6150	-113.789	10.0	49.7088	32.2903	P,S
2009:08:09	10:55:56	33.1670	137.941	297.0	24.2937	94.7802	P,SKS
2009:08:20	06:35:05	72.20	0.9440	6.0	272.153	44.7022	P,S
2009:08:28	01:52:06	37.6960	95.7180	13.0	355.629	96.5394	P,SKS
2009:08:31	10:15:30	37.6170	95.8360	10.0	355.711	96.6251	P
2009:09:05	03:58:39	-15.1210	-70.2480	210.20	353.795	61.1952	P,S
2009:09:07	22:41:37	42.660	43.4430	15.0	318.333	77.4643	P,S
2009:09:10	02:46:52	48.3170	154.192	36.0	34.7085	75.6208	P,S
2009:09:10	19:46:03	-29.9260	-111.824	10.0	23.6262	80.7273	P
2009:09:11	08:49:12	48.2940	154.230	35.0	33.3823	79.4529	P
2009:09:12	20:06:26	10.7050	-67.920	10.0	348.002	36.178	P,S
2009:09:17	23:21:38	-29.1440	-112.267	10.0	27.4712	80.3974	P,S
2009:09:21	08:30:14	71.0370	-7.670	10.0	270.349	41.879	P
2009:09:24	07:16:22	18.8220	-107.345	10.0	35.2224	36.2882	P,S
2009:09:30	19:03:17	-15.5510	-69.2930	255.40	356.271	58.7814	P
2009:10:06	17:22:54	18.7960	-107.527	10.0	35.3351	36.4089	P
2009:10:07	05:38:37	52.1570	178.052	138.40	48.9992	66.5311	P
2009:10:13	05:37:22	52.7540	-166.997	24.0	59.4026	55.472	P,S
2009:10:13	20:21:52	52.6040	-167.118	14.0	58.1435	58.9588	P,S
2009:10:15	00:13:33	52.8530	-166.750	20.0	58.493	58.6374	P
2009:10:15	17:48:21	3.2720	-103.823	10.0	24.0503	48.0983	P
2009:10:22	00:51:39	6.7290	-82.5790	18.0	10.7303	37.3414	P,S
2009:10:22	19:51:27	36.5170	70.950	185.90	334.958	94.2051	P
2009:10:29	10:52:56	18.9110	-95.5380	16.40	24.3184	30.6012	P
2009:10:29	17:44:32	36.3880	70.7250	202.20	338.761	93.3371	P
2009:11:04	18:41:44	36.1540	-33.8740	10.0	295.733	31.1975	P
2009:11:04	21:56:10	37.6450	95.7590	3.0	353.634	92.3849	P
2009:11:06	08:49:52	-23.4360	-64.4650	5.0	349.966	70.154	P
2009:11:13	03:05:58	-19.3940	-70.3210	27.0	357.251	62.5418	P,S

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Table C.1: *continued*

Date.	Time.	Lat.	Long.	Depth (Km)	Back. Azi.	Great Circle Arc.	Phases.
2009:11:13	07:27:13	-17.9170	-64.0950	608.0	349.285	64.8146	P,S
2009:11:14	19:44:29	-22.9650	-66.6410	220.40	354.412	66.3354	P,S
2009:11:17	15:30:46	52.1310	-131.397	3.30	78.5167	34.8282	P,S
2009:11:20	19:31:27	-0.180	-18.0020	5.0	319.888	69.7531	P
2009:11:22	22:07:52	-39.6360	-74.9720	4.0	0.958689	82.6477	P
2009:11:23	06:08:41	18.0830	-105.597	34.80	39.9602	36.8952	P
2009:11:26	19:08:10	13.5140	-89.9070	56.80	14.9094	33.8543	P
2009:12:10	02:30:52	53.4280	152.712	656.0	33.0056	75.6767	P,S
2009:12:17	01:37:47	36.4240	-9.9330	10.0	307.678	47.4703	P
2010:01:10	00:27:38	40.6448	-124.763	21.710	68.4793	37.5798	S
2010:01:10	00:27:41	40.6520	-124.692	29.30	61.5842	39.0974	P
2010:02:04	20:20:21	40.4120	-124.961	23.60	61.2724	39.3921	P
2010:02:27	06:34:13	-36.1220	-72.8980	22.90	1.69823	83.2337	P,S
2010:03:11	14:39:45	-34.290	-71.8910	11.0	1.16739	81.176	P
2010:03:11	14:55:29	-34.3260	-71.7990	18.0	1.10387	81.2104	P,S
2010:04:02	22:58:09	-36.2270	-72.8780	24.0	1.839	83.1291	P
2010:04:04	22:40:43	32.2970	-115.278	4.0	52.9617	37.1866	P,S
2010:04:11	22:08:11	36.9650	-3.5420	609.80	304.166	49.1182	P,S
2010:05:06	02:42:48	-18.0580	-70.5470	37.0	0.262528	64.9963	P
2010:05:23	22:46:51	-13.9280	-74.3520	101.40	3.23502	61.0029	P
2010:05:24	16:18:28	-8.0870	-71.5580	581.20	1.13	55.0975	P
2010:07:12	00:11:20	-22.1460	-68.2160	115.0	358.555	69.0857	P,S
2010:08:12	11:54:15	-1.2660	-77.3060	206.70	6.44292	48.7086	P
2011:03:06	12:31:58	-18.0210	-69.3620	118.0	359.371	64.9631	P,S
2011:03:11	05:46:23	38.2970	142.373	29.0	21.5184	90.0028	P,SKS
2011:03:11	06:25:50	38.0580	144.590	18.60	22.9101	89.3064	P
2011:03:11	11:36:40	39.2410	142.463	25.70	21.5767	89.1002	P
2011:03:22	07:18:46	37.2440	144.003	11.0	22.5196	90.493	P,SKS
2011:04:01	13:29:11	35.490	26.670	75.80	313.61	69.1107	P,S

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Table C.1: *continued*

Date.	Time.	Lat.	Long.	Depth (Km)	Back. Azi.	Great Circle Arc.	Phases.
2011:04:07	13:11:23	17.2080	-94.3380	166.20	28.334	35.9529	P
2011:04:07	14:32:44	38.2760	141.588	42.0	21.0328	90.2465	P
2011:04:11	08:16:12	37.0010	140.401	11.0	20.3033	91.7699	P,SKS
2011:04:23	10:12:46	39.0970	142.870	31.0	21.8274	89.1168	P
2011:06:03	00:05:01	37.350	143.960	20.50	22.493	90.4084	P
2011:06:22	21:50:51	39.990	142.510	41.30	21.6119	88.3913	P
2011:06:24	03:09:38	52.050	-171.836	52.0	49.9922	60.6048	P,S
2011:07:16	19:59:12	54.7870	-161.290	36.0	57.4672	53.9029	P,S
2011:08:04	13:51:36	48.70	155.280	43.50	29.9595	76.5989	P,S
2011:08:24	17:46:11	-7.6410	-74.5250	147.0	3.60812	54.7762	P,S
2011:09:02	10:55:54	52.1710	-171.708	32.0	50.1169	60.4665	P,S
2011:09:02	13:47:10	-28.3980	-63.0290	578.90	354.963	75.5657	P,S
2011:09:09	19:41:33	49.5350	-126.893	22.0	71.2732	36.9616	P,S
2011:10:23	10:41:22	38.7220	43.5130	16.0	319.992	76.0146	P,S
2011:11:07	22:35:25	11.560	-85.8610	177.0	17.4088	37.9403	P,S
2011:11:22	18:48:16	-15.3640	-65.090	549.90	356.078	62.4853	P,S
2012:01:30	05:10:59	-14.1680	-75.6350	43.0	4.22012	61.3219	P,S
2012:02:26	06:17:19	51.7080	95.9910	12.0	350.514	80.7178	P
2012:03:05	07:46:09	-28.2460	-63.2940	553.90	355.145	75.3947	P
2012:03:14	09:08:36	40.8870	144.944	12.0	23.1275	86.8558	P
2012:03:25	22:37:06	-35.20	-72.2170	40.70	1.51311	82.3217	P,S
2012:04:11	22:41:46	43.5840	-127.638	8.0	64.0085	39.7085	P
2012:04:11	22:55:10	18.2290	-102.689	20.0	35.2967	39.3036	P
2012:04:12	07:15:49	28.6960	-113.104	13.0	48.9004	37.9925	P
2012:04:17	03:50:15	-32.6250	-71.3650	29.0	0.808767	79.5081	P,S
2012:05:14	10:00:40	-17.6780	-69.5910	105.90	359.542	64.6199	P,S
2012:05:20	02:03:52	44.890	11.230	6.30	303.673	54.054	P,S
2012:05:28	05:07:23	-28.0430	-63.0940	586.90	355	75.2084	P,S
2012:06:02	07:52:53	-22.0590	-63.5550	527.0	355.165	69.2362	P

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Table C.1: *continued*

Date.	Time.	Lat.	Long.	Depth (Km)	Back. Azi.	Great Circle Arc.	Phases.
2012:08:02	09:38:30	-8.4140	-74.2590	144.60	3.35566	55.5268	P,S
2012:08:14	02:59:38	49.80	145.064	583.20	23.6542	78.6386	P
2012:09:05	14:42:07	10.0850	-85.3150	35.0	16.3331	39.1875	P
2012:09:30	16:31:35	1.9290	-76.3620	170.0	6.02372	45.6973	P,S
2012:10:28	03:04:08	52.7880	-132.101	14.0	72.0742	39.06	P,S
2012:11:07	16:35:46	13.9880	-91.8950	24.0	24.2315	38.0423	P
2012:12:07	08:18:23	37.890	143.949	31.0	22.268	89.7667	P
2012:12:14	10:36:01	31.0950	-119.660	13.0	51.9637	41.0325	P,S
2013:01:28	16:38:53	42.570	79.760	24.40	339.873	86.1383	P
2013:01:30	20:15:43	-28.090	-71.180	45.80	1.03988	75.4543	P,S
2013:02:02	14:17:35	42.850	143.240	105.30	21.6193	85.252	P,SKS
2013:02:09	14:16:07	1.1350	-77.3930	145.0	6.89504	46.597	P,S
2013:02:14	13:13:52	67.6310	142.508	11.0	24.2447	62.6294	P,S
2013:02:22	12:01:59	-27.9320	-63.0970	575.20	355.16	75.3102	P,S
2013:02:28	14:05:50	50.9540	157.283	41.0	31.2624	73.86	P,S
2013:03:25	23:02:13	14.4870	-90.4630	189.0	23.1862	37.0297	P,S
2013:04:19	03:05:52	46.2210	150.788	110.0	26.7416	80.0884	P,S
2013:05:19	18:44:10	52.340	160.065	18.0	33.2142	71.7615	P,S
2013:05:24	05:44:49	54.8920	153.221	598.10	29.3026	71.6943	P,S
2013:05:24	14:56:31	52.2350	151.444	624.0	25.3163	75.8014	P,S
2013:06:16	21:39:04	34.3470	25.1590	19.0	313.543	68.7729	P,S
2013:06:24	22:04:13	10.7010	-42.5940	10.0	332.922	43.327	P,S
2013:09:28	07:34:06	27.1825	65.5052	12.0	331.424	95.8433	SKS
2013:10:01	03:38:21	53.1995	152.786	573.0	28.9872	73.4522	P
2013:10:12	13:11:53	35.5142	23.2523	40.0	312.392	67.0546	P,S
2013:10:19	17:54:54	26.0913	-110.321	9.50	45.5779	37.9367	P,S
2013:10:25	17:10:19	37.1557	144.661	35.0	22.9197	90.3713	P
2013:10:31	23:03:59	-30.2921	-71.5215	27.0	0.924904	77.1846	P,S
2013:11:02	15:52:46	-23.6357	-112.596	10.0	27.7435	80.492	P

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Table C.1: *continued*

Date.	Time.	Lat.	Long.	Depth (Km)	Back. Azi.	Great Circle Arc.	Phases.
2013:11:12	07:03:51	54.6859	162.302	43.0	35.3266	69.2011	P,S
2013:11:17	17:37:14	-10.3290	-11.8343	10.0	324.086	77.1581	P
2013:11:25	05:56:50	45.5613	151.005	34.0	26.7689	80.3852	P
2013:12:08	17:24:54	44.4438	149.167	28.0	25.852	82.3312	P
2013:12:17	12:30:00	-25.5090	-70.5210	71.50	0.231013	72.4109	P
2013:12:28	15:21:04	36.0280	31.310	40.70	315.225	71.4311	P,S
2014:01:11	13:10:51	14.6437	-92.0592	78.0	25.2058	32.6921	P
2014:01:13	04:01:03	19.0428	-66.8096	20.0	355.122	28.2628	S
2014:01:16	07:33:09	51.1701	-179.225	27.0	45.481	64.6058	P
2014:01:26	13:55:42	38.2082	20.4528	8.0	310.475	63.5922	P
2014:01:27	16:42:37	-19.2592	-63.9189	600.20	355.337	66.4319	P
2014:02:03	03:08:46	38.2637	20.3897	5.0	310.432	63.5185	P,S
2014:02:12	09:19:49	35.9053	82.5864	10.0	341.804	93.7014	SKS
2014:02:18	09:27:13	14.6682	-58.9272	14.80	346.153	33.8092	P,S
2014:02:18	23:35:58	-14.1648	-75.5997	57.0	4.19298	61.3162	P,S
2014:02:26	21:13:40	53.6047	-171.821	265.0	50.6902	59.6037	P
2014:03:02	09:37:54	12.5559	-87.6879	60.0	12.6119	33.4107	S
2014:03:04	10:51:15	-33.6046	-71.9570	20.0	1.21522	80.4936	P,S
2014:03:05	09:56:57	-14.7378	169.823	638.0	43.4918	120.96	SKS
2014:03:10	00:38:15	16.0414	-98.3358	8.40	30.8099	38.8549	P,S
2014:03:10	05:18:13	40.8287	-125.134	16.60	61.6829	39.3089	P,S
2014:03:13	17:06:50	33.6842	131.825	79.0	14.9177	97.0425	SKS
2014:03:15	08:59:21	-14.0848	-76.3113	20.0	4.7457	61.2908	P,S
2014:03:16	21:16:29	-19.9807	-70.7022	20.0	0.373602	66.9096	P,S
2014:03:17	05:11:34	-20.0168	-70.8837	21.0	0.507894	66.9468	S
2014:03:18	21:26:45	-19.9262	-70.7961	3.30	0.65954	67.443	P,S
2014:03:22	12:59:59	-19.7625	-70.8740	20.0	0.501659	66.6938	S
2014:03:23	18:20:01	-19.6899	-70.8538	21.0	0.55922	67.0735	P,S
2014:03:24	11:26:39	-19.8239	-70.7651	22.0	357.624	62.9503	P,S

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Table C.1: *continued*

Date.	Time.	Lat.	Long.	Depth (Km)	Back. Azi.	Great Circle Arc.	Phases.
2014:03:24	11:40:43	-19.8288	-70.8824	15.20	0.507639	66.7598	P,S
2014:03:24	15:45:31	-19.5932	-70.8207	17.10	0.462707	66.5249	P,S
2014:03:27	23:50:52	-29.1646	-68.7459	10.0	353.301	75.2394	P
2014:03:28	14:36:08	11.6487	-86.5532	46.0	18.154	38.3257	P
2014:03:31	12:53:06	-19.5110	-69.1740	114.50	359.474	67.0281	P
2014:04:01	23:46:47	-19.6097	-70.7691	25.0	0.424326	66.541	P,S
2014:04:02	04:46:18	-20.0759	-70.8325	16.50	357.684	63.1982	P
2014:04:03	01:58:30	-20.3113	-70.5756	24.10	0.119843	67.4514	P,S
2014:04:03	02:43:13	-20.5709	-70.4931	22.40	351.86	62.5465	P,S
2014:04:03	05:26:15	-20.7969	-70.5865	25.0	0.286206	67.721	P,S
2014:04:04	01:37:50	-20.6426	-70.6540	13.70	0.336277	67.5678	P,S
2014:04:04	20:08:06	37.2833	23.8711	107.0	312.048	66.2379	P,S
2014:04:07	13:43:21	-20.1273	-70.8513	8.0	0.698803	67.6436	P,S
2014:04:08	10:14:31	-20.5124	-70.9226	6.0	0.534758	67.4402	S
2014:04:11	00:01:45	-20.6590	-70.6472	13.80	356.749	66.0808	P,S
2014:04:11	20:29:12	11.6420	-85.8779	135.0	17.4571	37.8675	P,S
2014:04:13	12:11:30	-20.5664	-70.7470	13.40	0.405048	67.4926	S
2014:04:13	12:36:19	-11.4633	162.051	39.0	40.3033	123.605	SKS
2014:04:18	18:44:18	67.7185	-162.668	23.40	65.7759	48.2834	P
2014:04:19	13:28:00	-6.75470	155.024	43.40	41.2931	123.913	SKS
2014:04:19	20:54:42	-20.0283	-70.9196	10.0	356.942	65.4402	P,S
2014:04:24	03:10:10	49.6388	-127.732	10.0	70.9985	37.4444	P,S

Table C.2

This table gives details of all stations used in the analysis of data from southeast Canada (Chapter 3), excluding the QM-III network and western Superior stations (see sections 5.2.2, 5.2.2).

Table C.2: Location and details of seismic stations used in the relative arrival-time study of Chapter 3.

Netwrk.	Sta.	Lat.	Long.	Elev. (Km)	Inst. Type.
CN	A11	47.2425	-70.1978	0.061	BB
CN	A16	47.4706	-70.0064	0.015	BB
CN	A21	47.7036	-69.6897	0.046	BB
CN	A54	47.4567	-70.4125	0.381	BB
CN	A61	47.6935	-70.0914	0.393	BB
CN	A64	47.8264	-69.8922	0.137	BB
CN	BATG	47.2767	-66.0599	0.336	BB
CN	CHEG	46.8084	-60.6745	0.446	BB
CN	CNQ	49.3022	-68.0744	0.200	SP
CN	CRLO	46.0375	-77.3801	0.168	SP
CN	DAQ	47.9644	-71.2425	0.939	SP
CN	DPQ	46.6804	-72.7774	0.167	BB
CN	DRLN	49.256	-57.5042	0.238	BB
CN	EEO	46.6411	-79.0733	0.398	SP
CN	GAC	45.7033	-75.4783	0.062	BB
CN	GBN	45.4077	-61.5128	0.038	BB
CN	GGN	45.117	-66.822	0.030	BB
CN	GRQ	46.6067	-75.8600	0.290	SP
CN	GSQ	48.9142	-67.1106	0.398	SP
CN	HAL	44.6377	-63.592	0.064	BB
CN	ICQ	49.5217	-67.2719	0.058	BB
CN	KAPO	49.4504	-82.5079	0.210	BB
CN	KGNO	44.2272	-76.4934	0.089	BB
CN	KILO	48.4972	-79.7232	0.314	BB
CN	LG4Q	53.6269	-74.0972	0.168	SP
CN	LMN	45.852	-64.806	0.363	BB
CN	LMQ	47.5485	-70.3258	0.429	BB

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Table C.2: *continued*

Netwrk.	Sta.	Lat.	Long.	Elev. (Km)	Inst. Type.
CN	MALO	50.0244	-79.7635	0.271	BB
CN	MNQ	50.5333	-68.7744	0.564	SP
CN	MNT	45.5003	-73.6228	0.112	BB
CN	MNTQ	45.5003	-73.6228	0.112	BB
CN	MOQ	45.3120	-72.2541	0.841	SP
CN	NATG	50.2872	-62.8102	0.000	BB
CN	OTT	45.3942	-75.7167	0.077	BB
CN	SADO	44.7694	-79.1417	0.243	BB
CN	SCHQ	54.8324	-66.8332	0.496	BB
CN	SJNN	47.5952	-52.6781	0.146	BB
CN	SUNO	46.6438	-81.3442	0.343	BB
CN	TRQ	46.2175	-74.5514	0.864	SP
CN	VABQ	45.9047	-75.6079	0.210	BB
CN	VIMO	52.8173	-83.7449	0.078	BB
CN	VLDQ	48.1901	-77.7573	0.093	BB
CN	WBO	45.0003	-75.2750	0.085	SP
CNSN	QCQ	46.7789	-71.2758	0.091	BB
LC	MDV	43.9992	-73.1812	0.134	SP
LD	ACCN	43.3843	-73.6678	0.340	BB
LD	ALLY	41.6492	-80.1448	0.390	BB
LD	BMNY	44.83987	-74.5065	0.115	BB
LD	BRNJ	40.6828	-74.566	0.050	BB
LD	BRNY	41.4141	-74.0119	0.248	BB
LD	CCNY	42.92732	-78.8551	0.194	BB
LD	CPNY	40.7911	-73.9602	0.027	BB
LD	CUNY	40.7335	-73.8178	0.020	BB
LD	FLET	44.72253	-72.95213	0.336	BB
LD	FMPA	40.0478	-76.3208	0.121	BB
LD	FOR	40.8603	-73.8852	0.024	BB

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Table C.2: *continued*

Netwrk.	Sta.	Lat.	Long.	Elev. (Km)	Inst. Type.
LD	FRNY	44.835	-73.5883	0.242	BB
LD	HBVT	44.36306	-73.06326	0.349	BB
LD	HCNY	42.6968	-74.3984	0.273	BB
LD	KSCT	41.72612	-73.48425	0.114	BB
LD	KSPA	41.55702	-75.76815	0.298	BB
LD	LUPA	40.5987	-75.3718	0.255	BB
LD	MCVT	43.95367	-72.99175	0.439	BB
LD	MMNY	42.7319	-77.9066	0.241	BB
LD	MSNJ	40.8841	-74.1815	0.132	BB
LD	NCB	43.9734	-74.2228	0.575	BB
LD	NPNY	41.7546	-74.1435	0.216	BB
LD	ODNJ	41.0829	-74.6056	0.187	BB
LD	PAL	41.0056	-73.9079	0.066	BB
LD	PANJ	40.3769	-74.7029	0.100	BB
LD	POTS	44.6634	-74.9732	0.112	BB
LD	PRNY	42.4666	-76.5361	0.205	BB
LD	TUPA	40.16625	-75.18642	0.112	BB
LD	UNH	43.09265	-70.86503	0.005	BB
LD	WCNY	43.981	-75.6549	0.245	BB
LD	WVNY	42.4062	-78.6042	0.490	BB
NESN	FFD	43.4701	-71.6533	0.131	BB
NESN	PQI	46.6710	-68.0168	0.175	BB
NESN	WVL	44.5648	-69.6575	0.085	BB
PE	PAGS	40.227766	-76.722122	0.120	BB
PE	PSDB	41.127333	-78.753144	0.437	BB
PE	PSWB	41.305475	-76.015183	0.398	BB
PE	UPAO	40.48206	-80.021592	0.274	BB
PE	WRPS	40.7941	-77.8653	0.320	BB
PO	ACTO	43.6087	-80.0624	0.360	BB

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Table C.2: *continued*

Netwrk.	Sta.	Lat.	Long.	Elev. (Km)	Inst. Type.
PO	AKVQ	60.80800	-78.19120	0.170	BB
PO	ALFO	45.6283	-74.8842	0.000	BB
PO	ALGO	45.9544	-78.0509	0.235	BB
PO	BANO	45.0198	-77.928	0.360	BB
PO	BASO	44.01331	-81.66482	0.210	BB
PO	BELQ	47.398	-78.6874	0.355	BB
PO	BMRO	44.59519	-81.21742	0.217	BB
PO	BRCO	44.2437	-81.4423	0.273	BB
PO	BUKO	45.4423	-79.3989	0.317	BB
PO	BWLO	44.11741	-81.13824	0.255	BB
PO	CHGQ	49.9105	-74.3748	0.406	BB
PO	CLWO	44.4491	-80.3013	0.485	BB
PO	DELO	44.5177	-77.6186	0.213	BB
PO	DMCQ	48.9646	-72.068	0.197	BB
PO	DRCO	43.8766	-78.7023	0.084	BB
PO	DREO	43.87244	-78.70396	0.094	BB
PO	DRWO	43.871897	-78.729983	0.059	BB
PO	ELFO	43.193	-81.3163	0.298	BB
PO	HANN	45.8818	-66.7677	0.102	BB
PO	INUQ	58.4509	-78.1186	0.012	BB
PO	KAJQ	58.694	-65.9305	0.046	BB
PO	KB1O	45.3981	-76.1768	0.108	BB
PO	KB3O	45.3783	-76.1509	0.095	BB
PO	KB4O	45.3743	-76.1633	0.102	BB
PO	KB5O	45.3931	-76.1453	0.098	BB
PO	KB6O	45.3905	-76.1668	0.090	BB
PO	KJKQ	55.2769	-77.7455	0.015	BB
PO	KLBO	45.3566	-80.2132	0.177	BB
PO	KNGQ	61.5849	-71.9483	0.054	BB

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Table C.2: *continued*

Netwrk.	Sta.	Lat.	Long.	Elev. (Km)	Inst. Type.
PO	KRSQ	60.0220	-69.9911	0.132	BB
PO	LATQ	47.3835	-72.7819	0.163	BB
PO	LSQQ	49.058	-76.9796	0.308	BB
PO	MATQ	49.7589	-77.6376	0.280	BB
PO	MEDO	43.1646	-78.4546	0.191	BB
PO	MKVL	55.0923	-59.184	0.079	BB
PO	NANL	56.5371	-61.6884	0.034	BB
PO	NMSQ	51.7133	-76.0237	0.275	BB
PO	NWRL	53.5253	-60.1424	0.006	BB
PO	ORHO	45.4563	-75.5367	0.051	BB
PO	ORIO	45.4515	-75.511	0.074	BB
PO	PECO	43.934	-76.9939	0.092	BB
PO	PEMO	45.6773	-77.2466	0.180	BB
PO	PKRO	43.9643	-79.0714	0.197	BB
PO	PLIO	41.7505	-82.6284	0.143	BB
PO	PLVO	45.0396	-77.0754	0.279	BB
PO	STCO	43.2096	-79.1705	0.096	BB
PO	TOBO	45.2257	-81.5234	0.169	BB
PO	TORO	43.6136	-79.3433	0.080	BB
PO	TYNO	43.095	-79.8702	0.205	BB
PO	WBHL	52.903	-66.8663	0.595	BB
PO	WEMQ	53.0535	-77.9737	0.172	BB
PO	WLVO	43.9236	-78.397	0.070	BB
PO	YOSQ	52.8666	-72.1998	0.649	BB
POLARIS	HSMO	47.3708	-79.6657	0.306	BB
POLARIS	MRHQ	45.8870	-74.2127	0.422	BB
POLARIS	MSNO	51.2913	-80.6113	0.004	BB
POLARIS	OTRO	50.1818	-81.6286	0.109	BB
POLARIS	RSPO	46.0734	-79.7602	0.264	BB

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Table C.2: *continued*

Netwrk.	Sta.	Lat.	Long.	Elev. (Km)	Inst. Type.
POLARIS	TIMO	48.4659	-81.3032	0.392	BB
POLARIS-UQAM	NEMQ	51.6837	-76.2576	0.197	BB
US	PKME	45.2644	-69.8917	0.109	BB
X8	EDEY	45.4409	-62.3190	0.139	BB
X8	HOLY	46.5259	-66.4640	0.193	BB
X8	JOSY	44.2636	-66.0128	0.076	BB
X8	MALY	45.7935	-66.3551	0.033	BB
X8	PENY	44.4084	-64.2694	0.014	BB
X8	SHEY	45.1324	-61.9859	0.009	BB
X8	SUSY	45.7174	-65.4320	0.099	BB
X8	WODY	45.0983	-64.6629	0.038	BB

Appendix D

AARM user guide

From relative to absolute teleseismic travel-times: the
Absolute Arrival-time Recovery Method (AARM)

AARM code user guide

The published manuscript

This user guide provides support for the absolute arrival-time recovery method (AARM) code distributed upon request used in the following manuscript, when using the code please cite the work as below:

Boyce A, Bastow I.B, Rondenay S and Van der Hilst R.D (2017) From relative to absolute teleseismic travel-times: the Absolute Arrival-time Recovery Method (AARM), Bulletin of the Seismological Society of America 107(5), pp 2511-2520, doi:10.1785/0120170021

The manuscript describes the methodology (Figure D.1) for recovery of absolute delay-times from relative arrival-time datasets found by using the multichannel cross-correlation code of *VanDecar and Crosson* (1990). Picks are compared to those catalogued by the International Seismological Centre (ISC) (*Di Giacomo et al.*, 2014; *International Seismological Centre*, 2016). However it can also be

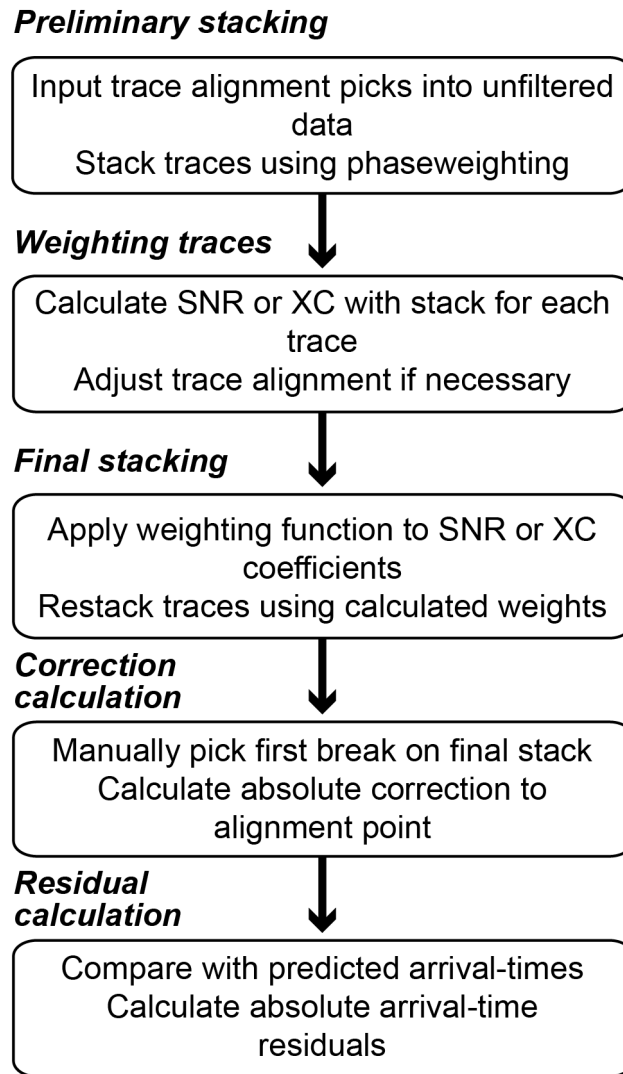


Figure D.1: **A summary of AARM**, the recovery of absolute arrival-times using a relative arrival-time dataset.

generalised easily for use following other trace alignment methods (e.g. *Rawlinson and Kennett, 2004*).

Before we start - required codes and formats

A variety of seismological tools are required for use of the AARM code. Due to bugs within my current versions of MacSac (10.6d-grh110/116) and IRIS SAC for linux (101.5c) I use both versions within the code written to work in the terminal in Mac OS X 10 Yosemite, El Capitan, Sierra etc. In instances where the variable `$mac` is used this is just to enable the codes to work across both laptop and desktop without changing the paths.

The data I use is initially processed to calculate accurate relative arrival-times using the Multichannel

cross correlation method of *VanDecar and Crosson* (1990) and the predicted arrival-times of the ak135 (*Kennett et al.*, 1995) using the TauP Toolkit (*Crotwell and Ritsema*, 1999). Please refer to *Boyce et al.* (2016) for further detail of this workflow (such as filtering parameters). However it should be easily achievable to generalise this method for use following another relative arrival-time determination scheme such as *Rawlinson and Kennett* (2004).

The codes I use in this toolkit are listed below:

- MACSAC - Version 10.6d-grh110/116 (www1.gly.bris.ac.uk/george/sac-download.html)
- IRIS SAC - Version 101.5c (ds.iris.edu/ds/nodes/dmc/software/downloads/sac/)
- The Generic Mapping Tools, Version 5.1.1 (r12968) - (gmt.soest.hawaii.edu/projects/gmt/wiki/Download)
- Sac miscellaneous tools:
 1. `sac1st` (geophysics.eas.gatech.edu/classes/SAC/)
 2. `sactosac` (www1.gly.bris.ac.uk/george/sac-bugs.html)
 3. `sac2xy` (web.utah.edu/thorne/software.html)

The AARM code outlined here accepts SAC data files in the following format: `YYYYMMDDHHMMSS_NETWORK_STATION.00.EHZ`, which corresponds to: `YEARMONTHDAYHOURMINUTESECOND_NETWORKCODE_STATIONCODE.LOCATION.COMPONENT`.

The component EHZ is used as instrument response is removed prior to the analysis to normalise all data to the shortest period response on the network (see *Boyce et al.*, 2016). The AARM code can be changed relatively easily to accept other component choices. Event directories containing these SAC files are also assumed to be named using the YYYMMDDHHMMSS 14 character format. The notable sac header placements are T0 for relative arrival-time derived alignment points (output of MCCC in my case (*VanDecar and Crosson*, 1990) and T1 for ak135 predicted arrivals. Note that these alignment points must be present in the unfiltered, instrument response corrected input data for this code.

Getting started

First unzip the toolkit:

```
>>tar -zxvf Absolute_arrival-time_toolkit.tar.gz
```

Move the contents of the resulting directory `Absolute_arrival-time_toolkit` to the location of your choice. The main script `calc_arr_times.sh` is written to work in the directory containing all event directories (YYYYMMDDHHMMSS) but others can be placed elsewhere. Open all scripts (`*.sh`) and edit all lines beginning with `/Users/` as these refer to your specific paths - make sure these are correct first! The SAC macro `normalize.m` must be placed within a directory which both your versions of SAC can find it. Also check the correct bash profile is being sourced in the first few lines. Finally make all script files executable using

```
>> chmod u+x *.sh
```

ISC reference picks

Within the directory `ISC_FILES`, the script (`TT_calculator_ISC.sh`) can be used to convert a file in ISC download format (e.g. `ISC_picks_2007-2014.txt`) using python to the format readable by the main script here (e.g. `ISC_reference_picks.txt`). This also uses a standard JWEED format event file for reference (e.g. `correct_evts_304.txt`) which are the events previously processed to obtain relative arrival-times. If this step is not done the ISC pick check part will be skipped.

Weighting function

Open the script `weighting_function.sh` and select the desired function by changing the parameter `$FUNC_NUM`. The default is 1, i.e. a linear weighting. As mentioned within the main manuscript, our testing showed that adding a non-linear weighting function to the final stack did not show any visible improvement and thus a linear weighting function is used. The functionality is left in here to aid in future applications of the method where weighting is found to be useful.

Plotting

In `Make_plots_Dataset.sh` edit the line `Dataset_name=` to reflect your own input dataset. This is used in the title and file name only. The script `Make_plots_Event.sh` will only produce a subplot of comparison with ISC picks if the correct files are in place following calculation in `calc_arr_times.sh`.

Quality control measures

The four quality control (QC) parameters can be found near the top of *calc_arr_times.sh*. These values cause the AARM code to notify the user that an individual file exceeds a particular QC parameter at the end of the implementation. Each QC parameter is explained below, please adjust in your own implementation as you see fit.

- **ISC_CUTOFF** - The maximum allowable difference between available ISC picks and those generated using the AARM code.
- **SNR_CUTOFF** - The lowest signal-to-noise ratio for an individual trace that is allowed to contribute to the stack. Traces with SNR values lower than this are assumed not to give reliable arrival-time measurements.
- **XC_CUTOFF** - The maximum allowable offset (in seconds) from zero in the cross correlation of each trace with the final stack. If an individual trace has a greater **XC_CUTOFF** it is not contributing constructively to a sharp pulse onset and should thus be removed.
- **AUTO_CUTOFF** - The maximum allowable autocorrelation-estimated pick error, traces with pick errors greater than this can be easily removed.

Using the code

Initiating the code

Your data should now be ready to use with the AARM code. As explained in the main manuscript (*Boyce et al., 2017*) there are two methodologies for weighting the individual traces in the final stack: Cross correlation (XC) and signal-to-noise (NOISE). The choice is input as the first argument (\$1) to the code. The second argument (\$2) indicates whether the plotting scripts for each event and the whole dataset are to be called (YES/NO). An optional third argument (\$3) enables the user to specify a rerun of the code (RERUN) for only the selection of events that contained traces that were removed in QC. These events will be listed in *bad_events.txt* after an initial run. Example implementations are shown below:

```
>> calc_arr_times.sh <SCHEME> <PLOTTING> <RERUN>
```

```
>> calc_arr_times.sh XC NO
>> calc_arr_times.sh NOISE YES RERUN
```

Manual input

For each event one user input is required. This is picking the first breaking onset time on the final stack. Once initiated the AARM code will run for the first event and sac will initiate an X11 window showing the final stack for the first event. The user is then required to pick the onset time. To do this simply hover the mouse over the X11 window in the desired location and press “a” on the keyboard:

```
>> a
```

You can adjust your pick by pressing **a** in a different location if required. As with viewing any files in SAC using **ppk** it is possible to zoom in and out using **x** and **o**. When you are happy with the pick, press “n” on the keyboard to move the script on to the next event:

```
>> n
```

The AARM code will then calculate the absolute arrival-times and produce the plot for the first event if this function is enabled. The process then repeats for the remaining events until the dataset plot is formed and the code exits.

Outputs

To terminal

The code outputs text to the terminal window indicating the files which have failed the quality control criteria in section 3.4. The files to be removed are contained within the file *bad_files.txt* within the directories found in *bad_events.txt*. Use the command:

```
>> while read line; do 'echo $line'; done<bad_files.txt
```

The files that failed the QC criteria will then be zipped and the analysis can be rerun, but only for the events which have had traces removed. To do this, rerun your initial implementation but add “RERUN” as the third argument in the command line. For example:

```
>> calc_arr_times.sh NOISE YES RERUN
```

To .sac file

There are three SAC file outputs from the AARM code for each event:

- *stack.sac* - The initial unweighted stack.
- *stack2.sac* - The final weighted stack (on which the onset time is picked).
- *stack2_auto.sac* - The autocorrelation of the final stack.

To .txt file

The AARM code outputs ten textfiles to record statistics, calculations and most importantly the absolute arrival-time residuals. Many of these are used by the plotting scripts.

- *auto_corr_errors2.txt* - The autocorrelation-estimated picking error.
- *Conv_stack-ISC_picks.txt* - The difference between calculated absolute arrival-times (from AARM) and those derived from the ISC (*Di Giacomo et al.*, 2014; *International Seismological Centre*, 2016). **Convention:** negative if calculated arrival-time is less than the ISC time.
- *correlation_sort2.txt* - The cross correlation coefficients between the stack and each trace and the offset in XC maximum from zero, sorted by XC max offset.
- *correlation_weightings.txt* - Record of calculation of stack weights for individual traces.
- *d.txt* - Absolute arrival-time residuals (s), arrival-time compared to ak135. **Convention:** negative if calculated arrival-time (from AARM) is less than predicted.
- *SNR.txt* - Estimated SNR for each trace contributing to the stack.
- *SNR_pick_error.txt* - Mean SNR and mean autocorrelation error for use in plotting.
- *stack2_SNR.txt* - Estimated SNR of the final stack.

- *TT_calc_results.txt* - Record of arrival-time calculations for each trace. Columns are: Filename, omarker, rel-arr alignment, Predicted arrival, absolute arrival-time, predicted arrival-time, absolute arrival-time residual.
- *XC_means2_.txt* - Mean SNR and mean cross correlation coefficient for use in plotting.

To postscript file

The individual plot for each event will be named as `YYYYMMDDHHMMSS.ps`. The dataset plot will be contained within the directory above, from which the main code is run (`$Dataset_name.ps`).

Example data

The directory `EXAMPLE_DATA` contains sample data used in the study of *Boyce et al.* (2016). This section outlines the processes required to calculate absolute arrival-times for these two events 20071008171037, 20090809105556. Navigate into the directory containing the example data and move the main script to that location:

```
>> cd EXAMPLE_DATA
>> mv ../calc_arr_times.sh .
```

Start the AARM code using either XC or NOISE weighting schemes then press **a** followed by **n** when the final stack appears for each event (two in this case):

```
>> calc_arr_times.sh XC YES
a
n
a
n
```

You will notice (depending on parameters used) that both directories (20071008171037 and 20090809105556) may have some traces that do not pass the default QC criteria as shown in the terminal window. These can be removed (zipped, to prevent further analysis) using:

```
>> while read line; do 'echo $line'; done<bad_files.txt
```

Then the AARM code can be rerun using your initial parameters but only operating on the events for which traces have been removed from the analysis:

```
>> calc_arr_times.sh NOISE YES RERUN
```

You will notice that the AARM code only reruns the analysis for the earthquakes that have bad individual traces, this is to save time when processing larger datasets. Now all data pass the QC criteria as the file *bad_file.txt* is not displayed as it is empty.

Go ahead and take a look at the text, SAC and postscript files produced by the code. You may need to adjust the X-Y scaling on some plots to make the data visible depending on dataset size.

Explanation of sections within each code - detailed comments

Calculate arrival-times - Shell Script - C1

```
calc_arr_times.sh
```

- C1.1 - Specifies path locations, removes files from a previous implementation, sets quality control limits and checks code has been called correctly by the user. The event loop criteria is also setup here, either all events, individual events or a rerun.
- C1.2 - Loop is started for each event in the \$file_list, old files removed, saclst function is used to read Omarker, T0, T1 headers from original sac files (*.EHZ)
- C1.3 - Prepare files before stacking: (MacSac) traces are cut around their alignment point to 60s (P-waves) in length (change to 120s for S-waves), synchronised (optionally integrated to displacement) and saved as .stk files These traces are then normalised using the normalize.m macro (Files *.stk are overwritten).
- C1.4 - (MacSac) read in normalised files from previous step and calculate RMS value for 25s noise and signal windows. Write out .EHZ.stk.s files. Read *.EHZ.stk files use signal stacking subprocesses within MacSac to form a phasewighted stack. Normalize the output stack.

- C1.5 - (Linux Sac) read the stack and each trace (**.EHZ.stk*). Correlate each trace with the stack throughout a 20 s window. Store time of min and max in T8 T9 headers, write **.EHZ.stk.corr*. Use SAC headers User0 and User5 to calculate estimate for SNR for each trace - output to file *SNR.txt*.
- C1.6 - Use preset trace SNR cutoff from C1.1 to copy name of low SNR traces to *bad_files.txt* this is used later to remove traces in a rerun of the script.
- C1.7 - Collect cross correlation coefficients and location of XC maximum. This is used for weighting and calculating any shifts that can be used to improve the alignment. Use GMT to compute mean of XC coefficients.
- C1.8 - When the noise weighting scheme is used: Calculate normalised SNR values, paste to *SNR_norm.out*. Use the normalised SNR values as input into weighting function (default is a linear distribution) (*weight_function.sh*). These weightings are then written to a SAC macro (*addstack.m*) to be used in the second phaseweight stack.
- C1.9 - When the XC weighting scheme is used: Calculate normalised XC values, paste to *correlation_norm.out*. Use the normalised XC values as input into weighting function (default is a linear distribution) (*weight_function.sh*). These weightings are then written to a SAC macro (*addstack.m*) to be used in the second phaseweight stack. Use the T9 header from *correlation.out* to adjust the T0 header in the files **.EHZ* thus slightly adjusting the alignments where necessary (*correct_T0.m*). Normalise and write out SAC macro *norm_stk2.m*. Use the above macros to new T0 adjusted files. Output normalised updated files for stacking **.stk2*.
- C1.10 - (MacSac) Use signal stacking subprocesses on **.stk2* add files to stack using weightings in *addstack.m*. Stack as before but saving an “improved” second stack (*stack2.sac*) and normalise. Prompt manual picking of first break on second stack (**USE a**). Again input RMS values into headers for SNR calculation.
- C1.11 - (Linux Sac) read the second stack (*stack2.sac*) and each updated trace (**.stk2*). Correlate each trace with the stack throughout a 20 s window. Store time of min and max in T8 T9 headers, write **.stk2.corr2*. Collect new cross correlation coefficients and location of XC maximum. This is sorted by XC maximum offset to give file *correlation_sort2.txt*
- C1.12 - Using preset XC offset maximum (\$XC_CUTOFF) paste files above this limit to *bad_files.txt* by running through *correlation_sort2.txt*.

- C1.13 - On the prior correlation of stack with itself (autocorrelation) in C1.11, mark position of the maximum cross correlation value between the stack and the trace for each station. This will be offset from zero and thus acts as a picking error estimate. Using preset autocorrelation pick error estimate limit (\$AUTO_CUTOFF) paste files above this limit to *bad_files.txt* by running through *auto_corr_errors2.txt*. Use GMT to paste mean trace SNR and mean pick error to *SNR_pick_error.txt*.
- C1.14 - Collect the amarker from *stack2.sac*, this is the correction to be applied T_{corr} in Eq 7 of *Boyce et al. (2017)*. Calculate the SNR for the second stack in *stack2_SNR.txt*. Collect Omarker T0, T1 headers from **.EHZ.new* to enable arrival-time calculations. Calculate arrival-time (TT), predicted arrival-time (ak135) and absolute arrival-time residual following Eq 7 of *Boyce et al. (2017)*.
- C1.15 - For each station check whether an ISC pick is available, calculate the difference between the official pick and the one calculated using this code, paste to *Conv_stack-ISC-picks.txt*. If this is above the preset value (\$ISC_CUTOFF) paste the file name to *bad_files.txt*
- C1.16 - use GMT to calculate statistical parameters for each event. Mean trace SNR and mean cross correlation coefficient for use in plotting.
- C1.17 - Call plotting scripts for individual event and dataset.
- C1.18 - Output the contents of *bad_files.txt* to the terminal. The user can now zip those individual files by calling:

```
>> while read line; do 'echo $line'; done<bad_files.txt
```

The weighting function - Shell script - C2

weight_function.sh

- C2.1 - This specifies input and output locations and checks whether code has been called with correct usage.
- C2.2 - Specify here which weighting function (\$FUNC_NUM) is to be used. Default is 1 as testing found this to be ineffective for our test datasets. The functionality remains for future use.

$$y1 = x \tag{D.1}$$

$$y2 = (\sin((\pi * x) - \pi/2) + 1)/2 \tag{D.2}$$

$$y3 = \exp(-0.5 * (\log(x)).^2) \tag{D.3}$$

$$y4 = \exp(-(\log(x)).^2) \tag{D.4}$$

$$y5 = \exp(-2 * (\log(x)).^2) \tag{D.5}$$

$$y6 = \exp(0.5 * (\log(x)).^3) \tag{D.6}$$

$$y7 = \exp((\log(x)).^3) \tag{D.7}$$

$$y8 = \exp(2 * (\log(x)).^3) \tag{D.8}$$

$$y9 = -((x - 1).^4) + 1 \tag{D.9}$$

$$\tag{D.10}$$

Produce summary plot for each event - GMT 5 Shell script - C3

`Make_plots_Event.sh`

- C3.1 - Sets up plotting standards, outlines required files and formats and specifies their location.
- C3.2 - Plots weightings for all traces along the chosen weighting function, displays % of traces with weighting > 0.6 (lower left).
- C3.3 - Uses *sac2xy* code to plot primary stack (blue) and secondary, weighted stack (red) and final stack SNR (upper left).
- C3.4 - Plots histogram of absolute arrival-time residuals (upper middle).
- C3.5 - Plots distribution of Absolute arrival-time residuals with respect to epicentral distance. Station names are plotted below (lower middle).
- C3.6 - Plots a histogram of autocorrelation errors (upper right).
- C3.7 - Plots the difference in residual between code-derived pick and the ISC pick for available stations. This will not appear in cases where picks are not available. The correct file must have the appropriate format and be specified in this section under `$ISC_FILE`.

Produce dataset plot for all events - GMT 5 Shell Script - C4

`Make_plots_Dataset.sh`

- C4.1 - Sets up plotting standards, outlines required files and formats and specifies their location.
- C4.2 - Plots absolute arrival-time residual distribution for the entire dataset including some statistics (lower left).
- C4.3 - Plots histogram and statistics for autocorrelation errors for the entire dataset (upper left).
- C4.4 - Plots the difference in residual between code-derived pick and the ISC pick for available stations for the entire dataset (upper middle).
- C4.5 - Plots the mean autocorrelation error for each event with the mean trace SNR ration from each event (lower middle).
- C4.6 - Shows the mean cross-correlation with of each trace with the stack against the mean trace SNR (lower right).
- C4.7 - Plots the SNR of the stack against the mean trace SNR (upper right).
- C4.8 - Computes statistics of the dataset and prints to terminal output.

Normalises individual sac files so the maximum amplitude is 1. - Sac macro - C5

`normalize.m`

Appendix E

Copyright

05/24/18

Dear Seismological Society of America,

I am completing my PhD thesis at Imperial College London entitled 'The Proterozoic building of North America: insights from broadband seismic tomography'.

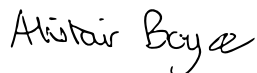
I seek your permission to reprint, in my thesis, the *accepted* version of a paper I published in: Boyce, A., I. D. Bastow, S. Rondenay, and R. Van der Hilst (2017), From relative to absolute teleseismic travel-times: the Absolute Arrival-time Recovery Method (AARM), Bull. Seis. Soc. Am., 107(5), 2511–2520, doi: 10.1785/0120170021.

I would like to include the paper in my thesis which will be added to Spiral, Imperial's institutional repository <http://spiral.imperial.ac.uk/>.

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Yours sincerely,

Alistair Boyce



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Signed:



Name:

Kathleen Gaffney

Organisation: Seismological Society of America

Job title: Publications Director

Date: May 30, 2018