New Constraints on Extensional Environments
through Analysis of Teleseisms

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ABSTRACT

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We apply a variety of teleseismic methodologies to investigate the upper mantle structure in extensional environments. Using a body wave dataset collected from a regional deployment in the Woodlark Rift, Papua New Guinea, we image anisotropic velocity structure of a rapidly extending rift on the cusp of continental breakup. In the process, we develop a technique for azimuthal anisotropy tomography that is generally applicable to regions of relatively simple anisotropic structure. The Cascadia Initiative ocean bottom seismometer (OBS) deployment provides coverage of an entire oceanic plate in unprecedented detail; we measure attenuation and velocities of teleseisms to characterize the temperature and melt structure from ridge to trench.

Our study of shear wave splitting reveals strong azimuthal anisotropy within the Woodlark Rift with fairly uniform fast directions parallel to extension. This observation differs markedly from other continental rifts and resembles the pattern seen at mid-ocean ridges. This phenomenon is best explained by extension-related strain causing preferential alignment of mantle olivine. We develop a simple relationship that links total extension to predicted splitting, and show that it explains the apparent dichotomy in rifts’ anisotropy.

Finite frequency tomography using a dataset of teleseismic $P$- and $S$-wave differential travel times reveals the upper mantle velocity structure of the Woodlark Rift. A well developed slow rift axis extending $>250$ km along strike from the adjacent seafloor spreading centers demonstrates the removal of mantle lithosphere prior to complete crustal breakup. We argue that the majority of this rift is melt-poor, in agreement with geochemical results. A large temperature gradient arises from
the juxtaposition of upwelled axial asthenosphere with a previously unidentified cold structure north of the rift that hosts well located intermediate depth earthquakes. Localization of upper mantle extension is apparent from the velocity structure of the rift axis and may result from the presence of water following recent subduction.

In order to resolve potential tradeoffs between anisotropy and velocity gradients, we develop a novel technique for the joint inversion of $\Delta V_S$ and strength of azimuthal anisotropy using teleseismic direct $S$-waves. This approach exploits the natural geometry of the regional tectonics and the relative consistency of observed splits; the imposed orthogonality of anisotropic structure takes care of the non-commutative nature of multi-layer splitting. Our tomographic models reveal the breakup of continental lithosphere in the anisotropy signal, as pre-existing fabric breaks apart and is replaced by upwelling asthenosphere that simultaneously advects and accrues an extension-related fabric. Accounting for anisotropy removes apparent noise in isotropic travel times and clarifies the velocity model. Taken together, our results paint a detailed and consistent picture of a highly extended continental rift.

Finally, we collect a dataset of differential travel time ($\delta T$) and attenuation ($\Delta t^*$) measurements of $P$- and $S$-waves recorded on OBS stations that span the Juan de Fuca and Gorda plates. We observe large gradients in $\Delta t^*$, with values as high as 2.0 s for $S$-waves at the ridge axes. Such high values of differential attenuation are not compatible with a purely thermal control, nor are they consistent with focusing effects. We assert that melt, grain size, and water enhance anelastic effects beneath the ridge. The combination of attenuation and velocity measurements enables us to place quantitative constraints on the properties of the upper mantle in the vicinity of the spreading axis.
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Introduction

Continental rifts lie on a spectrum between incipient extension and fully fledged mid-ocean ridge spreading. The maturation of divergent systems entails rheological, thermal, and compositional changes, the production of melt, crustal thinning, and the accrual of finite strains (with associated fabric development) as crust and mantle lithosphere thin and asthenosphere upwells.

Despite the fundamental importance of rifting to well understood plate tectonic cycles, there are several facets of the rifting process that we do not fully understand. The relative thinning rate of crust versus mantle lithosphere underpins the dynamics of extension. While simple models and kinematic constraints predict homogeneously extended continental lithosphere [McKenzie, 1978], this story is belied by observations of active rifts and passive margins that indicate preferential removal of mantle lithosphere [Brune et al., 2014; Driscoll and Karner, 1998; Keranen et al., 2009; Kington and Goodliffe, 2008; Lavier and Manatschal, 2006; Mohn et al., 2012; Rychert et al., 2012]. The role of magmatism differs between rifts [Bastow and Keir, 2011; Maccarferri et al., 2014; Muirhead et al., 2015]; the necessity of melt for weakening plates and localizing extension is conjectured but not well established [Buck, 2006; Havlin et al., 2013]. Asthenospheric flow patterns must evolve from small-scale divergence and upwelling in incipient rifts to the corner flow regime that underlies mid-ocean ridges [McKenzie and Sclater, 1969]. The efficiency with which this flow is established has not been previously constrained.

Seismological observations allow us to probe the deep structure of extensional
environments. Travel time tomography provides us with 3-D images of rifts’ interiors and the structure of spreading centers. We then rely on experimentally derived scaling relationships to translate inferred wave speeds ($V_P$, $V_S$) to tectonophysically relevant parameters like temperature, melt fraction, grainsize, and composition. Seismicity provides a proxy for the locus and depth extent of shallow, brittle deformation [Abers et al., 2016; Keir et al., 2006; Scholz, 1988; Weller et al., 2012], which itself relates to plate strength [Burov, 2010; Jackson et al., 2008]. Scattered phases (receiver functions) pick out sharp lithological boundaries, permitting high-quality mapping of crustal thinning. Seismic anisotropy depends on the time-integrated strain history (and/or current stress state and melt content) of mantle rocks. Thus, techniques such as shear wave splitting can reveal rifts’ temporal evolution and dynamic behavior. Anelastic behavior of rocks depends strongly on a variety of rheologically critical parameters; seismic attenuation observations offer a powerful complement to estimates of wave speed for discriminating the thermodynamic state of the Earth’s interior.

In this thesis, I incorporate each of these techniques in a multi-faceted examination of divergent tectonic environments. By applying multiple seismic measurements in concert, I attempt to resolve tradeoffs and arrive at a more complete understanding of upper mantle structure within, and adjacent to, the locus of spreading.

The inherently transient nature of rift systems, combined with their relative geographic scarcity, means that there are few accessible examples of continental breakup caught in media re. As such, geoscientists often rely on a few canonical examples to answer fundamental questions: the East Africa Rift, the Basin and Range Province, the Rio Grande Rift, the Baikal Rift, the Gulf of California, and select back-arc systems. The Woodlark Rift, in southeastern Papua New Guinea is one of the youngest (<8.4 Ma) and most rapidly extending (∼30 mm/yr) zones of rifting continent [Taylor et al., 1999]. The remarkable extension rates, coupled with a ∼1000 km along-strike
gradient in divergence [Goodliffe and Taylor, 2007], makes this rift uniquely suitable for studies of the transition from rift to drift.

The CDPAPUA project is part of the multi-institutional Continental Dynamics program to study the relationship between rifting and high/ultra-high pressure (HP/UHP) rock exhumation [Eilon et al., 2014; Ellis et al., 2011; Gordon et al., 2012; Little et al., 2011; Wallace et al., 2014; Zirakparvar et al., 2014]. In 2010-2011, scientists from LDEO deployed a temporary amphibious broadband array around the D’Entrecasteaux Islands and Papuan Peninsula, spanning the region of maximally extended continental crust within the Woodlark Rift. We sought to understand the along-strike character of lithospheric deformation adjacent to propagating spreading centers, the relationship between limited surficial volcanism and deeper structures, the process of ongoing gneiss dome exhumation, and the mechanism that produced the world’s youngest (U)HP rocks [Baldwin et al., 2008].

Mid-oceanic ridges (MORs) represent the ultimate fate of successful rift systems. Extensive mantle melting produces $56 \times 10^{12}$ kg per year of new ocean floor at ridges worldwide. Canonical cooling models robustly describe the macroscopic thermal structure of MORs and aging oceanic plates [McKenzie et al., 2005; Parsons and Sclater, 1977; Ritzwoller et al., 2004; Stein and Stein, 1992]. Such well defined temperature constraints hold promise for calibration of the relationship between seismic parameters (wave speeds, attenuation, anisotropy) and intrinsic thermodynamic variables [Priestley and McKenzie, 2006, 2013]. However, the complex system of silicate melt production, channelization, and infiltration or stagnation close to the ridge axis overprints the simple temperature field. Moreover, the location and extent of melt has salient geodynamical and seismological implications. The Cascadia Initiative’s Amphibious Array provides an unprecedented opportunity to study an entire oceanic

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1 Assuming ocean crust is 7 km thick, the global MOR system is 50,000 km long, average spreading is 50 mm/yr, and crustal density is 3.2 g/cm³.
plate, from MOR to trench, using data from a multi-institutional deployment of ocean bottom broadband seismometers.

Ocean bottom seismometry is a developing field that promises to revolutionize seismology by granting us access to the 70% of Earth’s surface covered by water. My research contributes to recent community efforts to extend and develop the suite of seismic methodologies applicable to ocean bottom seismometer data [e.g. Bell et al., 2014, 2015; Collins et al., 2012; Crawford and Webb, 2000; Janiszewski and Abers, 2015; Webb, 1998; Webb and Crawford, 1999, 2010]; few prior studies had utilised amphibious anisotropy measurements [cf. Bodmer et al., 2015; Karalliyadda et al., 2015; Martin-Short et al., 2015; Pozgay et al., 2007; Smith et al., 2001], and no other study has used ocean-bottom instruments to measure attenuation of teleseisms.

In Chapter 1, I present a shear wave splitting study that uses core-refracted $SK(K)S$ waves to interrogate seismic anisotropy in the mantle beneath the Woodlark Rift. Having applied and tested several splitting inversion methods to assess the stability of our results (Appendices A and B), we demonstrate that strong anisotropy underlies the Woodlark rift, with a preponderance of fast directions parallel to extension. This observation is unusual for continental rifts but resembles patterns of anisotropy seen at MORs. We conduct a simple exercise linking extension, strain, fabric, and anisotropy to demonstrate that one can predict the CPO\(^2\)-related splitting magnitude at a rift purely as a function of its total extension. We contend that spreading-parallel anisotropy is evidence for olivine CPO that results from high degrees of strain in the underlying mantle; the corollary is that the convecting mantle beneath the rift axis has established a flow field similar to a MOR, despite incomplete crustal breakup.

Building on these ideas, Chapter 2 comprises a teleseismic tomography study, where differential $P$- and $S$-wave travel times are inverted for 3-D structure beneath

\(^2\text{Crystallographic preferred orientation.}\)
our array. We use a finite frequency approach to relate model slowness perturbations to observed time residuals. Our models clearly identify a seismically slow rift axis that extends >250 km along strike from nearby seafloor magmatic centers, and coincides with a variety of crustal features that demarcate the locus of extension. These models also include a seismically fast structure north of the rift axis which contains the first well-located intermediate depth seismicity seen in this region. We discuss a variety of explanations for this cold pod of material at depth. We argue that temperature is the primary determinant of wave speed in this rift, and that large gradients in seismic velocity arise from the juxtaposition of cold, seismogenic material to the north with asthenospheric temperatures (∼1350°C) within the rift axis. This finding is consistent with mature asthenospheric flow field indicated by the splitting results, but implies almost total asthenospheric removal significantly ahead of the oceanic spreading centers and prior to completion of crustal thinning.

Anisotropic effects on travel times can bias velocity models that assume isotropy. Having demonstrated that the Woodlark Rift exhibits large velocity perturbations and substantial anisotropy, we designed an innovative approach to jointly invert for these two parameters (Chapter 3). We exploit the natural geometry of the rift, as well as the relatively simple anisotropy measured in Chapter 1, to fix the anisotropic fast axis, enabling us to parameterize the fast and slow wave speeds of an arbitrarily incident shear wave in terms of fractional anisotropy and mean velocity (Appendices D, E, and F). This method provides otherwise-absent depth constraints on anisotropy. Our tomography reveals that the signature of rifting is present in the anisotropy signal. The implied asthenospheric strain field will help sharpen geodynamical simulations of the rifting process. The mean $V_S$ maps from the anisotropic inversion compare favorably with their isotropic counterparts, and validate the ability of a joint inversion to resolve tradeoffs that otherwise generate artefacts.

Chapter 4 extends our study of divergent environments to a mid-ocean ridge
and oceanic plate, with a focus on the seismological influence of melt at the ridge axis. We measure teleseismic $P$- and $S$-wave travel time residuals and differential attenuation from ridge to coastline using amplitude and phase spectra (Appendix G). The combination of these constraints upon seismic velocity and intrinsic attenuation proves potent for identifying the signature of melt beneath the ridge. Our study implies extremely low values of $Q$$_\mu$(∼25) in the vicinity of the ridge, comparable to the most highly attenuating back-arc settings [Abers et al., 2014; Wei et al., 2014]. We conduct forward modelling to place quantitative bounds on the contribution of melt, water, and grainsize reduction to anelastic processes, with qualitative implications for sub-lithospheric mantle rheology.

Our approach has been to utilise high-quality regional datasets for thorough study of two particular extensional environments, using teleseisms. We show how a multiplicity of datatypes, taken together, provide robust information about the physical state of the Earth’s interior. Not only do these studies advance our knowledge of particular field regions: they contribute to our general understanding of extensional processes.
Note on publication status:
Chapters 1 - 3 have already been published or accepted to peer-reviewed journals and are reproduced here without abridgement. The only edits made to these chapters are updates of references (e.g. from conference abstracts to published papers), and minor alterations for consistency of shorthands (e.g. “DEI” versus “DI”). Unfortunately, these chapters therefore include some necessary recapitulation of tectonic context.

Note on nomenclature:
The multifarious absolute/relative/travel/splitting/attenuation times being measured and discussed for different portions of this thesis make things quite confusing. I have attempted to establish an internally consistent system as follows:

- $\delta \tau$ refers to splitting time: the time separation of ideal fast and slow quasi-shear pulses for a split shear wave.

- $\delta T$ refers to differential travel time: the difference between arrival times at different stations (having accounted for path length differences, etc.). This is also referred to as residual travel time.

- $\Delta t^*$ refers to differential attenuation: deviations in integrated attenuation ($Q^{-1}$ multiplied by travel time) recorded at different stations. The units are in seconds.
Chapter 1

Anisotropy beneath a highly extended continental rift

Co-authors: G. A. Abers, G. Jin., and J. B. Gaherty

This chapter details a $SK(K)S$ splitting study of seismic anisotropy within the Woodlark Rift\(^1\). We relate inferred strain to the magnitude of extension and consider the Woodlark in the context of global rifts. We argue that the underlying mantle has organized ahead of seafloor spreading.

Abstract

We have employed shear wave splitting techniques to image anisotropy beneath the D’Entrecasteaux Islands, in southeastern Papua New Guinea. Our results provide a detailed picture of the extending continent that lies immediately ahead of a propagating mid-ocean ridge tip; we image the transition from continental to oceanic extension. A dense shear wave splitting dataset from a 2010-11 passive-source seismic deployment is analyzed using single- and multichannel methods. Splitting delay times of 1-1.5 s are observed and fast axes of anisotropy trending N-S, parallel to rifting direction, predominate the results. This trend is linked to crystallographic preferred orientation of olivine, primarily in the shallow convecting mantle, driven by up to 200 km of N-S continental extension ahead of the westward-propagating Woodlark Rift. This pattern differs from several other continental rifts that evince rift-strike-parallel fast axes


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and is evident despite the complex recent tectonic history. We contend that
across most of this rift, the unusually high rate and magnitude of extension has
been sufficient to produce a regime change to a mid-ocean-ridge-like mantle
fabric. Stations in the south of our array show more complex splitting that
might be related to melt or to complex inherited structure at the edge of the
extended region.

1.1 Introduction

The ability to constrain ongoing deformation and time-integrated stress fields makes
shear wave splitting an ideal tool to interrogate dynamically evolving environments
such as continental rifts. Several studies \cite{Gao1997, Kendall2005} have applied shear wave splitting analysis to zones of active continental extension,
or back-arc basins \cite[e.g.][]{Smith2001} Most prior studies have focused on slow,
intra-continental rifts that have undergone relatively little (up to a few tens of km)
accumulated extension. By contrast, the Woodlark Rift in southeastern Papua New
Guinea (Figure 1.1) exhibits the full range of extensional regimes, from ocean-floor
spreading in the east to the first stages of continental thinning in the west, at rates
reaching several cm/yr. The D’Entrecasteaux Islands (DEI) lie on relatively thinned
continental crust at the center of this continuum, immediately to the west of the
youngest spreading centers. These islands are cored by topographic domes that expose
ultra-high pressure (UHP) metamorphic rocks at the surface, including the youngest-
known coesite eclogite. Peak UHP metamorphism has been dated to 5-6 Ma \cite{Gordon2012} or 7 Ma \cite{Zirakparvar2011}, implying average uplift rates of up
to 20 km/My (based on pressure estimates from Baldwin et al. [2008]) coeval with
130-200 km of continental extension across these islands \cite{Taylor1999}. We use
the term “continental crust” to mean non-oceanic, differentiated, metamorphosed
crust, with thicknesses and seismic velocities characteristic of the continents, akin to
Several outstanding questions motivate investigation of this area. It is not clear how the unusually high rate and magnitude of extension is accommodated. Different mechanisms have been invoked to explain the exhumation of the UHP rocks, notably the end member models of Rayleigh-Taylor instabilities [Ellis et al., 2011; Little et al., 2011] and low-angle unroofing [Hill et al., 1992; Webb et al., 2008].

Shear wave splitting offers an understanding of time-integrated 3-D flow fields, thereby constraining mantle response to extension, as well as providing a sub-lithospheric framework for models of exhumation and extension. Several continental rifts (e.g. the Main Ethiopian Rift, MER [Kendall et al., 2005]) show significant seismic anisotropy (splitting times ≤2.5 s) with fast axes parallel to the strike of the rift. This is attributed to the strong anisotropic signal of melt within the lithosphere. In magma-poor rifts, where plate stretching accommodates extension, a spreading-parallel lithospheric fabric is expected [e.g. Tommasi et al., 1999], but this signal is likely to be weak due to tradeoffs between strain and thickness [Silver and Chan, 1991]. At mid-ocean ridges, by contrast, large splitting times are observed, with fast axes parallel to the spreading. This anisotropy arises from flow in the asthenospheric mantle that produces lattice preferred orientation (LPO) of individually anisotropic crystals [e.g. Blackman and Kendall, 1997].

Our data provide the most detailed seismological information on the underlying structure of this region to date and our dense shear wave splitting dataset contributes to the relatively few measurements of this sort from an extensional regime on the cusp between continental rifting and oceanic spreading. Our results imply that the unusually high strain rates and magnitudes of extension have created a strong unidirectional fabric at mantle depths that dominates any remnant structure from previous tectonism. This pattern differs from most other continental rifts, where extension and strains are much lower, suggesting that some critical strain has been reached whereby
the mantle fabric aligns with flow.

### 1.2 Tectonic and geologic context

As many workers have detailed [Baldwin et al., 2012; Hill and Hall, 2003; Pegler et al., 1995; Wallace et al., 2004], Papua New Guinea is a unique natural laboratory for understanding tectonic processes. Since the Eocene, the tectonics have been controlled by the oblique convergence of the Pacific and Australian plates at \( \sim 110 \text{ mm/yr} \), accommodated by a broad deformational belt consisting of several microplates [Tregonning et al., 1998; Wallace et al., 2004]. In the late Miocene, the Woodlark microplate (WLK, which probably behaves as a rigid block together with the Solomon Sea plate [Wallace et al., 2004]) began rotating away from the Australian continent (AUS); driven by slab pull at the northern boundary of the Solomon Sea [Wallace et al., 2004; Weisell et al., 1982].

Taylor et al. [1999, Goodliffe and Taylor, 2007 and] used magnetic anomalies in the Woodlark Rift east of 152\(^{\circ}\)E to describe the relative WLK-AUS motion since \( \sim 8.4 \text{ Ma} \) as anti-clockwise rotation about an Euler pole at (9.3\(^{\circ}\)S, 147\(^{\circ}\)E) at -4.022 \(^{\circ}\)/My. New seafloor has been produced diachronously at westward propagating spreading centers since 4-6Ma and the DEI lie just west of the rift tip (151.7\(^{\circ}\)E), along the extrapolated line of the youngest spreading center. Seismic tomography shows an abrupt transition from slow continental crust to fast oceanic crust to the immediate east of the DEI [Ferris et al., 2006]. GPS data shows that the current relative motion between the Trobriand Block and AUS is -2.74 \(^{\circ}\)/My about a pole at (9.43\(^{\circ}\)S, 147.5\(^{\circ}\)E) [Wallace et al., 2014]. The difference in poles results in a \( \sim 30\% \) discrepancy between long-term geologic rates of divergence (\( \sim 30 \text{ mm/yr} \)) and geodetic rates (\( \sim 15-20 \text{ mm/yr} \)) across the DEI [Wallace et al., 2014] possibly due to a tectonic rearrangement at \( \sim 0.52 \text{ Ma} \) [Taylor et al., 1999].
Simple calculations using the geologic poles require \( \sim 190 \) km of extension somewhere at the longitude of Normanby Island (151°E) and \( \sim 140 \) km across Goodenough Island (150°E) [consistent with *Kington and Goodliffe, 2008*] (Figure 1.2). However, the absence of obvious structures that have accommodated this divergence has led to concern over whether the whole WLK block is deforming rigidly at rates predicted by the seafloor anomalies. Left-lateral slip on the Nubara fault (which at present has a right-lateral sense) or significant displacement on the Trobriand Transfer fault [*Little et al., 2007*] could decouple the DEI from the seafloor to the east, making
the above estimates of extension upper bounds, although there is little evidence of large offset on the Trobriand platform where a Trobriand strike-slip fault has been postulated [e.g. Goodliffe et al., 1999]. Extension around the DEI and within the Papuan Peninsula is connoted by ongoing tectonism, including low-angle normal-faulting earthquakes [Abers et al., 1997] on structures that follow steep scarps bounding the islands. Raised terraces along the north coast of the Papuan Peninsula, as well as fluvial incision profiles, confirm that extension is currently accommodated on the Goodenough and Mai’iu fault structures (Figure 1.2) in concert with exhumation of the Suckling-Dayman Massif [Mann et al., 2009; Miller et al., 2012].

The DEI include several topographically dominant gneiss domes [Ollier and Pain, 1980], often called metamorphic core complexes (MCCs) [Davies and Warren, 1988], which contain the youngest-known UHP metamorphic rocks [Baldwin et al., 2004]. The mechanism for dome exhumation is beyond the scope of this paper but there is consensus that it is tied to tectonic extension [e.g. Ellis et al., 2011; Webb et al., 2008]. These domes are bounded by numerous faults and mylonitized shear zones, and are intruded by plutons. Recently, workers have argued that there is a radial pattern to microstructural lineations in the carapace, inferred to represent diapiric exhumation [Little et al., 2011]. Exhumation timescales for these domes are constrained by CA-TIMS dating, which yielded ages of 5.82±0.20 Ma to 4.78±0.17 Ma [Gordon et al., 2012] or 7.1±0.7 Ma [Zirakparvar et al., 2011] for acoesite-bearing eclogite that has since ascended >100km, and put the earliest intrusions at 3.49±0.01 Ma [Gordon et al., 2012]. Clastic sediments in the Trobriand basin sediments to the north indicate that the domes broke sea level ~3 Ma [Francis et al., 1987], while raised coral reefs and stream-profile modeling [Miller et al., 2012], demonstrate ongoing uplift of these edifices since mid-Pliocene to their present elevations of ≤2.5 km.

Abers et al. [2002] used teleseismic P-wave (\(V_P\)) tomography to show a broad low-velocity region beneath the DEI and extending some distance beneath the Woodlark
Basin. The \( V_p \) anomalies are up to 5% slow within the upper mantle and indicate significant lithospheric thinning in a swath from the westernmost spreading center to Normanby Island. Associated upwelling and melting is represented by widespread intrusive and extrusive igneous activity; the DEI contain Pliocene-Pleistocene granodiorite intrusions and Quaternary calk-alkaline extrusive volcanism, with variably deformed leucosome dykes cross-cutting the domes \cite{Gordon et al., 2012}. Historically active volcanic stratovolcanoes lie in a rough E-W line along strike from the present-day spreading centers, and include Mt. Victory and Goropu on the Papuan mainland, and the Walilagi Cones, Iamelele, and the Dawson Strait group on the D’Entrecasteaux themselves. The lavas produced at these volcanic centers have geochemical signatures indicative of a subduction-enriched melt source region, including Ba/La ratios greater than 20 and LREE enrichment \cite{Smith and Johnson, 1981}. The Trobriand arc and the geochemistry suggest that subduction may have been important to recent dynamics, although the eastern volcanics trend to more alkalic compositions with a more rift-like character \cite{Stolz et al., 1993}. Abers et al. \cite{2002} found no evidence for a subducted plate, and a regional tomography study clearly imaging fast New Britain and Solomon slabs shows no fast anomaly associated with the Trobriand arc \cite{Hall and Spakman, 2002}.

### 1.3 Data and methods

#### 1.3.1 Deployment and data

We installed an array of land and ocean-bottom broadband seismic stations over a \(~250\times250\) km area just west of the propagating rift tip, approximately centered on the DEI (Figure 1.2). The seismic deployment comprised 8 ocean-bottom broadband seismographs (Trillium 240 with 240s-corner seismometers, with differential
Figure 1.2: Map showing deployment of broadband seismic stations. Green symbols indicate stations for which shear wave splitting results were obtained, black symbols are stations with no splitting results. Circles are ocean bottom seismometers, point-upward triangles are land stations in this experiment, inverted triangles are stations that were occupied during the 1999-2000 Woodseis experiment (overlapping triangles therefore denote re-occupied sites, BBU, ESA and KIR); gneiss domes/metamorphic core complexes indicated in green; red arrows indicate cumulative vectorial extension calculated over the last 6 Ma, accounting for the changes in rate/pole at 500 ky and 80 ky. Recent GPS results indicate that approximately 10 mm/yr of extension is taking place on structures within Goodenough Basin and the northern coast of the D’Entrecasteaux Islands is moving northwards away from stable Australia at an additional 15 mm/yr [Wallace et al., 2014].

Inset: shear wave splitting results from the Port Moresby GSN station, plotted at lower-hemisphere pierce points. Blue triangles indicate back azimuth of good quality null measurements; red lines indicate measured splitting.

Pressure gauges not used in this analysis) and 31 land-based PASSCAL broadband instruments (Guralp CMG-3T with 120s-corner sensors) that were installed by direct burial. Orientations of all stations were checked (see below) and corrections made where necessary. All instruments recorded at 50 sps, and we did not decimate for the
splitting analysis. The array operated variously between March 2010 and late July 2011, with 86% data recovery, including retrieval of all 8 OBS instruments in January 2011. Several stations had intermittent power problems, and broadband seismic channels for OBS “I” failed, but most stations had sufficient data for splitting analysis. We also used data from several of the stations deployed during the 1999-2000 Woodseis experiment [Abers et al., 2002; Ferris et al., 2006].

### 1.3.2 Station orientation

For the purposes of shear wave splitting it is essential to know the sensor orientation accurately. Ocean-bottom seismometer (OBS) instruments are self-levelling but the azimuths of the horizontal channels depend on how the instrument descends to the seafloor. We determined the orientations of our broadband OBS stations using Rayleigh wave polarization across several teleseismic events [Stachnik et al., 2012], specifically by maximizing the cross correlation at zero lag between the Hilbert transform of the vertical component (to correct for the -90° phase shift involved in the elliptical retrograde particle motion) and the resolved radial component [Baker and Stevens, 2004]. We ascertained correction angles for all seven functioning OBS instruments with confidence bounds (on average ±5°) produced using a non-parametric linear bootstrap, with the balanced resampling method. These uncertainties may be compared favorably to the average uncertainty in land-based station orientation and are on the order of variation in arrival azimuth due to scattering/diffraction along the raypath. Four land stations were also found to have ~10-20° misalignments, probably due to erroneous declination corrections during installation, and were corrected (see supplementary material).
1.3.3 Single-channel splitting methods

For the shear wave splitting analysis we used earthquakes with Mw ≥ 6.25 within the relevant SK(K)S distance window (≈85° to ≈140°), giving a total of 21 earthquakes. The back-azimuthal distribution of these earthquakes was highly non-uniform, dominated by events from the Andean margin in 2010 (Figure 1.3).

The SplitLab software package [Wüstefeld et al., 2008] was used to make individual shear wave measurements, using a bandpass filter of 8-50 s and manually picked windows (approx. 15-25 s long) around the shear wave arrivals. Measurement quality was manually designated ‘good’, ‘fair’, or ‘poor’ on a case-by-case basis, based on signal-to-noise ratio (SNR), minimal associated energy on the vertical component, impulsivity of arrival, minimal energy on horizontal components before the predicted arrival, and small estimated errors in splitting parameters, following standard practise [e.g. Long, 2010]. We discarded traces with SNR < 3, arrivals for which the cross-correlation between the transverse and the time-derivative of the radial was less than 0.6 (following Chevrot [2000]), and likely null measurements for which the estimated fast azimuth was closer than 15° to the measured polarization. We inverted the seismograms for the two splitting parameters, fast azimuth (ϕ) and time delay (δτ), by grid searching through model parameter space for -90 < ϕ ≤ 90° and 0 ≤ δτ ≤ 4.
Figure 1.4: SC measurement for a SKS arrival at station BAYA, on the north coast of Cape Vogel, from a MW 7.2 earthquake was 91° away from the station. Thin lines are 12.5% energy contours spaced evenly between the maximum and minimum, and the shaded region is the 95% confidence region for the energy minimum. The signal had a measured SNR of 16.7 and the traces were bandpass filtered from 8 to 50 seconds. The linearized particle motion and near-identical fast/slow waveforms indicates well resolved splitting. This measurement was designated “good” quality (Section 1.3.3).

We stacked weighted error surfaces over individual measurements, where possible, to improve the robustness of the splitting measurement at each station [Wolfe and Silver, 1998] (Figure 1.5). Picks with qualities of ‘good’, ‘fair’ and ‘poor’ were given weights of 2, 1 and 0, respectively, and weights were multiplied by the measured SNR [Restivo and Helffrich, 1999] based on synthetic tests which showed that fidelity of estimated model parameters to ‘true’ inputs increased with SNR (see supplementary material). Null measurements were included in the stacks with weights scaled down to 10%, to prevent deep minima in null energy surfaces from dominating the stacks. Uncertainties were calculated assuming the errors are $\chi^2$ distributed and using the inverse F-distribution to find $2\sigma$ bounds, with estimates of the degrees of freedom calculated for each trace using the first zero crossing of the autocorrelation function [Yang et al., 1995].
Figure 1.5: Stacked splitting result at station PEMM, with contoured energy surfaces for each individual arrival (peripheral) and the stack (central) from the minimum energy (SC) method [Silver and Chan, 1991]. Relative energy for each plot is shown by 2% contours, scaled to the energy range separately for each plot, and colored from high energy (red) to low energy (blue). Best fit: $dt = 1.68 \pm 0.72$ s and $\phi = -4.0^\circ \pm 8.0^\circ$ (2$\sigma$ errors). Weights (Wt) of individual measurements in the stack are calculated as the product of the measured SNR [after Restivo and Helffrich, 1999] and the quality (1 for fair, 2 for good). Splitting parameter measurements are made at the energy minima (red dots), where splitting time (0-4 s) is indicated by radial distance and fast axis by angle (0-360$^\circ$). 95% error bounds are shown by thick black lines, calculated according to the method of Silver and Chan [1991]. N.b. the splitting model parameters are evenly spaced in $\phi$ and $\delta \tau$ so these are conic projections.

1.3.5 Multi-channel method

To complement the individual results, we use the approach of Chevrot [2000], which measures anisotropy beneath each station in a different sense. This method relies upon the predicted 180$^\circ$ periodicity (a ‘2-$\theta$’ pattern) in the back-azimuthal variation of splitting intensity (SI), defined as $s = -2Tr/|r|^2$ where $r = dR/dt$, the time-
derivative of the radial component, and $T$ is the transverse component seismogram (Figure 1.6). The success and reliability of this technique depends on a wide backazimuthal distribution and so we augmented the $SK(K)S$ data, described above, with direct $S$ arrivals (from $40^\circ$ to $80^\circ$), using only those from deep (>450km) source events in an attempt to minimize the influence of source-side splitting [e.g. Fouch and Fischer, 1996]. This added three earthquakes to our dataset that had clear $S$ arrivals. We calculated the polarization of direct $S$ arrivals by stacking horizontal components across the array and taking the inverse cosine of the first eigenvector of the 2-D covariance matrix [Vidale, 1986]. Arrivals were discarded if measured polarization was too dissimilar (>15°) to the CMT solution prediction or if (for non-nulls) the cross-correlation between the transverse and the time-derivative of the radial was less than 0.7. Although Foley and Long [2011] noted significant source-side splitting from deep events in the Tonga slab, our deep-source direct $S$ measurements agreed well with the $SK(K)S$ results, so were retained. Splitting parameters were calculated using a weighted least-squares fit to the splitting intensity, weighting by the formal uncertainties [Chevrot, 2000] of the measurements.
1.3.6 Method comparison

Some authors [e.g. Long and Van Der Hilst, 2005; Vecsey et al., 2008] have sought to qualitatively compare the performance of the three principal individual-arrival methods (rotation-correlation [Bowman and Ando, 1987], the minimum energy method used here [Silver and Chan, 1991], and eigenvalue minimization [e.g. Silver and Chan, 1991]) for measuring shear wave splitting. In terms of consistency, reliability, and robustness (see supplementary material) we prefer the SC method of the single-channel techniques, in agreement with Vecsey et al. [2008]. In the case of simple structure, we find that the SI method is extremely successful (although it overestimates splitting times in some cases) but note that this technique obscures some of the hallmarks of complicated structure at depth [cf. Silver and Long, 2011].

1.4 Results

1.4.1 Single-channel methods

Splitting parameters were estimated for $SK(K)S$ arrivals at each station. Following the results of our synthetic tests, we use results from our preferred method (SC) that meet criteria defined in Section 1.3.3. We obtained results from the majority of our stations, including all seven functioning OBS. Figure 1.7 shows individual measurements from a representative selection of stations across the array. The results of one station-stack are shown in Figure 1.5 where polar projections display how individual measurements contribute to the stack. All stacked results from the SC method are shown in Figure 1.8, and full results are given in Table 1.1. An analysis of frequency dependency at stations KIR and AGAN showed no significant difference in splitting for bandpass filters of 1-5 s, 5-10 s, and 10-100 s, although splitting measurements were much poorer for the highest-frequency inputs.
Table 1.1: Splitting results for all stations.

<table>
<thead>
<tr>
<th>Station</th>
<th>avSNR</th>
<th>dof</th>
<th>N&lt;sub&gt;SC&lt;/sub&gt;(null)</th>
<th>δτ&lt;sub&gt;SC&lt;/sub&gt;</th>
<th>φ&lt;sub&gt;SC&lt;/sub&gt;</th>
<th>N&lt;sub&gt;SI&lt;/sub&gt;</th>
<th>δτ&lt;sub&gt;SI&lt;/sub&gt;</th>
<th>φ&lt;sub&gt;SI&lt;/sub&gt;</th>
</tr>
</thead>
<tbody>
<tr>
<td>AGAN</td>
<td>7.56</td>
<td>126</td>
<td>9(6)</td>
<td>1.0 ± 0.3s</td>
<td>-44 ± 8</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>BASI</td>
<td>8.12</td>
<td>116</td>
<td>8(5)</td>
<td>2.4 ± 0.4s</td>
<td>-22 ± 3</td>
<td>13</td>
<td>1.4 ± 0.2s</td>
<td>26 ± 6</td>
</tr>
<tr>
<td>BAYA</td>
<td>16.72</td>
<td>9</td>
<td>1(0)</td>
<td>2.1 ± 0.4s</td>
<td>12 ± 3</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>BBU</td>
<td>11.81</td>
<td>63</td>
<td>5(1)</td>
<td>1.2 ± 0.2s</td>
<td>2 ± 8</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>DIOD</td>
<td>8.39</td>
<td>98</td>
<td>7(3)</td>
<td>0.9 ± 0.1s</td>
<td>36 ± 8</td>
<td>11</td>
<td>0.9 ± 0.4s</td>
<td>21 ± 29</td>
</tr>
<tr>
<td>ESA2</td>
<td>9.33</td>
<td>115</td>
<td>7(7)</td>
<td>0.1 ± 0.1s</td>
<td>32 ± 38</td>
<td>5</td>
<td>1.0 ± 2.1s</td>
<td>32 ± 18</td>
</tr>
<tr>
<td>GARU</td>
<td>11.23</td>
<td>26</td>
<td>2(1)</td>
<td>1.2 ± 0.6s</td>
<td>-76 ± 89</td>
<td>7</td>
<td>2.3 ± 1.7s</td>
<td>60 ± 52</td>
</tr>
<tr>
<td>GOGO</td>
<td>10.09</td>
<td>82</td>
<td>6(5)</td>
<td>0.2 ± 0.2s</td>
<td>-58 ± 48</td>
<td>12</td>
<td>1.4 ± 0.9s</td>
<td>57 ± 47</td>
</tr>
<tr>
<td>GUMA</td>
<td>8.08</td>
<td>62</td>
<td>4(3)</td>
<td>1.2 ± 0.4s</td>
<td>-4 ± 17</td>
<td>12</td>
<td>1.4 ± 0.9s</td>
<td>47 ± 47</td>
</tr>
<tr>
<td>IAME</td>
<td>3.85</td>
<td>10</td>
<td>1(0)</td>
<td>3.3 ± 1.2s</td>
<td>-8 ± 18</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>JONE</td>
<td>6.05</td>
<td>69</td>
<td>5(0)</td>
<td>0.5 ± 0.3s</td>
<td>78 ± 89</td>
<td></td>
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<tr>
<td>KAPEP</td>
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<td>54</td>
<td>4(4)</td>
<td>0.6 ± 0.3s</td>
<td>-46 ± 8</td>
<td>12</td>
<td>1.2 ± 0.8s</td>
<td>50 ± 19</td>
</tr>
<tr>
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<td>0.8 ± 2.3s</td>
<td>357 ± 38</td>
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<td>1.0 ± 1.4s</td>
<td>-20 ± 48</td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>KIR</td>
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<td>106</td>
<td>9(0)</td>
<td>1.1 ± 0.1s</td>
<td>-6 ± 6</td>
<td>13</td>
<td>0.6 ± 0.8s</td>
<td>4 ± 34</td>
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<tr>
<td>MAPM</td>
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<td>28</td>
<td>2(2)</td>
<td>3.0 ± 1.0s</td>
<td>-50 ± 3</td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>MAYA</td>
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<td>85</td>
<td>6(4)</td>
<td>0.4 ± 0.4s</td>
<td>52 ± 16</td>
<td>11</td>
<td>1.4 ± 0.8s</td>
<td>65 ± 39</td>
</tr>
<tr>
<td>MENA</td>
<td>5.19</td>
<td>34</td>
<td>3(1)</td>
<td>1.6 ± 0.6s</td>
<td>18 ± 15</td>
<td></td>
<td></td>
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</tr>
<tr>
<td>PEMM</td>
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<td>116</td>
<td>10(2)</td>
<td>1.7 ± 0.3s</td>
<td>-4 ± 3</td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>PMNN</td>
<td>9.17</td>
<td>88</td>
<td>7(3)</td>
<td>1.0 ± 0.2s</td>
<td>8 ± 10</td>
<td>13</td>
<td>1.2 ± 0.6s</td>
<td>47 ± 26</td>
</tr>
<tr>
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<td>29</td>
<td>2(1)</td>
<td>1.2 ± 0.2s</td>
<td>10 ± 7</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>SEHA</td>
<td>5.21</td>
<td>15</td>
<td>1(0)</td>
<td>2.1 ± 1.3s</td>
<td>-30 ± 17</td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>SINA</td>
<td>8.49</td>
<td>71</td>
<td>6(1)</td>
<td>1.4 ± 0.2s</td>
<td>-14 ± 6</td>
<td></td>
<td></td>
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<tr>
<td>SIRI</td>
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<td>1.6 ± 1.1s</td>
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<tr>
<td>VAKU</td>
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<td>132</td>
<td>9(7)</td>
<td>1.8 ± 0.4s</td>
<td>-22 ± 2</td>
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<tr>
<td>WAIB</td>
<td>9.40</td>
<td>61</td>
<td>5(2)</td>
<td>2.0 ± 0.7s</td>
<td>36 ± 7</td>
<td>9</td>
<td>1.7 ± 1.3s</td>
<td>352 ± 32</td>
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<tr>
<td>WANI</td>
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<td>63</td>
<td>4(3)</td>
<td>0.3 ± 0.1s</td>
<td>-46 ± 30</td>
<td>12</td>
<td>1.5 ± 0.5s</td>
<td>58 ± 40</td>
</tr>
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<td>WAPO</td>
<td>4.56</td>
<td>14</td>
<td>1(0)</td>
<td>1.4 ± 1.7s</td>
<td>8 ± 47</td>
<td></td>
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</tr>
<tr>
<td>B</td>
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<td>6(4)</td>
<td>1.2 ± 0.8s</td>
<td>-50 ± 14</td>
<td></td>
<td></td>
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<tr>
<td>D</td>
<td>7.48</td>
<td>56</td>
<td>5(2)</td>
<td>0.6 ± 0.3s</td>
<td>-34 ± 28</td>
<td></td>
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<tr>
<td>E</td>
<td>11.45</td>
<td>43</td>
<td>4(3)</td>
<td>1.4 ± 0.3s</td>
<td>32 ± 2</td>
<td></td>
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<tr>
<td>F</td>
<td>6.91</td>
<td>11</td>
<td>2(0)</td>
<td>1.8 ± 1.1s</td>
<td>86 ± 89</td>
<td></td>
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<tr>
<td>G</td>
<td>10.07</td>
<td>55</td>
<td>5(1)</td>
<td>1.1 ± 0.4s</td>
<td>-54 ± 8</td>
<td></td>
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<tr>
<td>H</td>
<td>5.76</td>
<td>46</td>
<td>3(0)</td>
<td>1.2 ± 0.4s</td>
<td>-74 ± 89</td>
<td></td>
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<tr>
<td>J</td>
<td>7.15</td>
<td>19</td>
<td>2(0)</td>
<td>1.2 ± 0.5s</td>
<td>46 ± 10</td>
<td></td>
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Columns are: (1) station name; (2) average SNR within time windows used for the SC method; (3) total degrees of freedom after stacking; (4) the number of measurements in the SC stack, with the number that were null in brackets; (5) estimated splitting time of the stacked SC result, in seconds, ±2σ; (6) estimated fast azimuth of the stacked SC result, ±2σ; (7) number of splitting intensity measurements used in the SI method calculation; (8) estimated splitting time using the SI method, in seconds, ±2σ; (9) estimated fast azimuth using the SI method, ±2σ. For the SI method, errors were estimated a balanced resampling bootstrap method, and a weighted L2 fit. Few stations had good SI measurements. N.B. only stations with at least one single channel measurement that passed quality control criteria (see Section 1.3.3) are shown.
The splitting in the north and on the DEI is clear, with fast directions approximately north-south and splitting times on the order of 1-1.5 s (Figure 1.8). Importantly, this observation is consistent across arrival, back-azimuth, and station, indicating that a simple fabric pervades the northern region of our array. This region was also characterized by agreement between results of the three single-channel methods, which is a good indicator of simple structure [Long and Van Der Hilst, 2005].
Figure 1.8: Stacked splitting results from the SC method (red) and SI method (magenta), where the orientation of the lines indicates the azimuth of measured fast axes and their length is the associated time delay. Line thickness corresponds to the uncertainty in the fast azimuth; we emphasize less uncertain measurements with thicker lines (see key). For the SC results, formal uncertainties in splitting parameters are estimated from the energy surface (Section 1.3.4), while for the SI results we use a bootstrap - see Table 1.1 for numerical values. Stations with only null SC results are circled in red. Yellow lines are small circles about best-fitting WLK-AUS pole derived from GPS data [Wallace et al., 2014] with present-day rates (in mm/yr) relative to AUS indicated. Station symbols are the same as Figure 1.2.

The individual measurements in Goodenough Basin and the southeastern Papuan Peninsula are more complex, particularly a broad swath running approximately parallel to the rift axis to the south of the D’Entrecasteaux. We observe rapidly varying or contradictory splitting between stations and between different arrivals at a given station. Several stations (e.g. B, D, G) exhibit two populations of measured fast axes, depending on back-azimuth, some trending approximately N-S and others trending ENE-WSW, parallel to the rift zone. Station stacks within this region show large ($\leq 90^\circ$) changes in direction of apparent fast axis between nearby stations. Inter-
estingly, the three single-channel methods return similar results in this region (see Supplementary Figure 1.7), which is not normally the case for complex anisotropy [Levin et al., 2004; Long and Silver, 2009a].

Most of these apparently anomalous measurements have high SNR and clear minima in the error surfaces. Although many of the anomalous results are measured at the OBS sites, which traditionally have low SNR on the horizontal channels used for shear wave splitting [Webb, 1998], we consider these measurements to be fairly robust given the good quality of the arrival waveforms and the fact that some of the adjacent land stations (e.g. KEIA, MAYA) show similar results. The OBS’s all lie in enclosed, isolated basins, so the long-period tilt noise commonly associated with infragravity waves may be low compared with more typical open-ocean sites [e.g. Webb and Crawford, 2010].

In the northeastern Papuan Peninsula, stations in the northwest near Cape Vogel evince fast directions approximately north-south. This appears to be a continuation of the structure on the DEI. A Rayleigh wave phase velocity map calculated at 32 s period [Jin et al., 2015] (Figure 1.9) shows a continuous region of low velocities beneath the DEI extending westwards beneath Cape Vogel, at least as far west as 149°E. Surface waves of this period are sensitive to the uppermost mantle, and we attribute this low-velocity region to elevated temperatures along the axis of the rift where the lithosphere has been thinned. This result supports previous evidence that lithospheric thinning within this rift is localized to the axis [Abers et al., 2002]. The low-velocity swath also coincides with locations of recent volcanism [Smith and Milsom, 1984]. We observe strong (>1 s) splitting, at stations overlying the region of slow phase velocities, and consistently rift-perpendicular fast axes.

Results from PEMM and MENA, on Cape Vogel itself, showed significant discrepancies between the different single-channel splitting methods. Stations further to the east and on the southern coast of the Papuan Peninsula generally returned very
few splitting results. While some of these were noisy stations, several results from GOGO, a good station on the southern coast, were null measurements. This might be a happenstance of the incident polarizations, or it might imply that the magnitude of the coherent anisotropy decreases going southeastwards across the peninsula (Section 5.1.4.). For reference, we analyzed splitting at the Port Moresby global seismic network station (PMG), about 200 km to the west of our field area, and found that the overwhelming majority of arrivals at this station are clear nulls (Figure 1.2).

1.4.2 SI method

The SI method requires several shear-wave arrivals to produce a measurement and so yielded results from fewer stations than the single-channel methods, even with the added direct S arrivals. We show only results that are obtained by fit to at least five splitting intensity measurements. Using an F-test we show that only six of the station measurements fit the observed splitting intensities significantly (at $p = 5\%$) better than a null hypothesis of no splitting. Despite this, many of the results with less statistically significant fits are consistent with the more robust SI results so in Figure 1.8 we include the 12 results that give a better fit than null at $p = 33\%$ ($1\sigma$).

The SI observations indicate that structure in the north, beneath the DEI and Trobriand Islands is $\sim$NNE-SSW, consistent with SC method results. On Cape Vogel, whereas the $\phi_{SC}$’s align roughly N-S, the $\phi_{SI}$’s trend NE-SW. This $\sim45^\circ$ disagreement between the two methods might imply complex structure, or it may be that the SI results are being dominated by the (often noisy) arrivals from a back-azimuth of $\sim140^\circ$. Arrivals with anomalously energetic transverse components from one azimuth will produce large splitting intensities that force the $2-\theta$ sinusoid to fit a peak at that azimuth. To explore this possibility we also fitted the splitting vector using the L1 norm, which is less weighted towards outliers, and obtained almost identical $\phi$’s with marginally smaller (by 0.1-0.2 s) $\delta\tau$’s.
The splitting magnitudes estimated by the SI method are on the order of 1s, which qualitatively agrees with single channel methods. Discrepancies of ~0.5 s between $\delta\tau_{SI}$ and $\delta\tau_{SC}$ are probably not significant, given similar uncertainties in $\delta\tau_{SI}$ indicated by synthetic tests (Figure A.6d). Measurements of $\phi$ at individual stations differ between the SI and SC methods by an average of $21 \pm 13^\circ$, not including the two stations with significant discrepancies, GOGO and WANI. These two stations have unclear or null SC results, but $\delta\tau_{SI} \approx 1-1.5$ s, $\phi_{SI} \approx 40-60^\circ$.

1.5 Discussion

1.5.1 Anisotropy beneath the D’Entrecasteaux Islands

Simple, extension-related fabric

$SK(K)S$ shear wave splitting indicates seismic anisotropy somewhere on the receiver side of the raypath beneath a station. Our splitting results overwhelmingly indicate N-S trending azimuthal anisotropy beneath this region. At stations on the Papuan
Peninsula north of \(\sim 10^\circ S\) as well as on the DEI and the northern islands, N-S splitting is observed across different phases and polarizations, and splitting analyses using different methods produce consistent results (Figure 1.8 and Appendix A). Such coherency of measurements makes a strong case for simple, horizontal seismic anisotropy beneath the northern two-thirds of our array, probably in the upper mantle [Long and Silver, 2009a; Long and Van Der Hilst, 2005].

Anisotropy is commonly attributed to crystallographic preferred orientation (CPO) of individually anisotropic mineral grains, whereby bulk alignment evolves as a response to finite strain [Christensen, 1984]. The fast axis of anisotropy in olivine-dominated mantle is widely observed to align with inferred mantle flow direction [Becker et al., 2008; Gaboret et al., 2003; Soedjatmiko and Christensen, 2000], in agreement with experimental and xenolith data for olivine deforming by dislocation creep from a wide range of P,T,fO2 conditions [Blackman et al., 2002; Ismail and Mainprice, 1998; Tommasi et al., 2000]. Thus the anisotropic structure observed in our field region suggests a dominant component of N-S flow or distributed shearing in the mantle related to the N-S extension. Despite the youth of this extensional regime, the rate and magnitude of the divergence have been sufficient to establish a strong flow-parallel CPO that dominates the anisotropy.

While the relationship between microplate motions and convecting mantle flow is unclear, there is a striking agreement between ongoing surface deformation evinced by GPS [Wallace et al., 2014] and our shear wave splitting results. Fast axes of the anisotropy mostly align closely with the geodetic velocity vectors and the region in the southeast where the lowest GPS velocities are observed is also the area where splitting is anomalous, weak, or unclear. Overall, the mantle fabric predicted by the modern strain field (Figure 1.10a) is extremely similar to the anisotropy that we observe.

Rayleigh-wave phase velocities show a swath of slow upper-mantle material that
we infer to represent the rift axis, in agreement with heat flow [Martinez et al., 2001], body wave tomography [Abers et al., 2002] and the location of the volcanoes. The lowest velocities likely represent flowing asthenosphere. Coherent, spreading-parallel fast axes observed on the DEI indicate that an CPO has developed in the shallow convecting mantle in response to extension beneath, and adjacent to, the rift axis.

**Asymmetry in apparent anisotropy**

There is a distinct asymmetry to the splitting results; splitting in the north and beneath the DEI is coherent and strong, while some anomalous splitting is observed to the south of the DEI (Figure 1.8). This pattern may be enhanced by uneven distribution of the sensors, but overall the region is well sampled. It is unlikely to be a product of differential station noise; the stations in the interior of the Papuan Peninsula that show complex anisotropy are not noisier than the island stations which show simple anisotropy.

Rather, we interpret that north of ∼10°S large strains at depth have recently produced a strong extension-dominated fabric, while the relatively un-deformed continent to the south of the parallel has much less clear structure. This is strongly supported by surface wave phase velocity maps, which indicate thinned lithosphere in an E-W swath between 9°S and 10°S that extends from the DEI beneath Cape Vogel, where splitting is predominantly N-S (Figure 1.9). A corollary is that significant extension has taken place within the Papuan Peninsula north of ∼10°S, roughly corresponding to the active Goodenough and Mai’iu fault structures (Figure 1.2), which explains the appreciable splitting times observed at stations on Cape Vogel. It is not clear whether this represents a wholesale southward jump of the crustal extension, or if it simply relates to distributed deformation within the weak continent.

An alternative cause of the asymmetry, which would also explain the significant splitting at the rift axis itself, is asymmetric mantle upwelling at depth due to rela-
tive motion of the rift axis in the hotspot reference frame (Figure 1.10b). The south flank of the rift is moving northwards with the Australian plate at \( \sim 50 \text{mm/yr} \), but the north flank moves northwards faster. Thus, material upwelling beneath the axis must have come from the north, resulting in a thicker anisotropic region beneath the northern flank [Conder et al., 2002; Toomey et al., 2002]. A similar interpretation has been suggested for shear wave splitting observations across the East Pacific Rise [Hammond and Toomey, 2003]; like our dataset, those results evince significant splitting beneath the spreading axis and asymmetric splitting times, with larger \( \delta \tau \) in the direction the rift is moving. This configuration may explain our large splitting times beneath the rift axis and clearer anisotropy in the north than the south. Asymmetric spreading rates are also shown to produce asymmetric dips in the lithosphere-asthenosphere boundary (LAB) on the flanks of the rift [Hammond and Toomey, 2003], with a steeper LAB on the slower-spreading side; Holtzman and Kendall [2010] suggest that melt organized along the flank with a steeply dipping LAB could cause shear wave splitting with a fast axis parallel to the rift. This mechanism provides a possible explanation for the rift-parallel splitting we observe within and around the Goodenough Basin, on the slower, southern margin of the rift. However, this conjecture is difficult to test, and there is no evidence in the body wave tomography or the surface wave inversions for asymmetric LAB gradients.

**Alternative contributions to anisotropy**

Shear wave splitting observations suffer from a lack of depth resolution. We have ascribed observed splitting to olivine CPO arising from dislocation creep in the shallow, convecting mantle, but must rule out alternative sources of anisotropy. We observe consistent splitting from different back-azimuths and between \( SKS \) and SKKS arrivals, implying the source of anisotropy in our field area is not in the deep mantle. We discuss some possibilities here (Figure 1.10).
Figure 1.10: Cartoons depicting various possible contributions to the anisotropy; arrows qualitatively indicate predicted magnitude of $\delta\tau$ and azimuth of $\phi$. Putative rift axis given by red lines. 

a) Anisotropy due to shearing beneath diverging plates, here depicted as asymmetric. 
b) Anisotropy from LPO due to shear between the shallow convecting mantle and plates moving in the hotspot reference frame. 
c) Anisotropy (fabric unknown, extent indicated by green shading) frozen into lithosphere would be expected to vary with lithospheric thickness. 
d) Anisotropy expected from aligned melt pockets at axis of rifting and volcanism, cf. Ethiopia Rift. 
e) Anisotropy due to radial perturbations to mantle flow field around diapirs implicated in formation of gneiss domes [e.g. Little et al., 2011]. 
f) Anisotropy due to superposition of westwards along-axis flow of diapiric material (blue shading) and axis-normal corner flow [after Ellis et al., 2011]. The majority of our splitting measurements agree with predictions for plate divergence, except in the south.
Debayle et al. [2005] suggest that the ∼50 mm/yr northward motion of Australia in the hotspot reference frame could produce a N-S fabric in sub-lithospheric mantle due to basal shear (Figure 1.10b). However, $SK'(K)S$ studies of the Australian continent yield little to no splitting [Clitheroe and Van Der Hilst, 1998; Özalaybey and Chen, 1999]. Robustly null splitting measured at the nearby PMG station (Figure 1.2) implies that any shearing-related fabric beneath the leading edge of Australia does not produce coherent splitting and bolsters our hypothesis that organized anisotropy within our array is the result of local structure.

There may also be an anisotropic contribution from the mantle lithosphere, resulting from plate stretching [cf. Tommasi et al., 1999] or “frozen-in” from previous deformation (Figure 1.10c). The robustly null splitting at PMG indicates that coherent lithospheric fossil fabric does not pervade across most of this region. Moreover, the absence of thick lithosphere beneath the DEI inferred from mantle seismic velocities [Abers et al., 2002] (Figure 1.9) indicates that splitting observed throughout much of the array cannot have a lithospheric component.

Despite presumed upwelling and active volcanism, melt does not appear to have a great influence on the overall pattern of anisotropy. Melt pocket SPO or melt-modified CPO are predicted to result in extension-perpendicular fast axes (Figure 1.10d), strongest where asthenosphere comes closest to surface and solidus temperatures are most likely to be exceeded [Blackman and Kendall, 1997; Holtzman et al., 2003], whereas our results evince a dominantly extension-parallel fast direction beneath much of the DEI, in the volcanically active region. As discussed above, melt channeled along a dipping LAB may contribute to complex fabrics south of 10°S, but any such melt zone is offset from volcanoes and otherwise not imaged.

Brownlee et al. [2011] anticipated that crustal anisotropy might be a diagnostic test for the mechanism of UHP exhumation because diapirism and low-angle detachment would produce distinct patterns of foliation within the domes. Our observations
do not have the resolution to differentiate between the two alternatives, and predicted crustal $\delta \tau$ of $\leq 0.4$ s (based on velocities from Brownlee et al. [2011] and thicknesses from Ferris et al. [2006]) would likely be overwhelmed by the mantle splitting signal.

Present-day subduction at the New Britain and San Cristobal trenches $\sim 500$km to the north of the rift axis might affect local mantle dynamics, but these slabs dip northwards, so associated return flow is not likely to complicate the signature of extension-related flow within the Woodlark Rift.

Notwithstanding complex recent tectonics, and ongoing UHP exhumation, the observed splitting is mostly simple. Beneath the DEI, we do not find evidence for more exotic contributions to splitting, such as radial flow around exhuming diapirs (Figure 1.10e) [cf. Behn et al., 2007] or along-axis flow of a felsic nappe such as that proposed in the models of Ellis et al. [2011] (Figure 1.10f). Although such flow may take place, it is not sufficient to alter the predominant extension-parallel fabric caused by the large-scale extension-driven flow.

Second-order complexities in the south

Some stations, particularly in the south of our array, do not agree with the overall extension-parallel trend we have identified. On west Normanby Island the active Dawson Strait volcanic field may be obscuring signal from mantle anisotropy at stations ESA2 and MAYA, which show null or unclear splitting. Similarly WANI, which shows a large discrepancy between SI and SC results, is adjacent to the active strato-volcano Mt. Victory. Several stations in and around the Goodenough Basin show a bimodal distribution of $\sim$N-S and $\sim$ESE-WNW fast axes (Figure 1.7) although the station stacks give roughly ESE-WNW axes (Figure 1.8). Complex splitting such as this may arise from a layered anisotropic structure beneath the stations [Silver and Savage, 1994]. We have insufficient measurements to quantitatively model any layering, but can qualitatively assert a shallow contribution from melt (perhaps organized
along a dipping LAB, see above) or along-axis flow as proposed by Ellis et al. [2011]. GOGO, which evinced large $\delta \tau_{SI}$ despite several null individual arrivals, lies further from the rift axis, where extension is presumed to be minimal. We speculate that complex splitting here may be a result of pre-existing fabric in continent that has not experienced large extension.

### 1.5.2 Broader implications

**A quantitative treatment of strain-induced anisotropy**

Experimental studies indicate that the strength of CPO fabric (defined by the dimensionless fabric strength index factor $J$, which quantifies the spread of the grain orientations, [Bunge and Morris, 1982]) is related to the magnitude of strain (axial strain, $\epsilon$, or shear strain, $\gamma$) a rock has undergone. At low strains, this produces an approximately linear increase in anisotropy with strain [Mainprice and Silver, 1993]. At larger strains ($\gamma = 200-300\%$) seismic anisotropy asymptotes to a maximum due to diminishing increase in both fabric strength with strain and anisotropy with fabric strength [İsmail and Mainprice, 1998; Tommasi et al., 2000]. İsmail and Mainprice [1998] showed that 5-10% anisotropy on the hand-sample scale is produced by moderate fabrics ($J$-factor $\geq 5$) resulting from $\sim 50\%$ equivalent strain in experiments [e.g. Mainprice and Silver, 1993] or viscoplastic self-consistent (VPSC) simulations [Tommasi et al., 2000]. These studies calibrate the relationship between anisotropy strength and $J$, and between $\epsilon$ (or $\gamma$) and $J$, providing an approximate relationship between finite strain and anisotropy.

We assume the extension ($X$) across the D’Entrecasteaux has caused a simple shear over some depth in the low-viscosity sub-lithospheric mantle (Figure 1.11). The magnitude of shear strain ($\gamma$) will vary inversely with the thickness of the layer ($\Delta z$) over which $X$ is accommodated, since $\gamma = X/\Delta z$. Hence, we expect a larger $\Delta z$ will give rise to a less strong fabric for a given $X$, and weaker CPO. However, for given
Figure 1.11: Calculation of anisotropy and splitting time as a function of surface displacement, $X$, at different thicknesses, $\Delta z$, of layer being sheared. (a) The relationship between $\gamma$ and J-index [Tommasi et al., 2000]. (b) The relationship between J-index and sample $V_S$ anisotropy observed in naturally deformed olivine aggregates (blue dots [Ismail and Mainprice, 1998]) and a quadratic polynomial fit to them (red line, dashed line indicates extrapolation) constrained to pass through (0,0). (c) Strength of anisotropy within the layer increases with extension ($X$, in km) but decreases with layer thickness $\Delta z$. Line colors indicate $\Delta z$ following key in (d). Anisotropy described as $\delta V_S/V_S$, the fractional difference between fast- and slow- $S$ wavespeed, assuming vertically propagating shear waves and horizontal fast axes (d) Predicted splitting times, $\delta \tau$, due to LPO as a function of extension. Dashed lines indicate extrapolation of $\delta V_S$ (J) beyond limits of data. Approximate values of extension and the predicted splitting times from LPO alone are shown for Baikal (BKL), the Main Ethiopian Rift (MER) and the D’Entrecasteaux Islands (DIs). Fits in (a) and (b) used to generate models in (c).

anisotropy, a larger $\Delta z$ will cause a larger accumulated splitting time for subvertical ray paths. The tradeoff between these two effects depends on the gradients of the strain-fabric-anisotropy relationships.

For several possible values of $\Delta z$ we calculate $\gamma$ as a function of $X$, and then use results from VPSC simulations of simple shear [Tommasi et al., 2000] to estimate J-factor (Figure 1.11a). We convert J-factor to $V_S$ anisotropy (Figure 1.11b) using the relationships in the Olivine Fabric Database [Ismail and Mainprice, 1998], making
the assumption that hand-sample anisotropy is typically three times than that of km-scale volumes of rock [Christensen, 2004], to estimate anisotropy as a function of $X$ at different $\Delta z$ (Figure 1.11c). We then compute predicted $\delta \tau$, assuming an average shear-wave velocity of 4.5 km/s (Figure 1.11d). For shear strains between $\sim$30% and 130% and $X \leq 200$ km, there is little dependency of $\delta \tau$ on $\Delta z$, such that the curves of $\delta \tau(X)$ at different values of $\Delta z$ collapse onto a common line. In other words, the decrease in fabric strength is balanced by the increase in raypath length. Therefore, we can predict the magnitude of splitting without good constraints on the thickness of the anisotropic low-viscosity region. We calculate that 140-190 km of extension at the longitude of the DEI should give rise to splitting times of 1 - 1.4 s. This independent constraint agrees well with the splitting magnitudes we observe, and provides quantitative support for the hypothesis that the observed anisotropy can be explained wholly by the recent extension.

**Comparison with other rifts**

We may compare our splitting observations to those from other regions of continental rifting, such as Baikal, the MER, or the Rio Grande Rift. Baikal evinces complex splitting close to the rift axis [Gao et al., 1997] and the MER is categorized by almost exclusively rift-parallel splitting fast directions [Kendall et al., 2005]. These two patterns are attributed, respectively, to small-scale mantle convection [Gao, 2003] and oriented melt pockets (OMP) [Bastow et al., 2010] or segregated melt bands [Holtzman and Kendall, 2010] within the lithosphere. Fast axes of splitting within the Rio Grande Rift are also parallel to the rift axis [Sandvol et al., 1992]. Our results do not resemble any of these rifts: while there is a small region with complex splitting, the majority of results indicate fast directions parallel to spreading.

Both Baikal and the MER have undergone relatively little total extension, at slow rates. Baikal has accumulated $\sim$10.5 km [Lesne et al., 2000] to 9 km [Zorin and
Cordell, 1991] of extension at 3-5 mm/yr since the mid-Pliocene, while the MER has accommodated ∼30-40 km of Nubian-Somalian separation at rates of no more than 7-8 mm/yr [Corti, 2009] and broadly distributed extension across Rio Grande is ongoing at just ∼1 mm/yr [Berglund et al., 2012]. As a result, lithospheric structures dominate seismic anisotropy. By contrast, 140-190 km of extension has taken place across the DEI in <6 My, setting it apart from the other two rifts; seismic anisotropy beneath this rift is controlled by asthenospheric flow fabrics. Calculations (Figure 1.11) indicate that as a result of the disparity in extension, shear wave splitting times caused by extension-induced CPO at Baikal or the MER would be difficult to measure (<0.4 s), while they would be appreciable across the DEI (Figure 1.11d). Plate stretching in this relatively magma-poor setting might add a lithospheric component to the anisotropy, but thin lithosphere beneath the rift axis limits this contribution (Section 1.5.1).

If small-strain continental rifts represent one end-member extensional regime, the other end-member is a mid-ocean ridge (MOR). Shear wave splitting studies at MOR’s have shown spreading-parallel fast directions [Nowacki et al., 2012; Wolfe and Solomon, 1998] albeit with some complexity at the axis. At some point on the continuum of continental breakup, the dominant anisotropic fabric within the rift must make the transition from the spreading-perpendicular trend observed at small-strain rifts to the spreading-parallel trend seen at MORs; Figure 1.11 offers a way to quantify this transition. Our interpretation is that the Woodlark rift has evolved to the spreading-parallel fabric, and provides an upper bound on the minimum extension required for a predominating CPO.

The relatively narrow surface expression of the Woodlark Rift indicates that well-understood strain localizing processes are at work [Buck, 1991] and that extension in the lithosphere seems to be localized [Abers et al., 2002]. However, within the Trobriand Platform, our results indicate strong, coherent spreading-parallel anisotropy
Figure 1.12: Cartoon depicting lithospheric extension and mantle flow beneath the D’Entrecasteaux Islands, including potential asymmetric flow pattern at depth due to relative motion of spreading axis in hotspot reference frame. Surface elevation taken from topography, crustal thickness is approximated from receiver functions [Abers et al., 2016] and lithospheric thinning estimated from $\delta V_P/V_P$ from body wave tomography [Abers et al., 2002]. Arrows indicate notional streamlines, in the frame of reference of the axis [cf. Conder et al., 2002]. Column at right shows 1D strain field from $X$ horizontal displacement accommodated over depth $\Delta z$ (see Figure 1.11). In both the north and south, the influence of past subduction on the lithosphere geometry at depth is uncertain, and structures with question marks are unconstrained. Possible organized melt pockets result from porosity/viscosity gradients at the inclined base of the lithosphere [Holtzman and Kendall, 2010]. Vertical scale changes at 5 km; below this depth there is no vertical exaggeration.

at distances >100 km from the rift axis notwithstanding a lack of crustal faulting. This implies that organized structure at depth precedes surface extension. We infer that a region of convecting mantle wider than the surficial rift is affected by the extension and that tectonic strain becomes more diffuse as it is distributed to depth.

1.6 Conclusions

Shear wave splitting reveals a preponderance of extension-parallel fast-axes in this young, continental rift system, contrary to results from less-extended rifts, such as Ethiopia or Baikal. The Woodlark-D’Entrecasteaux Rift represents an intermediate stage in the evolution between a small-extension intra-continental rift and an oceanic spreading center. The observed anisotropy indicates that the concomitant transition in mantle fabric has already taken place, from a fabric possibly dominated by melt or pre-existing structure to an CPO controlled by the extension. The fabric change is a consequence of the magnitude of extension and demonstrates how efficiently deformation mechanisms establish CPO in the mantle as it accumulates shear strain. The observation of anisotropy beneath the undeformed Trobriand Plateau indicates that appreciable flow of convecting mantle may precede crustal deformation. Despite the proximate volcanic centers, our results do not require melt to make a significant contribution to the structure, although unexplained complexity in the southern part of the rift (where there is no volcanism) might be influenced by deep melt migrating along the rift margin or pre-existing structure. Any model for UHP exhumation and extension across the D’Entrecasteaux Islands that is eventually preferred must be consistent with a simple, spreading-parallel fabric at depth.
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Imaging continental breakup using teleseismic body waves:
the Woodlark Rift, Papua New Guinea

Co-authors: G. A. Abers, J. B. Gaherty, and G. Jin.

This chapter comprises teleseismic body wave imaging of the upper mantle structure of the Woodlark Rift\textsuperscript{1}.

Abstract

This study images the upper mantle beneath the D’Entrecasteaux Islands, Papua New Guinea, providing insight into mantle deformation beneath a highly rifted continent adjacent to propagating spreading centers. Differential travel times from \(P\)- and \(S\)-wave teleseisms recorded during the 2010-2011 CDPapua passive seismic experiment are used to invert for separate \(V_P\) and \(V_S\) velocity models of the continental rift. A low-velocity structure marks the E-W axis of the rift, correlating with the thinnest crust, high heat flow, and a linear trend of volcanoes. This slow region extends 250 km along strike from the oceanic spreading centers, demonstrating significant mantle extension ahead of seafloor breakup. The rift remains narrow to depth indicating localization of extension, perhaps as a result of mantle hydration. A high-\(V_P\) structure at depths of 90-120 km beneath the north of the array is more than 6.5% faster.

than the rift axis and contains well-located intermediate depth earthquakes. These independent observations place firm constraints on the lateral thermal contrast at depth between the rift axis and cold lithosphere to the north that may be related to recent subduction, although the polarity of subduction cannot be resolved. This geometry is gravitationally unstable; downwelling or small-scale convection could have facilitated rifting and rapid lithospheric removal, although this may require a wet mantle to be realistic on the required timescales. The high-V structure agrees with the maximum P,T conditions recorded by young ultra-high pressure rocks exposed on the rift axis and may be implicated in their genesis.

2.1 Introduction

2.1.1 Overview

The Woodlark Rift, in southeastern Papua New Guinea, is one of the youngest and most rapidly extending continental rifts in the world. It hosts the youngest known ultra-high pressure (UHP) rocks, exhuming at $\sim$20 mm/yr by a mechanism that remains unclear [Baldwin et al., 2008; Ellis et al., 2011; Gordon et al., 2012; Little et al., 2011; Webb et al., 2008]. This region is one of the few active examples of a mid-ocean ridge tip propagating into thinning continental crust, spanning the transition from lithospheric extension to seafloor spreading (Figure 2.1).

The CDPapua experiment images the mantle of this continental rift in detail. Body wave tomography reveals the mantle structure of the rifted continent just west of the spreading center. From these images, we address the following questions: How does the mantle behave during continental rifting? Is seismologically-observed structure explicable by temperature variation alone, or are additional mechanisms (e.g. melt, lithology, volatiles) required? Can we image structures that are candidate
sources for UHP rocks? Is there a relationship between this region’s history of recent subduction and the present rifting?

2.1.2 Tectonic setting

Since the late Miocene, the Woodlark and Solomon Sea (SOL) microplates have been rotating northward relative to Australia (AUS), about a nearby Euler pole to their west [Taylor et al., 1999; Wallace et al., 2004, 2014]. The resultant north-south-opening basin evinces maximal extension in the east, where seafloor spreading developed ∼6 Ma, and grades towards minimally thinned continental crust in the west. With time, the spreading centers have propagated westwards into the continent. Magnetic lineations in the young seafloor to the East imply up to 140-190 km of extension at the longitude of the D’Entrecasteaux Islands (DEI) [Taylor et al., 1999]. While others have challenged these estimates as incompatible with surficial expressions of extension [Kington and Goodliffe, 2008], in the absence of an obvious transfer structure it is not clear how else to accommodate the far-field motions. GPS data, reflection seismology and geologic observations [Fitz and Mann, 2013; Kington and Goodliffe, 2008; Wallace et al., 2014] reveal that extension is distributed at the surface across a number of structures, including strain within Goodenough Basin and possibly on faults bounding the DEI.

Quaternary volcanism along an east-west line of volcanoes extends approximately along strike from the youngest seafloor spreading centers. These volcanoes produce calc-alkaline lavas with $K$ and $Ba/La$ signatures indicative of a subduction-enriched source [Smith and Milsom, 1984] although no proximate subduction is presently evident, and no slab has previously been identified from seismicity or regional tomography. During the Cenozoic, the Papuan orogenic belt developed on the mainland and along the Papuan Peninsula as ophiolites, arcs, and subduction complexes accreted to the leading edge of the Australian plate on north-dipping thrust faults [Davies and
Figure 2.1: Tectonic map of southeastern Papua New Guinea, modified after Baldwin et al. [2008]. Red arrows: present day plate motions with respect to Australia (AUS) according to best fitting GPS model [Wallace et al., 2014]. Yellow circle: Euler pole for 4.02°/My rotation of the Woodlark Plate (WLK) with respect to AUS from 3.6-0.5 Ma [Taylor et al., 1999]. Oceanic crust shown by gray shading, Brunes chron indicated by white dashed line, and recent shift in tectonics evident from the obliquity of present spreading ridges to magnetic isochrons. Purple shaded: Papuan Ultramafic Belt; light blue shaded: Owen-Stanley Metamorphics. DIs: D’Entrecasteaux Islands; TTF: Trobriand Transfer Fault, shown with geologic left-lateral slip although recent GPS results indicate right-lateral motion [Wallace et al., 2014]; OSFZ: Owen Stanley Fault Zone. Blue box shows area of field deployment and later figures, Holocene volcanic centers in this region only are depicted. Inset: simplified regional tectonics showing role of WLK and South Bismarck Plate in mobile belt between obliquely converging Australian and Pacific plates.

Jaques, 1984]. Scattered intermediate depth seismicity extending from the northern coastal ranges (146°E) southeastward beneath the Papuan Peninsula near 100-150 km depth is related to subduction [Pegler et al., 1995] or lithospheric instability [Abers and Roecker, 1991].

The DEI are cored by high pressure metamorphic rocks at the surface, including a UHP coesite-eclogite that records conditions of ~3 GPa as recently as 5-8 Ma [Baldwin et al., 2004; Gordon et al., 2012]. Physical mechanisms suggested for accommodating uplift at ~20 mm/yr include low-angle detachment zones [Webb et al., 2008], eduction [Petersen and Buck, 2015], and diapirism [Ellis et al., 2011; Little et al., 2011]. Abers et al. [1997] found that shallowly dipping faults bounding the easternmost domes...
are actively accommodating exhumation via low-angle normal faulting earthquakes, synchronous with extension.

2.2 Data

2.2.1 Deployment and measurement of travel times

The CDPapua passive seismic experiment comprised 31 on-land broadband PASSCAL stations, as well as 8 broadband OBS, deployed for ~15 months between March 2010 and July 2011 (Figure 2.2, Table C.1). The aperture of the array was approximately 250 km with station spacing of 20-50 km, suitable for teleseismic tomography between 50 and 250 km depth.

We measured body wave arrivals from 218 teleseismic earthquakes between 5th March 2010 to 30th July 2011. This dataset included 3554 direct \( P \) arrivals from 192 events and 2810 direct \( S \) arrivals from 172 events in a distance range of 30-90° from the center of our array (with the exception of four events >90° away that had
strong diffracted phases). 91 $PKP$ travel times for six selected events were also used; these near-vertically incident rays improve the lateral resolution of our model. A dataset of $> 600$ teleseismic $P$ arrivals from 90 events measured on the 1999-2000 Woodseis array [Abers et al., 2002] supplement our data, extending the region of coverage further to the east.

We use a modern implementation of the ACH tomography approach [Aki et al., 1977], accounting for expected move-out across the array to measure differential travel times of $P$ and $S$ phases. This approach relies on the assumption that moveout-corrected travel time variations are due to velocity heterogeneities within our model volume, while an event time residual captures integrated effects of velocity variations outside our model volume.

Each set of body wave arrivals was filtered and windowed by eye to preserve as much high frequency information as possible, including just the first couple of oscillations. The mean center frequency used for the $P$ waves was 0.5 Hz and for $S$ waves was 0.12 Hz; the precise center frequency used for each set of arrivals was later used to calculate finite-frequency kernels. Multi-channel cross correlation [VanDecar and Crosson, 1990] of the filtered, windowed traces was performed to obtain arrival times relative to predictions calculated with TauP (using velocity model IASP91) [Crotwell et al., 1999].

Eilon et al. [2014] established that this region contains strong anisotropy with N-S fast geometry within the rift. Recent work [Eilon et al., 2016] indicates heterogeneous anisotropy throughout this region; the degree to which this influences $S$ wave first-arrival times will depend on polarization and is not straightforward to remove. We separately measured $S$ relative delay times on BHN and BHE channels and for this isotropic inversion used delay times of $S$-waves on whichever component had more arrivals or higher mean values of cross correlation maxima - the preferred component is generally related to the $S$-wave polarization. This procedure prioritises the highest
quality travel time measurements, while the ‘noise’ introduced by anisotropy will be averaged out over multiple events and will contribute to inversion misfit; alternative approaches, such as using the transverse component, would have similar distortion due to anisotropy but in general lower signal-to-noise level.

In total, the dataset comprised 4257 $P$ arrivals and 2810 $S$ arrivals with RMS differential travel times of 0.396 s and 0.983 s respectively. The earthquakes were well distributed with respect to azimuth and distance around our region of interest (Figure 2.2), although there was a relative paucity of arrivals from the NE and SW octants and disproportionate illumination from the north due to aftershocks of the great Tohoku earthquake in March 2011.

Following Schmandt and Humphreys [2010a] we describe the spatial coverage of the data by the “hit quality” parameter, a combined measure of the hit count and the back azimuthal coverage for each voxel. This function is calculated as the number of rays with sensitivity in each voxel from each back azimuthal hexant, with the count saturating at a maximum of 6 rays per hexant, giving a total score out of 36 that is normalized to a value between 0 and 1. Within the array our spatial coverage is excellent, indicating that models should be able to faithfully resolve lateral heterogeneities. Outside the array, and at greater depth, hit quality deteriorates.

### 2.2.2 Travel time residuals

Both $P$ and $S$ differential travel time maps (Figure 2.3) show an E-W swath of positive delay times beneath the center of the array, indicating that a relatively slow region of upper mantle underlies the DEI, Goodenough Basin, and Cape Vogel, compared to faster structure to the north and south.

This dataset also reveals marked back-azimuthal variation of delay time at several stations, for example on the NW coast of Goodenough Island, or on the northern coast of the Papuan Peninsula at $150.8^\circ$E. Such strong variation is an indicator of
lateral velocity gradients directly beneath the station. While anisotropy is strong in this region, isotropic velocity heterogeneities better explain the spatial pattern of the sites with strong back-azimuthal variation.

In the far north of the array several stations have large travel time deficits, implying fast structure. By contrast, several of the OBS stations (particularly to the NW of Goodenough Island, and in the SW of Goodenough Basin) have marked travel time delays compared to stations close to them. These stations lie on some of the thickest sediments in this region (between 3.0 and 5.0 s two-way travel time to basement) [Fitz and Mann, 2013] and have large station static terms in the inversion.
2.3 Tomographic method

2.3.1 Finite frequency kernels

We apply a finite frequency approach to inverting for seismic velocities, following the approach of Schmandt and Humphreys [2010a] who use a first Fresnel zone paraxial approximation to the Born theoretical kernel [Dahlen et al., 2000], based on 1D ray tracing in the AK135 reference model [Kennett et al., 1995]. These simplified kernels are advantageous in terms of computational efficiency, and are suitable given imprecise knowledge of the true ray path in a 3-D heterogeneous Earth.

The kernel is calculated by finding the ray-normal first Fresnel zone radius, $R_f$ at 15 km increments along the ray, as a function of velocity($v$), distance along the ray ($D_R$), frequency ($f$), and total path length ($\Delta$):

$$R_f = \sqrt{\left(\frac{v}{f}\right) \frac{D_R (\Delta - D_R)}{\Delta}}$$  \hspace{1cm} (2.1)

These radii are interpolated along the ray path and then used to compute a ray-normal sensitivity function that approximates the bi-modal Born kernel, without side-lobes [Schmandt and Humphreys, 2010a, equation 2]. The kernel is normalized to the ray theoretical sensitivity along the path [Schmandt and Humphreys, 2010a, equation 3] such that for velocity heterogeneities wider than the sensitivity kernel the predicted travel time anomaly using the finite frequency approach is identical to the ray theoretical prediction. Finally, ray-normal smoothing is applied (calibrated by the results of Saltzer and Humphreys, 1997), where smoothing increases with distance from the station.

2.3.2 Parameterization and regularization

Differential velocities are computed on an irregular rectangular mesh of nodes. The node spacing increases with distance from the center of the array, from 30 km (within
the array) to 45 km, and increases with depth from 30 km in the shallowest mantle to 40 km at the base of the model domain, at 290 km.

We apply both spatial smoothing and model norm damping to regularize the inverse problem, which then minimizes the cost function:

$$E = ||\omega(Gm - d_{obs})||^2 + \gamma||Lm||^2 + \epsilon||m||^2$$  \hspace{1cm} (2.2)

where \(m\) is the vector of perturbations to the initial velocity model, \(d_{obs}\) is the vector of differential travel times, and \(G\) is the matrix with \(G_{ij} = \partial d_i / \partial m_j\). \(\omega\) is a diagonal matrix of data weights, \(\gamma\) is the smoothing parameter, \(L\) is a smoothing matrix that applies an exponentially decaying covariance between nearby grid points [Menke and Eilon, 2015], and \(\epsilon\) is the damping parameter.

We analyse the tradeoff between model roughness and data misfit (Figure C.1) to determine regularization parameters where model roughness and data misfit are mutually minimized [Menke, 1984]. \(\gamma\) and \(\epsilon\) are given by the minimum of the penalty function, \(\chi = ||Gm - d_{obs}||^2 + 2||m||^2\), where our choice to penalize model roughness by twice as much as data misfit is arbitrary and selected to give a model with the most realistic structure. For close-to-optimal values of damping (\(\epsilon\) between \(\sim 1\) and \(\sim 10\)) the strength of smoothing (\(\gamma\)) does not have a large effect on the penalty function demonstrating that \(\epsilon\) has more influence over the location of the L-curve corner. We favor slight over-damping to yield only the most robust features, which are not dependent on nodal geometry.

The model parameter vector, \(m\) consists of a vector of velocity perturbations at each node in the model (of which there are \(n_x \times n_y \times n_z\)), as well as \(n_{sta}\) static station terms and \(n_{evt}\) static event terms. The static terms are intended to account for travel time residuals caused by local site effects and velocity heterogeneities outside of the model volume. \(P\) and \(S\) models are determined separately. Large static terms decrease effective time delays that must be accounted for within the model volume and thus mute model structure. In order to avoid these static corrections
spuriously removing velocity heterogeneity from the model, we apply limited damping to these station and event terms equal to 50% and 10% of the velocity damping values, respectively.

A series of tests allows us to ascertain the influence of an *a priori* crustal correction versus station static terms solved for in the inversion. The prior crustal terms are computed using a Moho surface from receiver functions and a surface wave $V_S$ model [Jin et al., 2015] with mean $V_P/V_S$ of Ferris et al. [2006]. The data are inverted with and without any prior crustal correction, and for the cases of high, moderate, and zero station static damping. These cases correspond to permitting zero, moderate, and unconstrained variation to prior crustal corrections, respectively. Without station static terms the variance reduction for both $P$ and $S$ models is diminished by $\sim 10\%$.

The most important result was that the overall structure of the rift below 50 km (where the rays begin to cross) is unaffected by the prior crustal model used; models vary by an average of just 2.5% between cases employing a prior crustal correction and those without. Changing crustal model affects only the uppermost (40 km) layer in the model and we therefore do not interpret this layer. The station static terms in the inversion absorb the effect of the crust, giving a similar final result whether or not a prior crustal model is imposed.

### 2.3.3 Solving the inverse problem

The linear inverse problem is solved by applying the LSQR algorithm [Paige and Saunders, 1982] to the weighted, damped least squares problem by minimizing $E$ in Equation 2.2. The data weights are based on the maximum value of the maximum cross-correlation ($C_i$) between each trace and the stack for that event, where $\omega_i = 10^{(C_i - 0.7)/0.3}$. The 1D ray tracing used to compute travel time kernels ignores focusing/defocusing caused by 3-D velocity heterogeneity, with the result that fast features will tend to be artificially broadened, slow features to be artificially thinned,
and side lobes to appear [Bijwaard and Spakman, 2000]. However, given the coarse grid spacing of >30 km and the small model volume this factor is not likely to introduce large spurious structure into our models. Some of these problems are also obviated by the use of finite-frequency kernels, which are only computed for 1D reference structures [Dahlen et al., 2000]. The goodness of fit to our data is quantified by the variance reduction:

$$\Delta \text{Var} = \left(1 - \frac{\text{Var}(d_{\text{obs}} - d_{\text{pred}})}{\text{Var}(d_{\text{obs}})}\right)$$

(2.3)

where Var is the variance function, normalized by $n - 1$, $d_{\text{obs}}$ is the observed data and $d_{\text{pred}}$ is the data predicted from the model, i.e. $d_{\text{pred}} = Gm$. This measure of the variance reduction is somewhat inflated as it includes data from poorly resolved regions that are underdetermined and hence bound to be well-fit. Following Schmandt and Humphreys [2010b] we also compute the variance reduction ($\Delta \text{Var}_{hq}$) using data predicted only from the part of the model domain that is well resolved (Section 2.4.1).

### 2.3.4 Synthetic tests

A suite of synthetic tests demonstrates the ability of these data to usefully resolve structure. Checkerboard tests (Figure 2.4) demonstrate that our data can accurately resolve structure on the order of 90 km in the upper 150 km of both $V_P$ and $V_S$ models. These tests indicate that vertical resolution remains adequate down to $\sim$150 km depth for $V_P$ and $\sim$100 km depth for $V_S$. Northward smearing evident deeper than 150 km is likely the result of a preponderance of arrivals from this azimuth (Figure 2.2).

Our models have a maximum amplitude recovery of $\sim$80% ($P$) and $\sim$75% ($S$) in the shallower 150 km of the domain, deteriorating to $\sim$60% ($P$) and $\sim$50% ($S$) in the lower 150 km. Following Zelt [1998] we compute the semblance between the input and output checkerboard models (contours in Figure 2.4). Values of semblance
Figure 2.4: Results of checkerboard tests. Left column: input structure, center column: *P* model output structure, right column: *S* model output structure. Top three rows: horizontal tomograms at different depths, bottom row: a north-south vertical section at 148.8°E. The output models are contoured by model semblance (a measure of agreement between the input and output checkerboards) following Zelt [1998].
greater than 0.7 indicate good model fidelity.

Checkerboard tests suffer from the weakness that they simultaneously accentuate smearing along diagonals and underestimate smearing along cardinal directions; as such, they do a poor job of identifying artifacts in the output of the inversion. We performed tests inverting artificial data from a model comprising structures of interest: an E-W slow “ridge axis” and a deeper fast structure in the north of the model (Figure C.2).

These tests demonstrate that the data can resolve $V_P$ structure of a narrow rift, with little horizontal smearing in the upper 150 km. Vertical smearing increases with depth from $<50$ km in the upper 150 km of the model to $\sim100$ km in the lower half of the model, with along-ray smearing becoming particularly problematic deeper than 160 km. The $V_S$ inversion gives much the same result, with somewhat greater vertical smearing and poorer amplitude recovery.

Since the velocity perturbations within each layer average to zero, these inversions may lead to artificial side lobes adjacent to velocity anomalies. Synthetics indicate these artifacts are likely to be low amplitude (Figure C.2). However, apparent vertical velocity gradients may arise from our lack of constraints on absolute velocity [cf. Lévéque and Masson, 1999].

2.3.5 Squeezing tests

The steep incidence angle of teleseisms results in poorer vertical resolution than horizontal resolution. A squeezing test is used to ascertain the minimum depth extent of anomalies required by the data, as follows:

In step (i), we solve

$$G' m_1 = d$$

using the constrained least squares inversion. $G'$ includes strong damping that permits velocity perturbations, $m_1$, only in the upper part of the model, above the “squeezing
depth” ($z_{sq}$). The residual from this inversion,

$$ r = d - G' m_1 $$

then becomes the data for step (ii), solving this time for the whole model space:

$$ G m_2 = r $$

By linearity,

$$ m = m_1 + m_2 $$

In this way we obtain a model, $m$, that is not minimum length but fits as much of the data as possible with structure above the squeezing depth. If none of the data require deep structure, then the residual from step (i) will be small, $m_2$ will be small, and the final model will have very little structure below the squeezing depth and the same variance reduction as the constrained inversion.

We perform squeezing tests for the $V_P$ model with $z_{sq}$ of varying from 120 km to 250 km, quantifying the test results by the fraction of the final $\Delta$Var that is achieved in step (i) and the norm of $m_2$. For values of $z_{sq} > 180$ km, $m_1$ produces $>95\%$ of the final variance reduction and $\|m_2\|_2$ is $85\%$ reduced, i.e. the great majority of the signal can be fit with structure shallower than 180 km and only weak features are required below this depth. This finding does not preclude still deeper structure but any deeper features are not necessitated by our measured time delays. A similar set of tests for the $V_S$ model also shows that features deeper than 180 km are also not required by those data. For $z_{sq} \leq 150$ km the appearance of low velocities along the rift in $m_2$ indicate that the deep extension of low velocity features in our model is a robust feature.
2.4 Results

2.4.1 Tomographic models

Based on the results of L-curve tests (Section 2.3.2) and the squeezing tests (Section 2.3.5) our preferred $P$ and $S$ models use $\gamma_P = 2$, $\epsilon_P = 5$ and $\gamma_S = 4$, $\epsilon_S = 6$, respectively. The model uncertainty is estimated by bootstrapping using 100 resampled datasets randomly drawn from the full dataset. In regions deeper than 50 km with hit-quality greater than 0.8 the bootstrapped average standard deviation is 0.53% for $\delta V_P$ and 0.9% for $\delta V_S$. This analysis demonstrates that the main features of the model are well resolved deeper than 50 km. We also note that the region of high checkerboard semblance ($> 0.7$) closely coincides with the region of high hit-quality and low bootstrapped standard deviation (Figure 2.5); the fact that these three measures of expected model resolution agree, and that the well resolved region contains much of the interesting structure that we observe, permits us to confidently interpret the images. The consistency of the $P$ and $S$ models (which are derived from independent datasets) is evident from comparison of Figures 2.6 and 2.7 and the robust linear trend in $\delta \ln V_S / \delta \ln V_P$.

The $V_P$ model achieves $\Delta \text{Var} = 82.3\%$ and $\Delta \text{Var}_{hq} = 79.0\%$, while the $V_S$ model, with fewer data and stronger regularization, achieves $\Delta \text{Var} = 76.4\%$ and $\Delta \text{Var}_{hq} = 71.3\%$. Given that the $V_P$ model has 50% more data, with slightly greater spatial coverage, higher frequency information and superior variance reduction, we focus on this model to interpret the structure and dynamics of this rift. The $V_S$ model complements the $V_P$ results and gives additional constraints on physical mechanisms of velocity heterogeneities.

Final station terms (computed by adding the prior crustal correction to the station statics from the inversion) strongly correlate with mean differential travel times at each station (Figure 2.3, Table C.1). These results agree with published moho maps.
Figure 2.5: Different measures of model resolution at selected depths. Top pair of rows: hit quality (Section 2.2.1) for $P$ (upper) and $S$ (lower). Bottom pair of rows: bootstrapped estimate of standard deviation (Section 2.4.1) of percentage $\delta V$ for $P$ (upper) and $S$ (lower) models.

showing shallow crust beneath the DEI grading to thicker crust on the rift shoulders [Abers et al., 2002, 2016]. Exceptions include stations on the Papuan Peninsula south of -10°S, which may not have thick crust, and station KIR on the Trobriand Islands where shallow fast velocity structure must be present.
Figure 2.6: Variation in $P$-wave velocity plotted at different depths in the model domain, contoured for hit quality between 60 and 90%. Regions with hit quality <60% are masked out.

$V_P$ model

The $V_P$ model (Figures 2.6 and 2.8) shows an E-W swath of low velocities in the upper mantle beneath the DEI and extending westwards towards the Papuan Peninsula, beneath Cape Vogel. The E-W slow structure has clear north and south limits and the $\delta V_P < -1\%$ region remains narrower than 100 km down to $\geq 180$ km depth, beyond which synthetic tests indicate loss of resolution. These images are consistent with a previous $P$-wave tomography study centered to the east of our array [Abers et al., 2002] and with surface waves [Jin et al., 2015].

The steep southwards dip of the slow material in the western part of the model (Figure 2.8, section A-A’) may be present but is not a well constrained feature. Synthetic tests indicate that the apparent east-west bifurcation of the slow structure deeper than 200 km is an artifact resulting from diverging ray paths below the depth.
Figure 2.7: As for Figure 2.6 but for $S$-wave velocity.

of good resolution. However, two interesting features that do appear to be robust are the broadening of the slow structure and the increase in amplitude of the slow anomaly going from west to east, in the direction of increasing extension (Figure 2.8, B-B’ and C-C’).

To the north of this slow rift there is a marked (2-3%) positive $V_P$ anomaly between 8.5° and 9.0°S. The lateral continuity of the fast structure is less apparent in $V_P$ than $V_S$ but it seems to extend along strike (E-W) in the uppermost ~100 km. There is a clear fast velocity region to the NW of Goodenough Island at 120 km depth that is in a region of the model that is well resolved. While this fast velocity structure is visible in the tomograms down to 180 km it becomes muted deeper than 135 km. Given the magnitude of this positive velocity anomaly, its well-defined boundaries, and the results of synthetic tests, the fast feature cannot be an side-lobe artifact of the tomography but represents faster-than-ambient mantle material. At a depth of 90 km
there is a 6.5% peak-to-peak $V_P$ variation; this magnitude of velocity heterogeneity is significantly greater than the typical value of <3% observed in other continental rifts [Achauer and Masson, 2002] and approaches values observed in back-arc settings [Conder and Wiens, 2006], or in the region of subducting plates [Lay, 1997]. The $V_S$ model shows the same features, although slightly broader (Figure 2.7). Based on synthetic tests (Section 2.3.4) and the fact that it is strongly dependent on the crustal
model used, the discontinuous fast structure in the less-well resolved region south of the rift axis seems to be a side-lobe artifact.

**$V_S$ model**

The $V_S$ model (Figures 2.7 and C.3) displays similar features to the $V_P$ model in that there is a clear E-W trending low-velocity in the center of our model region and fast structure beneath the north of the array. In general, the $V_S$ model shows less organized structure at depths greater than 120 km and squeezing tests demonstrate that no structure beneath 180 km is required by the data. The apparent southward step of the rift axis from east to west at $\geq 150$ km depth is not well resolved.

In contrast to the $V_P$ model, the low shear velocities appear to be much more concentrated at shallow depths beneath the DEI and, in particular, Goodenough Basin. Another distinction is that there is no clear southern boundary to the slow velocity structure along the rift axis. The fast structure beneath the Trobriand Platform is more laterally contiguous in $V_S$ than in the $V_P$ model, although squeezing tests indicate that it is not required to persist deeper than 120 km.

### 2.5 Local seismicity

Earthquakes that occurred around the network were detected and located following procedures similar to Li et al. [2013], giving a catalog of 795 well-located hypocenters [Abers et al., 2016]. These earthquakes are located in a one-dimensional velocity model determined from the 1999-2000 WOODSEIS seismic array overlapping with the eastern edge of the present study region [Ferris et al., 2006], and have formal location errors less than 7 km vertically and 5 km horizontally. Improved locations were then obtained using the double-difference algorithm HYPODD [Waldhauser and Ellsworth, 2000]. Together with a previous seismicity study farther east [Ferris et al.,
Figure 2.9: Plot of earthquakes in the region of our array. Earthquakes observed in this study (filled circles) or the Woodseis deployment (triangles) [Ferris et al., 2006] were relocated using the HYPODD method; only events with estimated 3-D location uncertainties less than 2.5 km are plotted. Open circles: events from the EHB catalog [Engdahl et al., 1998] for the period 1964-2002 (plotting only events constrained with at least one depth phase). The cross section shown has no vertical exaggeration.

These earthquakes outline a belt of crustal deformation extending westward from the oceanic rift tip at 151.7°E (Figure 2.9).

Most earthquakes ring Goodenough and Fergusson Islands where faults bounding the metamorphic core complexes have been previously inferred [e.g. Davies and Warren, 1988; Little et al., 2011]. Seismicity demonstrates that these faults are actively accommodating ongoing dome exhumation, both on the north and south sides of these domes, in agreement with GPS data [Wallace et al., 2014]. The catalog includes diffuse seismicity all along the Papuan Peninsula at crustal depths, including a southward step from the DEI to the Papuan mainland that indicates the Dayman-Suckling massif is the along-strike continuation of the chain of domes exhumed as part of the shallow extension process. Despite the good network coverage and geodetic evidence for motion here [Wallace et al., 2014], few earthquakes are observed along the Goodenough Fault (Figure 2.2). These results will be discussed in more detail in a subsequent paper.

One of the most unexpected findings in the seismicity catalog is a group of 6-8
well-located earthquakes at intermediate depths, 90-120 km, near (8.7°S, 149.7°E), with others farther west outside of the network. These earthquakes occur at the same depths as those from which nearby UHP rocks were exhumed at 5-8 Ma and are the first well-located subcrustal earthquakes located within the Woodlark-D’Entrecasteaux Rift region, although intermediate-depth earthquakes have been documented 200 km farther west beneath the center of the Papuan Peninsula [e.g. Abers and Roecker, 1991; Pegler et al., 1995]. Modern catalogs such as EHB [Engdahl et al., 1998] show that the intermediate-depth earthquake zone extends from the Papuan Peninsula ESE and shallows to the region of the earthquakes we report here (Figure 2.9). These earthquakes almost directly underlie the western part of the array and are well-recorded by the OBS’s above them; all have at least one station within 40 km of the epicenter and are located by at least 10 phases, leading to formal errors in depth less than 1.5 km and horizontal errors less than 3.0 km. Thus, it is highly unlikely that these are mislocated in depth significantly.

2.6 Discussion

2.6.1 The rift axis

We have imaged the axis of the rift represented by an E-W structure of low seismic velocities in the upper mantle. This body is co-located with the shallowest Moho observed in receiver functions [Abers et al., 2002, 2016], the region of highest heat flow [Martinez et al., 2001], and Quaternary surface volcanism [Smith, 1982]. The low velocities in the center of the rift lie directly along strike from the youngest spreading center at 151.5°E, and continue at least 250 km to 149°E. The amplitude of the δVₚ anomaly suggests almost total lithospheric removal, implying mantle divergence and adiabatic upwelling of asthenosphere considerably ahead of the propagation of seafloor spreading (in agreement with spreading-related anisotropic fabric [Eilon et al., 2014].
and some other continental rifts [e.g. Rychert et al., 2012]).

The velocity models indicate that the rift remains relatively narrow: the $\delta V_P < -1\%$ region remains <100 km wide along strike and to the maximum depth it is resolved. Its width at 60 km depth increases from $\sim$60 to 100 km going from 149.5 to 151.0°E, in the direction of increasing extension. The crustal seismicity overlies the low velocities down to >150 km depth, demonstrating the localized and vertical nature of the rifting (Figure 2.8). Squeezing tests indicate that the velocity anomaly is not required by the data deeper than $\sim$180 km.

The narrow profile of the low-$V$ region to depth is similar to that observed in the Main Ethiopian Rift (MER) [Bastow et al., 2005; Hammond et al., 2013] but more localized than Rio Grande [Gao, 2004; Wilson et al., 2005] or Baikal [Gao, 2003] and contrasts with the wide triangular zone of upwelling thought to exist beneath mid-ocean ridges undergoing classic corner flow [Hammond and Toomey, 2003]. Buck [1991] demonstrated that narrow continental rifts are favored by cold geotherms, thin (<40 km) crust, and high strain rates - all features exhibited by this rift system. They also found that model core complexes form in areas of high geothermal gradient, raising interesting questions about their formation here where mean heat flow is low for a rift ($\sim$70 mW m$^{-2}$ [Martinez et al., 2001]). Very low rift velocities at all depths can be reconciled with low heat flow in Goodenough Basin [Martinez et al., 2001] if they are localized to the immediate north of the basin, and/or if the mantle rifting is significantly younger than the crustal conductive heating timescale.

Strain localization may also arise as a result of prior viscosity heterogeneity; perhaps previous subduction has resulted in a weak corridor of mantle that could be exploited by incipient extension (e.g. in the hydrated mantle wedge and/or beneath the active arc [cf. Dixon et al., 2004; Kirby et al., 2014]). The rift may no longer be wet; our data do not require the large $\delta \ln V_S/\delta \ln V_P$ expected for water in nominally anhydrous minerals (Section 2.6.2). Furthermore, the presence of basaltic volcanism
along the rift axis probably requires temperatures too high to sustain water stored this way at depth [Hirth and Kohlstedt, 1996]. Despite this, Ruprecht et al. [2013] found >3 wt% H$_2$O primary melt content in Goodenough Island lavas - these are derived from the shallow region where we note large $\delta \ln V_S$ excursions (Section 2.6.2).

Another local factor that might facilitate strain focussing is the presence of a cold, high-viscosity structure adjacent to the rift (Section 2.6.3). In addition, high topography resulting from the Eocene Papuan orogen could have provided potential energy responsible for accelerating extension.

2.6.2 Physical mechanisms for velocity heterogeneities

Plotting the relative perturbations in $V_P$ and $V_S$ for each node in our model yields arrays of points (Figure 2.10) with a slope $\delta \ln V_S/\delta \ln V_P$ that is diagnostic of physical mechanism, given assumptions regarding composition, pore structure, and anelastic effects. Steeper slopes indicate a stronger variation in shear modulus than bulk modulus, where $1.2 \leq \delta \ln V_S/\delta \ln V_P \leq 2.0$ is compatible with thermal heterogeneity alone [Anderson et al., 1992] while $\delta \ln V_S/\delta \ln V_P > 2.0$ requires some mechanism (e.g. partial melt, water) to preferentially depress the shear modulus.

Our data (Figure 2.10) give a $\delta \ln V_S/\delta \ln V_P$ of 1.58 for the entire model volume, with an estimated 95% uncertainty of ±0.56 and a correlation coefficient ($R$) of 0.69. Subdividing the model space above and below 100 km shows that the upper part of the model has more scatter ($R = 0.67$) than the part deeper than 100 km ($R = 0.79$) and a steeper slope (1.80 ± 0.99 versus 1.40 ± 0.45), although the difference is not statistically significant.

While slope analysis provides only a lower bound on $\delta \ln V_S/\delta \ln V_P$ (because $V_S$ is more smoothed than $V_P$ both intrinsically and by regularization), the observed slope matches predicted temperature-dependent velocity variations (Figure 2.10). Applying the scaling relationships of Jackson and Faul [2010] we compute $\delta \ln V_S$ and $\delta \ln V_P$
Figure 2.10: Left: Plot of $S$ versus $P$ model velocity variations at common nodes, subdivided into the upper and lower parts of the model. Only nodes with hit quality $> 0.7$ are plotted, and the extreme 5% of velocity heterogeneities are ignored. Triangles (on both plots): shallow nodes beneath the D’Entrecasteaux Islands and Goodenough Basin (White boxes, Figure 2.10). The best-fit slope for all the nodes, constrained to (0,0), is found by orthogonal regression, minimizing weighted misfit to both $\delta V_S$ and $\delta V_P$, giving: $m = 1.58 \pm 0.56$ (2σ bounds), with a correlation coefficient of 0.69. Right: Predicted covariation of anelastic velocity variations with temperature at 95 km depth, assuming $T_{pot} = 1250^\circ C$ and $dT/dz = 0.4$ K/km. Different curves correspond to different grain sizes (see legend), as well as the anharmonic prediction. Full details in Section 2.6.2.

over a range of temperatures below $T_{adiabat}$ at 95 km depth for relevant grain sizes (between 1mm and 100mm [Behn et al., 2009]), using the mean center frequencies of our measurements. We assume a mantle potential temperature of 1250°C based on DEI tephra [Ruprecht et al., 2013] and use anharmonic $V(P, T)$ for pyrolite computed using HeFESTo [Stixrude and Lithgow-Bertelloni, 2011]. For 10 mm grains the predicted $\delta \ln V_S/\delta \ln V_P$ is 1.62, similar to the mean value (over a range of depths) from our model, supporting a predominantly thermal control on $\Delta V$ in this area.

We may not formally reject the hypothesis that the shallow part of the model is also recording only temperature heterogeneity but other factors may also be important. Strong anisotropy in this region [Eilon et al., 2014] may bias estimates of
\[ \delta \ln V_S / \delta \ln V_P \] [Hacker and Abers, 2012] as might compositional heterogeneities associated with recent subduction (Section 2.6.3). Moreover, several model nodes scatter towards large values of \(|\delta \ln V_S / \delta \ln V_P|\). In particular, shallow nodes lying beneath the DEI and Goodenough Basin (Figure 2.8, white box; Figure 2.10, triangles) plot up to 4% below the main trend in \(\delta \ln V_S / \delta \ln V_P\) suggesting the presence of melt or volatiles, in spatial agreement with the surface volcanism, and consistent with a surface wave study showing very low \(V_S\) immediately below the crust in the rift axis [Jin et al., 2015]. Nonetheless, our results do not support the presence of large volumes of melt within the rift axis, in contrast to the MER [Bastow et al., 2005; Rychert et al., 2012]. The relative paucity of melt in this rift (despite likely hydration) may be due to the low mantle potential temperature [Ruprecht et al., 2013] compared to, for example, the plume-influenced convecting mantle beneath the MER [Bastow et al., 2008] or the western U.S. [Wang et al., 2013].

### 2.6.3 Intermediate depth seismicity and fast structure

Intermediate depth seismicity in this region (Section 2.5) is definitive evidence of cold, seismogenic material in the upper mantle; earthquakes in subducting slabs are thought to occur at temperatures less than 700-800°C [Hacker et al., 2003; Peacock, 2001]. The hypocenters that fall within our array, and that have the smallest formal errors, are situated within the fast velocity structure in the northeast of the \(V_P\) model at the same depth, near 8.7°S, 149.8°E. We infer that temperatures in the high-velocity body do not exceed 700-800°C and that the positive \(\delta V_P\) structure is not only fast relative to the average model velocity at that depth but is cold and seismically fast in an absolute sense. Our calculations show that a \(\sim 700°C\) structure adjacent to \(\sim 1300°C\) adiabatically rifted mantle would produce the observed large \((\Delta V_P \sim 6.5\%)\) velocity perturbations (Figure 2.10) irrespective of whether the fast mantle structure is depleted [Schutt and Lesher, 2006]. The independent observations
of seismicity and $\Delta V$ provide an unusually specific constraint on steep lateral thermal gradients in a rifting context.

The southern boundary of the fast structure is well resolved and observable from back azimuthal variation in the differential travel times (Figure 2.3). The northern, western, and eastern limits of the fast region at the edge of our array are not well constrained. This body appears somewhat discontinuous in the $V_P$ model but to have greater along-strike contiguity in the $V_S$ model (which has stronger smoothing). Our models constrain only deviations from average layer velocity but, assuming the rift axis material is at similar velocity with depth (along a mantle adiabat), the diminished relative amplitude of the fast structure deeper than 130 km is persuasive evidence of its limited depth extent.

Interpretations for this structure include: a subducted slab, the cold nose associated with previous subduction, or a lithospheric instability (Figure 2.11). Unusually cold structure at depth likely extends over a large region. The $>70$ km deep earthquakes within our array lie within a band of diffuse intermediate depth seismicity extending $\sim 500$ km WNW beneath the Papuan Peninsula, although dip and lateral contiguity of this band are unclear [Abers and Roecker, 1991]. Similarly, our tomographic images do not show a clear dip to the fast region, nor do the earthquakes in our study exhibit a Wadati-Benioff zone with discernible directionality. This young and tectonically active region is far from any cratonic continental interior and so the pre-rifting lithospheric thickness is not thought to have been large but to have formed in Cenozoic accretionary events.

**A relic of subduction**

The fast structure may represent a hitherto unrecognized subducted slab or cold nose of a mantle wedge related to subduction at either the Trobriand Trough (TT) or Aure-Moresby-Pocklington Trough (AMPT). Both candidate subduction boundaries
Figure 2.11: Cartoons of the rift showing different interpretations of the fast structure in the north of the models as representing (a) a slab fragment from southward subduction at the Trobriand Trough, (b) a slab fragment broken off by rifting of underthrust lithosphere from northward subduction at the Aure-Moresby-Pocklington Trough, or (c) a lithospheric instability either left over from the Papuan Orogen or associated with rifting. Note 3.5× vertical exaggeration shallower than 5 km. GB: Goode-nough Basin, other abbreviations in text.
are roughly parallel to the scattered trend of intermediate depth seismicity beneath the Papuan Peninsula and (in the southeast) a roughly E-W trend of Holocene volcanism on the DEI that bears hallmarks of a subduction-enriched source [Smith and Milsom, 1984]. With no resolvable dip the earthquakes corroborate neither scenario [Abers and Roecker, 1991]. The main faults on the Papuan Peninsula are all north-dipping [Daczko et al., 2009; Davies, 2012] but these largely reflect the Eocene orogeny and do not preclude subsequent south-vergent thrusting. Regional studies have not found tomographic evidence for subduction in the vicinity [Hall and Spakman, 2002] although their resolution in the uppermost mantle is likely to be poor away from seismically active slabs.

Several authors [Cooper and Taylor, 1987; Davies et al., 1987; Fitz and Mann, 2013; Pegler et al., 1995; Yan and Kroenke, 1993] have postulated southward subduction of the oceanic Solomon Sea at the TT, which may have terminated with the onset of northward subduction at New Britain and the San Cristobal trenches [Yan and Kroenke, 1993] leaving a dangling slab (Figure 2.11a) at shallow depth. At present the TT is largely aseismic, with just a single (normal faulting) \( M_W > 6 \) earthquake instrumentally recorded, and plate motion estimates are difficult to reconcile with any convergence at this boundary [Wallace et al., 2004, 2014]. Thus, evidence for ongoing subduction at this boundary is weak, although the trough may have been the site of significant convergence at some time in the past.

Alternatively, Cenozoic convergence at the AMPT has been linked to northward subduction of the leading edge of the Australian margin. Subducted material could have been rafted \(~200 \text{ km}~\) northwards by recent extension to now lie north of the rift (Figure 2.11b). This scenario is attractive because it involves the subduction of continental material into the source region of the continentally-derived UHP rocks found within the DEI. The timing and dynamics of UHP exhumation would then be dictated by breakup of underthrust Australian lithosphere during rifting; incomplete
breakup or underplating of this felsic material where less extension has occurred could explain low sub-Moho velocities west of Goodenough Island [Jin et al., 2015]. However, Eocene AMPT subduction [Yan and Kroenke, 1993] may be incompatible with observed seismogenicity and low temperatures because a 100 km thick lithospheric slab heated from both sides would warm with an e-folding timescale of $\sim 32$ My. Webb et al. [2014] and Mutter [1975] suggest AMPT subduction in the mid-Miocene ($\sim 12$-$13$ Ma) but symmetric magnetic anomalies in the Coral Sea to the south permit only limited AMPT subduction since $56$ Ma [Weissel and Watts, 1979].

Lithospheric instability?

Whatever its origin, our observations provide an unusually well imaged juxtaposition of hot rift axis and cold dangling lithosphere, where the combination of velocity information and seismicity delineates a sharp lateral gradient of $\sim 700$ K over $<70$ km. Such large thermal (and, hence, density) contrast implies the system may be gravitationally unstable.

Mareschal [1983] proposed that rapid zones of extension such as this one require lithospheric delamination, a process which has been implicated in the development of metamorphic core complexes [Lachenbruch et al., 1994; Molnar, 2015] and which could contribute to rapid UHP exhumation during rifting [Ellis et al., 2011; Little et al., 2011]. Abers and Roecker [1991] postulated that the intermediate depth seismicity beneath the Papuan Peninsula north of $\sim 8.5^\circ$ S could be explained by lithospheric convective instability (drip) just as well as any proposed subduction, suggesting that volcanism on the DEI represents a region where a continental lithospheric root has already detached (Figure 2.11c); this interpretation is similar to that suggested for intermediate depth earthquakes and fast mantle structures in the Carpathians [Ren et al., 2012]. However, assuming mantle lithosphere viscosity of $10^{23}$ Pa s [Burov, 2009] convective instabilities would grow with a characteristic timescale of 500-800
My, about \(10-100\times\) longer than available here, permitting long-lived slab fragments [Wang et al., 2013]. Stratigraphy provides an important constraint on this process: sediments younger than 8.4 Ma on the Trobriand Platform do not permit more than \(\sim 500\) m of recent vertical displacement [Fitz and Mann, 2013], precluding significant dynamic or isostatic topography associated with the fast structure and indicating slow or negligible instability growth.

Still, some mechanism must exist to remove the lithosphere beneath the rift axis prior to the onset of sea-floor spreading. Small scale convection associated with the horizontal temperature gradient could facilitate rifting [Buck, 1986] and result in large velocity contrasts [Hieronymus et al., 2007] as seen in our imaging. It is possible that the introduction of water by prior subduction could hasten convective processes by lowering mantle viscosity and facilitating lithospheric removal.

If lithospheric instability or small scale convection have not influenced rifting it is not clear how the hot shallow temperature at the rift axis has been achieved so rapidly. The discrepancy between apparently limited crustal thinning and lithospheric removal in this rift is not consistent with pure shear extension models [McKenzie, 1978] and perhaps necessitates lower crustal flow, a process which has been frequently linked to development of metamorphic core complexes [Buck, 1988].

2.6.4 Relationship to UHP rocks

The inferred temperatures of \(\leq 700^\circ\text{C}\) at \(\sim 100\) km depth are very similar to the maximum pressure and temperature recorded by the UHP rocks. The intermediate-depth seismicity lies along strike to the east, and somewhat north, of the metamorphic core complexes. Speculatively, the UHP rocks could have formed in a continuation or extension of the same cold body at depth, prior to extension. Parts of this body closest to the rift axis could have been entrained in exhumation processes related to the rifting, while other fragments of the same body were not. Alternatively, no direct
connection exists between this slab fragment and the UHP rocks at the surface; the 
P, T coincidence reflects the Earth’s common subduction geotherms rather than a 
causal link. That said, the presence of this deep, cold material close to the along-
strike continuation of the present metamorphic core complexes (not including the 
Suckling-Dayman massif) is circumstantial evidence that the fast structure in our 
models is related to the UHP source. Petersen and Buck [2015] model rapid UHP 
exhumation in the Woodlark Rift by extension and eduction (reversal of subduction) 
of continental material on a north-dipping subduction plane. If the high-\(V_P\) structure 
we image is a slab fragment, their geodynamical model may explain the cessation 
of subduction and exhumation of material from UHP depths within a rift that has 
asymmetry similar to our tomographic cross sections, although with opposite polarity.

2.7 Conclusions

We have conducted a detailed seismic investigation of the Woodlark Rift beneath the 
DEI, imaging the rifted continent. A narrow low-velocity structure extends along 
strike from the nearby spreading centers >250 km into the extended continent; this 
structure correlates with surficial volcanism and the thinnest crust seen in receiver 
functions and indicates marked thinning of lithosphere along the rift axis far ahead 
of sea floor spreading.

The continental rift remains narrow to depth. We posit that high volatile content 
in a relict mantle wedge, as well as the contrast in strength between asthenosphere 
and a dangling slab, facilitated strain localization processes that are expected in 
rapidly-extending rifts such as this one.

Well-located intermediate depth earthquakes in this region are co-located with a 
fast velocity structure in our \(V_P\) and \(V_S\) tomographic models. Several alternatives 
exist that would allow the subduction of cold material to these depths at some time
prior to the onset of rifting; the models do not clearly define a dip to the fast region. The juxtaposition of this cold, fast lithosphere with hot, slow rifted mantle explains the large peak-to-peak velocity contrast in our images. The longevity of this feature is an open question, given the gravitational instability implied by the strong thermal contrast between dangling lithosphere and upwelling asthenosphere.

Both seismicity and velocity heterogeneities indicate that the fast structure in our models is at similar P,T conditions to those recorded by the coesite-eclogite sample, along strike from the present-day core complexes. This may represent un-exhumed material awaiting sufficient overlying crustal extension before beginning ascent, or simply a remnant of the subduction system that advected the UHP rocks’ protoliths to depth.

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Chapter 3

A joint inversion for shear velocity and anisotropy: the Woodlark Rift, Papua New Guinea

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This chapter builds on the isotropic tomographic study and observation of strong, organized anisotropy. By simultaneously inverting for azimuthal anisotropy and isotropic velocity heterogeneities, we resolve potential tradeoffs between the two.

Abstract

Tradeoffs between velocity and anisotropy heterogeneity complicate the interpretation of differential travel time data and have the potential to bias isotropic tomographic models. By constructing a simple parameterization to describe an elastic tensor with hexagonal symmetry, we find analytic solutions to the Christoffel equations in terms of fast and slow horizontal velocities that allow us to simultaneously invert differential travel time data and splitting data from teleseismic $S$ arrivals to recover 3-D velocity and anisotropy structure. This technique provides a constraint on the depth-extent of shallow anisotropy, otherwise absent from interpretations based on $SKS$ splitting alone. This approach is well suited to the young Woodlark Rift, where previous studies have

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found strong velocity variation and substantial SKS splitting in a continental rift with relatively simple geometry. This study images a low velocity rift axis with $\leq 4\%$ spreading-parallel anisotropy at 50-100 km depth that separates regions of pre-existing lithospheric fabric, indicating the synchronous development of extensional crystallographic preferred orientation and lithospheric thinning. A high-velocity slab fragment north of the rift axis is associated with strike-parallel anisotropic fast axes, similar to that seen in the shallow mantle of some subduction zones. In addition to the insights provided by the anisotropy structure, the improvement in fit to the differential travel time data demonstrates the merit to a joint inversion that accounts for anisotropy.

3.1 Background

Seismic anisotropy is a widespread feature of the uppermost mantle and, in many places, is strong enough to cause travel time anomalies on the same order as isotropic velocity variations [Anderson, 1989; Bezada et al., 2014]. In addition to removing spurious velocity artefacts, a consideration of anisotropy has the potential to offer dynamical and structural insights by providing a proxy for time-integrated strain history and internal fabric. Yet most tomographic models neglect the effect of anisotropy because of the complexity it adds to the inverse problem and the inability of standard differential travel time measurements to sufficiently resolve the additional parameters it requires. Here we present a method for simplifying the anisotropic parameterization and collecting data that is sensitive to anisotropy and velocity heterogeneity in order to invert jointly for velocity and anisotropy structure.

The complex elastic anisotropy of natural rocks and the non-linearity of anisotropic effects mean that anisotropic body wave tomography poses a difficult problem. Teleseismic shear wave splitting studies image vertically-integrated anisotropic structure with good horizontal resolution but they lack the ability to re-
solve more than one or, occasionally, two layers of anisotropy [Silver and Savage, 1994], due to a paucity of data, near-vertical ray paths, lack of crossing rays, and the non-commutativity of the splitting operator [Silver and Long, 2011]. By making simplifications to parameterize weak anisotropy [Mensch and Rasolofosaon, 1997; Thomsen, 1986] 3-D anisotropic travel time inversions have been attempted (e.g. Pratt & Chapman 1992) but the problem is ill-conditioned and mostly limited to $P$-wave anisotropy, often in subduction settings [Zhao et al., 2015] and assuming hexagonal symmetry [Wang and Zhao, 2013]. Shear wave splitting tomography from local earthquake signals has been carried out in the well-illuminated mantle wedge of a subduction zone [Abt et al., 2009; Pozgay et al., 2007] but the method has not proved straightforward to generalise. Anisotropic tomography using the “multichannel method” [Chevrot, 2000], which measures the commutatively additive splitting intensity, has been developed in theory [Chevrot, 2006; Favier and Chevrot, 2003] but the necessity for excellent back-azimuthal coverage has limited its application in practice. Other workers have decomposed data based on polarization perpendicular or parallel to expected fast azimuth to estimate the effect of anisotropy, including depth resolution [Boyd et al., 2004; Hammond and Toomey, 2003]; this idea is the foundation for the method we present.

Continental rifts manifest strong seismic velocity variations as crust thins, cold lithosphere breaks up, and hot, seismically slow asthenosphere replaces it. Several rifts have also been shown to contain measurable anisotropy, variously attributed to small scale convective flow [Gao et al., 1997], aligned melt pockets [Kendall et al., 2005], large scale deformation [Tommasi et al., 1999; Vauchez et al., 2000] or mantle flow [Montagner et al., 2007], organized melt along steep lithospheric gradients [Holtzman and Kendall, 2010], pre-existing lithospheric fabric [Kendall et al., 2006], or a combination of the above [Hammond et al., 2014]. SKS splitting in rifts such as Rio Grande and the Main Ethiopian Rift (MER) is dominated by spreading-perpendicular
fast axes [Gok et al., 2003; Kendall et al., 2005] in contrast to mid-ocean ridges, where models and observations indicate spreading-parallel fast axes where strain has been sufficient to establish crystallographic preferred orientation (CPO) of olivine [Blackman and Kendall, 2002; Wolfe and Solomon, 1998]. Detailed imaging of anisotropy within continental rifts has the potential to discriminate between, and inform our understanding of, different modes and dynamics of extension [Vauchez et al., 2000].

3.1.1 Tectonic setting

The Woodlark Rift lies in a zone of broad oblique convergence between the Australian and Pacific plates, where a belt of microplates manifest complex tectonics including rapid rotation, translation, multiple recent subduction episodes, and associated volcanism. The Woodlark is one of the youngest and most rapidly extending rifts known, with 70-190 km of opening Petersen and Buck [2015]; Taylor et al. [1999] since the late Miocene, and contains ultra-high pressure (UHP) metamorphic rocks that ascended from their maximum P,T equilibration at ≥100 km depth since 5-8 Ma Baldwin et al. [2008]; Gordon et al. [2012]. Lithospheric removal and consequent asthenospheric upwelling along the rift axis has given rise to strong velocity heterogeneities. Hot axial material is juxtaposed with a seismically fast and seismogenic body to the north of the rift that may be a relict slab, possibly related to UHP burial Eilon et al. [2015]. SKS measurements indicate strong anisotropy within this rift, with spreading-parallel fast axes that can be attributed to mantle CPO developed during extension Eilon et al. [2014]. However, without depth constraints on the anisotropy it is difficult to reconcile the tomographic images with SKS data, and to separate lithospheric from asthenospheric effects. The 2-D, orthogonal geometry and well defined boundaries of this rift, together with the probability of tradeoffs between isotropic velocity structure and strong anisotropy, make this an ideal proving ground for a joint velocity-anisotropy imaging study.
3.2 Anisotropic parameterization

A full description of anisotropic velocities within a model volume would involve too many free parameters for a tractable inverse problem; defining the anisotropic elastic tensors of natural rocks requires between 5 and 21 independent parameters per node. Although previous workers have implicitly assumed CPO by summing elastic tensors scaled to those for various known mineralogies to construct aggregate anisotropic tensors [Abt and Fischer, 2008; Pozgay et al., 2007], the parameterization established in the present study is agnostic as to the mechanism of the anisotropy. It is not clear that the data are sufficiently good to warrant a more complex parameterization, and the introduction of assumptions regarding mantle mineralogy and CPO would add unnecessary uncertainty. Instead, we describe the anisotropy and velocity within a model volume by the fewest possible parameters, by making a number of simplifying assumptions.

Consistent with natural and experimental petrofabrics of deformed olivine aggregates [Christensen, 1984; Hansen et al., 2014; Ismaıl and Mainprice, 1998], we assume an elastic tensor with hexagonal symmetry and horizontal symmetry axis; this geometry captures anisotropy imaged for horizontal CPO [Tommasi et al., 1999] or aligned melt pockets [Kendall et al., 2005]. We link $P$ and $S$ velocities by assuming equal $P$ and $S$ anisotropy and fixing the ratio, $\nu$, between average $V_P$ and $V_S$. In this simple framework, a vertically-incident shear wave would be split into two quasi-shear pulses, orthogonally polarized parallel to, and perpendicular to, the symmetry axis, with velocities $V_\parallel$ and $V_\perp$, respectively (Figure 3.1). These are related to the two physically relevant parameters, average shear velocity, $V_{Sav}$, and fractional anisotropy, $\alpha$, as follows:

$$V_\parallel = V_{Sav} (1 + \alpha)$$
$$V_\perp = V_{Sav} (1 - \alpha) \quad (3.1)$$
so:

\[
V_{\text{Sav}} = \frac{V_{\parallel} + V_{\perp}}{2} \quad \alpha = \frac{V_{\parallel} - V_{\perp}}{V_{\parallel} + V_{\perp}}
\] (3.2)

An arbitrarily incident shear wave will be split into two orthogonal pulses which may have different velocities. Because of the hexagonal symmetry, these velocities are purely a function of the angle (\(\zeta\)) between the propagation direction and the symmetry axis (Figure 3.2). Assuming that the value of the anisotropic parameter \(\eta \approx 1\) in the upper mantle (Appendix D) \([Anderson, 1989; Tommasi et al., 2000]\), we can then show (Equations D.2 - D.8) that the two velocities into which any shear
wave will be split are:

\[
V_{SH}(\zeta) = \sqrt{V_\perp^2 \sin^2 \zeta + V_\parallel^2 \cos^2 \zeta}
\]

\[
V_{SV}(\zeta) \approx \sqrt{V_\parallel^2 \cos^2 2\zeta + \left(\nu^2 \left[V_\parallel^2 - V_\perp^2\right] / 4 + V_\parallel^2\right) \sin^2 2\zeta}
\]  

(3.3)

and

\[
\cos \zeta = \sin i \left(\cos \theta \cos \phi + \sin \theta \sin \phi\right)
\]  

(3.4)

where \(i\) is the incidence angle measured from the vertical, \(\theta\) is the ray propagation azimuth, and \(\phi\) is the azimuth of the symmetry axis (Figure 3.1). In this context, \(V_{SV}\) denotes the shear velocity perpendicular to ray propagation that is in the plane containing the propagation vector and the hexagonal symmetry axis, and \(V_{SH}\) is the shear velocity perpendicular to this plane. In the case of vertical incidence, these reduce to \(V_{SH} = V_\perp\), and \(V_{SV} = V_\parallel\). The nomenclature comes from the common assumption of radial anisotropy (“transverse isotropy”) where \(V_{SV}\) is vertical and \(V_{SH}\) is horizontal [e.g. Maupin and Park, 2007].

By comparison to the solutions to the Christoffel equations, we show that the \(V_{SH}\) vector is always perpendicular to the symmetry axis, and that the \(V_{SV}\) vector is approximately parallel to the symmetry axis for upper mantle teleseisms (\(\sim 47^\circ \leq \zeta \leq 85^\circ\)), reaching a maximum angular error of 22° for incidence angles relevant to our data (Figure 3.2).

Finally, the E-W striking, N-S opening Woodlark rift and the simple, predominantly N-S fast anisotropy measured here from \(SK(K)S\) splitting [Eilon et al., 2014] suggests a natural geometry for the anisotropic fabric in this particular setting. We therefore assume that the fabric is orientated such that the symmetry axis is oriented north-south. This fixes the permitted fast direction to N-S (for \(\alpha > 0\), and \(V_\perp < V_\parallel\)) or E-W (for \(\alpha < 0\), and \(V_\perp > V_\parallel\)). We have tested the choice of fixing the symmetry axis N-S versus E-W and found that this choice has a negligible effect on the result of the inversion. Using the notation \(u = \sin i \cos \theta = \cos \zeta\) (since \(\phi = 0\)), Equation
Figure 3.2: $P$ (a) and $S$ (b) wave velocity and (c) $V_S$ anisotropy as a function of angle between the propagation vector and the symmetry axis for precise and approximate cases. (d) Error in $V_{SV}$ and anisotropy ($\alpha$) and difference between true fast azimuth and North for $V_{SV}$. The errors in polarization and amplitude of anisotropy are computed as the difference between the solution to the Christoffel equations and the solutions to our simplified approximation to the anisotropy (Equation 3.5).
(3.3) simplifies to:

\[ V_{SH} = \sqrt{u^2 V_\parallel^2 + (1 - u^2) V_\perp^2} \]
\[ V_{SV} \approx \sqrt{V_\parallel^2 + \nu^2 u^2(1 - u^2)} \left( V_\parallel^2 - V_\perp^2 \right) \] (3.5)

where \( V_{SV} \) is the velocity of waves polarized north-south, and \( V_{SH} \) is the velocity of waves polarized east-west; these velocities determine relative timing of shear wave arrivals recorded on the N-S and E-W components of seismic instruments, respectively (Appendix E). Notice that \( V_{SH} \) and \( V_{SV} \) are each sensitive to both \( V_\perp \) and \( V_\parallel \) for non-vertical incidence.

By making these assumptions, we have derived analytically differentiable expressions relating travel times of orthogonally polarized \( S \)-wave arrivals to anisotropy strength, polarity (N-S or E-W) and isotropic velocity variations. These expressions form the basis of an inverse problem for 3-D anisotropy and velocity structure. The parameterization limits such an inversion to imaging N-S fast versus E-W fast anisotropy; azimuthal anisotropy with true fast azimuth intermediate to these directions will be projected onto this orthogonal basis, introducing artefacts and making our approach non-ideal for regions with more diverse or smoothly varying anisotropy. However, other orthogonal choices of symmetry axis azimuth can be easily accommodated by setting \( \phi \neq 0 \) in Equation (3.4).

### 3.3 Data

Data were collected during the CDPapua passive seismic experiment, a March 2010 to July 2011 deployment comprising 31 land-based broadband PASSCAL seismometers and 8 broadband OBS stations (Figure 3.3). The instruments were distributed on and around the D’Entrecasteaux Islands (DEI) over the \( \sim 250 \times 250 \) km region of extended continent ahead of the propagating spreading centers, with station spacing of 20-50
Figure 3.3: Maps of study area. a) Deployment and geography, including station locations and types. Magenta lines: opening paths for Euler poles derived from seafloor magnetic anomalies to the east, with 0.5 Ma tick marks; along-strike extrapolation of western spreading centers also shown. b) Results of synthetic splitting tests (Section 3.5.2) showing comparison between individual SKS observations Eilon et al. [2014] and synthetic values obtained by propagating SKS waves through our final model structure. Only high-quality, non-null splitting observations were used. Best-fitting fast azimuth and splitting time are indicated by line orientation and length, respectively, and are plotted at 50 km depth piercing points. 1000 m elevation contours plotted.

S-wave arrival times were measured for earthquakes $M_W \geq 5.5$ events between $30^\circ$ and $90^\circ$ from the array (as well as three events $\geq 90^\circ$ with large diffracted phases): 335 earthquakes in all. For each arrival, a 200 s excerpt of data centered on the theoretical S-wave arrival (from IASP91) was detrended, tapered and padded with zeros, before being bandpass filtered. A hand-selected 20 s window of data around the predicted S arrival was excerpted for cross-correlation, applying a 20% Tukey window. Only events with a distinct S phase were used.

“Intermediate arrivals” will be partitioned between the $V_{SV}$ and $V_{SH}$ principal directions, and will propagate along each with its respective velocity, accruing a time difference $\delta \tau_{N-E}$ (the definition of shear wave splitting). As a result, arrivals at a given station measured on the NS and EW channels should have near-identical
waveforms but arrive at different times. For these events we perform a three-way inversion, cross correlating NS waveforms between all stations, cross-correlating EW waveforms between all stations, and cross-correlating NS with EW waveforms at each station to compute splitting beneath that station. We adapt the least squares approach of VanDecar and Crosson [1990] to simultaneously invert for $\delta T_E$, $\delta T_N$, and $\delta \tau_{N-E}$ (Appendix F).

We emphasize that it is merely a convenient coincidence that the particular geometry of this rift is such that the anisotropic structure is aligned with cardinal seismometer components. The method we have presented here is general; if the geometry of the tectonics were different, a vertical-axis rotation would achieve the coordinate transformation required to measure arrivals on the appropriate orientations.

Filter frequencies were hand-selected for each set of arrivals to preserve as much high-frequency energy as possible while recovering a high-amplitude signal. The median corner frequencies were 0.125 Hz and 0.02 Hz. Only arrivals with signal-to-noise-ratio (SNR) greater than 3 and with cross-correlation coefficient ($c_{max}$) greater than 0.7 were used. In order to avoid the $\delta T$ values dominating the structure, we upweight the splitting measurements by the ratio of the number of $\delta T_N: \delta \tau_{N-E}$ measurements ($\sim 3$).

### 3.3.1 Out-of-volume effects

Anisotropy measurements from direct $S$-waves provide better depth resolution than core-traversing phases but may include splitting signal from anisotropy close to the source [Foley and Long, 2011; Lynner and Long, 2014] or in the mid-mantle [Fouch and Fischer, 1996; Nowacki et al., 2015]. The non-linearity and non-commutativity of anisotropic effects [Silver and Long, 2011] make the problem of out-of-volume structure more complicated for splitting measurements than for differential teleseismic travel times. A good back azimuthal distribution and the inclusion of event $\delta \tau_{N-E}$
Figure 3.4: Source-side splitting terms (Section 3.3.1) estimated from the inversion ($\delta \tau_{INV}$: top) and measured at PMG ($\delta \tau_{PMG}$: bottom) plotted by mean back azimuth and ray parameter. The cluster of E-faster-than-N events from $\sim 350^\circ$N are Tohoku aftershocks. Inset: plot of $\delta \tau_{PMG}$ versus $\delta \tau_{INV}$ showing best fit line with $R^2=0.54$, where filled symbols are from the Tohoku region.

Terms serve to cancel out or capture the effects of anisotropy outside the model volume, respectively. Complex waveforms, identified by $<0.8 \ c_{max}$ between N-S and E-W arrivals, may have encountered multiple/complex splitting and are discarded. Station $\delta \tau_{N-E}$ terms are included as model parameters in the inversion to account for anisotropy in the crust, although this term is damped to ensure that crustal splitting times are within the limits of what is generally observed ($<0.5 \ s$) [Yang et al., 2015].

As a check, $\delta \tau_{N-E}$ values were measured directly from cross correlation of North and East channels at the Port Moresby GSN station (PMG) 250 km to the west of the CDPapua array - this proximity means that teleseisms recorded at this station and recorded on our array have near-identical paths except in the upper mantle beneath the region. Null splitting measured from $SK(K)S$ arrivals at PMG over a variety of back azimuths [Eilon et al., 2014] demonstrates that the upper mantle beneath this station is essentially isotropic. Therefore, any splitting of direct $S$-waves measured at this station must arise from anisotropy close to the source or along the ray path, offering an independent estimate of out-of-volume splitting of waves impinging on the array.
There is a moderate correlation between the event $\delta \tau_{N-E}$ terms solved for in the inversion and those measured at PMG ($R^2 = 0.54$). Plotting the $\delta \tau_{N-E}$ terms by back azimuth and ray parameter (Figure 3.4), there is a clear spatial agreement between the two datasets. In particular, there is a preponderance of $\sim 1.5$ s positive (E faster than N) travel times from arrivals from the north (Japan) that are tightly grouped in back azimuth and ray parameter. This signal may arise from anisotropy in the region of the Marianas and Izu-Bonin slabs where Nowacki et al. [2015] and Lynner and Long [2015] record strong mid-mantle anisotropy. Importantly, the marked E-faster-than-N event terms from these earthquakes are in strong agreement with the PMG splitting measurements, demonstrating that this signal is attributable to source-side structure and can be removed.

Following this analysis, we use the $\delta \tau_{N-E}$ values measured at PMG as a priori estimates of event splitting terms for the inversion, with moderate damping constraining deviation from these prior values.

### 3.3.2 Travel time residuals

In total, the dataset comprises 1601 $\delta T_N$ measurements, 1300 $\delta T_E$ measurements, and 969 $\delta \tau_{N-E}$ measurements from 145 earthquakes. The RMS residuals are 0.984 s, 1.340 s, and 1.167 s respectively, and the demeaned values (removing the effect of mean splitting) are 0.984 s, 0.971 s, and 0.573 s. The similarity between de-meaned RMS $\delta T_N$ and $\delta T_E$ implies that the majority of the differential travel time signal is attributable to heterogeneities in average velocity. The $\sim 0.6$ ratio between average $\delta \tau_{N-E}$ and de-meaned RMS differential travel times indicates that the magnitude of anisotropy heterogeneities is expected to be roughly 0.6 times that of velocity heterogeneities.

The measured differential arrival times (Figure 3.5) on both components show travel time delays in an E-W swath beneath the DEI, in agreement with isotropic
Figure 3.5: (top) Examples of three-way cross correlation (Appendix F) for an earthquake in Java sampling mostly N-S fast anisotropy (left) and an earthquake in Tohoku that carries a source-side signature of E-W fast anisotropy. (bottom) Data used for this inversion, de-meaned to show only differential values of travel time and splitting. Area-proportional rose diagrams indicate backazimuthal coverage, with concentric contours at 1000, 500, and 100 arrivals.

tomography models [Eilon et al., 2015] that show the slow rift axis extending westwards from the seafloor spreading center at $\sim 151.5^\circ$. Travel time deficits seen at northern stations indicate a fast structure beneath the north of the array. Splitting data are noisier (and sparser) than differential travel times but show predominantly negative values (i.e. N-S arriving earlier than E-W) beneath the DEI.

3.4 Tomographic problem

3.4.1 Inversion approach

Following the approach of Schmandt and Humphreys Schmandt and Humphreys [2010a], we relate travel time data to model parameters using finite frequency velocity kernels with a first-fresnel zone approximation (also described in Eilon et al.
Finite frequency kernels for anisotropic media have been calculated [Favier and Chevrot, 2003; Long et al., 2008] but by describing the model in terms of orthogonal velocities, the simpler velocity sensitivity kernels may be used. Our method also lends itself to incorporation of frequency-dependent splitting measurements [Long, 2010] but that application is beyond the scope of this paper. The data include differential travel times, which are sensitive only to horizontal deviations in velocity, and splitting times, which are also sensitive to absolute velocity.

The anisotropic $V_S$ model is defined on an irregular rectangular mesh with eight layers between 40 and 240 km depth (based on station spacing and array aperture) and horizontal node spacing increasing from 30 km at the center to 40 km at the edge of the array.

The nodes are parameterized in terms of absolute orthogonal velocities $V_{\perp}$ and $V_{\parallel}$, where the mean $V_{S_{av}}$ at each depth is obtained from a local 1D model for the top 40 km [Ferris et al., 2006] transitioning smoothly to the Voigt-averaged $V_S$ from a global 1D model [Kustowski et al., 2008] down to 400 km. We fix the $V_P/V_S$ ratio, $\nu$, equal to 1.81 [Ismail and Mainprice, 1998] although this inversion, without $P$-wave data, is very weakly sensitive to this value. Tests varying $\nu$ between 1.6 and 2.0 yielded RMS variations of just 0.03% and 0.06% in isotropic velocity perturbations and anisotropy, respectively, when compared to the preferred $\nu$ value. The model parameter vector then comprises nodal velocities, isotropic delays ($\delta T$) and splitting terms ($\delta \tau$) for each station and event: $m = \{V_{\perp}, V_{\parallel}, \delta T_{\text{evt}}, \delta T_{\text{sta}}, \delta \tau_{\text{evt}}, \delta \tau_{\text{sta}}\}$.

The forward model, $g(m)$, and Fréchet kernels, $G = \partial g(m)/\partial m$, are themselves functions of the model parameters (Appendix E), so the inverse problem is non-linear and is solved iteratively by Newton’s method with partial derivatives re-calculated at each iteration [Menke, 2012, Eq. 9.11]:

$$m_{k+1} = m_k + G_k^{-g} (d_k - g(m_k)) \quad (3.6)$$

where $G_k^{-g}$ denotes the “generalised inverse”, the solution to the weighted, damped
least squares problem that minimises the generalised error:

$$\Phi(m) = \| \omega^2 (d^{obs} - g(m)) \|^2 + \gamma \| Lm \|^2 + \epsilon \| m \|^2$$  \quad (3.7)

where $\omega^2$ is a diagonal matrix of data weights ($= \sigma^{-2}_d$) defined by $\sigma^{-1}_d = 0$ for $c_{max} < 0.7$ and $\sigma^{-1}_d \propto 1 \rightarrow 2$ for $c_{max}$ in the range $0.7 \rightarrow 1$. The constraints of first-derivative smoothing (imposed by $L$) and damping are weighted by $\gamma$ and $\epsilon$, respectively [Menke and Eilon, 2015]. The solution is the least squares solution to $Fm = f$, i.e. $G^{-g} = [F^T F]^{-1} F^T$, where:

$$F_k = \begin{bmatrix} \sigma^{-1}_d G_k \\ \gamma L \\ \epsilon I \end{bmatrix} \quad \text{and} \quad f_k = \begin{bmatrix} \sigma^{-1}_d (d - g(m_k)) \\ 0 \\ 0 \end{bmatrix}$$  \quad (3.8)

The inversion is terminated after 20 iterations, or when the residual decreases by less than 1% between iterations.

Regularization of average velocity and anisotropy at a given node, $i$, is achieved by damping the values of $(V^i_\perp + V^i_\parallel)$ and $(V^i_\perp - V^i_\parallel)$, respectively. The ratio of vertical to horizontal smoothing is 0.4 and the sparser $\delta\tau$ data require the anisotropy to be smoothed 3× more than velocity. Based on “L-tests” (Figure 3.6) that seek to mutually minimise model roughness and data misfit [Menke, 1984], our preferred inversion uses $\gamma = 3$, $\epsilon = 3$.

### 3.4.2 Resolution and squeezing

Resolution of the model is estimated by computing the hit quality (a combination of hit count and back-azimuthal coverage [Eilon et al., 2015]) and the semblance (a quantitative measure of agreement between input and output checkerboard structures from synthetic tests [Zelt, 1998]). Regions expected to be well resolved have hit quality and semblance greater than 0.7; the product of these two values describes the estimated uncertainty, where values greater than 0.5 are considered acceptable.
Figure 3.6: Results of “L-tests” to ascertain the combination of regularization parameters that mutually minimize misfit and model $l^2$-norm. Curves show results for values of damping ($\epsilon$) between 0.3 and 30 and four different choices of smoothing. Dotted line: anisotropy model only ($\delta\tau_{N-E}$ variance reduction vs. $\alpha$-norm); dashed line: velocity model only ($\delta T$ variance reduction vs. V-norm); solid line: total variance reduction vs. total norm. For consistency of units, the V-norm and $\alpha$-norms are calculated in seconds from the differential travel time and splitting time, respectively, accrued by a vertical ray passing through each node. The preferred values of $\gamma = 3$, $\epsilon = 3$ are indicated.

(Figure 3.7). Synthetic tests demonstrate that not only are we able to recover input structure but that we can independently resolve velocity and anisotropy heterogeneity with good fidelity (Figure 3.7).

We ascertain the depth of heterogeneity *required* by the data by introducing a squeezing depth into the first two iterations of the inversion [Eilon et al., 2015]. If the data require structure below $z_{sq}$ it will be introduced by later, unconstrained iterations, the final model will have appreciable model norm below $z_{sq}$, and this deeper structure will significantly contribute to variance reduction. We find that for tests that include squeezing to depths ($z_{sq}$) of 170 km or greater, >95% of the final variance reduction is achieved with structure above $z_{sq}$. This indicates that the data requires models to contain velocity and anisotropy heterogeneities down to a depth of at least 170 km but that deeper structure does not improve the fit to the data and is not required. In the final inversion, we therefore include a single squeezing step to 170 km depth. By comparing the norm of the structure deeper than $z_{sq}$ before and after squeezing to 170 km we find that 16.4% of the anisotropy heterogeneities, but only 9.05% of the velocity heterogeneities, appear deeper than this depth, suggesting more power in anisotropy variation than velocity variation in the deeper part of the domain. This observation agrees with our expectation that passive rifts do not evince deep velocity signatures, but may accrue strain (and anisotropic fabric) at large depths.
Figure 3.7: Checkerboard tests for a joint $V_{S_{an}}$ and anisotropy inversion, contoured by semblance. In the well imaged region with hit quality greater than 0.7 the mean semblance is 0.73, indicating good amplitude and structure recovery. Note the staggering of the $V_S$ and $\alpha$ checkers is preserved in the output model, demonstrating that the two parameters are independently constrained.
3.5 Results

3.5.1 Tomographic models

Our preferred models have an overall variance reduction of 87.5% after 8 iterations, beyond which point there is negligible improvement in the fit. The reduction in variance for the differential travel time data and splitting data is approximately equal and the final RMS error is 0.41 s. RMS station static differential travel time is 0.50 s and the distribution of station static values is consistent with receiver function data that show thinned crust beneath the DEI compared to the Papuan Peninsula [Abers et al., 2002, 2016]. Crustal time excesses at OBS stations on the western Trobriand Platform are due to thick sediments [Fitz and Mann, 2013], and < 1 s time deficits at stations on the Amphlett Islands and Trobriand Islands (in agreement with the isotropic inversion) indicate fast shallow structure. RMS crustal splitting times is 0.17 s, indicating negligible anisotropy in the crust (although this parameter is damped, even when damping is reduced by 10\times the RMS crustal splitting is just 0.28 s).

2-D models

We first consider models where along-strike variation in velocity and anisotropy is damped to zero (Figure 3.8). The resultant models (which achieve a variance reduction of 83.7%) represent a 2-D average of the rift that includes only the most robust structure required by the data. These north-south cross sections reveal the low-velocity rift axis extending down to \(\sim 170\) km depth beneath the DEI and to their south. There is also a well resolved high velocity feature in the upper 140 km beneath the north of the array. Within the rift axis delineated by the low velocities, a N-S fast anisotropic fabric extends from depths of \(\sim 250\) km to the top of the model, directly beneath the DEI. To the north and south, on the shoulders of the rift, the models reveal shallow E-W fast structure extending off-axis to at least the edges of
Figure 3.8: North-south sections through velocity/anisotropy model where heterogeneities are constrained to vary only in 2-D (Section 3.5.1). Black dashed line encloses slow rift axis ($V_S^{av} < -2\%$), blue arrows and red circles (denoting arrows “into the page”) show N-S-fast or E-W-fast fabric, respectively. 10% hit quality contours shown. White circles: seismicity observed during Woodseis and CDPapua experiments, relocated using hypoDD Abers et al. [2016]. The Moho is derived from receiver functions along a N-S transect at 150.5°E. Note scale change at 5 km depth so topography is $3.5 \times$ exaggerated. PP: Papuan Peninsula; TP: Trobriand Platform; TI: Trobriand Islands.

By comparison with the 3-D models, the apparent necking of N-S fast anisotropy at ~100 km depth is likely an artifact arising from along-strike smoothing combined with slightly heterogeneous ray coverage.

3-D models

The velocity model prominently contains a rift-parallel swath of low velocities beneath Goodenough Basin and the southern DEI that extends westward beneath the Papuan Peninsula (Figure 3.9). Within the well-resolved region, the axial material (defined within the $d\ln V_S^{av} < -1\%$ contour) is, on average, $\sim 3\%$ slower than the layer mean and with increasing depth widens from 120 km to 160 km. There is a sharp boundary between the slow rift axis and a large, high velocity structure in the north of the model; at depths of 60-85 km there is a maximum of 11% velocity contrast between these structures, and a sharp horizontal velocity gradient at the boundary (8% over just 60
km) matching strong back-azimuthal variations of differential travel times recorded at the latitude of Goodenough Island. The loss of resolution beyond the boundaries of the array means that the northern and western boundaries of the high velocity feature are poorly constrained but its reduction in amplitude further east than 150° suggests that it does not extend as far as the edge of our array. Overall, the imaged structure agrees well with isotropic tomography results but is more coarsely resolved owing to the greater regularization in this study.

The anisotropy model reveals a ~150 km wide region of 1-2% N-S fast anisotropy, with maximum strength ≤4.1%, co-located with the low velocities in the axis of the rift (Figure 3.11). This structure spans the width of the model and increases in width and amplitude down to 170 km depth. In the upper 100 km of the model, the N-S fabric extends from eastern Fergusson/Normanby Island northwards to the Trobriand Islands, approximately coinciding with the eastern limit of the high velocity structure. Elsewhere, on the shoulders of the rift we image a predominant E-W fabric, although to the south of the rift this is not in a well-resolved region. The strong E-W fabric north of Goodenough Island is co-located with the high-velocity structure.

3.5.2 Synthetic splitting

As a further check on our tomographic model, we compare measured individual SKS splits [Eilon et al., 2014] to synthetic SKS waveforms propagated through the imaged structure, as follows: rays are traced through the model space using a 1-D velocity profile from (IASP91). Each ray is divided into 10 km segments; for each segment the mean velocity and anisotropy are interpolated from the tomography model and used to calculate the appropriately oriented elastic tensor (Equation D.6). We solve the Christoffel equations using the ray propagation vector in each segment, computing the fast and slow velocities and polarizations in the ray-based coordinate system. We then propagate an 8 s gaussian wavelet along this ray, starting with P-SV polarization...
and resolving this energy onto the fast and slow polarizations in each segment, with respective propagation velocities. Finally, this arrival is resolved onto north and east channels, and filtering and windowing is applied prior to standard splitting analysis using the Minimum Energy method [Silver and Chan, 1991].

We find that the initial polarization is preserved and there is good agreement between our simplified anisotropy approximation (Equation 3.5) using a finite frequency approach and the full solution to the Christoffel equations using a ray-based approach.

Figure 3.9: Horizontal and vertical slices through the isotropic shear velocity model (Section 3.5.1) with 10% hit quality contours shown.
Splitting time ($\delta \tau$) calculated with finite frequencies is, on average 26% smaller than that calculated with rays, because of intrinsic smoothing. The fast direction ($\varphi$) error is very small, as expected for the small incidence angle of $SK(K)S$ phases (Figure 3.2). This agreement is possible despite the multiple anisotropic layers because of the constraint on the geometry of the anisotropy to be orthogonal (NS-fast or EW-fast), leading to linearly additive splitting. This condition is unlikely to be true in the real Earth [Silver and Savage, 1994] and will be a source of error in the analysis.

The mean angular misfit (weighted by measured $\delta \tau$) between predicted and observed $SKS$ $\varphi$ is $28.2^\circ \pm 25.9$ (1$\sigma$) demonstrating that the anisotropic velocity model — obtained using only direct $S$ arrivals — is in good agreement with the $SKS$ measurements (Figure 3.3). The mean predicted $\delta \tau$ values systematically underestimate the measured values from SKS splitting; this discrepancy likely arises from the larger uncertainty in measured $\delta \tau$ (compared to $\varphi$) [Silver and Chan, 1991], over-damped
model anisotropic structure, and the limited depth extent of the well resolved model space (Section 3.4.2). The only location in which there is systematic angular misfit between observed and measured $\phi$ is in the northwest of the model where measured splitting ($\varphi \approx 0$) is orthogonal to the E-W fabric in the tomographic model. Much of this misfit comes from arrivals at stations on the Lucansay Islands, close to the edge of our model domain, where the $SKS$ rays may be sampling deeper N-S structure.

Figure 3.11: Horizontal and vertical slices through the anisotropy model (Section 3.5.1) with 10% hit quality contours shown. Positive values (blue) indicate N-S fast, and negative values (red) indicate E-W fast.
associated with overall mantle flow [Eilon et al., 2014]. Overall, SKS data — which were not used in the inversion — offer an independent confirmation of the structure in our models.

3.6 Discussion

3.6.1 Rift velocity and anisotropy structure

The models contain a slow shear velocity region demarcating the axis of the rift, consistent with structure in isotropic $V_P$ and $V_S$ models [Eilon et al., 2015] and aligned with the thinnest Moho [Abers et al., 2002, 2016], Holocene volcanic centers, and the along-strike extrapolation of the oceanic spreading centers immediately east of the array. A fast feature north of the rift axis is also seen in isotropic models; the large velocity contrast ($\delta V_{S_{av}} = 13\%$ in these models) with the axial material and the observation of intermediate depth seismicity in this structure are evidence for its being cold lithosphere, possibly a relict slab fragment left over from recent subduction [Eilon et al., 2015].

This study shows that the rift axis also contains strong N-S fast fabric with spatial extent very similar to the low-$V_{S_{av}}$ region. The pattern of anisotropy agrees well with shear wave splitting measurements of spreading-parallel fast azimuth on the DEI. We infer that the region of mantle at the center of the rift has experienced lithospheric removal and asthenospheric upwelling and has also developed a spreading-parallel anisotropic fabric: we are observing the process of lithospheric breakup in the anisotropy signal. As the continent breaks apart, increasing extensional strain and lithospheric thinning result in mantle CPO being simultaneously accrued and advected upwards within the rift axis [Tommasi et al., 1999]. A positive feedback between temperature, anisotropic viscosity, and strain may facilitate this process and explain the congruence of the slow velocity and N-S fast structures.
Models [Blackman and Kendall, 2002] and observations [Gaherty et al., 2004] have established that spreading-parallel azimuthal anisotropy dominates the shallow upper mantle in the ocean basins. It is an open question how quickly corner flow at ridges implements a horizontal fabric. Vertical flow of ascending material at the ridge axis should produce radial, rather than azimuthal, anisotropy. However, at the rift axis we observe strong, shallow azimuthal anisotropy consistent with spreading-parallel fabric (Figure 3.8). This observation, from a region on the cusp of full seafloor spreading, is in agreement with strong azimuthal anisotropy measured on the East Pacific Rise [Wolfe and Solomon, 1998] and elsewhere [Nowacki et al., 2012]. Even accounting for intrinsic smoothing associated with finite frequency fresnel zones, our results suggest that corner flow efficiently establishes a spreading fabric through extensional shear close to the spreading axis.

This process may be accelerated in continental rifts; Nielsen and Hopper [2004] demonstrate that dehydration and depletion due to melting should increase the viscosity and buoyancy of uppermost mantle within the rift axis. This phenomenon could enhance the shallow establishment of azimuthal anisotropy as it suppresses upwelling and favors predominantly horizontal strain, explaining the observed anisotropic structure.

The N-S fabric may extend below the well resolved part of the model, which would explain the synthetic SKS $\delta\tau$ underestimates (Section 3.5.2) and higher norm in the deeper layers following squeezing (Section 3.4.2). Our preferred models suggest that anisotropic structure persists deeper than velocity heterogeneities. This observation could result from deteriorating depth resolution of differential travel time data compared to SKS data. Alternatively, it may be evidence for extensional strain being accommodated in the convecting mantle where temperature changes little and deformation yields only small horizontal temperature (and hence, velocity) gradients. Deep fabric may arise from asymmetric mantle flow beneath the moving ridge seg-
ment [cf. Conder et al., 2002]. The Woodlark Rift is translating northwards at \( \sim 65 \text{ mm/yr} \) in a hot spot reference frame, resulting in asymmetric flow that could explain deep N-S fabric [Eilon et al., 2014].

A limitation of our method is the inability to resolve dip of the anisotropic symmetry axis, which is fixed to be horizontal (implicitly assuming that gradients of strain rate are vertical [Tommasi et al., 1999]). Dip can be detected from the ratio of the first two azimuthal harmonics of the splitting intensity [Chevrot, 2000]; this measurement is noisy at almost all stations in this study, but there is no evidence suggesting a dip to the symmetry axis [Eilon et al., 2014]. In principle the dip could be included in the inversion, but tests demonstrate that our data cannot discern this parameter. Fabric dip would introduce a back-azimuthally varying source of error, and (for \(<30^\circ\) dip) on average lead to overestimation of isotropic velocity and underestimation of anisotropy strength (Figure 3.2). A non-horizontal symmetry axis within this rift may result from corner-flow fabric near the spreading axis [Blackman and Kendall, 2002], a vertical component of strain as the lithosphere thins, or melt-rich layers at a dipping lithosphere-asthenosphere boundary [Holtzman and Kendall, 2010]. Despite these possibilities, shallow or negligible dip was observed at the East Pacific Rise [Hammond and Toomey, 2003].

Unlike other continental rifts such as Rio Grande [Gok et al., 2003] or the MER [Bastow et al., 2010; Hammond et al., 2010; Kendall et al., 2005], this rift is not dominated by shallow rift-parallel fabric, often inferred to result from melt lenses aligned normal to spreading [e.g. Kendall et al., 2005]. Isotropic tomography indicates that this rift is relatively magma-poor [Eilon et al., 2015], consistent with low mantle potential temperature [Ruprecht et al., 2013] and volumetrically modest calc-alkaline Holocene volcanism [Smith, 1982]. The relative paucity of melt and the greater magnitude and rate of extensional strain explain the predominance of spreading-parallel CPO in this rift compared to its tectonic analogues [Eilon et al., 2014].
3.6.2 Rift shoulder structure

The N-S fabric appears to have pushed apart a shallow E-W fabric that now lies north and south beneath the rift shoulders. This geometry indicates that prior to rifting there was an E-W “frozen-in” lithospheric fabric, likely developed during Eocene-Miocene northward convergence, terrane accretion, and lithospheric shortening [Baldwin et al., 2012; Davies and Jaques, 1984]. Observations [Huang et al., 2000; Kaviani et al., 2009] and simulations [Tommasi et al., 1999] indicate development of lithospheric anisotropy with fast azimuth parallel to the strike of orogenic belts — for the Papuan Peninsula this would correspond to an E-W fabric, as observed.

The E-W-fast structure northwest of the rift axis is greater in amplitude and depth than the conjugate feature to the south. This structure co-locates with the high-\(V_{S_{av}}\) relict slab and the boundary between the slab and the rift axis. A north- or south-dipping slab [Eilon et al., 2015] would explain the strong E-W fabric: trench-parallel splitting is commonly observed in subduction zones [Long and Silver, 2009b] and may arise from along-strike flow, b-type fabric [Karato et al., 2008], tilted radial anisotropy [Song and Kawakatsu, 2012], or even intra-slab anisotropy [Eakin et al., 2015]. Although this structure is strong and well resolved in the anisotropy model, it gives rise to the only major discrepancy between synthetic and observed SKS measurements (Section 3.5.2); perhaps because core refracted phases are mostly sampling deeper N-S fabric not traversed by direct-\(S\). Alternatively, the disagreement could arise from locally intermediate-azimuth (e.g. NE-SW) anisotropic structure in this region being mapped onto our orthogonal basis; this possibility reinforces the importance of using multiple data types to interrogate complex structure. That said, the strength of E-W anisotropy we model in this region indicates that any intermediate-azimuth structure here must have a strongly dominant E-W component, clearly differentiable from anisotropy within the rift axis.

Tectonic constraints imply a S-dipping, E-W striking subduction boundary to
the north [Yan and Kroenke, 1993], suggesting a fundamentally 2-D geometry is applicable. Despite this, the presence of a relict slab in the northwest may complicate shallow flow fields beneath the rift’s northern shoulder. Isotropic $V_P$ tomography and intermediate earthquake locations [Eilon et al., 2015] and $V_S$ constraints from this study hint that cold slab temperatures becomes less pronounced east of 150.5°E (Section 3.5.1). This may result from the E-W gradient in extension.

Strong, shallow N-S fast anisotropy in the northeast of the model, apparently coincident with the edge of the high-velocity body, could reflect toroidal flow around an eastern edge of the slab or simply extensional strain that extends further north when uninhibited by the presence of a viscous/rigid slab.

The sharp and relatively vertical velocity boundaries demarcating the rift edges imply strain localization processes involving water, chemical depletion, or mechanical controls (Eilon et al. 2015). Our results indicate asymmetry between the rift shoulders. The $\sim$5% $\delta \ln V_S$ between the rift axis and the well-resolved part of the southern shoulder is similar to that observed in other rifts [Bastow et al., 2008] and ridges [Nettles and Dziewonski, 2008]. The velocity gradient between rift axis and shoulder is much greater in the north, due to the large temperature contrast of upwelling asthenosphere juxtaposed against a cold lithospheric fragment.

### 3.6.3 Anisotropy - velocity tradeoffs

Tomographic studies routinely interpret differential travel times as a function of velocity heterogeneity alone. A key component of this study is the quantification of tradeoffs between the contributions of velocity and anisotropy to observed travel times. Inversions in which anisotropy is damped to zero provide a significantly poorer fit: the variance reduction to $\delta t$ data achieved by isotropic velocity heterogeneity alone is 69.3% (compared to 86.8% with anisotropy). Inversions in which velocity heterogeneities are damped to zero fit the $dT_{N-E}$ data only marginally less well (84.6%
compared to 87.1% for the full model) as expected from the second order control of $\Delta V_{S_{av}}$ on splitting. However, these inversions do reduce the differential travel time variance by 34% despite no velocity heterogeneity, demonstrating that a substantial portion of the signal in the $\delta t$ data is due to anisotropy alone.

We have previously argued that strong velocity perturbations in this region (as observed in the isotropic inversion - Figure 3.10) are explicable by unusually large temperature heterogeneities [Eilon et al., 2015]. We believe the more muted $\Delta V_{S_{av}}$ in the anisotropic model underestimates true variations. This phenomenon results from the increased regularization required by the anisotropic inversion, which contains more model parameters. The loss in amplitude is consistent with our synthetic tests (Figure 3.7).

The importance of a joint inversion is not only evident from an improved fit to data, but from the differences in observed structure. In models with only isotropic velocity variations, the southern rift shoulder is not clear in $V_S$, but well-defined in the $V_P$ model [Eilon et al., 2015]. For azimuthally-homogeneous data distribution, the effects of azimuthal anisotropy should, to first order, cancel out. However, by computing the average polarization of waves interacting with each node in the model, we find that the region south of the rift axis is disproportionately illuminated by waves that happen to be polarized N-S, which move slower through the E-W fabric here, counteracting the effect of colder (isotropically faster) structure. In this inversion, once the effects of anisotropy and velocity structure are separated, we retrieve a model with a clearly-defined southern boundary to the slow rift axis (Figure 3.10) emphasising the localized extension to depth.
3.7 Conclusions

We have developed a simple parameterization of anisotropy and mean shear velocity by describing elastic tensors relevant to upper mantle fabrics in terms of two orthogonal shear velocities ($V_\parallel$ and $V_\perp$). This parameterization yields approximate solutions to the Christoffel equations that are analytically differentiable, allowing efficient calculation of Fréchet derivatives that make up the data kernel of a non-linear inverse problem. By assuming the orientation of the anisotropy, on the basis of prior splitting information and the geometry of the tectonic environment, the anisotropic velocity structure is related to teleseismic $S$-wave differential travel times and splitting times measured simultaneously with three-way cross correlation. These data are iteratively inverted for 3-D velocity and anisotropy heterogeneity, using finite frequency sensitivity kernels.

This method is applied to the Woodlark Rift, Papua New Guinea, imaging a low-velocity rift axis extending westward from the nearby mid-ocean ridge tip more than 250 km into the continent congruent with shallow Moho, Holocene volcanism, and high heat flow. This study reconciles previous isotropic tomography and $SKS$ splitting measurements in this rift, providing an important depth control on anisotropy. In agreement with isotropic $P$, $S$ and surface wave images, there is a large velocity contrast between the $\sim 6\%$ slow rift axis and $\sim 5\%$ fast cold lithospheric material at $>100$ km depth to the north. Up to $4\%$ anisotropy with N-S fast azimuth has developed from 50-150 km depth within the rift axis, breaking apart a pre-existing lithospheric fabric still evident on the rift shoulders. The congruency of the N-S fabric and the low-velocity structure indicates the co-development of lithospheric thinning and extension-related CPO in the upwelling asthenosphere. The observation of strong, shallow azimuthal anisotropy within the rift axis implies efficient establishment of fabric in a corner flow regime. Jointly inverting for shear velocity and anisotropy allows us to account for, and quantify, tradeoffs between the two parameters.
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Chapter 4

*Attenuation structure of an oceanic plate from ridge to trench*

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This chapter concerns the measurement of differential attenuation using teleseisms recorded on the Cascadia Initiative’s Amphibious Array. This approach broadens our investigation of extensional environments. The complementary measurements of attenuation and velocity inform us about the physical state of the mantle beneath a mid-ocean ridge.

**Abstract**

Measurements of seismic wavespeeds and attenuation provide complementary constraints on the thermal and compositional character of the Earth’s interior. Anelastic processes reduce effective wavespeeds and cause intrinsic attenuation of seismic energy. These processes are thermally activated, but it remains an open question as to whether temperature exerts the primary control on physical dispersion in the shallow mantle. We have analysed a unique dataset of teleseismic body waves recorded on the Cascadia Initiative’s Amphibious Array, with detailed spatial sampling of an entire oceanic plate. We compute relative phase and amplitude spectra for $P$- and $S$-wave arrivals and calculate differential integrated attenuation ($\Delta t^*$) between stations on the seafloor. There is a strong age-dependency to the apparent attenuation, with particularly high attenuation ($\Delta t^*_S \sim 2.0$ s) observed at stations close to the
axes of the Juan de Fuca and Gorda ridges. The small variation in $\Delta t^*$ seen at seafloor ages $>2$ Ma is consistent with a thermal control on attenuation. The increase in $\Delta t^*$ within 2 Ma of the spreading centers requires a large, localized contribution from alternative dissipative processes. We argue that non-intrinsic sources of apparent attenuation, such as focusing and scattering, cannot produce the majority of the observed signal. Several lines of evidence point to the importance of deep melt beneath the spreading centers. We use synthetic seismic rays to probe a model mid-ocean ridge that includes elevated sub-axial water or melt. These forward models place quantitative bounds on the contributions to anelastic processes at the ridge axis.

4.1 Background

4.1.1 Attenuation at spreading centers

Mid-ocean ridges (MORs) comprise the largest and most long-lived volcanic systems worldwide. Yet the quantity, distribution, and ascent pathways of melt at MORs are poorly understood. The Juan de Fuca and Gorda ridges produce a minimum of 420 km$^3$ of melt\(^1\) per km of ridge length, per Ma. Passive upwelling models [McKenzie and Sclater, 1969] imply decompression melting within a roughly triangular melt region with a basal width of 100-200 km [Katz, 2010]. Despite this, the majority of surficial volcanism emanates from within just 1-2 km of the Juan de Fuca ridge axis [Canales et al., 2005; Wilson, 1992] and surveys of ridge structure show that full oceanic crustal thickness is achieved within 1-3 km of the axis [Carbotte and Nedimovic, 2008; Cudrak and Clowes, 1993], implying efficient localization processes.

The scale and mechanism of these localization processes have not been precisely characterized. Two dominant paradigms have been proposed: narrow mantle up-

\(^1\)Assuming crustal thickness of 7 km, full spreading rate of 60 mm/yr, and 100% efficient transfer of melt to an igneous crust with no mantle intrusions.
welling due to dynamic effects, or wide mantle upwelling with lateral melt transport. The former requires low-viscosity and positive-buoyancy beneath the ridge axis, putatively maintained by substantial retained melt [Morgan and Parmentier, 1987; Scott and Stevenson, 1989; Su and Buck, 1993]. Mechanisms suggested for the latter include: flow along a compaction or dehydration boundary [Karato, 2012; Sparks and Parmentier, 1991; Spiegelman, 1993], porosity-driven pressure gradients [Katz, 2010] related to grain size [Turner et al., 2015; Wark and Watson, 2000], shear banding [e.g., Holtzman and Kendall, 2010; Holtzman et al., 2003; Katz et al., 2006], and reaction-infiltration instabilities [Spiegelman et al., 2001]. Dunite channels in ophiolites have been taken as evidence that segregated melt is funnelled from depth towards the ridge axis [Kelemen and Shimizu, 1995]. To date, however, we lack direct observations to support this inference.

Seismological studies of oceanic spreading centers show a seismically slow zone down to ≥200 km [Dalton et al., 2014; Hammond and Toomey, 2003; Nettles and Dziewonski, 2008; Webb and Forsyth, 1998] with shear velocities depressed 3–6 % relative to >10 Ma mantle off axis. Magnetotelluric studies at separate locations along the East Pacific Rise (EPR) show a highly conductive feature in the upper mantle directly beneath the ridge axis [Baba et al., 2006; Key et al., 2013]. This feature is approximately 40 km wide, and extends from 20–40 km down to 80–130 km depth. The anisotropic inversion of Baba et al. [2006] demonstrated that subridge vertical resistivity is markedly smaller than resistivity in horizontal directions, indicating predominantly vertical connectivity of electrically conductive pathways. These findings are suggestive of significant in situ melt beneath the ridge. Despite this, a narrow, melt-rich zone focussed beneath the ridge axis has not been previously imaged at mantle depths.

Global models of upper mantle attenuation have observed low Q in the upper 200 km beneath MORs, with $Q_S$ as low as 50-60 in young oceans inferred from 50-250
Rayleigh waves [Dalton et al., 2008, 2009]. These studies have resolution at wavelengths of several thousands of km and so provide lower bounds for extrema of $dQ^{-1}$.

Yang et al. [2007] conducted the most detailed regional study of attenuation at a MOR to date. They used surface waves to simultaneously solve for $V_S$ and $Q_S$ (which have related Rayleigh wave sensitivity kernels [Dalton and Ekström, 2006]) at the EPR and found that $Q_S$ is consistent with asthenospheric values ($\sim 70 - 110$) in the upper 120 km beneath the ridge axis. Using a classic functional form for $Q(\omega, T, P)$ and $V_S(\omega, T, P)$ [Yang et al., 2007, Equations 2,3] they showed that no combination of parameters simultaneously fit their velocity and attenuation measurements, attributing the apparent wavespeed deficit to the presence of melt and asserting that melt does not cause attenuation within the seismic frequency band. However, this careful study was limited by coarse resolution inherent to the use of long-period Rayleigh waves. An alternative possibility is that intrinsic spatial averaging caused Yang et al. [2007] to overestimate shallow $Q_S$; it is, perhaps, surprising that their observations ($Q_S \geq 70$ at periods of 16-67 s)) so close to the ridge axis imply a less attenuating upper mantle than do global best fitting values for 0–25 Myr oceans ($Q_S \geq 60$ [Dalton et al., 2008]) within a degree 12 model.

Simultaneous observations of velocity and attenuation can help discriminate the importance and fraction of melt. Attenuation at mid-ocean ridges has received relatively little study, with the primary attention on shallow attenuation from crustal magma/hydrothermal systems [e.g., White and Clowes, 1994; Wilcock et al., 1995, 1992]. Different studies have asserted that the upper mantle at MORs is highly attenuating [Sato and Sacks, 1989; Solomon, 1973] or similar to ambient asthenosphere [Ding and Grand, 1993; Yang et al., 2007]. Here we use a dense dataset of travel times and integrated attenuation to place bounds on the temperature and melt character of a mid-ocean ridge and cooling oceanic plate.
4.1.2 Anelasticity within the Earth

The passage of seismic waves through the Earth’s interior is determined by the elastic and anelastic behavior of rocks. A variety of physical mechanisms contribute towards dissipation of seismic energy (attenuation) producing a deterministic, if complex, dependency of attenuation on temperature, grain size, composition, and the presence of melt and/or water \((T, d, X, \phi, C_{H_2O})\). Simultaneous measurements of seismic travel times and integrated attenuation elucidate the state of the Earth’s interior, where we can discriminate between variations in temperature, melt, composition, etc. by exploiting their distinct influences on velocity versus attenuation [e.g., Dalton et al., 2009].

Quantitative parametrizations of anelastic scaling relationships \(V_S(P, T, d, X, \phi, C_{H_2O})\) and \(Q_\mu(P, T, d, X, \phi, C_{H_2O})\) rely largely on empirical fits made to laboratory creep and forced-torsion oscillation experiments [e.g., Cooper, 2002; Jackson and Faul, 2010; Kampfmann and Berckhemer, 1985; McCarthy et al., 2011; Xu et al., 2004]. By necessity, these experiments are conducted at normalized frequencies\(^2\) and equilibrated grain sizes approximately three orders of magnitude removed from the conditions pertinent to the propagation of seismic waves through the mantle. Thus, derived scaling relationships require extrapolation; good agreement between body wave, surface wave, and predicted attenuation measurements has proved elusive [Dalton et al., 2008; Faul and Jackson, 2015]. The discrepancy requires improved observations of \(V_S, Q\) made at body wave frequencies, particularly in the oceans where we have good constraints on thermal structure [McKenzie et al., 2005; Priestley and McKenzie, 2013].

Jackson and Faul [2010] offer one formalism for the computation of anelastic effects, describing the complex compliance (the inverse of the shear modulus, \(\mu\)) as

\(^2\)“Normalized frequency” is defined with respect to the Maxwell frequency: \(f_N = f \cdot \tau_M\) [McCarthy et al., 2011].
the sum of its elastic and non-elastic\(^3\) parts: \( J^\ast(\omega) = J_1(\omega) + iJ_2(\omega) \). These can be calculated using a Burgers model:

\[
J_1(\omega) = J_U \left\{ 1 + \Delta \int_{-\infty}^{\infty} \frac{D(\tau)d\tau}{(1 + \omega^2\tau^2)} \right\} \tag{4.1}
\]

\[
J_2(\omega) = J_U \left\{ \omega \Delta \int_{-\infty}^{\infty} \frac{\tau D(\tau)d\tau}{(1 + \omega^2\tau^2)} + \frac{1}{\omega\tau_M} \right\} \tag{4.2}
\]

The unrelaxed modulus, \( J_U \), is a function of pressure, temperature, and composition, \( \tau_M \) is the Maxwell time, \( \Delta \) is the strength of the Burgers element (i.e. the fractional weakening of the anelastic response compared to the anharmonic response). \( D(\tau) \) is the relaxation spectrum, a combination of a “high-temperature background” (HTB) spectrum caused by transient diffusion creep [Cooper, 2002; Jackson and Faul, 2010; McCarthy et al., 2011] and a semi-empirical high-frequency absorption peak [e.g., Jackson and Faul, 2010; Sundberg and Cooper, 2010; Xu et al., 2004], \( D(\tau) = D_B(\tau) + D_P(\tau) \) where:

\[
D_B(\tau) = \frac{\alpha \tau^{\alpha - 1}}{\tau_H^{\alpha - 1} - \tau_L^{\alpha - 1}} \tag{4.3}
\]

\[
D_P(\tau) = \frac{1}{\sigma \sqrt{2\pi}} \exp \left\{ -\frac{\ln((\tau/\tau_P)^2)}{2\sigma^2} \right\} \tag{4.4}
\]

and \( \alpha \) is the frequency dependency. The relevant timescales have Arrhenius-type behavior and are defined\(^4\) with respect to reference conditions (subscript “\( R \)”):

\[
\tau_i = \tau_{iR} \left( \frac{d}{d_R} \right)^{m_i} \exp \left[ \left( \frac{E^*}{R} \right) \left( \frac{1}{T} - \frac{1}{T_R} \right) + \left( \frac{V^*}{R} \right) \left( \frac{P}{T} - \frac{P_R}{T_R} \right) \right] \tag{4.5}
\]

assuming all viscous and viscoelastic processes scale similarly [Abers et al., 2014; Olugboji et al., 2013]. Although we employ the form of attenuation spectrum, \( D(\tau) \), from Jackson and Faul [2010], the relaxation time scaling (Equation 4.5) is applicable

\(^3\)Where “non-elastic” comprises anelastic response plus viscous response.

\(^4\) Where \( \tau_i \) are characteristic timescale, \( E^* \) is activation energy, \( V^* \) activation volume, \( R \) is the Boltzmann constant, \( m \) is the grain size exponent. Subscript “\( i \)” refers to each of the timescales for which this functional form is applied: the high and low bounds of the HTB spectrum (\( \tau_H; \tau_L \)), the Maxwell time (\( \tau_M \)), and the peak time (\( \tau_P \)). Some argue that the peak has a different grain size exponent (\( m_i \)) than the HTB [Jackson and Faul, 2010].
to all formalisms (Andrade, master curve, etc.). Some data hint that the height and width of the low-temperature absorption peak depend on temperature and grain size [Takei et al., 2014], but experimental data confirming and characterizing this phenomenon for olivine samples are not yet available. While the high-frequency peak has been identified in laboratory experiments [Jackson and Faul, 2010; Sundberg and Cooper, 2010; Xu et al., 2004], it has never been confirmed by seismic observations.

Attenuation, or energy loss per period, is described by the quality factor, $Q$, equal to the ratio of the imaginary and real parts of the complex compliance: $\Delta W/W = Q^{-1} = J_2/J_1$. In addition to amplitude reduction, anelasticity leads to modulus dispersion according to the Kramers-Kronig relation for $Q \gg 1$ [e.g., Anderson and Minster, 1979]:

$$\mu(\omega) = \mu(\omega_0) \left[1 + \frac{1}{\pi Q} \ln \left(\frac{\omega}{\omega_0}\right)\right]^2$$

(4.6)

where frequency independence of $Q$ is assumed (the consequences of relaxing this assumption are shown in Section 4.3.1 and Appendix G). Thus, attenuated seismic waves are modified in both amplitude and phase. The effects of attenuation can be described by the delayed attenuation operator ($\mathcal{D}$) in the frequency domain, composed of an amplitude term and a phase term: $\mathcal{D}(\omega) = A(\omega) \exp(\imath \phi(\omega))$, where:

$$A(\omega) = \exp\left(-\frac{1}{2} \omega t^*\right)$$

(4.7)

$$\phi(\omega) = \frac{1}{\pi} t^* \ln\left(\frac{\omega}{\omega_0}\right)$$

(4.8)

and $\omega_\infty = 2\pi e^\pi$; see Appendix G. Here we have introduced the useful parameter $t^*$, which describes the integrated attenuation along a raypath:

$$t^* = \int V^{-1} Q^{-1} ds$$

(4.9)

We may consider a waveform, $g(t)$ that passes through two structures with different attenuation and velocity structure to be recorded at different stations. If we narrow-
band filter the recorded waveforms to $\omega_i$, we will obtain (Appendix G):

$$g'_1(t)\big|_{\omega_i} = S_1 A_1(\omega_i)\ g(t - \phi_1(\omega_i) - T_1 + f(\Delta_1))\big|_{\omega_i} \quad (4.10a)$$

$$g'_2(t)\big|_{\omega_i} = S_2 A_2(\omega_i)\ g(t - \phi_2(\omega_i) - T_2 + f(\Delta_2))\big|_{\omega_i} \quad (4.10b)$$

That is, the two waveforms that are shifted and scaled versions of the original, multiplied by station amplification terms ($S_i$). Part of the time lag is due to anelasticity ($\phi$), and part due to anharmonic wave speed differences within the region of interest (where $T_2 - T_1 = \delta T$). We account for distance-related travel times, $f(\Delta)$, by subtracting predicted travel times using the IASP91 1-D reference Earth model, but global 3-D heterogeneity will introduce a frequency independent phase delay that we must solve for by inversion. We can simultaneously exploit the relative amplitude and phase delay spectra at this pair of stations to compute differential attenuation, $t^*_2 - t^*_1 = \Delta t^*$. We combine Equations 4.7, 4.8, and 4.10 to obtain:

$$R(\omega) = \frac{S_2 A_2(\omega)}{S_1 A_1(\omega)} = \frac{S_2}{S_1} \exp\left(-\frac{1}{2} \omega \Delta t^*\right) \quad (4.11)$$

$$\Delta \psi(\omega) = (\phi_2(\omega) - \phi_1(\omega)) + \delta T = \frac{1}{\pi} \Delta t^* \ln\left(\frac{\omega_\infty}{\omega}\right) + \delta T \quad (4.12)$$

These form a system of equations that can be solved by least squares methods for best fitting $\Delta t^*$ (Figure 4.1).

### 4.2 Data and methods

The Amphibious Array deployment of the Cascadia Initiative provides an unprecedented opportunity to study an entire ocean plate from ridge to trench. 70 broadband ocean bottom seismometers (OBS) were deployed in four year-long phases from 09/2011 - 09/2015, spending at least 2 years at each location in a $\sim$70 km spaced grid. We present data recorded at 87 unique station locations. Global body wave teleseisms recorded on these instruments allow us to study the Juan de Fuca and Gorda plates with excellent lateral resolution.
Figure 4.1: Example of a station-station differential $t^*$ measurement for which best fitting $\Delta t^* = 2.0$ s. Top: displacement waveforms (T-component) showing time relative to predicted S-wave arrival at stations 1 (on ∼4.9 Ma seafloor) and 2 (close to the ridge). Blue lines indicate window for spectra calculation. Middle: logarithm of amplitude ratio measured for each of the narrow-band filters (Figure G.1). Symbol size inversely proportional to cross correlation of optimally shifted and scaled traces at each frequency. Black line: amplitude spectral ratio calculated using Thomson multitaper method Thomson [1982]. Green lines: best fit predictions from slope analysis (solid line for $\alpha = 0$ and dashed assuming $\alpha = 0.27$). Arrow: frequency at which signal spectra cross the pre-arrival noise spectrum. Bottom: differential phase spectrum; symbols as above. Based on synthetic tests, we penalize amplitude and phase misfit in the ratio 5:1, but both spectra are well fit.

We collect $P$ and $S$ wave data from all $M_W > 6.25$ teleseismic ($\Delta = 25 - 90^\circ$) earthquakes recorded on Cascadia Initiative OBS stations (IRIS Network codes 7A, 7D, X9\(^5\)) between 2011-08-24 and 2014-06-23. Instrument response is removed and

\(^{5}\)X9 corresponds to the Plate Boundary Evolution and Physics at an Oceanic Transform Fault System experiment instrumenting the Blanco Transform - only 5 of these stations had open data.
seismograms are integrated to displacement before we rotate horizontal channels to true North/East orientation by using best-fitting correction azimuths determined by maximizing receiver function radial components [Janiszewski and Abers, 2015, and pers. comm.] or Rayleigh-wave polarization (OBSIP Cascadia Horizontal Orientations Reports 2011-12, 2012-2013, 2013-2014).

For each earthquake in our catalog, we measure differential travel times ($\delta T$) using cross-correlation [VanDecar and Crosson, 1990] with a $0.2 - 1$ Hz filter ($P$-waves) or $0.083 - 1$ Hz filter ($S$-waves) applied to a hand-picked window around the first arrival (Figure 4.2). We discard all traces with signal-to-noise ratio smaller than 10.

Differential $t^*$ is computed pair-wise between all combinations of stations by selecting a 35 second window starting 5 seconds before the predicted body wave arrival and applying a comb of 30 narrow-band filters spaced equally in center-frequency from 0.05 - 1 Hz (Figure G.1). At each frequency, the phase shift ($\Delta \psi_{ij}$) and amplitude ratio ($R_{ij}$) that best transform trace $i$ to match trace $j$ are computed, to construct the relative phase and amplitude spectra, $\Delta \psi(\omega)$ and $R(\omega)$. We also obtain a vector of weights equal to the final agreement between $i$ and optimally transformed $j$ (Figure 4.1). Frequencies for which the final cross-correlation coefficient is less than 50% are discarded completely. We also discard frequencies for which the signal does not exceed the pre-event noise. If fewer than four frequencies in the comb satisfy the criteria, that station-station pair is ignored.

By assuming that effects of attenuation may be expressed in the frequency domain by the delayed attenuation operator we can show (Equations 4.11 and 4.12):

$$\ln (R) = k_1 - \pi f \Delta t^*$$  \hspace{1cm} (4.13)

$$\Delta \psi = k_2 + \frac{1}{\pi} \ln (f) \Delta t^*$$  \hspace{1cm} (4.14)

where $f$ are the center frequencies of each filter in the comb, $k_1$ is a constant related to the difference in frequency-independent amplification between the two stations, and
Figure 4.2: Example of waveforms and differential spectra recorded at Juan de Fuca OBS stations from an event on 03-04-2014 at a distance of \( \sim 84^\circ \), from a back azimuth of \( \sim 129^\circ \). a) S-waves (displacement) recorded on the T-component, aligned by cross correlation, colored/arranged by distance to ridge, and plotted with a 0.05-2 Hz 4th order Butterworth filter. Dashed lines: data window for calculation of spectra. b) Differential amplitude (upper) and phase (lower) measurements as a function of frequency. Values at each station are computed relative to the event stack and then averaged across stations in 1 Ma age bins from 0 – 8 Ma. Colors are by bin center age.

\( k_2 \) is a constant related to differences in anharmonic wave speed or global propagation delays unaccounted for by a 1-D model.

We compute \( \Delta t^* \) at each station by a one-step constrained least squares fitting of all of the pairwise differential amplitude and phase measurements, weighting as above and solving for each station’s gain term (Figure 4.3). We have also tested the approach of solving individually for \( \Delta t^* \) at each pair of stations by linear regression (Figure 4.1) and then using constrained least squared minimization to calculate best-fitting \( \Delta t^* \) at each station for that earthquake [Hwang and Ritsema, 2011]. The two
Figure 4.3: Example of differential attenuation recorded at Juan de Fuca OBS stations for the same event as Figure 4.2. a) Map view of station $\Delta t^*_S$ computed by constrained least squares from all pairwise combinations of station-station spectral ratios. 1 Ma seafloor age contours interpolated from [Wilson, 1993]. b) Top: vertically exaggerated topography along section, with relevant tectonic features (DF: deformation front). Middle: station $\Delta t^*_S$ projected onto the 2-D section. Green dots show spectral ratio estimates for comparison. Bottom: $\delta T_S$ obtained by cross correlation. Red line: predicted contribution to $\delta T$ from topography and sediments alone.
approaches yield extremely similar results, but synthetic tests indicate that the former approach is more stable. We do not further interpret the station amplification terms. The non-co-temporal northern (years 1 and 3) and southern (year 2) deployments are linked by land stations that record throughout.

On the OBS stations, we record a total of 435 and 414 individual $\Delta t^*$ values for $S$- and $P$-waves, respectively. We measure 679 and 419 $\delta T$ values for $S$- and $P$-waves, respectively.

$t^*$ measurements have been shown to vary with the frequency range over which spectral slope fits are computed [Cafferky and Schmandt, 2015; Li et al., 2015; Roth et al., 1999]. Variation may arise from true frequency-dependency of $Q$, or instabilities in slope fitting of noisy data. Our frequency range is limited by the noise (which equals the signal at $f > 1$ Hz for our teleseisms) and the window length (35 s). We tested several choices of frequency ranges between 1–0.033 Hz (1–30 s) by computing station averaged $\Delta t^*_S$ values for 10 most widely recorded events and plotting the average and standard deviation of the results as a function of age (Figure 4.4). The narrower passbands at both low frequency (0.067–0.033 Hz and 0.1–0.033 Hz, i.e., 15–30 and 10–30 s) and high frequency (1–0.1 Hz, i.e., 1-10 s) ends of the spectrum yield much more unstable results, with large inter-event variation and highly discontinuous age trends. This pattern is also evident for $\Delta t^*_P$ measurements, which evince even more instability at short periods. Based on these tests, we select 1–0.05 Hz (1–20 s) as the optimal measurement band; this range provides the most stable results between different events, and smooth profiles of $\Delta t^*_S$ and $\Delta t^*_P$ with age.

4.2.1 Tilt and compliance tests

Ocean noise comprises a significant portion of the signal on some OBS stations; frequency-dependent noise could bias measurements of apparent attenuation. We conduct tests wherein we apply tilt and compliance corrections [Bell et al., 2014;
Crawford and Webb, 2000] to the seismometer vertical components. This treatment exploits the coherence between vertical and horizontal and/or pressure channels for signal that arises from tilt or compliance noise. The noise correction is only possible on the vertical component, so we apply it to all vertical channels and re-calculate ∆t_p^* using corrected Z-components.

The correction itself may introduce bias: vertical-pressure coherence is predicted to deteriorate at frequencies higher than $\sqrt{g/2\pi h_w}$, where $g$ is gravitational acceleration and $h_w$ is water depth (in m) [Bell et al., 2014; Crawford and Webb, 1998]. Therefore, for a 200 m deep OBS station, the compliance correction may remove noise signal at longer periods than 11 s, while leaving higher frequencies unchanged. As a consequence, the signal would be artificially enriched in high-frequency energy and appear less attenuated.

Despite these potential pitfalls, our tests with and without noise correction on five well-recorded events did not significantly affect the distribution or pattern of measured ∆t_p^* on deep water OBS stations. This is partly because these events already had high signal-to-noise ratio and partly because the noise correction mainly applies at noisy shelf stations that already have apparent artefacts. Given these results, we do not believe ocean noise imparts strong bias for deepwater stations.
This result is not too surprising; our measurement technique is designed to only use portions of the signal spectrum that are coherent (in the time domain) between stations. Random noise will not satisfy this criterion, so noisy traces are automatically eliminated by our quality control criteria. Thus, noise corrections might provide us with more data, but are not expected to greatly change the results. Importantly, noise corrections are only available for vertical channels, and so we cannot remove noise affecting $Q_S$ measurements. The results presented below use verticals that have not been corrected for tilt or compliance noise.

### 4.3 Results

Patterns of back-azimuthal and inter-station variations indicate spatial coherency to observed $\Delta t^*$ (Figure 4.5). $\Delta t^*_S \sim 2.0$ s is recorded across a range of back-azimuths at all stations close to the ridge axes, while stations on older crust are uniformly less attenuating ($\Delta t^*_S < 2.0$ s). This phenomenon is observed for both Juan de Fuca and Gorda ridges. Juan de Fuca ridge stations evince a slight (27%) bias towards higher $\Delta t^*_S$ for arrivals from the west, suggestive of ridge asymmetry that has been noted previously [Davis and Karsten, 1986, and Bell et al., submitted]. The highest values of $\Delta t^*$ are seen on the axis of the Gorda ridge and close to the southern part of the axis of the Juan de Fuca ridge. In comparison, stations on the northern part of the Juan de Fuca ridge record noticeably less attenuation (by $\sim 1.0$ s), although they are still more attenuating than off-ridge stations.

Several stations in the ‘focus arrays’ on the continental shelf record high values (>3.0 s) of apparent attenuation, with >2.0 s variations in measured $\Delta t^*$ between stations separated by just 15 km. This signal cannot originate in the mantle beneath the stations, but is likely the product of shallow scattering or soft-sediment absorption processes.
Figure 4.5: Maps of differential attenuation, $\Delta t^*$, recorded at OBS station for $S$-waves (a) and $P$-waves (b). Radial spokes show individual arrivals at their incoming azimuth, while central circles show least squares station average terms, having accounted for event $\Delta t^*$ values. Open circles show land stations used to link Juan de Fuca and Gorda arrays. Boxes in (b) show three areas: north Juan de Fuca (mauve), south Juan de Fuca (red), and Gorda (blue).

Figure 4.7 shows the systematic relationship between $\Delta t^*$ and distance from the ridge. In general, the most highly attenuating stations are close to the ridge axis; these stations also evince the maximum travel time delays. With increasing crustal age, stations record decreasing attenuation and faster travel times. $S$-wave station-averaged attenuation values show a gradual trend from 10 Ma to 2 Ma. Around 2 Ma (mean ridge distance $\sim$70 km) there is a steeper increase to uniformly large values of attenuation close to the ridge on $\lt$2 Ma plate. This step is most clear for southern Juan de Fuca stations (red), while no northern Juan de Fuca stations
Figure 4.6: Maps of differential travel time recorded at OBS station for S-waves (a) and P-waves (b). Radial spokes show individual arrivals at their incoming azimuth, while central circles show least squares station average terms, having accounted for event terms. Travel times are corrected for station elevation assuming crustal velocity of $V_P = 6.5$ km/s or $V_S = 3.4$ km/s. We do not correct for sediment thickness. Open circles show land stations used to link Juan de Fuca and Gorda arrays.

(magenta) attain such high values. The Gorda stations (blue) have a more gradual decrease with age, although in map view (Figure 4.5) the discontinuous increase east of the ridge axis is apparent.

In map view, $\Delta t_P^*$ shows a less clear spatial pattern than the S-wave data (Figure 4.5). However, when plotted by age, P-wave attenuation measurements are consistent with their S-wave counterparts. The majority of more attenuating ($\Delta t_P^* \leq 0.5$ s) stations are close to the ridge axes, while older abyssal plain stations show a gradual or absent trend with age. Although this trend contains more scatter (and the majority
of stations on the abyssal plain had fewer $P$ attenuation measurements than $S$), the same age trends are evident including a discontinuous step in $\Delta t^*$ at $\sim$2 Ma.

Unexpectedly, OBS stations record mean $\Delta t^*_{S}$ that is $\sim$1.5 s lower than their land-based counterparts (Figure H.1). This observation suggests that the continental plate and/or arc system contribute substantially to $t^*$. This may be a local feature; other studies show the Cascadia margin is unusually attenuating [Dalton et al., 2008; Ma et al., 2016]. Interpretation of attenuation on land will be the focus of a later study.

Differential travel times (Figure 4.6) are highly consistent with the attenuation results. The most delayed arrivals are recorded close to the ridge axes, with negative travel times residuals (indicating faster structure) at greater plate ages. As with attenuation, the southern Juan de Fuca ridge stations are moderately ($\sim$0.5 s) slower than their counterparts north of $\sim$47°N. The clearest discrepancy between $\Delta t^*_{S}$ and $\delta T^*_{S}$ patterns is that whereas Gorda attenuation is high at the ridge but decreases

Figure 4.7: Station-averaged $\Delta t^*$ for $S$-waves (top) and $P$-waves (bottom) measured at OBS stations, plotted as a function of age of oceanic crust [Wilson, 1993]. Black curves indicate theoretical predictions from [Faul and Jackson, 2005] computed at three different grain sizes, for plate cooling model with $T_p = 1350^\circ$C. Attenuation averages are corrected for event terms, as above. Mauve dots: north Juan de Fuca stations; red dots: south Juan de Fuca stations; blue dots: Gorda stations (see Figure 4.5 for area designation). Symbol size $\propto 1/N_{\text{obs}}$. Only stations with $N_{\text{obs}} \geq 5$ plotted.
with distance, travel time residuals throughout the Gorda deformation zone are large, and barely decrease with age (Figure 4.8). For both Juan de Fuca and Gorda plates, the $\delta T$ values do not mimic the sharp step at 2 Ma seen in the attenuation results.

As expected, the $S$-waves display greater travel time variations than $P$-waves, consonant with anelastic effects more strongly affecting the shear modulus. Similar to the attenuation results, the $P$ arrival times are not noticeably delayed compared to stations on the abyssal plain.

The influence of shelf sediments is evident from extremely delayed arrivals of both $S$- and $P$-waves on the continental shelf, with short wavelength scatter in average station $\delta T$. The implied slow shallow wave speeds confirm that we should not give much credence to high apparent attenuation measured at shelf stations.
4.3.1 Frequency dependency

Hitherto we have assumed that \( Q \) is constant with frequency. This need not be the case. Theory \([Minster and Anderson, 1981]\), experiments \([Jackson and Faul, 2010; Jackson et al., 2002]\), and observations \([Lekic et al., 2009; Sipkin and Jordan, 1979]\) indicate that the Earth’s absorption spectrum may have non-zero slope at seismic frequencies. This should be resolvable from \( t^* \) measurements \([Bellis and Holtzman, 2014]\). The frequency dependency can be described by \( \alpha \), where we posit \( Q(\omega) = Q_0 (\omega/\omega_0)^\alpha \) such that \( \log(Q) \) approximates a linear slope in frequency space, with \( \alpha \) measured in the range 0 - 0.5 \([Faul and Jackson, 2005; Lekic et al., 2009; Stachnik, 2004]\). \( Q_0 \) is defined at the reference frequency, \( \omega_0 \). This approximation may be locally reasonable over small \( \omega \) range, but break down over large frequency ranges \([McCarthy et al., 2011]\). If the slope of Earth’s absorption spectrum is markedly non-linear in the seismic frequency band then observed \( \alpha \) will depend on the depth, spatial sensitivity and frequency content of data with which it is constrained. If \( Q \) in the Earth is strongly frequency dependent (with \( \alpha > 0 \)), it would heighten the discrepancy in measured attenuation between our results and those of \([Yang et al., 2007]\), made at surface-wave frequencies.

Using a frequency-domain attenuation operator that accounts for frequency-dependent \( Q \) \([Anderson and Minster, 1979; Karato, 1993; Minster and Anderson, 1981]\) (Section G.3) we modify Equations 4.13 and 4.14 for non-zero \( \alpha \):

\[
\ln(R) = k_1 - \pi f^{1-\alpha} \Delta t^*_0 \tag{4.15}
\]
\[
\Delta \psi = k_2 + \frac{1}{2} \cot\left(\frac{\alpha \pi}{2}\right) f^{-\alpha} \Delta t^*_0 \tag{4.16}
\]

Here \( \Delta t^*_0 \) refers to the value of differential attenuation corrected to 1 Hz by accounting for non-zero \( \alpha \), such that \( \Delta t^*(\omega) = \Delta t^*_0 \omega^{-\alpha} \). \( k_1 \) and \( k_2 \) are constants related to the difference in gain between the two stations, and any difference in anharmonic wave speed, respectively.
These expressions are non-linear functions of the frequency exponent, $\alpha$; we grid search over values of $\alpha$ in the range 0-0.9, seeking the value that minimizes the global misfit to all station-station differential amplitude and phase measurements (including both land and OBS stations). We find that best-fitting $\alpha$ is approximately 0.2 (Figure 4.9), although only $\alpha > 0.6$ is ruled out at the 1$\sigma$ significance level and only $\alpha > 0.9$ is ruled out at the 95% significance level. $\alpha$ between 0 and 0.3 is consistent with previous observations [Lekic et al., 2009] and close to experimentally estimated $\alpha$ values [Jackson and Faul, 2010; Jackson et al., 2002; McCarthy et al., 2011]. Since we cannot show that $\alpha > 0$ is statistically required by the data, we proceed with interpreting the results assuming $Q$ is frequency independent.

The above approach constrains the minimum-misfit $\alpha$ value across the entire measurement frequency range (1 - 0.05 Hz) but does not reckon with possible changes in frequency dependency within that passband [Sipkin and Jordan, 1979]. The presence of a high-frequency absorption peak could produce such a slope change, which might increase apparent $Q$. This possibility is explored in Section 4.4.4.
Figure 4.10: Station averaged $\Delta t^*_S$ versus $\Delta t^*_P$ (a) and $\delta T_S$ versus $\delta T_P$ (b), with trend lines fits by orthogonal regression to $\leq 7$ Ma stations, using an L1 norm and weighting by the number of $P$ and $S$ observations contributing to each station mean. Shaded regions indicate $1\sigma$ uncertainties. Only non-shelf stations with $\geq 5$ $S$ and $P$ observations are included. a) Best fitting $Q_P/Q_S = 2.40 \pm 1.43$. b) Best fitting $\Delta V_P/\Delta V_S = 1.62 \pm 0.95$. Light shaded line shows fit to $\geq 7$ Ma station averages.

4.4 Discussion

4.4.1 Implications for attenuation mechanisms

The ratio of $P$ to $S$ wave attenuation depends on the relative values of shear and bulk attenuation, $Q_\mu$ and $Q_\kappa$ (as well as the $V_P/V_S$ ratio). In the case that $Q_\kappa$ is extremely high, as is often assumed, a Poisson solid will have $Q_P = 2.25 Q_S$ [Anderson et al., 1965]. Despite this, values from 1 – 1.75 are often observed in practice [Pozgay et al., 2009; Roth et al., 1999]. Figure 4.10a shows that linear fitting by orthogonal regression to values of $\Delta t^*_S$ versus $\Delta t^*_P$ observed on our OBS stations yields an estimated $Q_P/Q_S = 2.40$ (with $Q_P/Q_S \geq 1.35$ to $1\sigma$) indicating that the assumption of negligible bulk attenuation is appropriate in the oceans.
Comparisons of $S$- and $P$-wave differential travel times provides some proxy for the physical mechanism responsible for velocity heterogeneities (cf. Section 2.6.2). The relative perturbations to $V_P$ and $V_S$ may be estimated from the slope of $\delta T_S/\delta T_P$ (divided by the $V_P/V_S$ ratio, assumed to be 1.73). If temperature alone drives variations in velocity beneath the array, we expect a shallow slope ($1.2 \leq \delta \ln V_S/\delta \ln V_P \leq 2.0$ [Anderson et al., 1992]), while the presence of melt disproportionately affects the shear modulus and produces larger $\delta T_S/\delta T_P$. In the limit of small velocity variations, a perturbation to only the shear modulus predicts $\delta \ln V_S/\delta \ln V_P = 2.25$ for a Poisson solid. The best fit value from station-averaged travel time residuals (using only $\leq 7Ma$ stations) is $\delta \ln V_S/\delta \ln V_P = 1.62$ (Figure 4.10b). This value is within the range expected for temperature but its $1\sigma$ upper bound of 2.30 does not rule out melt playing an important role beneath the array. The smaller $\delta \ln V_S/\delta \ln V_P$ for stations at greater ages implies little influence of melt further from the ridge axis.

Another way of testing this is to consider relative perturbations to $\delta T_S$ and $\Delta T^*_S$ (Figure 4.11). We overlay station average $\delta T_S$ and $\Delta T^*_S$ with the expected variation in these parameters for thermal variations between $1150^\circ C$ and $1400^\circ C$ at 1 GPa and for 2 mm grains. These conditions should be relatively favorable for shear attenuation. The slope of the prediction from thermal perturbations is too shallow to explain the large relative values of $\Delta T^*_S/\delta T_S$. This exercise demonstrates either that there is some contribution to attenuation in addition to thermally activated anelastic processes, or that our characterization of these processes (based on experimentally derived scaling relationships [Jackson and Faul, 2010]) is highly inaccurate.

The observed values of $\Delta t^*_S$ at the ridge (up to $\sim 2.0$ s) are surprisingly large. Global and local mantle velocity models indicate that the seismic signature of passive spreading ridges extends down to $\sim 200$ km depth [Hammond and Toomey, 2003; Nettles and Dziewonski, 2008], in accordance with predicted perturbations to temperature and the onset of melt [Dasgupta et al., 2013]. Flanagan and Wiens [1994]
show that $sS-S\Delta t^*$ largely saturates for earthquake foci greater than 200 km depth.

We surmise that contributions to ridge-related attenuation also originate above this depth. Figure 4.12 illustrates that in order to achieve $\Delta t^*_S = 2.0$ s over <200 km path lengths we require mean $Q_S \approx 25$. This value of mean $Q$ is a upper bound because measured $\Delta t^*$ represent deviations above some non-zero value of absolute $t^*$ accrued in the shallow mantle beneath all stations. If we were to assume frequency dependency with $\alpha = 0.3$, the implied $Q_S$ at 1 Hz would be greater ($\approx 32$).

Some of the lowest previously observed values of mantle $Q_S$ are at the Lau back-arc, where Wei et al. [2014] imaged $Q_S < 25$ in the sub-arc mantle wedge using intraslab-earthquake $t^*$ tomography sensitive to absolute $Q$ values at frequencies of 0.1-2 Hz. Our results require similarly low $Q$ values, apparently over a greater spatial scale. Such a low-$Q$ region would be consistent with the finding of [Solomon, 1973] who used back-azimuthal variation in $t^*$ from an oceanic fracture zone earthquake to argue for a $\geq 50$ km wide zone of $Q \leq 10$ beneath the mid-Atlantic ridge.
Figure 4.12: Predicted values of $\Delta t^*$ as a function of mean $Q$ and path length ($L$) assuming an average $V_S$ of 4.1 km/s. For reference, other relevant upper mantle estimates of $Q$ are shown. LQZ: ‘Low-$Q$ Zone’; UM: ‘Upper Mantle’

### 4.4.2 Extrinsic contributions to apparent attenuation

Our approach measures only apparent attenuation. If a significant proportion of high-frequency energy is elastically scattered then we will overestimate intrinsic anelastic dissipation [e.g., Ricard et al., 2014; Richards and Menke, 1983]. Similarly, frequency-dependent focusing or defocusing effects could cause us to overestimate or underestimate intrinsic attenuation, respectively [Allen et al., 1999; Ma et al., 2016; McKenzie and Priestley, 1998; Wielandt, 1987, 1993]. For instance, if a deep, localized low-velocity zone beneath the ridge axis were to preferentially focus longer wavelengths then stations closer to the ridge would appear to record more attenuation.

Allen et al. [1999] found that diffraction effects can dominate observations of apparent attenuation, demonstrating that a narrow, slow conduit acts as a lens that
defocuses high-frequency energy relative to low-frequency energy. They showed that this effect can produce $\Delta t^{*}_{\text{diff}}$ up to 5 s. Similarly, McKenzie and Priestley [1998] modelled scattering in the vicinity of a cylindrical plume with particularly strong focusing for $S$-waves. However, these studies both illustrate that the effect of focusing is strongly dependent on the geometry of the incident wavefield with respect to the scatterer. While there is some back-azimuthal variation to our observed $\Delta t^{*}$ at the ridge (Figure 4.5), it does not appear to be systematic between stations, and does not evince a clear pattern with respect to ridge geometry.

The influence of long-period focusing should be evident from a comparison of the absolute amplitudes of long-period waves at all stations (Figure 4.13). We find that in the 20-13 s period band ridge stations have moderately lower amplitudes than abyssal plain stations – not higher amplitudes, as focussing might predict. At shorter periods, the amplitude increase as a function of age steepens. Overall, ridge stations have lower amplitudes at all frequencies, but are increasingly diminished at
higher frequencies. This observation is consistent with arrivals at these stations being attenuated, and demonstrates that measured $\Delta t^*$ arises from a comparative deficit of high-frequency energy, rather than a comparative excess of (focused) long-period energy.

The alternative possibility is that short-period energy has been defocused. $>0.2$ Hz $S$-waves, with a wavelength $\leq 20$ km, could be markedly refracted [McKenzie and Priestley, 1998]. However, scattering studies show that such strongly refracted energy would be highly delayed (by up to $\sim 5$ s) relative to un-scattered arrivals [Allen et al., 1999; McKenzie and Priestley, 1998]. We not observe extraordinarily delayed arrivals at the ridge; in fact measured travel time residuals at the ridge are relatively modest (Figures 4.8 and 4.11). Moreover, short-period defocusing would predict negative dispersion, whereby higher frequencies arrive later than lower frequencies. Instead, we observe positive dispersion (Figures 4.1 and 4.2), reflected in good fits to both amplitude and phase spectra.

Measured attenuation may be strongly influenced by shallow structure beneath individual stations, likely as a result of scattering in the low-velocity sediment layers. This effect may explain signals at shallow-water OBS stations on the continental shelf, which exhibit large variations in $\Delta t^*$ between sites separated by $<30$ km (Figure 4.5) and reduced amplitudes at all frequencies (Figure 4.13). In the 1-20 s period band, teleseisms at nearby stations are sensitive to almost identical volumes of upper mantle, meaning that discrepancies in apparent attenuation must arise from site effects. The thick, imbricated sediments of the accretionary prism introduce scatter, and we treat large $\Delta t^*$ measured at shelf stations very skeptically.

However, we do not find it likely that sediment-related scattering gives rise to the large values of $\Delta t^*$ close to the ridge, or the overall trend in $\Delta t^*$ with plate age. Sediments are thinnest at the ridge axis, and sediment thickness varies inversely with $\Delta t^*$ on the abyssal plain. Moreover, the estimated $Q_P/Q_S = 2.40$ (Figure
4.10a) provides indirect evidence that intrinsic attenuation, rather than scattering, dominates our results. Scattering would affect $P$- and $S$-waves roughly equally, and would produce a $Q_P/Q_S$ ratio close to 1.0 [Richards and Menke, 1983].

These arguments rule out the possibility that focusing or scattering effects produce the entirety of measured $\Delta t^\ast$. However, it is plausible that these phenomena contribute some component of the observed signal. Future work will entail modelling of non-intrinsic contributions to apparent attenuation. Forward modelling the passage of plane waves through complex shallow structure (e.g., using propagator matrices) will place quantitative bounds on the portion of measured $\Delta t^\ast$ attributable to scattering. 2-D SPECFEM simulations [Komatitsch and Vilotte, 1998; Tromp et al., 2008] of focusing at MORs will allow us to test the size and amplitude of heterogeneity that could introduce measurable focusing effects.

### 4.4.3 Anelastic scaling relationships: effect of temperature

To first order, we expect that mantle attenuation structure is controlled by temperature and grain size. In the oceans, temperatures are relatively well constrained by our robust understanding of young (<50 Ma) plate cooling, ground-truthed by heatflow and geochemical arguments. Equilibrium grain sizes have been computed using upper-mantle flow laws with applicable stress-strain conditions, using a paleowattmeter approach [Austin and Evans, 2007] to quantify the competition between diffusional grain growth and dynamic recrystallization [Behn et al., 2009]. Trace volatiles (e.g., lattice-bound H$_2$O) may complicate matters by reducing effective diffusion constants and promoting anelastic effects. However, Abers et al. [2014] demonstrate that the presence of water also enhances grain growth, thereby increasing diffusional length scales and largely cancelling out any viscosity reduction. In a polyphase crystalline aggregate, this grain growth effect could be limited by pinning, but this is only likely to be relevant at high water content.
We implement anelastic scaling relationships derived from laboratory experiments, with equilibrium grain sizes and effect of water as described above, to calculate $Q_S^{-1} = J_2/J_1$. We use expressions modified from *Jackson and Faul* [2010] (Equations 4.1 – 4.5), where constants are best-fitting values to experimental torsion forced-oscillation data for olivine. For now, we do not include the effects of water. Grain size is assumed to be constant, and we show results using a range of plausible grain sizes *[Behn et al., 2009]*.

This treatment allows us to predict $Q_S$ beneath an oceanic plate as a function of depth and plate age. We also calculate shear velocity, where anharmonic shear modulus ($= 1/J_U$) at relevant $P, T$ is obtained using the HeFesto package *[Stixrude and Lithgow-Bertelloni, 2013]*, assuming a pyrolite composition. Absent data on aggregates or other mineral phases, we assume that anelasticity for the bulk composition follows that of olivine. By propagating vertically incident rays through this plate model, we compute predicted differential travel time and $\Delta t^*$. Figure 4.7 shows the comparison between the predicted and observed values of differential attenuation as a function of ridge distance/age of seafloor. The two datasets have been shifted such that the baseline roughly corresponds to the values measured at the oldest seafloor.

At ages >2 Ma, the values of measured $\Delta t_S^*$ are not wholly inconsistent with the predictions from a thermally controlled trend; the shallow gradient of observations qualitatively matches the expected gradient. The data scatter arises from uncertainties inherent to the measurement method and a relative paucity of data. However, close to the ridge the data evince a clear departure from the thermal trend. For 1 mm grains, thermal predictions yield a differential $t_S^*$ of just 0.2 s between 0 and 15 Ma crust, an order of magnitude less than observed variation. The large attenuation at the ridge cannot be explained by simple ridge-plate thermal structure.

Some component of the signal might arise from an increase in mantle potential temperature close to the ridge. The Cobb thermal anomaly associated with Axial
Seamount has been estimated to have 30°–40°C elevated temperatures \cite{Hooft and Detrick, 1995}. It is perhaps no coincidence that the lowest $\Delta t^*$ close to the Juan de Fuca is measured at OBS stations immediately adjacent to Axial Seamount, with an apparent decrease going northwards along the ridge (Figure 4.5). However, since this thermal configuration would smooth out diffusively, resulting in elevated off-axis $\Delta t^*$ in this vicinity (not seen), a putative temperature spike would have to be localized and recent. Moreover, a modest (30°–40°C) temperature variation would produce less than 5% the magnitude of the observed signal $\Delta t^*$. Finally, the similarity between the Juan de Fuca and Gorda ridge measurements implies a structure that is common to ridge axes (at least in this region). Some alternative mechanism must produce the strong, localized attenuation close to the rift axis.

### 4.4.4 Melt at the ridge axis?

Low seismic wave speed \cite[e.g., Forsyth et al., 1998; Hammond and Toomey, 2003; Nettles and Dziewonski, 2008]{Forsyth et al., 1998} and high electrical conductivities \cite[Baba et al., 2006; Key et al., 2013]{Baba et al., 2006} indicate that melt plays an important role in controlling physical properties of the mantle at mid-ocean ridges (MORs). Anyhdrous mantle ascending isentropically melts at 60 km depth \cite{Langmuir et al., 1993} but the presence of water or CO$_2$ may initiate partial melting as deep as 70-120 km \cite{Asimow and Dixon, 2004} or 300 km \cite{Dasgupta et al., 2013}. The importance of melt is not strongly ruled out by $\delta \ln V_S/\delta \ln V_P$ values estimated for our dataset (Figure 4.10b). Could melt explain the large attenuation signal we observe at stations close to the ridge? This interpretation has two central challenges: Why have previous studies not observed similar signals at other MORs? What is the mechanism by which melt produces such strong attenuation?

Several workers have explored the mechanism by which melt contributes to seismic attenuation. \textit{McCall and Guyer} [1994] show that a cracked, jointed, and pore-riddled
elastic medium modelled as a hysteretic nonlinear material will have attenuation proportional to seismic wave amplitude and wavenumber. Hammond and Humphreys [2000] argue that seismic dissipation caused by melt squirt is negligible in the seismic frequency band for certain distributions of grain sizes and pore shapes. However, they demonstrate that melt squirt could have seismically detectable dissipative effects if pores are elongated and preferentially aligned into networks with anisotropic permeability. Such a geometry has been invoked at MORs as a result of shear banding [Holtzman et al., 2003; Katz et al., 2006; Spiegelman, 2003] or reaction-infiltration instabilities [Spiegelman et al., 2001], and explains strongly anisotropic electrical conductivity [Baba et al., 2006].

A melt squirt dissipation peak confined to the high-frequency part of the seismic frequency band would have a large effect on our measurements. By primarily attenuating high-frequency energy (i.e. having a local $\alpha < 0$, where $Q \sim \omega^\alpha$), this mechanism would markedly steepen the spectral slope in the range of our observation, boosting the apparent attenuation. However, a large high-frequency peak would substantially lower the effective shear modulus [Olugboji et al., 2013].

Alternatively, Holtzman [2015] contend that even small fractions of melt would diminish the effective Maxwell time by promoting diffusional processes at the grain boundaries [Takei and Holtzman, 2009a], thus enhancing seismic attenuation according to ‘master curve scaling’ arguments [Abers et al., 2014; McCarthy et al., 2011]. A corridor of hydrous melt beneath the rift axis surrounded by dry, off-axis restite [cf. Morgan, 1997] could have highly localized weakening effects.

The high attenuation signal that we see is confined to a narrow region about the ridge axis. It is conceivable that previous surface wave studies have not had sufficient lateral resolution to adequately image this short-wavelength feature. Our body wave data are better suited to discern a narrow, attenuating zone.
Forward modelling: a plausibility argument

We model the $Q$ and velocity structure at a MOR as in Section 4.4.3 but with modifications in the region of the ridge to test the plausibility of a high-attenuation zone at the ridge axis. We test two approaches: Firstly, we simulate the effect of melt in the region of the rift axis by introducing a large high-frequency absorption peak into the dissipation spectrum to simulate melt squirt (the “SQ approach”). Secondly, we introduce a region below the ridge axis that has distinct water and melt content by accounting for their influence on viscosity and grain size (the “MV approach”).

Grain size, $d$ is computed using the paleowattmeter Austin and Evans [2007]; Behn et al. [2009], augmenting grain growth parameters to account for water concentration [Abers et al., 2014; Behn et al., 2009]. We compute strain rates assuming a half-spreading rate of 35 mm/yr is accommodated over a 100 km thick layer, giving $\dot{\gamma} \sim 10^{-14}$. Drawing on the models of Turner et al. [2015], we set strain rates in a vertical column beneath the ridge, and in two 40 km thick side lobes beneath the plates, to $10 \times$ this value and fix strain rates in regions of $T < 800$°C to $10^{-19}$.

We propagate linear seismic ‘rays’ through the ridge models to stations evenly spaced at the surface, computing $Q(\omega, P, T, d, \phi, C_{H_2O})$ and $V(\omega, P, T, d, \phi, C_{H_2O})$ at 3 km increments along each raypath. $Q(\omega)$ is used to calculate the amplitude reduction in each increment as a function of frequency between 0.25-0.05 Hz (roughly the same as our observations), where $A'/A_0 = \exp(L\omega/2VQ)$. Finally $t^*(\omega)$ for each raypath is calculated by fitting a spectral slope to the amplitude spectrum, assuming $\alpha=0$; This approach is designed to match our measurement technique as closely as possible, to facilitate more robust comparison between synthetic and observed data. This method approximates the approach of Bellis and Holtzman [2014]. We also report 1 Hz $\delta T_{P,S}$.

In each model, the modified ‘ridge-axis’ region comprises a box 200 km deep and 140 km wide, centered on the ridge. We use a mantle potential temperature of 1350°C with a 2-D plate cooling model [Parsons and Sclater, 1977].
Table 4.1: SQ model results

<table>
<thead>
<tr>
<th>Model</th>
<th>$\Delta P$</th>
<th>$\sigma$</th>
<th>$\Delta t_S^*$ (s)</th>
<th>$\Delta t_p^*$ (s)</th>
<th>$\delta T_S$ (s)</th>
<th>$\delta T_P$ (s)</th>
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<td>4</td>
<td>0.05</td>
<td>0.01</td>
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<td>0.04</td>
<td>3.12</td>
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<td>0.07</td>
<td>2.57</td>
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<td>0.10</td>
<td>10.53</td>
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</tr>
<tr>
<td>P×10,W×2</td>
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<td>8</td>
<td>0.79</td>
<td>0.20</td>
<td>8.92</td>
<td>2.01</td>
</tr>
</tbody>
</table>

Columns are: (1) model run name, indicating peak height (P) and width (W) increase; (2) peak height; (3) peak width; (4) maximum differential $t_S^*$ between the ridge axis and furthest flank; (5) same for $P$-waves; (6) maximum differential $S$-wave travel time between the ridge axis and furthest flank; (7) same for $P$-waves.

SQ approach

The first approach is to explicitly account for melt in the absorption spectrum. We posit that melt squirt produces a Debye peak at high frequencies [Hammond and Humphreys, 2000; Xu et al., 2004]. The parametrization of Jackson and Faul [2010] already contains a low-temperature (high-frequency) peak. This peak is more likely attributable to crystal impurities rather than melt: it is also seen at low temperatures, and in melt-free samples [Sundberg and Cooper, 2010]. However, Takei et al. [2014] present experiments in analogue materials that indicate the height and width of this peak may vary with temperature, broadening close to the solidus. [Xu et al., 2004] observe a high-frequency peak in olivine+basalt samples that increases in height with increasing melt fraction. Therefore, for the sake of simple tests, we explore the possible impact of a significant melt-squirt peak by modifying the height ($\Delta P$) and width ($\sigma$) of the existing peak from Jackson and Faul [2010] (Table 4.1) [cf. Olugboji et al., 2013]. While we do scale the peak relaxation time to account for changes in grain boundary viscosity (Equation 4.5), we do not explicitly vary the peak position to account for other changes possibly brought about by melt, like the widening of grain boundaries. Unlike the constant-grain-size models in Section 4.4.3, these models have equilibrium grain sizes.
These tests (Table 4.1) are highly unsuccessful in reproducing the observations. None of these models yields sufficient integrated attenuation. Moreover, the enlarged peaks produce extremely slow velocities beneath the ridge (by reducing the relaxed shear modulus), resulting in enormous differential travel times that are clearly incompatible with seismological observations.

Although our precise results are contingent on choices of peak parametrization and other small assumptions, the overall result is general. Our joint constraints of attenuation and travel time allow us to rule out a large, high-frequency peak by demonstrating that it causes unrealistically large modulus dispersion (velocity decrease) in the seismic band, while failing to provide observed attenuation. We infer that melt squirt can, at most, contribute < 5% percent of observed attenuation.

MV approach
The alternative is to introduce the effects of melt and water by considering their influence on the diffusion creep shear viscosity. We scale all of the timescales (modifying Equation 4.5) involved in the anelasticity calculation by the relevant multiplicative constants:

\[
\tau_i = \frac{\tau_R}{\eta_{\text{dry}}} \left( \frac{C_{OH}}{C_{OH_R}} \right)^r \left( \frac{d}{dR} \right)^{m_i} \exp \left[ \left( \frac{E^*}{R} \right) \left( \frac{1}{T} - \frac{1}{T_R} \right) + \left( \frac{V^*}{R} \right) \left( \frac{P}{T} - \frac{P_R}{T_R} \right) \right]
\] (4.17)

where we have parametrized hydrous effects as in Hirth and Kohlstedt [2003], with \( C_{OH_R} = 50 \text{ ppm H/Si} \) and \( r = 1.0 \) [Abers et al., 2014; Behn et al., 2009]. Melt is assumed to alter the effective viscosity from \( \eta_{\text{dry}} \) to some reduced value, \( \eta_\phi \), as a function of melt fraction, \( \phi \) [Holtzman, 2015]. While there is some steady decrease in \( \eta_\phi/\eta_{\text{dry}} \) as a shallow exponential function of \( \phi \), experiments [Faul and Jackson, 2007; Hirth and Kohlstedt, 1995; McCarthy and Takei, 2011] demonstrate that a significant portion of the melt-related weakening occurs at extremely small melt fractions when grain boundary melt tubules first form [Holtzman, 2015; Takei and Holtzman, 2009a]. We therefore simplify the effect of melt by imposing a constant \( \eta_\phi/\eta_{\text{dry}} \) within the
sub-ridge region.

We conduct four models, with increasing influence of melt and water. The tested models are as follows:

1. **DR+NoM**: This is the control case, with dry (i.e. nominal 50 ppm H/Si) mantle everywhere and no melt. Unlike the constant-grain-size models in Section 4.4.3, this model has equilibrium grain sizes.

2. **WR+NoM**: We impose 840 ppm H/Si at the ridge axis up to the dry solidus (≈60 km depth), based on observations of the MORB source [*Hirth and Kohlstedt, 1996*]. Here we make the assumption that replacement flux and the low melt fraction allow the rock to remain water-saturated to this depth despite moderate hydrous melting [*Mei et al., 2002; Morgan, 1997*]. In reality the olivine water content will decrease sharply at the wet solidus, asymptotically declining to the nominally anhydrous state at the dry solidus [*Hirth and Kohlstedt, 1996*]. Off-axis, we assume complete dehydration. No melt is included in this model.

3. **WR+200M**: We include the effect of melt below the ridge axis by assuming that $\eta_{\text{dry}}/\eta_\phi = 200$, based on fits to experimental results that include both chemical and geometrical contributions to viscosity at $\phi < 2\%$ [*Hirth and Kohlstedt, 2012*].
Figure 4.14: Example of calculation of synthetic $\Delta t^*$ and $\delta T$ for custom ridge model WR+200M. Main panel: raypaths through the mid-ocean ridge structure with ray increments colored by $Q_S$. Background colors indicate temperature, with 100°C contours indicated. The region with raised $C_{\text{H}_2\text{O}}$ and $\eta_{\text{dry}}/\eta_0$ is indicated by white dashed lines (and depressed $Q$). Right: Profiles of 1 Hz $Q_S$ and $V_S$ along the raypaths to stations at 0 Ma (red) and 10.3 Ma (orange). Bottom: computed values of $\Delta t^*$ and $\delta T$ at each station (crosses show anharmonic $\delta T$ predictions).

1995; Holtzman, 2015]. As before, at the ridge axis we assume 840 ppm H/Si up to the dry solidus.

4. WR+1000M: We include the effect of melt below the ridge axis assuming that $\eta_{\text{dry}}/\eta_0 = 1000$. This value is more illustrative than realistic; it exceeds experimentally constrained melt effects and its reasonableness is discussed below. As before, at the ridge axis we assume 840 ppm H/Si up to the dry solidus.

These models show (Table 4.2) that compared to the control case reasonable quantities of water (model WR+NoM) do not produce markedly greater attenuation, despite a profound effect on viscosity [Hirth and Kohlstedt, 1996]. This observation
confirms the buffering effect of grain growth in the presence of water [Abers et al., 2014]. This effect could be mitigated if secondary phases inhibit grain growth (pinning). We tested this possibility by fixing a maximum grain size of 10 mm, but in this test the maximum $\Delta t^*_S$ increased by an insignificant 0.02 s. If the mantle here contains more than 840 ppm H/Si [e.g., Asimow and Dixon, 2004] water could have a stronger influence, but crustal thicknesses are inconsistent with too hydrated a source [Asimow and Langmuir, 2003; Carbotte and Nedimovic, 2008].

Melt has a more potent effect on attenuation. We show that small fractions of melt (model $WR+200M$) produce up to $\Delta t^*_S \sim 1$ s. While this is an appreciable signal, it only accounts for half of the $\Delta t^*_S = 2.0$ s in our data. The model $WR+1000M$ is able to achieve values that are significantly closer to the observations (especially if frequency-dependence reduces the estimated $\Delta t^*_0$ to $\sim 1.6$). This model does a fair job of reproducing the observations for both $S$ and $P$ attenuation and differential travel times (although the latter are slightly over-predicted).

A $1000 \times$ reduction in viscosity seems geodynamically implausible and exceeds values for $\eta_{\text{dry}}/\eta_\phi$ constrained by experiments. However, if grain sizes were smaller beneath the ridge (e.g., because of strain localization resulting from melt-related viscosity reduction) then the viscosity reduction due to melt would need to be less. Additional models (not shown) demonstrate that if grain sizes below the ridge are reduced by a factor of just 3 then we predict $\Delta t^*_S = 1.60$ s for the $WR+200M$ case. More sophisticated grain size modelling that accounts for the influence of melt on dislocation creep is beyond the scope of this work.

Isolation of any one factor is made more complex by the linkages between these parameters: water promotes melting, melt dehydrates the solid matrix, and viscosity controls grain sizes through strain rate. Our separation of melt, water, and grain size effects is somewhat artificial; we have placed bounds on the product $(\eta_{\text{dry}}/\eta_\phi) (C_{OH}/C_{OH_R})^r (d/d_R)^m$ but remain somewhat agnostic as to the precise com-
combination of effects that produces the signal.

A more sophisticated treatment of this problem was conducted by Goes et al. [2012], who incorporated the effects of chemical depletion, dehydration, water, and melt to model the velocity and attenuation structure beneath a ridge. Their preferred anelasticity model\(^6\) matched observed EPR velocity constraints and predicted \(Q_S < 20\) over >100 km depth beneath the ridge axis. This finding bolsters our argument that a highly attenuating upper mantle at the ridge axis is consistent with petrological, geodynamical, and experimental constraints.

Small increases in mantle potential temperature can produce large increases in melt content [White and McKenzie, 1992]. The observation of particularly elevated \(\Delta t_S^*\) at OBS stations close to Axial Seamount (underlain by the Cobb thermal anomaly) is consistent with excess temperatures producing abundant melt in this locale, reflected also by thicker igneous crust [Hooft and Detrick, 1995]. The Cobb thermal anomaly was only captured by the Juan de Fuca ridge within the last 0.5 Ma [Carbotte and Nedimovic, 2008; Karsten and Delaney, 1989]; prior to this time any Cobb-related melt would have been transported westwards by the divergent system. Hypothetically, the limited eastward transport of melt since 0.5 Ma could help explain the step in \(\Delta t_S^*\) east of the ridge as the product of a sharp gradient in melt content.

4.4.5 The role of anisotropy

Abundant, aligned MOR melt is predicted to have strong influence on anisotropic S-wave velocities, producing strong shear wave splitting at the ridge [Blackman and Kendall, 1997; Blackman et al., 1996; Kendall, 1994]. There is limited evidence for ridge-parallel fast axes on the Juan de Fuca ridge axis, quickly rotating into a shear-

\(^6\)Goes et al. [2012] used anelastic scaling relationships based on the parametrization of Jackson et al. [2002] and Faul and Jackson [2005], which predicted more attenuation than the later study [Jackson and Faul, 2010], which we employ. Thus, Goes et al. [2012] may overestimate \(Q^{-1}\).
parallel direction off axis [Bodmer et al., 2015; Martin-Short et al., 2015]. No such 
signal is seen at the Gorda ridge, and other studies have shown semi-horizontal ridge-
perpendicular fast axes across the whole ridge system [Hammond and Toomey, 2003;
Nowacki et al., 2012; Wolfe and Solomon, 1998]; sub-ridge strains are thought to 
induce crystallographic-preferred orientation that dominates any signal from aligned 
melt.

The organization of melt pockets and alignment of orthorhombic crystals could 
also produce anisotropic viscosity and attenuation. Simulations indicate that a frac-
tured or layered medium with hexagonal symmetry can have greater anisotropy in 
attenuation than in velocity [Wenzlau et al., 2010; Zhu et al., 2006]. Takei and 
Holtzman [2009b] show that sheared melt-rich aggregates can develop geodynami-
cally consequential viscous anisotropy and layers alternately rich and poor with melt.

Analogies from fractured media indicate that $SH$-waves with particle motion par-
allel to the hexagonal symmetry axis would be maximally attenuated, and that $P$
-waves perpendicular to this direction (i.e., travelling along the layers) would be much 
less so [Carcione et al., 2012]. Beneath the ridge axis, we anticipate sub-vertical 
melt foliation with symmetry perpendicular to the spreading direction. Therefore, 
teleseismic $P$-wave attenuation at the ridge axis will be diminished, while teleseismic 
$SH$-wave attenuation (depending on polarization) could be enhanced. Off axis, where 
melt may stall in sub-horizontal pockets [Kawakatsu et al., 2009] or flow along a shal-
low permeability barrier [Sparks and Parmentier, 1991], this pattern would reverse. 
This conceptualization fits with radially anisotropic velocity structure established off 
axis [Beghein et al., 2014; Nishimura and Forsyth, 1989]. Such a change in melt ge-
ometry could contribute towards the transition from high on-axis $\Delta t^*$ to lower values 
at greater ages (Figure 4.7). Moreover, anisotropic attenuation would partly account 
for discrepancy between our attenuation results and prior Rayleigh wave estimates.
4.5 Conclusions

We have conducted the first body wave study of the attenuation structure of an entire oceanic plate, constraining the underlying mantle’s velocity and \( Q \) structure in unprecedented spatial detail. This work provides a comparison to previous surface wave attenuation studies, probing the Earth in a different frequency range and at finer geographic resolution.

We find a systematic and significant trend of decreasing attenuation and travel time delays with increasing age. The most highly attenuated teleseisms (\( \Delta t^*_S \sim 2.0 \) s, \( \Delta t^*_P \sim 0.5 \) s) are recorded at stations close to the Juan de Fuca and Gorda ridge axes, where the largest travel time delays are also observed (\( \delta T_S \sim 2.5 \) s, and \( \delta T_P \sim 0.6 \) s). The similarity in \( \Delta t^* \) trends for both ridges is suggestive of a common trait to the MOR system. The large values of \( \Delta t^*_S \), in particular, far exceed the predictions based on simple thermal models and imply upper mantle \( Q_S < 25 \) beneath the ridge axis; this value is at the lowest end of what has been previously observed worldwide. While apparent \( \Delta t^* \) could have some contribution from extrinsic attenuation mechanisms (focussing and scattering), we show that travel times, frequency dependency of phase delays, and estimated \( Q_P/Q_S \) ratios are more consistent with intrinsic attenuation in shear (\( 1/Q_\mu \)).

We posit that this phenomenon could be explained by melt in the upper mantle beneath the ridge axis, consistent with high estimated \( \delta \ln V_S/\delta \ln V_P \), low wave speeds, and (at least at other ridges) elevated electrical conductivity. We model the propagation and attenuation of synthetic seismic rays through a mid-ocean ridge structure, testing the influence of a sub-ridge region enriched in water and melt either affecting viscosity or expressed as an enlarged absorption peak. A viscosity reduction beneath the ridge of \( >1000 \times \) is indicated by our data; this could be achieved by widespread melt \textit{in situ} at grain boundaries, together with reduced grain sizes that result from strain localization feedbacks.
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Concluding remarks

This dissertation extends our understanding of how the upper mantle behaves in extensional environments. We show that a multiplicity of observations – splitting, $P$ and $S$ travel times, seismicity, receiver functions, and integrated attenuation – complementarily strengthen seismological analysis.

Our measurement of azimuthal anisotropy in a highly extended continental rift broadens the suite of rift splitting datasets and demonstrates a qualitative shift in mantle behavior with increasing degree of extension. We provide testable predictions of splitting in zones of extension and show that they explain the apparent discrepancy between more- and less-extended rifts.

Tomographic models of the Woodlark Rift advance our picture of continental breakup. We image substantial mantle deformation hundreds of kilometers along strike from propagating oceanic spreading centers. The identification of intermediate-depth seismicity permits unusually well constrained absolute temperature estimates of the rift axis, establishing the removal of mantle lithosphere markedly ahead of complete crustal thinning. This observation expands global studies of rift systems, reinforcing concepts of preferential lithospheric degradation (possibly requiring water or melt) and providing new details on the continental-to-oceanic spreading transition.

Our results augment seismological constraints on the complex Woodlark Rift, revealing the influence of recent subduction and identifying a possible source for the world’s youngest UHP rocks. In combination with geologic and geodynamical data, this study refines theories for rapid UHP exhumation, contributing towards the goal
of the CDPAPUA collaboration.

We have developed a novel methodology for jointly inverting $S$-wave travel time residuals and splitting times for velocity perturbations and azimuthal anisotropy. This technique is well suited to the Woodlark Rift. Our joint $\Delta V_S$ and anisotropy inversion improves on traditional isotropic models by resolving potential tradeoffs in the data. We find that the process of continental breakup is evident from the anisotropy structure, as a pre-existing lithospheric fabric gives way to extensional fabric at the axis of spreading.

The Cascadia Initiative’s Amphibious Array presents a unique opportunity to study an oceanic plate and spreading center. Our dataset of body-wave differential attenuation data is collected using an innovative technique inspired by surface wave analysis. The measurements sample the mantle beneath an entire oceanic plate in unprecedented spatial detail. This study enhances ongoing community efforts to characterize anelasticity within the Earth. Unexpectedly high attenuation beneath the MOR axes implies that melt, water, or grain size heighten anelastic effects in the region of upwelling. By linking $\Delta t^*$ measurement to differential travel times, and exploiting both $P$ and $S$ results, we obtain quantitative bounds on the physical state of the upper mantle at the locus of spreading.
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Appendices

A Quantitative comparison of SKS splitting measurement techniques


A.1 Single-channel methods

The single-channel methods for measuring shear-wave splitting use the pulse shape of the incident shear waves, as recorded on the transverse and radial components, to invert for splitting parameters $\phi$ (fast azimuth) and $\delta \tau$ (splitting time). The rotation correlation (RC) method seeks to maximize the cross-correlation between fast and slow waveforms once the inverse splitting operator has been applied, with the assumption that these should be identical in the absence of noise, each with the form of the un-split initial waveform, with time displacement of $\delta \tau$ [Bowman and Ando, 1987]. The eigenvalue (EV) method seeks to minimize the smaller eigenvalue in the cross-correlation tensor of rotated horizontal traces, which is equivalent to best linearizing the initial waveform - here making the assumption that the initial particle
motion of $SK(K)S$ phases is linear [Silver and Chan, 1991]. The minimum energy (SC, after Silver and Chan) method differs from the EV method in that it solves for the splitting parameters which minimize the energy on the transverse component of the initial waveform - making the assumption that core-traversing phases should be polarized in the vertical- radial plane before they encounter anisotropy [Silver and Chan, 1991]. All of these methods implicitly assume that the waveforms are only distorted due to a single, horizontal layer of uniform anisotropy.

**Approach for testing single-channel methods**

We define several criteria for choosing the ‘best’ method:

1. **robustness** - how well the method does for arrivals with small $\delta \tau$, low signal-to-noise ratio (SNR), and close to the null azimuths.

2. **consistency** - whether the same structure and ray paths yield the same splitting measurements across different arrivals. For the purpose of synthetic testing, this is tantamount to measuring the same splitting parameters for different realizations of the random noise.

3. **accuracy** - how faithfully results of synthetic tests recover the ‘true’ input structure ($\phi_{true}$ and $\delta \tau_{true}$).

Synthetic tests were conducted for all three of the individual-arrival methods; firstly a simple structure was tested by fixing a $\phi_{true} = 0^\circ$ and varying $\delta \tau_{true} = 1,2,3$ s. The back-azimuthal dependence of the methods was examined by varying the initial polarization ($\theta$) (Figure A.1) of the synthetic waveform in $10^\circ$ increments from $0^\circ - 180^\circ$ (the $2-\theta$ dependence on splitting predicts identical results in all four quadrants for vertically incident waves). A 10 second gaussian pulse within a 50 second window was propagated through the synthetic structure assuming ideal splitting and recombined prior to adding noise. The incidence angle was set at $10^\circ$ and the sampled anisotropy
calculated according to the Christoffel equations [Szeg, 1975], using an orthorhombic fabric for olivine elasticity from Abramson et al. [1997], with the plunge of the “a” and “b” axes set to zero.

Artificial noise was generated by adding a noise transform to the data transform in the Fourier-domain. Real noise spectra were taken from a 50s window before a representative $P$ arrival, then smoothed using a moving average, and the phase of the noise was randomized for each realization. At the periods sampled the background, microseisms, not $P$ coda, dominate noise. Noise spectra from various events, and from OBS (station J), island (station KIR) and mainland (station AGAN) sites were used, with little difference in results (results with noise from event 218, station KIR are shown in Figures A.2-A.6).

$SNR$ is not formally defined for a non-stationary time series such as the single pulse that comprised our synthetic seismogram. We defined $SNR$ as the maximum amplitude of the original (i.e. pre-split) pulse on the radial ($R_{\text{max}}$) divided by two standard deviations of the amplitude of the noise on the tangential:

$$A'_i = \frac{A_i N R_{\text{max}}}{2 A_0 SNR}$$  \hspace{1cm} (A.1)

where

$$(A_0)^2 = \sum_{i=1}^{N} A_i^2$$  \hspace{1cm} (A.2)

and

$$SNR = \frac{R_{\text{max}}}{2 \sigma_T} = \frac{R_{\text{max}}}{2 \sqrt{\frac{\sum T^2}{N}}} = \frac{R_{\text{max}}}{2} \sqrt{\frac{N}{P_T}}$$  \hspace{1cm} (A.3)

where $N$ is the number of samples, and $R$ and $T$ refer to the radial and tangential traces. $P_T$ is the power in $T$. Using Parseval’s theorem (and accounting for the IFFT’s amplitude scaling) we may then construct a simple scaling for the amplitudes $A_i$ of the noise in the frequency domain to obtain the right $SNR$ (above). The measured $SNR$ (after filtering) is contingent on the random noise added, so varies about the input $SNR$. For consistency with the actual data, we use the $SNR$ re-measured.
in the same manner as with actual data, rather than the nominal input SNR, to discuss robustness of methods. By taking different random realizations of the phase of the noise we obtained 10 traces for each $\delta \tau$ and azimuth. This synthetic data was then filtered using the same filter as the real data (8-50 s zero-phase bandpass) and then inverted for splitting parameters using each of the three single-channel methods. We experimented with varying the number and distribution (systematically spaced, randomly spaced, and clumped) of synthetic arrivals, exploring how well the stacks return the input model parameters. Further tests were conducted with two-layer anisotropic structures.

**Results of single-channel tests**

**Robustness**

As expected, the single channel methods do a poor job of retrieving true splitting parameters at, and close to, the null directions (Figure A.1). As the null polarization is approached, the RC method tends towards measuring low $\delta \tau$ and $\phi \sim 45^\circ$ from the back azimuth; this feature arises because close to nulls, little or no splitting means there is negligible correlation between radial and tangential traces, so the cross-correlation on orthogonal components is maximized by fitting a rotated version of the radial with itself, at zero lag. This significantly limits the window of back azimuths for which this method works. As Wüstefeld and Bokelmann [2007] note, the $4\theta$ (i.e. $90^\circ$) periodicity of this feature could potentially be of use for identifying $\phi_{true}$, but such a treatment fails for a limited back azimuthal distribution or for of complex, layered structure. The window of accurate RC measurements widens for larger $\delta \tau_{true}$. By contrast, the SC and EV methods successfully resolve splitting close to nulls even for low $\delta \tau_{true}$.

In general, the results are better at large SNR (Figure A.2). The error in measured splitting parameters ($E$, defined as the difference between true and estimated...
splitting parameters) diminishes with increasing SNR. This finding motivates us to weight our results by measured SNR (taken as the ratio of the largest amplitude to the r.m.s. amplitude over the selected time window) and to eliminate results with SNR lower than 5.

$\delta \tau_{\text{true}}$ is observed to have a significant effect on the accuracy of the results, with larger splitting times yielding measurements of both $\phi$ and $\delta \tau$ that are closer to the input values (Figure A.3). This is explained because larger $\delta \tau_{\text{true}}$ produce more significantly displaced fast and slow waveforms, so after re-combination and filtering the splitting is better preserved. We note that for lower (1-2 s) $\delta \tau_{\text{true}}$, the SC and EV methods tend to slightly over-predict the magnitude of the splitting, if the SNR is poor. As SNR increases, the effect of the $\delta \tau$ diminishes.

Consistency
The ability of the methods to resolve the same structure across different ‘arrivals’ (for synthetic testing, these comprise traces split identically with different random noise added as well as different incoming azimuth) is, to a large extent, contingent on the random noise added. However, for a given SNR, the RC method is the most consistent at low $\delta \tau_{\text{true}}$, although there is a backazimuthal dependence, as noted above (Figure A.3). The method that performs next best is SC, followed by EV. For low $\delta \tau_{\text{true}}$ and SNR, individual arrivals show significant scatter about the average value - indeed for the worst method (EV), measurements at a single backazimuth have average 2$\sigma$ scatter of $\pm 74^\circ$ (Figure A.3). This finding motivates us to trust station stacks much more than individual measurements, which depend on the idiosyncrasies of the noise, especially at low $\delta \tau_{\text{true}}$.

Accuracy
For tests with large $\delta \tau_{\text{true}}$ and high SNR, all three methods do a good job of returning the input $\phi$ and $\delta \tau$. At lower $\delta \tau_{\text{true}}$ and SNR, the SC method most consistently achieves the most accurate results, with mean errors in less than 10$^\circ$ and mean errors
in $\delta \tau$ less than 0.4 s (Figure A.3). Given the similarity in the procedures, we expect SC and EV to perform similarly and it is not surprising that EV does marginally worse, as an extra source of error is introduced in leaving the polarization as a free parameter. One consistent feature is that the SC and EV methods tend to overestimate $\delta \tau$ when $\delta \tau_{true}$ is small. By contrast, the maximum $\delta \tau$ measured by the RC method is always fairly close to $\delta \tau_{true}$. It might be fruitful to combine the splitting methods, whereby azimuths are estimated using the SC method, and times estimated using the RC method. This approach relies on having a good backazimuthal distribution; as we have noted, the RC method is susceptible to underestimating splitting times close to nulls.

We conclude that based on the full range of tested conditions, the SC method performs best, but with the caveat that measured $\delta \tau$ up to 2 s may be overestimates of $\delta \tau_{true}$. A practical approach may profitably entail comparison of the estimated $\delta \tau$ from RC and SC.

**Single-channel with stacking**

To improve single-arrival splitting results, many workers have stacked error surfaces [e.g. Restivo and Helffrich, 1999; Wolfe and Silver, 1998] and used global minima to resolve the best fitting structure beneath each station. When we apply this approach to the synthetic results we are able to get extremely good fits to the theoretical inputs. In particular, this approach resolves the above-mentioned tendency of the SC method to overestimate $\delta \tau$.

Real data rarely affords an even distribution of back azimuths. To investigate the importance of evenly distributed back azimuths to the stacked result, we computed splitting for 10 arrivals with systematically spaced as well as random distributions of polarizations (in the range 0-360°) and used the results to calculate stacked estimates of the splitting parameters. For the SC method, the systematic distribution of
polarizations showed no significant improvement over the random distribution.

Figure A.4 shows the results from 10 synthetic tests, each with 10 arrivals: each point is the stacked minimum from 10 arrivals arriving at random backazimuths, with formal 2σ error bars (Section 1.3.4). We found that stacked results of the RC method have more scatter about the true values while the SC and EV methods both produced stacked results which consistently fell close to the true minimum (Figure A.4). Moreover, SC and EV measurements which fell further from $[\delta \tau_{\text{true}}, \phi_{\text{true}}]$ generally had larger error bars, implying these two methods do a better job of estimating uncertainty. That said, several (>50%) results across our tests had formal 2σ error bounds which did not include $\delta \tau_{\text{true}}$ and $\phi_{\text{true}}$, implying a systematic underestimate of errors by these methods.

The stacking approach consistently returns better estimates of the true splitting parameters than individual measurements and has significantly smaller uncertainties than simple averages of individual results, as expected [Wolfe and Silver, 1998]. The SC method including stacking appears to be the most robust and reliable, and improves greatly on the accuracy of estimated $\delta \tau$ from the individual SC results.

**Note on single-channel with complex structure**

The stacking approach obscures the predicted back azimuthal dependence of measured splitting parameters predicted for complex structure. Stacking measurements of waves that have propagated through 2-layer structure produces a summed error surface with a clear minimum; the form of the resultant error surface is not differentiable from that produced by stacking measurements from simple structure. We conclude that the result of a stack on its own, at face value, may give misleading and incorrect results. It is helpful to complementarily analyse the individual measurements as well as the stacked result.

The SC and EV methods allow qualitative identification of layered structure,
and quantitative modeling of this structure by fitting individual observations to the complex 4-θ curves. By contrast, the RC method offers no information on complex structure. As shown in Figure A.5, the RC measurements from a two-layer structure with fast axes 0° (bottom) and -60° (top), each contributing δτ = 1 s, are indistinguishable from a one-layer structure with φ = -30° and δτ = 1. Therefore, we caution against using this method alone, as it obscures the presence of complex anisotropy.

We show (Figure A.5) that the majority of individual \( \phi_{est} \) using the SC method are biased towards the upper layer [cf. Silver and Savage, 1994] while the \( \phi_{est} \) from the stack is close to the average \( \phi_{true} \) of the layers (where the average is weighted by their respective \( \delta\tau_{true} \)). This phenomenon may allow us to constrain the order of anisotropic layers. If we observe two populations of individual \( \phi_{est} \) at a given station, but have insufficient splitting measurements to quantitatively model multi-layer anisotropy (by fitting the \( \sim 4-\theta \) variation in splitting parameters with back-azimuth), it is most likely that the majority of individual measurements (which may, therefore, dominate the stack) are controlled by the shallower anisotropy [Saltzer et al., 2000]. Poor back-azimuthal sampling would make this assessment difficult.

### A.2 Multi-channel method

The SI method [Chevrot, 2000] relies on the 2-θ sinusoidal variation in the magnitude of splitting with back-azimuth. In the case of a dipping layer, there will also be a component of 1-θ sinusoidal variation. The calculation of splitting parameters that describe the anisotropy is dependent on fitting a distribution of splitting intensity measurements with back-azimuth.

**Approach for testing multi-channel method**

The SI method requires several arrivals from a range of back-azimuths. Robustness may be defined as a description of how well the method performs for few arrivals
and for limited back-azimuthal distributions. Synthetic tests of this method are done with the same parameters as the single channel tests (incidence angle = 10°, orthorhombic symmetry, no plunge to a or b axis, and same filtering scheme), varying $\delta\tau$ and $SNR$ as well as exploring the effect of multiple layers.

**Multi-channel test results**

**Robustness**

The results of the SI method should be sensitive to back-azimuthal distribution, noise, and the number of arrivals. The splitting estimates are derived from a curve-fit to points which follow a $2\theta$ sinusoid, so more arrivals mean more degrees of freedom. Thus error in any one splitting intensity measurement becomes less significant to the overall station measurement, depending on the data distribution.

Figure A.6 shows tests with 5 and 10 arrivals, and with back-azimuthal distributions that vary from random to clumped to very clumped. It appears that for moderately limited back-azimuthal ranges the number of arrivals has a much more significant effect on the quality of the SI measurements (both in terms of consistency and accuracy) than the distribution of back azimuths. However, if the back-azimuthal range is extremely limited, the quality of the SI measurements deteriorates. If several arrivals with limited back-azimuthal range and low $SNR$ dominate the dataset (as is the case for many of the stations in this study) then the SI results are low quality, and often not statistically significant.

At the extrema of the splitting intensity (45° from the null directions) the SI technique tends to over-estimate the splitting intensity. In the case, then, that most arrivals come from an azimuth close to this maximum, the measured splitting magnitude may be an overestimate. However, this same effect leads to a more faithful constraint on the fast azimuth, by exaggerating the extreme points that force the maxima of the fit to the azimuths 45° from null. In order to explore the effect of
these points on the measured splitting time, we recommend fitting splitting intensity data with both L1 and L2 norms, to check if a few outlier points are dominating the measured $\delta \tau$.

**Consistency**

We observe good consistency of results between runs even at low $SNR$, implying perhaps that the SI method is less sensitive to noisy traces because it relies not on the shape of the waveform but the ratio of transverse to radial energy. Noise on the two components is seldom completely independent, so noisy traces do not exhibit drastically reduced cross-correlation between the R and T traces. Further, the SI approach makes use of information from multiple arrivals to compute splitting parameters, tacitly incorporating more information in each measurement than any of the individual methods.

**Accuracy**

The SI method tends to yield estimates of splitting which closely echo the input structure. The caveat to this finding is that when there are few arrivals that are from only a small range of back-azimuths, the uncertainty of the fit suffers (Figure A.6). In the case of few arrivals, comparison of L1 and L2 fits, as well as bootstrap error estimates, affords a quantification of how robust the fit is. In order to improve the fit, we may weight each datapoint by the estimated uncertainty. Our synthetic tests show that a simple unweighted least squares fit obtains a decent result, but that weighting by uncertainty of splitting intensity measurements often yields better estimates of anisotropic parameters.

**Multi-channel with complex structure**

In the case of multiple layers of anisotropy, the splitting intensity contributions from each layer should commutatively add, such that the net observed splitting intensity variation is the summation of the $2\theta$ sinusoids associated with each of the individual
Figure A.1: Cartoon showing relationship between polarization and null directions. The synthetic tests are done at a range of back azimuths, $\theta$. Splitting measurements of waves initially polarized in either of the null directions will exhibit no splitting. We show that there is a finite range of angles around the null directions that each method produces unreliable splitting measurements, and the size of this range is a measure of the methods’ robustness.

Figure S2. Plots of errors (deviations from true splitting parameters) for individual measurements at a range of measured SNRs, using the SC method. Blue points are for $\delta \tau_{true} = 1\text{s}$ red points are for $\delta \tau_{true} = 3\text{s}$. Errors clearly decrease with increasing SNR. Errors are also smaller for greater $\delta \tau_{true}$, although some of this effect for $|E_{\delta \tau}|$ is because close to nulls, the SC method gives $\delta \tau = 4 \text{s}$ (the maximum allowed), which causes the the clipping at 3 and 1 s for $\delta \tau_{true} = 1$ and 3 s, respectively.

layers [Silver and Long, 2011]. Our results confirm this finding, and we suggest the possibility of gaining more information about complex structure by comparing the results of (stacks of) individual methods (which are biased towards shallower structure) and the SI results, which are equally sensitive to structures at all depths.
Figure S3: Individual results averaged over 10 realizations of the random noise, in each case with $SNR \approx 10$ and $\phi_{true} = 0^\circ$, $\delta \tau_{true} = 1$ s (blue) 2 s (cyan) and 3 s (red). The incoming azimuths are 0, 10, 20, .. 180, although measurements with different $\delta \tau_{true}$ are plotted $\pm 2^\circ$ from these so as to not overlap. Error bars show one standard deviation in the scatter of measurements (using the circular standard deviation in the case of $\phi$) for the 10 realizations at that backazimuth. Dashed lines show predicted splitting parameters using Cristoffel equation.

Figure A.3: Individual results averaged over 10 realizations of the random noise, in each case with $SNR \approx 10$ and $\phi_{true} = 0^\circ$, $\delta \tau_{true} = 1$ s (blue) 2 s (cyan) and 3 s (red). The incoming azimuths are 0, 10, 20, .. 180, although measurements with different $\delta \tau_{true}$ are plotted $\pm 2^\circ$ from these so as to not overlap. Error bars show one standard deviation in the scatter of measurements (using the circular standard deviation in the case of $\phi$) for the 10 realizations at that backazimuth. Dashed lines show predicted splitting parameters using Cristoffel equation.
Figure S5 (next page). Results of one test run with 19 evenly-spaced synthetic ‘arrivals’ propagating through a two-layered anisotropic structure with $\phi_{\text{upper}} = -60^\circ$, $\delta_{\text{t,upper}} = 1\text{ s}$ and $\phi_{\text{lower}} = -60^\circ$, $\delta_{\text{t,lower}} = 1\text{ s}$. The average SNR was $\approx 20$. In the upper two rows of plots, dashed lines are predicted splitting parameters from the individual layers, solid lines are predicted splitting parameters [Silver and Long, 2011]. The third row contains the multichannel method back azimuthal fit and power in each azimuthal harmonic [c.f. Chevrot, 2000]. The bottom row contains the stacked energy surfaces and estimates of splitting parameters using these minima of these stacks. The 4-θ pattern to the individual measurements reveals the complex structure at depth, but the multichannel method and stacked results in the lower two rows of plots are indistinguishable from simple structure with $\phi = -30^\circ$ and $\delta_{\text{t}} = 1.0\text{ s}$.
Figure A.5: Results of one test run with 19 evenly-spaced synthetic ‘arrivals’ propagating through a two-layered anisotropic structure with \( \phi_{\text{upper}} = -60^\circ \), \( \delta\tau_{\text{upper}} = 1 \) s and \( \phi_{\text{lower}} = -60^\circ \), \( \delta\tau_{\text{lower}} = 1 \) s. The average \( \text{SNR} \) was \( \approx 20 \). In the upper two rows of plots, dashed lines are predicted splitting parameters from the individual layers, solid lines are predicted splitting parameters [Silver and Long, 2011]. The third row contains the multichannel method back azimuthal fit and power in each azimuthal harmonic [cf. Chevrot, 2000]. The bottom row contains the stacked energy surfaces and estimates of splitting parameters using these minima of these stacks. The 4-\( \theta \) pattern to the individual measurements reveals the complex structure at depth, but the multichannel method and stacked results in the lower two rows of plots are indistinguishable from simple structure with \( \phi = -30^\circ \) and \( \delta\tau = 1.0 \) s.
Figure S6: Results of synthetic tests of the SI method. For each set of inputs, 10 runs were performed, using different random realizations of the noise from station KIR, with $SNR \approx 10$, and $\phi_{true} = 0^\circ$, $\delta_{\tau true} = 1$ s. Mean results ($\pm 2\sigma$) using the SI method are shown at the top of the plots. Plots a, b, c have 5 datapoints per run, plots d, e, f have 10 datapoints per run. The distribution of back azimuths was varied from random (a,d) to limited ($\gamma = 40^\circ$) (b,e) to very limited ($\gamma = 20^\circ$) (c,f), where clumping was achieved by taking back azimuths normally distributed about random means with standard deviations $\gamma$. Changing the number of arrivals seems to have a much greater effect on the success of this method than how back-azimuthally clumped the arrivals are, although very limited back azimuthal ranges did yield significantly poorer results than random or moderately limited ranges.
B Supplementary splitting results

This appendix contains further supplementary materials submitted along with the SKS splitting study: Eilon, Z., G. A. Abers, G. Jin, and J. B. Gaherty (2014) “Anisotropy beneath a highly extended continental rift”, Geochem. Geophys. Geosyst., 15, doi:10.1002/2013GC005092. This material includes examples of different quality single-channel splitting measurements, as well as a comparison between the station-averaged results from the three single-channel methods.

Figure B.1: Stacked splitting results from the SC method (red), EV method (blue), and RC method (green). The orientation of the lines indicates the azimuth of measured $\phi$ and their length is the associated $\delta\tau$. Line thickness corresponds to the uncertainty in the fast azimuth, where formal uncertainties in splitting parameters are estimated from the energy/correlation surface (Section 1.3). We emphasize less uncertain measurements with thicker lines (see key)
Figure B.2: Examples of splitting measurements designated “good” quality, with a land station (top) and OBS (bottom) for comparison. The bottom trace was designated “null”. Fewer results were obtained at OBS stations, but as these figures show, the OBS data quality was often comparable to the land stations, after filtering.
Figure B.3: Same as B.2, but showing measurements designated “fair” quality.
Figure B.4: Same as B.2, but showing measurements designated “poor” quality.
C Supplementary material for isotropic tomography study

This appendix contains supplementary materials submitted along with the isotropic $V_P$ and $V_S$ study: Eilon, Z., G. A. Abers, J. B. Gaherty, and G. Jin (2015) “Imaging Continental Breakup using Teleseismic Body Waves: The Woodlark Rift, Papua New Guinea”, Geochem. Geophys. Geosyst., 16, doi:10.1002/2015GC005835. Figure C.1 shows the results of “L-curve” tests (described in the main text) to optimize regularization parameters for the inverse problem. Figure C.2 contains the results of synthetic tests described in the main text. Figure C.3 shows cross sections through the $V_S$ tomographic model, complementing cross sections through the $V_P$ models included with the main text. We also include details of seismic stations and crustal travel times solved for in the inversion (Table C.1).

![Graphs showing L-curves and contour plots for $V_P$ and $V_S$ models]  

Figure C.1: L-curves for the $V_P$ model (left) and the $V_S$ model (right). In the upper row, the lines show the value of the norms with varying damping at constant smoothness. The lower row shows the contoured penalty surface (see text). The dot indicates the values of $\epsilon$ and $\gamma$ corresponding to the penalty minimum.
Figure C.2: Results of synthetic tests with ‘realistic’ input structure. Left column: input structure, center column: $P$ model output structure, right column: $S$ model output structure. Top three rows: horizontal tomograms at different depths, bottom row: a north-south vertical section at 148.8°E.
Figure C.3: Three cross sections of $S$-wave velocity variation contoured for hit quality between 60 and 90%, where regions with hit quality < 60% are masked out. Locations of cross sections shown on 90 km horizontal slice. The moho plotted is to scale and derived from receiver functions. Note the change in vertical scale at 5 km: the topography is $3.5 \times$ exaggerated. Circles: seismicity from this study, relocated using HYPODD; plotted earthquakes have estimated 3-D location uncertainties $<2.5$ km and intermediate depth earthquakes are plotted in magenta. For the vertical slices only earthquakes within 25 km of the section are shown. White box is region with possible melt, indicated by triangles on Figure 2.10.
Table C.1: Station details for CDPAPUA stations

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Columns are: (1) station name; (2) latitude; (3) longitude; (4) elevation, in m, where OBS depth is indicated by negative elevations; (5) correction angle, $\Delta \phi$, (in degrees) – this angle is the clockwise angle from true north to the azimuth of the station north (Section 1.3). To correct data from these stations, horizontal channels be rotated anti-clockwise by the quoted angles; (6) de-meaned posterior estimate of vertical $P$-wave travel time to 40 km, in seconds; (7) de-meaned posterior estimate of vertical $S$-wave travel time to 40 km, in seconds; (8) Number of $P$-wave arrivals used per station; (9) Number of $S$-wave arrivals used per station. Not shown: details for Woodseis stations.


D Approximation of anisotropic velocities

Here we derive an approximation to the velocities of a split shear wave, as a function of propagation direction ($\zeta$) and just two parameters ($V_{S_{av}}$ and $\alpha$) describing the mean shear velocity and the magnitude of the anisotropy in terms of the orthogonal velocities $V_\perp$ and $V_\parallel$. First, we create an elastic tensor with hexagonal symmetry from just these two parameters, making some assumptions. Second, we compute the solutions, $V_S(\zeta, V_\perp, V_\parallel)$ to the Christoffel equations for this elastic tensor. Finally, we find approximate solutions as functions of $V_\perp$ and $V_\parallel$ that closely match the true solutions. These approximate solutions are chosen for their simple functional form, permitting analytic expressions of the gradients of velocities (and, hence, travel times) to be found. For the purposes of notational simplicity, we will define $x \equiv V_\perp$ and $y \equiv V_\parallel$.

Consider an anisotropic medium defined by a stiffness matrix using the Love parameters, $A$, $C$, $F$, $L$, and $N$ [Love 1927]. These 5 independent parameters are sufficient to define the elastic tensor in Voigt notation:

$$
\begin{pmatrix}
A & (A - 2N) & F & 0 & 0 & 0 \\
(A - 2N) & A & F & 0 & 0 & 0 \\
F & F & C & 0 & 0 & 0 \\
0 & 0 & 0 & L & 0 & 0 \\
0 & 0 & 0 & 0 & L & 0 \\
0 & 0 & 0 & 0 & 0 & N
\end{pmatrix}
$$

The Love parameters are simply related to the velocities of $P$ and $S$ waves travelling along, or perpendicular to, symmetry axes:
and $F = \eta (A - 2L)$. Assuming $\eta = 1$, $V_{P_{av}} = \nu V_{S_{av}}$ and equal $P$ and $S$ anisotropy (consistent with natural and synthetic samples Anderson [1989]; Ismail and Mainprice [1998]; Tommasi et al. [2000]), we define $x \equiv V_{S_{av}} (1 - \alpha)$ and $y \equiv V_{S_{av}} (1 + \alpha)$, giving

\begin{align*}
A &= \rho \nu^2 x^2 \\
C &= \rho \nu^2 y^2 \\
L &= \rho y^2 \\
N &= \rho x^2 \\
F &= \rho (\nu^2 x^2 - 2y^2)
\end{align*}

These give precisely the same results as the treatment of Panning and Romanowicz [2006], where their anisotropic parameters are related to ours by:

\begin{align*}
\xi &= \frac{V_{S_{av}}^2}{V_{S_{av}}^2} = \left( \frac{1 - \alpha}{1 + \alpha} \right)^2 \\
\phi &= \frac{V_{P_{av}}^2}{V_{P_{av}}^2} = \left( \frac{1 + \alpha}{1 - \alpha} \right)^2 = \frac{1}{\xi} \\
V_{S_{voigt-av}}^2 &= \frac{2V_{S_{\parallel}}^2 + V_{S_{\perp}}^2}{3} = \frac{V_{S_{av}}^2 (1 + \frac{2}{3} \alpha + \alpha^2)}{3} \\
V_{P_{voigt-av}}^2 &= \frac{V_{P_{\parallel}}^2 + 4V_{P_{\perp}}^2}{5} = \nu^2 V_{S_{av}}^2 \left( 1 - \frac{6}{5} \alpha + \alpha^2 \right)
\end{align*}

and whereas their tensor is symmetric about the (3) direction, ours is symmetric about the (1) direction. (N.B. their “$\phi$” is not the same as the direction of the fast azimuth, as $\phi$ is defined elsewhere in this paper. Their $\eta$ is identical to ours.)
The above expressions define velocities of waves propagating along symmetry axes. We substitute the Love parameters into expressions for $V_{SV}$ and $V_{SH}$ as a function of angle, $\zeta$, between the propagation direction and the hexagonal symmetry axis Mavko et al. [2009]:

$$V_{SH}(\zeta) = \sqrt{(N \sin^2 \zeta + L \cos^2 \zeta) / \rho}$$

$$V_{SV}(\zeta) = \sqrt{\left(\frac{A \sin^2 \zeta + C \sin^2 \zeta + L}{\sqrt{(A - L) \sin^2 \zeta - (C - L) \cos^2 \zeta}}\right)^2 + (F + L)^2 \sin^2 2\zeta} / 2\rho$$

(D.5)

to get precise formulae for the solutions to the Christoffel equations. The equation for $V_{SH}$ is straightforward and may be computed precisely. However, it is desirable to simplify the expression for $V_{SV}$; we approximate $V_{SV}$ as a function sinusoidally varying with $90^\circ$ periodicity between $y$ at $0^\circ$ and $90^\circ$ and $V_{SV}|_{\zeta=45^\circ}$ at $45^\circ$:

$$V_{SH} = \sqrt{(x \sin \zeta)^2 + (y \cos \zeta)^2}$$

$$V_{SV} \approx \sqrt{(y \cos 2\zeta)^2 + (V_{SV}|_{\zeta=45^\circ} \sin 2\zeta)^2}$$

(D.6)

We calculate the value of $V_{SV}|_{\zeta=45^\circ}$ as a function of $x$ and $y$ and $\nu$:

$$V_{SV}|_{\zeta=45^\circ} = \sqrt{\left(A + C + 2L - \sqrt{(A - C)^2 + 4(A - L)^2}\right) / 4\rho}$$

$$= \frac{1}{2} \sqrt{\nu^2 x^2 + \nu^2 y^2 + 2y^2 - \sqrt{\nu^4 (x^2 - y^2)^2 + 4(\nu^2 x^2 - y^2)^2}}$$

$$\approx \frac{1}{2} \sqrt{\nu^2 x^2 + \nu^2 y^2 + 2y^2 - 4(\nu^2 x^2 - y^2)^2}$$

$$= \frac{1}{2} \sqrt{\nu^2 (y^2 - x^2) + 4y^2}$$

(D.7)

Where the approximation made in the third step is valid for values of anisotropy less than 10%. Substituting back into equation D.6, we have:
\[ V_{SH} = \sqrt{x^2 \sin^2 \zeta + y^2 \cos^2 \zeta} \]
\[ V_{SV} \approx \sqrt{y^2 \cos^2 2\zeta + (\nu^2 [y^2 - x^2] / 4 + y^2) \sin^2 2\zeta} \]  

(D.8)

Where this approximation introduces an error of <0.3% for \( \zeta > 47^\circ \) or \( \zeta < 15^\circ \) (Figure 3.2). For a horizontal symmetry axis, \( \zeta = 90^\circ - \text{incidence angle} \), and so the approximation introduces negligible error for teleseismic rays in the uppermost mantle. Note that at the extremum of incidence perpendicular to the symmetry axis (\( \zeta = 90^\circ \)) the expressions simplify to \( V_{SH} = x \) and \( V_{SV} = y \), and at the extremum of incidence parallel to the symmetry axis (\( \zeta = 0^\circ \)), \( V_{SH} = V_{SV} = y \).
E Anisotropic travel times and derivatives

For a ray travelling through the anisotropic medium we have described, Equation 3.5 leads to the following expressions for differential travel times for shear waves polarised perpendicular (E-W) or parallel (N-S) to the symmetry axis:

\[
\delta T_E = \int \frac{ds}{(u^2 y^2 + (1 - u^2) x^2)^{1/2}} - \int \frac{ds}{V_{S_{ref}}} \\
\delta T_N \approx \int \frac{ds}{(y^2 + \nu^2 u^2 (1 - u^2) (y^2 - x^2))^{1/2}} - \int \frac{ds}{V_{S_{ref}}}
\]  

(E.1)

and \(\delta \tau_{N-E} = \delta T_N - \delta T_E\). Velocities \(x\) and \(y\) vary along the raypath \(s\), as does the parameter describing propagation direction, \(u\). By inspection it is evident that for vertically incident waves \((i=0\) and so \(u=0\)) the travel times reduce to \(T_E = \int ds/x\) and \(T_N = \int ds/y\), as required. The Fréchet kernels in terms of \(x\) and \(y\) are:

\[
\frac{\partial \delta T_E}{\partial x} = -\int \frac{(1 - u^2) x}{(u^2 y^2 + (1 - u^2) x^2)^{3/2}} \, ds \\
\frac{\partial \delta T_E}{\partial y} = -\int \frac{u^2 y}{(u^2 y^2 + (1 - u^2) x^2)^{3/2}} \, ds \\
\frac{\partial \delta T_N}{\partial x} = \int \frac{\nu^2 u^2 (1 - u^2) x}{(y^2 + \nu^2 u^2 (1 - u^2) (y^2 - x^2))^{3/2}} \, ds \\
\frac{\partial \delta T_N}{\partial y} = -\int \frac{(1 + \nu^2 u^2 (1 - u^2)) y}{(y^2 + \nu^2 u^2 (1 - u^2) (y^2 - x^2))^{3/2}} \, ds
\]  

(E.2)

\[
\frac{\partial \delta \tau_{N-E}}{\partial x} = \frac{\partial \delta T_N}{\partial x} - \frac{\partial \delta T_E}{\partial x} \\
and \frac{\partial \delta \tau_{N-E}}{\partial y} = \frac{\partial \delta T_N}{\partial y} - \frac{\partial \delta T_E}{\partial y}
\]
Three-way cross correlation for $\delta T$ and $\delta \tau$

A shear wave with intermediate polarization (e.g. $\psi = 45^\circ$) travelling through an anisotropic medium with principal directions (also known as “null directions”) N-S and E-W will be split into two quasi-shear pulses that arrive at station $i$ at distinct arrival times $T_N^i$ and $T_E^i$. The splitting time, $\delta \tau_{N-E}^i$, can be measured by cross correlating the arrivals on N-S and E-W components, which should have identical waveforms (although they may have different amplitudes, depending on $\psi$). Arrivals on a given component may also vary in time between different stations, reflecting local velocity heterogeneities. For a given component (e.g. N-S) the differential travel time between stations $i$ and $j$ is $T_N^i - T_N^j = \delta T_N^{i-j}$; this value can be measured by cross correlating arrivals on the N-S channel at different stations.

Therefore, for $M$ stations there are $2M$ unknowns (or model parameters): the times of N-S and E-W arrivals at each station. There are $M^2$ data measurements that can be made in total: $M(M-1)/2$ from cross-correlating N-S arrivals between all station-station pairs, the same number again for all E-W arrivals, and an additional $M$ splitting times. These $M^2$ measurements can be used to simultaneously and self-consistently solve for differential arrival times and splitting times. This is a mixed-determined problem Menke [1984]: differential arrival times are over-determined, but average arrival time is under-determined. Therefore, an appropriate inversion strategy is to use a least-squares approach with the constraint that the average of, say, the N-S differential travel times is zero, similar to VanDecar and Crosson (1990).
We construct the equation $\mathbf{Gm} = \mathbf{d}$:

$$
\begin{align*}
\delta T_{i-j}^N & \left\{ 
\begin{pmatrix}
1 & -1 & 0 & \cdots & 0 \\
0 & 1 & -1 & \cdots & 0 \\
1 & 0 & -1 & \cdots & 0 \\
\vdots & \vdots & \vdots & \ddots & \vdots \\
0 & 0 & 0 & \cdots & 0
\end{pmatrix}
\right. & \begin{pmatrix}
0 & 0 & 0 & \cdots & 0 \\
0 & 0 & 0 & \cdots & 0 \\
0 & 0 & 0 & \cdots & 0 \\
\vdots & \vdots & \vdots & \ddots & \vdots \\
1 & -1 & 0 & \cdots & 0
\end{pmatrix} \\
\delta T_{i-j}^E & \left\{ 
\begin{pmatrix}
0 & 0 & 0 & \cdots & 0 \\
0 & 0 & 0 & \cdots & 0 \\
0 & 0 & 0 & \cdots & 0 \\
\vdots & \vdots & \vdots & \ddots & \vdots \\
1 & 0 & -1 & \cdots & 0
\end{pmatrix}
\right. & \begin{pmatrix}
0 & 0 & 0 & \cdots & 0 \\
0 & 0 & 0 & \cdots & 0 \\
0 & 0 & 0 & \cdots & 0 \\
\vdots & \vdots & \vdots & \ddots & \vdots \\
1 & 0 & -1 & \cdots & 0
\end{pmatrix} \\
\delta \tau_{N-E}^i & \left\{ 
\begin{pmatrix}
1 & 0 & 0 & \cdots & 0 \\
0 & 1 & 0 & \cdots & 0 \\
0 & 0 & 1 & \cdots & 0 \\
\vdots & \vdots & \vdots & \ddots & \vdots \\
0 & 0 & 0 & \cdots & 1
\end{pmatrix}
\right. & \begin{pmatrix}
-1 & 0 & 0 & \cdots & 0 \\
0 & -1 & 0 & \cdots & 0 \\
0 & 0 & -1 & \cdots & 0 \\
\vdots & \vdots & \vdots & \ddots & \vdots \\
0 & 0 & 0 & \cdots & -1
\end{pmatrix} \\
\end{align*}

\begin{pmatrix}
T_N^1 \\
T_N^2 \\
T_N^3 \\
\vdots \\
T_N^M
\end{pmatrix}
= 
\begin{pmatrix}
\delta T_{N}^{1-2} \\
\delta T_{N}^{2-3} \\
\delta T_{N}^{1-3} \\
\delta T_{E}^{1-2} \\
\delta T_{E}^{2-3} \\
\delta T_{E}^{1-3} \\
T_N^1 \\
T_N^2 \\
T_N^3 \\
\vdots \\
T_N^M \\
\vdots \\
\vdots \\
\vdots \\
\delta T_{N-E}^{1} \\
\delta T_{N-E}^{2} \\
\delta T_{N-E}^{3} \\
\vdots \\
\delta T_{N-E}^{M-1} \\
\delta T_{N-E}^{M-2} \\
\delta T_{N-E}^{M-3} \\
\vdots \\
0
\end{pmatrix}

(F.1)
This problem has the classic least squares solution \( \mathbf{m}_{est} = [\mathbf{G}^T \mathbf{G}]^{-1} \mathbf{G}^T \mathbf{d} \) Menke [1984]. We choose to include a weighting function whereby each datum is weighted by the maximum value of the cross-correlation, \( c_{max} \), of that measurement in order to give more weight to measurements that may be more robust. The weighting matrix is a diagonal matrix of squares of \( c_{max} \) values: \( [\mathbf{W}_d]_{i,j} = a_i^2 \delta_{i,j} \) where \( \delta_{i,j} \) is the Kronecker delta. The weighted least squares solution is then: \( \mathbf{m}_{est} = [\mathbf{G}^T \mathbf{W}_d \mathbf{G}]^{-1} \mathbf{G}^T \mathbf{W}_d \mathbf{d} \) Menke [1984].
G  Attenuation measurement using a comb of filters

This appendix provides some background maths working through the use of a comb of filters to measure differential attenuation, first assuming a frequency-independent $Q$ and then introducing the case that $Q$ increases with frequency (as suggested by several observations and experiments).

G.1  Attenuation operator

From Fang and Müller [1991, eq. 1] we have the delayed attenuation operator in the frequency domain:

$$D(\omega, z) = \exp\left(\frac{-\omega z}{2c_{\infty}Q(\omega)}\right) \exp\left[-i\omega z \left(\frac{1}{c(\omega)} - \frac{1}{c_{\infty}}\right)\right]$$  \hspace{1cm} (G.1)

We express this in terms of a reference velocity at 1 Hz, and take the maximum velocity $c_{\infty} = c_0 (1 + Q_{\max}^{-1})$. Within the absorption band, the velocity at a given frequency relative to the reference velocity is given by the Kramers-Kronig dispersion relation for $Q \gg 1$ [e.g., Anderson and Minster, 1979]:

$$\frac{c(\omega_2)}{c(\omega_1)} = 1 + \frac{1}{\pi Q} \ln\left(\frac{\omega_2}{\omega_1}\right)$$  \hspace{1cm} (G.2)

assuming constant $Q(\omega) = Q_0 = Q$ (for now) and defining $c_0$ at 1 Hz, we have:

$$c_{\infty} = c_0 \left[1 + \frac{1}{Q}\right] \quad \text{(implying } \omega_{\infty} = 2\pi e^{\pi} \text{ rad/s)}$$  \hspace{1cm} (G.3a)

$$c(\omega) = c_0 \left[1 + \frac{1}{\pi Q} \ln\left(\frac{\omega}{2\pi}\right)\right]$$  \hspace{1cm} (G.3b)
Note: this construction leads to non causal arrivals at $\omega > 2\pi e^{\pi}$ rad/s and therefore breaks down at high frequencies. This problem is resolved if we account for frequency dependency of $Q$ (see below).

We will find it useful to define the important parameter $t^*$:

$$t^* = \int_L \frac{ds}{c(\omega)Q(\omega)} \approx \frac{z}{c_0Q} \quad \text{(G.4)}$$

in which the second equality is an approximation we will use to simplify equation G.5.

Substituting equations G.3a and G.3b into equation G.1, and using a Taylor series to expand the $1/c$ terms, discarding $(Q^{-2})$ and higher terms, we obtain:

$$\mathcal{D}(\omega, z) = \exp \left( -\frac{1}{2} \omega t^* \left[ 1 + \frac{1}{\pi Q} \ln \left( \frac{\omega}{2\pi} \right) \right]^{-1} \right) \exp \left( -\frac{1}{\pi} i\omega t^* \ln \left( \frac{\omega}{\omega} \right) \right)$$

$$\approx \exp \left( -\frac{1}{2} \omega t^* \right) \exp \left( -\frac{1}{\pi} i\omega t^* \ln \left( \frac{\omega}{\omega} \right) \right) \quad \text{(G.5)}$$

$$= Ae^{-i\omega \phi}$$

where

$$A(\omega) = \exp \left( -\frac{1}{2} \omega t^* \right) = \exp (-\pi ft^*) \quad \text{(G.6a)}$$

$$\phi(\omega) = \frac{1}{\pi} t^* \ln \left( \frac{\omega}{\omega} \right) = \frac{1}{\pi} t^* \left( \ln(f) - \ln(f) \right) \quad \text{(G.6b)}$$

$$= t^* \left[ 1 - \frac{1}{\pi} \ln \left( \frac{\omega}{2\pi} \right) \right] = t^* \left[ 1 - \frac{1}{\pi} \ln(f) \right]$$

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G.2 Differential amplitude and phase spectra

Now consider a waveform given by $g(t)$ that encounters an attenuating volume. If the Fourier transform of this waveform is $\hat{G}(\omega)$, where:

$$g(t) = \frac{1}{\sqrt{2\pi}} \int_{-\infty}^{\infty} \hat{G}(\omega) e^{i\omega t} d\omega \quad \text{(G.7)}$$

the attenuated waveform, $g'(t)$, is given by:

$$g'(t) = \frac{1}{\sqrt{2\pi}} \int_{-\infty}^{\infty} \hat{G}(\omega) D(\omega, z) e^{i\omega t} d\omega$$

$$= \frac{1}{\sqrt{2\pi}} \int_{-\infty}^{\infty} A(\omega) \hat{G}(\omega) e^{i\omega (t - \phi(\omega))} d\omega \quad \text{(G.8)}$$

Where the prime notation indicates the post-attenuation waveform. If we apply a narrow bandpass filter to this waveform, we may describe this (in the ideal case) as a multiplication in the frequency domain with a boxcar centered at $\omega_0$ with width $\Delta \omega$. In the limit that $\Delta \omega \to 0$, this filter function $\to \delta(\omega - \omega_0)$, where $\delta$ is the Dirac delta function\(^7\). So:

$$g'(t)|_{\omega_0} \to \frac{1}{\sqrt{2\pi}} \int_{-\infty}^{\infty} A(\omega) \hat{G}(\omega) e^{i\omega (t - \phi(\omega))} \delta(\omega - \omega_0) d\omega$$

$$= A(\omega_0) g(t - \phi(\omega_0))|_{\omega_0} \quad \text{(G.9)}$$

i.e. a scaled, phase-shifted version of the original trace filtered to that frequency. For positive (i.e. physically plausible) $t^*$ and finite frequency: the amplitude factor, $A$, will be in the range $0 < A < 1$, will be smaller if $t^*$ is greater, and will have a negative slope with frequency (more attenuation at high-$f$). The phase shift, $\phi$, will be $> 0$ (i.e. some phase lag), will be greater if $t^*$ is greater, and will also have a negative slope with frequency (less dispersion at high-$f$).

\(^7\)In practice, the filter width is finite, so if phase or amplitude changes rapidly, slight biases will be introduced into the measurements.
Equation G.9 implies that if we have the un-attenuated waveform, we can apply a comb of narrow-band filters centered at $\omega_i$ to a given signal to build up the distribution of $A(\omega)$ and $\phi(\omega)$ by time-shifting and multiplying $g'(t)|_{\omega_i}$ to match $g(t)|_{\omega_i}$.

For teleseismic arrivals, we do not know the original un-attenuated waveform but instead consider two waveforms arriving at separate stations ($g'_1$ and $g'_2$) that have passed through structure with distinct velocity and attenuation values, $c_1$, $Q_1$ and $c_2$, $Q_2$. In theory, their initial waveforms were the same (albeit unknown) prior to interacting with the structure of interest. Filtered to any frequency, $\omega_i$, these two waveforms will then be given by:

$$g'_1(t)|_{\omega_i} = S_1 A_1(\omega_i) g(t - \phi_1(\omega_i) - \delta T_1 + f(\Delta_1))|_{\omega_i} \quad (G.10a)$$
$$g'_2(t)|_{\omega_i} = S_2 A_2(\omega_i) g(t - \phi_2(\omega_i) - \delta T_2 + f(\Delta_2))|_{\omega_i} \quad (G.10b)$$

where we have introduced constant station gain terms ($S_i$), and $\delta T_1$ and $\delta T_2$ are frequency independent differential travel times caused by the difference between $c_1$ and $c_2$, respectively, and the reference velocity, $c_0$: $\delta T_i = z(1/c_i - 1/c_0)$. $f(\Delta_{1,2})$ are additional travel time effects due to structure outside the proximate volume (i.e. the region in which we are measuring lateral $t^*$ discrepancies) and are unknown functions of the source-receiver distance, $\Delta$. We attempt to remove these terms using travel times from a 1-D Earth model, but discrepancies between this model and true Earth structure will contribute errors. We drop the $f(\Delta)$ terms in the analysis below.
It follows that:

\[ g'_2(t)|_{\omega_i} = \Delta A(\omega_i)g'_1(t - \Delta \phi(\omega_i) - \delta T)|_{\omega_i} \]

\[ = \Delta A(\omega_i)g'_1(t - \Delta \psi(\omega_i))|_{\omega_i} \]  \hspace{1cm} (G.11)

where

\[ \Delta A(\omega) = S_2 A_2(\omega)/S_1 A_1(\omega) = (S_2/S_1) \exp \left( -\frac{1}{2} \omega \Delta t^* \right) \]  \hspace{1cm} (G.12a)

so

\[ \ln \Delta A(\omega) = \ln (S_2/S_1) - \pi \Delta t^* f \]

\[ \Delta \psi(\omega) = (\phi_2(\omega) - \phi_1(\omega)) + \delta T = \frac{1}{\pi} \Delta t^* \ln \left( \frac{\omega\infty}{\omega} \right) + \delta T \]  \hspace{1cm} (G.12b)

\[ = -\frac{1}{\pi} \Delta t^* \ln f + \Delta t^* + \delta T \]

and we have defined

\[ \delta T = \delta T_2 - \delta T_1 = z \left( \frac{1}{c_2} - \frac{1}{c_1} \right) \approx \frac{z}{c_{av}} \delta c \]  \hspace{1cm} (G.13a)

\[ \Delta t^* = t^*_2 - t^*_1 = z \left( \frac{1}{c_2 Q_2} - \frac{1}{c_1 Q_1} \right) \approx \frac{z}{c_{av}} \left( \frac{1}{Q_2} - \frac{1}{Q_1} \right) \]  \hspace{1cm} (G.13b)

To be clear, positive \( \delta T \) means \( c_1 \) faster than \( c_2 \); arrives later at station 2. Positive \( \Delta t^* \) means \( Q_1 \) greater than \( Q_2 \); more attenuated at station 2.

Therefore,

- a plot of \( \ln(\Delta A) \) vs. \( f \) will have slope equal to \( -\pi \Delta t^* \) and an intercept equal to the logarithm of the ratio of station amplification terms (and any other frequency-independent terms like geometric spreading/radiation pattern).

- a plot of \( \Delta \psi \) vs. \( \ln(f) \) will have slope equal to \( -\Delta t^*/\pi \) and intercept equal to the sum of \( \Delta t^* \) and \( \delta T \) (plus a “nuisance factor” term arising from using an imprecise 1-D model to correct for distance effects).  

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G.3 Frequency-dependent $Q$

Thus far, we have proceeded under the assumption of frequency independent $Q$. We will now explore the consequence of frequency dependent $Q(\omega) = Q_0 (\omega/\omega_0)^\alpha$, where $Q_0$ is now defined at $\omega_0$. The anharmonic velocity, $c_0$, is defined at infinite frequency (assuring causality). This assumption modifies equation G.3 to:

\[ c_\infty = c_0 \]  
\[ c(\omega) = c_0 \left[ 1 - \frac{1}{2Q_0} \cot \left( \frac{\alpha \pi}{2} \right) \left( \frac{\omega}{\omega_0} \right)^{-\alpha} \right] \]

[Karato, 1993; Minster and Anderson, 1981]. For notational convenience, we will henceforth use $\gamma = \cot (\alpha \pi / 2) / 2$.

The frequency-dependent attenuation operator is now:

\[ \mathcal{D}(\omega, z) = \exp \left( -\frac{1}{2} \omega t_0^* (\omega/\omega_0)^{-\alpha} \right) \exp \left( -i\omega \frac{z}{c_0} \left[ 1 - \gamma \frac{\omega}{Q_0 (\omega_0)} \left( \frac{\omega}{\omega_0} \right)^{-\alpha} \right]^{-1} - 1 \right) \]

Applying a Maclaurin expansion, $[1 - x]^{-1} \approx [1 + x + x^2 + x^3 + x^4 + \cdots]$ with

\[ x = \frac{\gamma \omega}{Q_0 (\omega_0)^{-\alpha}} \ll 1 \]

we obtain

\[ \mathcal{D}(\omega, z) \approx \exp \left( -\frac{\omega_0^\alpha}{2} t_0^* \omega^{1-\alpha} \right) \exp \left( -i\omega \left[ x + x^2 + x^3 + x^4 + \cdots \right] \right) \approx \exp \left( -\omega \gamma (\omega_0)^\alpha \right) \exp \left( -i\omega^{1-\alpha} t_0^* \gamma (\omega_0)^\alpha \right) = Ae^{-i\omega \phi} \]

where

\[ A(\omega) = \exp \left( -\frac{1}{2} \omega_0^\alpha \omega^{1-\alpha} t_0^* \right) \]
\[ \phi(\omega) = \gamma \omega_0^\alpha \omega^{\alpha} t_0^* \]

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and the approximation in Equation G.17 neglects \( x^2 \) and higher terms\(^8\). \( t_0^* \) refers to the value of \( t^* \) at the reference frequency (1 Hz).

Considering differential values of amplitude and phase, and applying the same logic as for Equations G.12a and G.12b, we obtain:

\[
\Delta A(\omega) = S_2 A_2(\omega) / S_1 A_1(\omega) = S \exp\left( -\frac{1}{2} \omega_0^\alpha \omega^{1-\alpha} \Delta t_0^* \right) \tag{G.19a} \\
= S \exp\left( -\pi f_0^\alpha f^{1-\alpha} \Delta t_0^* \right) \\
\Delta \psi(\omega) = (\phi_2(\omega) - \phi_1(\omega)) + \delta T = \gamma \omega_0^\alpha \omega^{-\alpha} \Delta t_0^* + \delta T \tag{G.19b} \\
= \gamma f_0^\alpha f^{-\alpha} \Delta t_0^* + \delta T
\]

Therefore, if we define the reference frequency, \( f_0 \), at 1 Hz:

- \( f^{1-\alpha} \) versus \( \ln(\Delta A) \) should have a slope of \(-\pi \Delta t_0^* \) and an intercept equal to the logarithm of the ratio of station amplifications.

- \( f^{-\alpha} \) versus \( \Delta \psi \) should have a slope of \(0.5 \cot(\alpha \pi/2) \Delta t_0^* \) and an intercept of \( \delta T \) (plus a “nuisance factor” term arising from using an imprecise 1-D model to correct for distance effects).

Effects of incorrect assumptions about \( \alpha \)

One notable feature of this approach is that there is a certain tradeoff between \( \alpha \) and apparent \( \Delta t_0^* \). This leads to some predictable consequences for assuming incorrect \( \alpha \). For simplicity, here we will consider only the amplitude spectrum. We have shown that the key to measuring apparent \( t^* \) is the slope of the amplitude spectrum. Moreover, the value of reference attenuation, \( t_0^* \) is defined by the spectral slope at the

\[^8\text{From Equation G.16, the approximation is valid if } Q_0 \gg (T_{\text{max}})^\alpha, \text{ where } T_{\text{max}} \text{ is the maximum period utilized. This approximation breaks down at small values of } Q \text{ or at long periods. For the lowest frequency in our study (period = 30 s), the approximation contributes an error of less than 5\% for values of } Q \geq 12, \text{ assuming } \alpha = 0.27.\]
reference frequency, $\omega_0$. As above, we have:

$$\ln A(\omega) = -\frac{1}{2} \omega^{1-\alpha} t_0^* \omega_0^\alpha$$

(G.20)

Therefore the slope ($b$) at any point is given by:

$$b(\omega) = \frac{d \ln A(\omega)}{d\omega} = -\frac{1 - \alpha}{2} \left( \frac{\omega_0}{\omega} \right)^\alpha t_0^*$$

(G.21)

Now consider the slope at $\omega_m$, the center frequency of the band-limited measurements. Assuming measurements are made in a reasonably narrow frequency band, this is roughly the slope being fit by our inversion algorithm:

$$b(\omega_m) = -\frac{1 - \alpha}{2} \left( \frac{\omega_0}{\omega_m} \right)^\alpha t_0^*$$

(G.22)

and the estimated value of $t_0^*$ will be

$$t_0^{*\text{est}} = \frac{-2 b(\omega_m)}{1 - \alpha} \left( \frac{\omega_m}{\omega_0} \right)^\alpha$$

(G.23)

Now, consider if $\alpha$ truly is zero ($f$-independent $Q$): assuming $\alpha = 0$ will (absent noise) yield the ‘true’ $t_0^*$:

$$t_0^{*\text{true}} = -2 b(\omega_m)$$

(G.24)

while assuming $\alpha > 0$ will yield $t_0^{*\text{est}}$ as above. Thus, a faulty assumption of $\alpha > 0$ will lead to an underestimate of attenuation if the measurement frequency is lower than the reference frequency:

$$\frac{t_0^{*\alpha>0}}{t_0^{*\text{true}}} = \frac{1}{1 - \alpha} \left( \frac{\omega_m}{\omega_0} \right)^\alpha$$

(G.25)

where $\alpha$ here is the erroneously assumed value ($\alpha_{\text{est}}$) used in the calculations.

Conversely, if $\alpha$ truly is larger than zero ($f$-dependent $Q$), then the assumption of $\alpha = 0$ will lead to an overestimate of true attenuation:

$$\frac{t_0^{*\alpha=0}}{t_0^{*\text{true}}} = (1 - \alpha) \left( \frac{\omega_m}{\omega_0} \right)^{-\alpha}$$

(G.26)
where \( \alpha \) here is the true value (\( \alpha_{\text{tru}} \)) experienced by the seismic waves.

If the reference frequency is 1 Hz and the measurements are at longer period, then a good rule of thumb results:

- incorrectly assuming \( \alpha = 0 \) overestimates \( t_0^* \) by a factor of \((1 - \alpha_{\text{tru}}) T^{\alpha_{\text{tru}}}\)

- incorrectly assuming \( \alpha > 0 \) underestimates \( t_0^* \) by a factor of \( T^{-\alpha_{\text{est}}} / (1 - \alpha_{\text{est}}) \)

Where \( T \) is the mean period of the measurements. If the mean frequency of the measurements is greater than 1 Hz then the sense (i.e. over/under) of the estimation error will be the opposite, but the formulae still hold.
Figure G.2: Illustration of the effects of choosing the wrong $\alpha$ value. The true value of $t_0^*$ at 1 Hz is fixed to be 1.63. We then extrapolate the amplitude reduction as a function of frequency (blue curves) for both $\alpha = 0$ and $\alpha = 0.27$. The amplitudes are then sampled to give ‘data’ at a range of 30 frequencies between 0.05 and 0.333 Hz (approximately the same as the measurements used in the paper, and giving $T=6.6$ s). Red circles are the ‘data’ if $\alpha_{\text{true}} = 0.27$, green circles are the ‘data’ if $\alpha_{\text{true}} = 0$. Each of these data sets are then inverted for $t_0^*$ (and a constant amplification term), assuming either $\alpha = 0$ or $\alpha = 0.27$. The measured values for $t_0^*$ are shown at the reference frequency ($\omega_0 = 2\pi$) and are simple functions of the gradient at this point (Equation G.21).

Inversions for $t_0^*$ that assume correct $\alpha$ do a good job of recovering the true value of 1.63.

The inversion that overestimates true $\alpha$ (i.e. $\alpha_{\text{est}} = 0.27$ while $\alpha_{\text{true}} = 0$) underestimates true $t_0^*$ by a factor of 0.83 (for $T=6.6$ s, Equation G.25 predicts 0.82).

The inversion that underestimates true $\alpha$ (i.e. $\alpha_{\text{est}} = 0$ while $\alpha_{\text{true}} = 0.27$) overestimates true $t_0^*$ by a factor of 1.20 (for $T=6.6$ s, Equation G.25 predicts 1.22).

Inset: zoom in on region of ‘data’, showing that for both sets of data, either assumption about $\alpha$ yields almost identically excellent fits to the measurements – the upshot is that discriminating which $\alpha$ is correct from the data fit will be highly challenging over a limited frequency band.
Additional attenuation results

This appendix includes additional results figures for the Cascadia attenuation study that were not included in the main body of the text. This material comprises the attenuation and travel time results from on-land stations (Figures H.1 and H.2), as well as the attenuation results computed using frequency-dependent $Q$ (with frequency exponent, $\alpha = 0.27$ (Figures H.3 and H.4).

The frequency-dependent $\Delta t^*$ results were calculated from the same phase and attenuation spectra as the results in the main body of the text, but now assuming that $Q$ increases with frequency. $Q$ is defined relative to its value at a reference frequency ($\omega_0$) by a power law according to $Q(\omega) = Q_0(\omega/\omega_0)^\alpha$ (Section 4.3.1 and Appendix G.3). In this case, we assume $\alpha = 0.27$, in accordance with other groups’ observations [Faul and Jackson, 2005; Lekic et al., 2009; Stachnik, 2004] and experimental results [Jackson and Faul, 2010; Jackson et al., 2002].

The effect of assuming positive frequency dependence is, in general, to reduce $\Delta t^*$. This comes about because $\alpha$ and $\Delta t^*$ trade off to some extent, as a steeper sloped amplitude or phase spectrum can be fit (to some extent) by an increase in either of these parameters. See Section G.3).
Figure H.1: Maps of differential attenuation $\Delta t^*$, recorded at OBS station for $S$-waves (a) and $P$-waves (b), including land station results that are used to link Juan de Fuca and Gorda arrays. Radial spokes show individual arrivals at their incoming azimuth, while central circles show least squares station average terms, having accounted for event $\Delta t^*$ values.
Figure H.2: Maps of differential travel time recorded at OBS station for $S$-waves (a) and $P$-waves (b), including land station results that are used to link Juan de Fuca and Gorda arrays. Radial spokes show individual arrivals at their incoming azimuth, while central circles show least squares station average terms, having accounted for event terms. Travel times are corrected for station elevation assuming crustal velocity of $V_P = 6.5$ km/s or $V_S = 3.4$ km/s. We do not correct for sediment thickness.
Figure H.3: Maps of differential attenuation, $\Delta t^*$, recorded at OBS station for $S$-waves (a) and $P$-waves (b). These results (unlike in Figure 4.5) were computed assuming $Q(\omega) = Q_0(\omega/\omega_0)^{0.27}$. Radial spokes show individual arrivals at their incoming azimuth, while central circles show least squares station average terms, having accounted for event $\Delta t^*$ values. Open circles show land stations used to link Juan de Fuca and Gorda arrays. Boxes in (b) show three areas: north Juan de Fuca (mauve), south Juan de Fuca (red), and Gorda (blue).
Figure H.4: Station-averaged $\Delta t^*$ for $S$-waves (top) and $P$-waves (bottom) measured at OBS stations, plotted as a function of age of oceanic crust [Wilson, 1993]. These results (unlike in Figure 4.7) were computed assuming $Q(\omega) = Q_0(\omega/\omega_0)^{0.27}$. Black curves indicate theoretical predictions from [Faul and Jackson, 2005] computed at three different grain sizes, for plate cooling model with $T_p = 1350^\circ$C. Attenuation averages are corrected for event terms, as above. Mauve dots: north Juan de Fuca stations; red dots: south Juan de Fuca stations; blue dots: Gorda stations (see Figure H.3 for area designation). Symbol size $\propto 1/N_{obs}$. Only stations with $N_{obs} \geq 5$ plotted.