ANISOTROPIC STRUCTURAL TRANSITION IN THE MANTLE BENEATH WESTERN AND CENTRAL WYOMING BASED ON SHEAR-WAVE SPLITTING

by
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A Thesis Submitted to the Faculty of the DEPARTMENT OF GEOSCIENCES In Partial Fulfillment of the Requirements for the Degree of
MASTER OF SCIENCE In the Graduate College
THE UNIVERSITY OF ARIZONA

2011
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I would like to dedicate this thesis to my Lord, God, who makes all things possible, my loving family, for always supporting my passion for the arts and sciences, and to all my teachers, for providing me the knowledge to succeed academically.
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ABSTRACT

We have analyzed shear-wave splitting in central and western Wyoming to characterize mantle anisotropy. Data were used from the 63 stations of the LaBarge Passive Source Seismic Experiment (LPSSE) and 58 additional regional stations. Two anisotropic domains are observed beneath western Wyoming: a western domain located over relatively thin lithosphere with a NE-SW fast direction and an eastern domain situated over thicker lithosphere with a NW-SE to W-E fast direction. The spatial coherency of splits indicates that anisotropy near the wake of the Yellowstone hotspot is centered at 110-210 km depth; however, the alignment of fast directions with the absolute plate motion of North America and observations of low-frequency splitting energy at stations closest to the Eastern Snake River Plain indicate that the anisotropy is within the asthenosphere. In contrast, the main source of splitting in the Wind River Basin, roughly the Sweetwater subprovince, is lithospheric and is estimated at 100-130 km beneath the surface. Because the tectonic grain of the subprovince and the estimated fast directions are parallel, we further conclude that anisotropy from the eastern splitting domain is best explained by the late Archean fabric of the Sweetwater subprovince. Possible backazimuthal dependence of the splitting parameters measured at stations east of the Wind River Range demonstrates $\pi/2$ periodicity, which may be evidence of vertically changing anisotropy. This, along with results demonstrating the dependence of delay time with depth or across strike, may explain the sparse number of splits and considerable number of nulls found at stations within the Green River Basin, including the LPSSE,
which could imply that a transition in anisotropic structure exists between the Eastern Snake River Plain and the Sweetwater subprovince.
INTRODUCTION

Earthquakes give us the ability to remotely study the elastic properties of the Earth and, by extension, its chemistry, temperature, pressure, and mineralogy. Additionally, Earth’s internal deformation can be observed as a result of seismic anisotropy, revealed by the bipolarization of shear waves travelling through an ordered medium. Among the most useful tools available to image subsurface geology, shear-wave birefringence has become the preferred technique for investigating anisotropy at depth. Furthermore, shear-wave analysis has the capability of resolving past and present rock deformation (or flow), depending upon its origin in either the lithosphere or asthenosphere. The rich tectonic history of Wyoming makes it an ideal place for studying seismic anisotropy. Episodes of cratonization and orogenic cycling in present-day Wyoming resulted in a rapid transition in lithospheric thickness that roughly coincides with the Rocky Mountain (Laramide) Tectonic Front [Yuan et al., 2011]. This strategic location allows us to answer questions regarding the relative contribution of the lithosphere and asthenosphere to continental Shear-Wave Splitting (SWS), to what degree lithospheric thickness affects lateral heterogeneities in anisotropy, and how well these changes can be observed at very short distances. The finite frequency of a shear wavefield allows us the opportunity to measure anisotropy at different length scales. The densely spaced LaBarge Passive Source Seismic Experiment (LPSSE) was deployed to document the presence of small-scale anisotropic heterogeneities at depth. The first part of our study investigates variations in SWS at very close spacing (250 m). The second incorporates additional stations within the vicinity of the LaBarge Array to investigate the
regional anisotropy. The LaBarge Array is located between the strikingly coherent
splitting estimates across the Eastern Snake River Plain (eSRP) [Schutt et al., 1998;
Schutt and Humphreys, 2001; and Waite et al., 2005] and the more complicated
anisotropy documented in the Rocky Mountains [Schutt and Humphreys, 2001; Savage et al., 1996]. We have utilized three different analytical approaches to accomplish these
goals and ultimately improve the robustness of our interpretations. First we use standard
SWS waveform analysis, which employs a single frequency band and relatively wide
filter and allows us to compare our results with previous studies [cited above]. These
results are the basis of our additional approaches, frequency and ray-piercing-point
analysis, which are utilized to estimate the “true” depth of observed SWS and reveal the
potential existence of vertically and laterally dependent anisotropy, respectively [Long,
2010; Gao et al., 2010].

TECTONIC SETTING

I. Wyoming Craton

The Wyoming Province is an Archean craton underlying most of present-day
Wyoming and some parts of adjacent states [Frost and Frost, 1993] (Figure 1A). It is
surrounded by three Proterozoic orogens, including the Great Falls tectonic zone to the
north, the Dakota (Trans-Hudson) Orogen to the east, and the Cheyenne belt (Colorado
Orogen) to the south [Mueller and Frost, 2006; Snoke, 1993] (Figure 1A). The Wyoming
Province is primarily comprised of three rock types: potassic granitoids (mostly granite)
derived largely from reworked older gneiss, orthogneiss and paragneiss, and a
subordinate quantity of supracrustal metavolcanic-metasedimentary rocks [Sims et al.,
Sims et al. [2001] used magnetic contrasts between granitoids and gneisses to map the gross lithology of covered portions of the Precambrian Wyoming basement, classifying the two principal rock units as magmatic domains and gneiss domains, respectively. Their study also revealed the overall structural pattern of the basement unit to be roughly semicircular, centered at the Bighorn basin, which is similar in shape and subparallel to the outer margin of the Wyoming craton.

The Montana metasedimentary province (MMP) and Beartooth-Bighorn Magmatic Zone (BBMZ) are the earliest subdivisions of the province to cratonize (~2.8 Ga) [Mueller and Frost, 2006]. The Southern Accreted Terranes (SAT) lie to the south of the BLMZ and were cratonized during the late Archean after a collection of terranes accreted to the southern margin of the Wyoming Province [Mueller and Frost, 2006]. Subduction-related deformation and metamorphism produced by the docking of the SAT created the Sweetwater subprovince, a subdivision of the BBMZ [Mueller and Frost, 2006; Chamberlain et al., 2003]. Intense Proterozoic structural overprinting occurred along the craton’s southern and eastern margins during the Colorado (1.78-1.75 Ga) and Trans-Hudson Orogenies (2.0-1.8 Ga) [Sims et al., 2001]. The collision of the Green Mountain volcanic arc with the Wyoming craton occurred during the Colorado Orogen and created the Cheyenne belt, an Archean-Proterozoic suture, which greatly deformed and metamorphosed Archean rock at least 75km inboard of the Laramie Mountain suture [Yuan and Dueker, 2005; Sims et al., 2001; Jones et al., 2011]. The Archean rocks of the Hartville uplift and Laramie Mountains were similarly deformed by the Trans-Hudson orogen along the craton’s eastern margin creating the Hartville uplift-Black Hills block
(HU-BH), another subdivision of the Wyoming craton [Sims et al., 2001; Mueller and Frost, 2006].

II. Sevier-Laramide Orogeny

Wyoming’s modern mountainous terrain was formed as a result of Mesozoic-early Cenozoic regional deformation during the Sevier-Laramide orogeny [e.g. Brown, 1993; Royse, 1993; Sims et al., 2001]. The Sevier Orogeny of the Late Jurassic to early Tertiary created the classic central foreland fold and thrust belt of North America [Taylor et al., 2000; Livaccari, 1991]. The Wyoming salient of the orogeny comprises three major thrust systems, which underwent episodic tectonism until the Early Cretaceous [Royse, 1993]. Upper crustal shortening was accommodated by the transfer of stress eastward along the weak bedding planes of Paleozoic and Mesozoic sedimentary rocks, thus resulting in the predominantly thin-skinned deformation that characterizes Sevier thrusting [Armstrong, 1968; DeCelles, 2004]. Sims et al. [2001] similarly argue that the lack of magnetic anomaly disruption in western Wyoming is evidence that Sevier contractile deformation left basement rock largely undisturbed.

The Laramide Orogeny (~75 - ~45Ma) overlapped in time and space with the Sevier orogeny and is defined by a decrease in magmatism and significant tectonic activity far inboard of the convergence boundary separating North America and the Farallon plate [Dickinson and Snyder, 1978; Humphreys, 2009]. Flat-slab subduction of the Farallon Plate is a commonly cited cause of these two phenomena [Dickinson and Snyder, 1978; Humphreys, 1995; Saleeby, 2003; DeCelles, 2004]. Farallon shallow subduction is further thought to be responsible for the basement-cored uplifts that typify
Laramide deformation as well as for disrupting the immense Western Interior Basin [Dickinson and Snyder, 1978; Brown, 1993; DeCelles, 2004]. In Wyoming, the Laramide tectonic front progressed eastward through time, uplifting the Wind River Range during the Maastrichtian, then the Owl Creek Mountains in the Paleocene to early Eocene, and finally the Bighorns and Beartooth Range during the early Eocene, with deformation ceasing after the middle Eocene [Brown, 1993] (Figure 1B). These compression-induced structures are not isolated, instead creating a system of anastomosing “arches” [Erslev, 1993; Sims et al., 2001] (Figure 1B). Primary Laramide structures trend northwest-southeast but are overprinted by subordinate northeast, east, and north oriented grains, resulting in a block-like pattern of faulted folds [Brown, 1993]. It has been additionally suggested that basement (Precambrian) structural fabrics strongly influenced Laramide features in Wyoming [Brown, 1987; Tonsson, 1986; Chamberlain et al., 2003; Frost et al., 2006; Mueller and Frost, 2006]. Although there is a wealth of data on the surface geology of the region, less is known about the structure of the underlying mantle and how stress is transferred between the crust and mantle if at all. Our study addresses anisotropy in the upper mantle using SWS analysis.

BACKGROUND

I. Seismic Anisotropy

Anisotropy is simply the directional dependence of a material’s elastic properties. More specifically, seismic anisotropy is the change in seismic wave speed with propagation direction and polarization (vibration) through an anisotropic medium [Silver and Chan, 1991; Savage, 1999; Long and Silver, 2009]. Near-vertically propagating S-
phases travelling through anisotropic medium with a horizontal symmetry axis “spilt” into a fast and slow component polarized in their respective directions [Backus, 1965; Silver and Chan, 1991; Vinnik et al., 1992; Fouch and Rodenay, 2006]. This phenomenon, known as shear-wave splitting (SWS), yields two splitting parameters—delay time (δt), the lag time between the fast and slow components, and fast polarization direction (φ), the direction of maximum finite extension [Silver & Chan, 1988; Silver and Chan, 1991; Vinnik et al., 1992] (Figure 2B). Knowledge of the splitting parameters allows us to infer the strain-related properties of the medium at depth [Backus, 1965], including the direction of deformation or mantle flow, determined from φ, and the intensity and thickness of the source of anisotropy, represented by δt [Silver and Chan, 1991].

Core phases (SK(K)S, PKS, etc.) are the most popular phases used in SWS analysis for several reasons. First, the direction of polarization is known because polarization is controlled by the P-to-SV conversion at the Core-Mantle Boundary (CMB). Second, the observed splitting is constrained to the receiver side of their raypaths because source-side splitting is removed by P-to-SV conversion at the CMB. Finally, their steep incidence angles guarantee wave propagation within the “shear-wave window” [e.g. Savage, 1999; Silver and Long, 2009; Wüstefeld et al., 2008]. A schematic of travel paths for core phases utilized for our study is illustrated in Figure 2B.

II. Depth (Location) of Anisotropy

Because their angle of incidence is nearly vertical, teleseismic body waves have excellent lateral resolution but poor vertical resolution [e.g. Fouch and Rodenay, 2006;
Silver and Long, 2009; Savage, 1999]. Consequently, this makes the task of locating the depth of anisotropy challenging. Below are brief discussions concerning SWS observations at various layers within the Earth and their contribution to splitting measurements (Figures 3).

A. Crust

Crampin [1994] concluded in his review of global evidence for crustal anisotropy that the majority of near-surface birefringence results from the shape-preferred orientation (SPO) of stress-aligned cracks and microcracks in the uppermost crust. This tends to cause $\varphi$ to parallel cracks, which often coincides with the direction of maximum principle stress, with the slow direction orthogonal to this [Crampin, 1985; Crampin and Lovell, 1991]. Furthermore, cracks cause differences in S-wave velocity from 1.5-4%. It is, therefore, often assumed that $\delta t$ of crustal anisotropy is $< 0.3$ s, making it a minor contributor to observed $\delta t$ values of $\sim 1$ s for the majority of raypath geometries [e.g. Silver, 1996; Savage, 1999; Fouch and Rodenay, 2006]. The crust’s small effect on $\delta t$ is expected considering teleseismic waves propagate almost vertically through crustal rocks, which typically have horizontal foliation planes [Barruol and Mainprice, 1993]. At depths greater than 10-15 km, lab experiments indicate the closure of cracks occurs in the lower crust as a result of lithostatic pressure, effectively eliminating crack-induced anisotropy [Kern, 1990; Hrouda et al., 1993]. This additionally suggests that observed splitting in the lower crust must result from other mechanisms [Savage, 1999] including:

(a) pure shear or simple shear deformation, which aligns the seismic fast axes of
anisotropic minerals [Karato, 1998], or (b) the SPO of magmatic dikes, shear zones, or other structures [Crampin, 1985].

B. Upper Mantle

Despite poor vertical resolution, many studies agree that anisotropy is mostly present in the upper 400 km of the Earth, albeit with little crustal contribution, as noted above [e.g. Vinnik et al, 1992; Barruol and Mainprice, 1993; Vinnik et al., 1996]. The upper mantle is also the strongest source of anisotropy, having the greatest affect on \( \delta t \) of any layer within the Earth [Savage, 1999]. S-wave birefringence at this depth is generally believed to result from the lattice preferred orientation (LPO) of olivine [e.g. Savage, 1999; Long and Silver, 2009], as olivine constitutes the bulk of the Earth’s upper mantle [Ringwood, 1975]. A further reason for strong upper mantle anisotropy in olivine is that individual crystals are highly birefringent, with anhydrous olivine having an inherent anisotropy of \( \sim 24\% \) [Blackman, 2007]. When olivine plastically deforms in aggregate, LPO is generated, owing to dislocation creep (the dominant form of deformation at depth) [Karato & Wu, 1993]. Next to olivine the effect of enstatite and other secondary mantle minerals on anisotropy is often minor, except in cases where olivine-induced birefringence is weak [Karato et al., 2008].

Measurements of LPO in mantle xenoliths and experiments on single, anhydrous olivine crystals and aggregates suggest that the olivine a-axis (\( \phi \)) preferentially aligns with the maximum shear direction in a high temperature (\( 1300^\circ \) C), high strain, and simple shear environment [Blackman, 2007; Zhang and Karato, 1995] (Figure 3A). This implies that \( \phi \) directions approximately coincide with the direction of dislocation creep.
[Silver and Chan, 1988; Ribe and Yu, 1991; Ribe, 1992; Zhang and Karato, 1995]; i.e., fast splitting directions reveal the direction of mantle flow. However, new findings from the past decade challenge this simple interpretation and instead suggest that olivine LPO is dependent on water content, temperature, and stress, leading to the development of five different olivine fabrics [see Karato et al., 2008 and references therein]. Only one of these fabrics (B-type) significantly changes the geometrical relationship between φ, direction and flow direction; furthermore, its location is constrained to subduction zone mantle wedges [e.g. Jung and Karato, 2001; Mizukami et al., 2004; Karato et al., 2008].

Olivine LPO may also vary with pressure, [Couvey et al. 2004] but this idea is considered controversial [Long and Silver, 2009]. For all these reasons, most investigators continue to assume that mantle flow and φ are parallel, and it will be assumed for this study, as well.

C. Transition Zone, Lower Mantle, and D” Layer

The presence of anisotropy in the transition zone, the region of the mantle between the upper mantle (above the 410 km seismic discontinuity) and lower mantle (below the 660 km seismic discontinuity), is still debated, as there is little experimental data of transition zone mineral LPO [Karato, 2008]. Conversely, it is widely accepted that the lower mantle does not create SWS because diffusion creep is the dominant deformation regime at these depths, therefore eliminating or severely reducing anisotropy. There is, however, increasing evidence of anisotropy in the D” layer based on the analysis of several shear phases such as ScS and direct S [e.g. Vinnik et al., 1995; Kendall and Silver, 1998; Karato, 1998; Ford et al., 2005; Garnero et al., 2004].
Preliminary findings indicate that the mineral phases perovskite, ferropericlase, and post-perovskite are possible contributors to LPO-induced birefringence if dislocation creep exists at depth [Long and Silver, 2009]. SPO models have also been proposed, but they rely on the existence of sharply-contrasting elastic properties between deep mantle inclusions (e.g. subducted slabs, infiltrated iron from the core, and partial melts) and the surrounding matrix [Kendall and Silver, 1998; Moore et al., 2004; Hall et al., 2004].

III. Anisotropy beneath stable cratons

Poor vertical resolution has made it difficult to determine whether vertically coherent deformation (VCD) in the subcrustal lithosphere or simple asthenospheric flow (SAF) in the asthenosphere is the dominant contributor to continental anisotropy in the upper mantle [Silver, 1996; Fouch and Rondenay, 2006; Silver and Chan, 1991] (Figure 4). Based on global evidence of splitting patterns related to surficial tectonic structures, Silver and Chan [1991] proposed that the lithosphere undergoes VCD, a process that deforms the crust and mantle lithosphere coherently. In regions of multiple tectonic events, olivine fabrics from the last major episode of deformation are preserved. This “fossil” anisotropy is generated when LPO created from the heating and compression associated with mountain building is frozen as mantle temperatures fall below ~900 °C [Silver and Chan, 1991; Silver and Chan, 1988]. Low-temperature creep is the dominant stress regime in the mantle below this critical temperature, causing dislocations in olivine crystals to move along the c-axis (intermediate seismic axis), and not producing LPO, thus preserving anisotropy over geologic time scales [Hirth, 2002; Raterron et al., 2004; Savage, 1999]. This not only suggests that the VCD contribution can be determined if
significant changes in $\phi$ over small distances nearly match those predicted from the
surface geology, but that spatial variations in splitting measurements coincide with, and
thus reveal, past deformation [Silver, 1996]. Some regions display $\phi$ directions that do
not vary significantly over regional scales. Instead they have $\phi$ directions aligned with the
local absolute plate motion (APM) direction, resulting from the viscous drag of the
overlapping lithosphere [Silver, 1996; Vinnik et al., 1992]. LPO produced in the
asthenosphere is short lived, at its oldest a few million years, therefore revealing modern
mantle deformation [Silver, 1996 and references therein].

Extensive studies of westernmost North America demonstrate that mantle
anisotropy is complex and seldom corresponds with recent or past surface geology
[Schutt and Humphreys, 2003 and references therein] (Figure 5). The SAF model appears
to be the source of some splits in the western US, as argued by Silver and Holt [2002],
but this assertion does not explain the circular splitting pattern present in the Basin and
Range [Long and Silver, 2009]. Recent attempts to address the cause of this fast direction
pattern include the ascending plume model of Savage and Sheehan [2000], the toroidal
flow model of Zandt and Humphreys [2008], involving localized flow around the
southern edge of the Juan de Fuca slab, lithospheric delamination (downwelling
asthenosphere) model of West et al. [2009], and the interaction of shallow upper mantle
flow created by plate motion shear and deeper upward flow from the East Pacific Rise
[Yuan and Romanowicz, 2010].
DATA and METHODS

I. Splitlab

SWS analysis was performed using the matlab code SplitLab, developed by Wüstefeld et al. [2008]. SplitLab uses a controlled manual approach during repetitive data processing, therefore, allowing the user to concentrate on quality control and ultimately interpretation of the results [Wüstefeld et al., 2008]. To begin data analysis, users create a SplitLab Project by first entering station information into the provided graphic user interface. Then an earthquake catalog (Global CMT or NEIC) is selected, allowing SplitLab to search for earthquakes that meet our criteria (i.e. a time window containing the period, distance, magnitude, and depth of an event). After the software finds and links downloaded seismograms to its respective event in the database, seismic phases are analyzed and visually inspected within the Seismogram Viewer. This graphical user interface provides various functions (e.g. filters, zoom, rotations) to facilitate phase and time window selection. Some of these functions are summarized below.

It is important to note that seismograms must be rotated into the three-dimensional ray system before splitting measurements can be calculated. The Seismogram Viewer specifically rotates data into the right-handed LQT system for the incidence angle of a chosen phase [Wüstefeld et al., 2008]. The positive (longitudinal) L-component points along the raypath toward the station, the (radial) Q-component is orthogonal to the L-component and points toward the event, and the (transverse) T-component is perpendicular to the ray plane [Wüstefeld et al., 2008]. To remove background noise, a third-order Butterworth filter is applied to the seismic data (filter
range is user specified). Measurement quality can be inferred from particle-motion analysis (e.g. elliptical particle motion suggests a “good” split) after a time window is selected around the phase of interest. Splitlab estimates the delay time and fast direction by removing the effects of splitting from the transverse component of the seismogram using three different methods simultaneously: the rotation-correlation (RC), minimum energy (SC), and eigenvalue (EV) methods. Each technique performs a grid-search, which finds the splitting parameters that best remove (correct for) the effect of splitting (i.e. linearize particle motion in the Q-T plane) [Wüstefeld et al., 2008].

II. Shear-Wave Splitting Techniques

The RC method is based on the assumption of a single, horizontal layer (simple case) of anisotropy and searches for the best linearization by rotating and time-shifting the Q and T components of a seismogram [e.g., Fukako, 1984; Bowman and Ando, 1987]. More simply, this technique finds the maximum cross-correlation between the corrected Q and T components, but it can only be utilized if the initial polarization direction is known [Fukako, 1984]. The popular SC method was introduced by Silver and Chan [1991]. Similar to the RC method, it assumes a simple case of anisotropy and rotates and time-shifts the Q and T components before identifying the minimum displacement energy on the transverse component. Finally, the EV method [e.g. Silver and Chan, 1991] is a slight variant of the SC method, minimizing the eigenvalue, instead of the transverse component energy, of the corrected covariance matrix [Silver and Long, 2009]. The former two methods are only applicable to phases with known initial polarization directions (e.g. core phases), but the EV method may be used when the initial
polarization direction of a wave is unknown (e.g., direct S, ScS, etc.) [Wüstefeld and Bokelmann, 2007].

Using three independent techniques allow us to compare results for consistency and to assess the quality of individual measurements. Agreement between techniques increases confidence in our results; however, discrepancies between these methods may diagnose complex anisotropy beneath the stations, random noise, or a wave nearly polarized in the null direction (nulls will be discussed at length in Section III) [e.g. Long and Silver, 2009; Wüstefeld and Bokelmann, 2007; Vecsey et al., 2008]. As Menke and Levin (2003) demonstrated, waveforms are complicated by complex anisotropy and will not agree with predictions of the simple anisotropy assumed by the RC and SC methods. Furthermore, the SC method appears to be more sensitive to complex anisotropy than the RC and EV techniques [Long and van der Hilst, 2005b]. We may, therefore, argue complex anisotropy exists at depth if different measuring methods produce inconsistent, albeit well constrained, results [Levin et al., 2004].

III. Present Study

Seismic data employed for this study were acquired from the LaBarge Passive Source Seismic Experiment (LPSSE). The project consisted of 63 broadband seismometers deployed over a seven-month period (October 2008 to June 2009), 55 of which were arranged in a dense array (L-stations) with 250 m station spacing (Figure 6A). The remaining eight instruments (A-stations) traversed the Green River Basin in a northeast-southwest line (Figure 6B). The regional study subsequently incorporated an additional 58 instruments scattered throughout southwest Wyoming and far eastern
Idaho, shown in Figure 7. Data from the additional stations were obtained from several sources archived at the Incorporated Research Institutions for Seismology (IRIS) Data Management Center (DMC) and include the EarthScope Transportable Arrays, CD-ROM [Fox and Sheehan, 2005], Intermountain West, the U.S. National Network, the Big Horn Regional Array Deployment, the Yellowstone Intermountain Seismic Array, and Deep Probe.

To ensure a clear $XKS$ ($PKS$ and $SK(K)S$) arrival, teleseismic events were chosen for analysis only if they met the earthquake minimum cutoff magnitude of 6.0 and $SK(K)S$ were between distances of 80°-140° from each station (to avoid contamination by other phases, such as $S$ and $ScS$, and ensure a strong arrival). Most events within this distance range come from a northwest backazimuth (Figure 8). All the events used in this study were band-pass filtered with corner frequencies of 0.03 and 0.4 Hz to eliminate low-frequency and high-frequency noise and a time window of 20-40 seconds was selected around phases, depending on the presence of other arrivals (Figure 9). Splitting measurements were rated using a data-base criterion similar to Wüstefeld and Bokelmann [2007]. An XKS pulse with similar waveforms on the radial and transverse seismogram components and a corrected linear particle motion are ranked as “good” [Wüstefeld and Bokelmann, 2007]. Measurements were defined as “good” quality splits if the difference in fast axis estimates ($\Delta \phi$) between the RC and SC or RC and EV methods were $\leq 8^\circ$. Figure 9A provides an example of a good splitting measurement. “Fair” quality splits required $\Delta \phi \leq 15^\circ$ between the RC and SC or EV method while splits with $\Delta \phi > 15^\circ$ were
categorized as “poor”. A visual inspection of the ranking was also conducted to verify and adjust measurement quality if necessary.

Measurements which had an initial linear particle motion and little to no energy on the tangential component were termed “nulls” [e.g. Wüstefeld and Bokelmann, 2007; Long and Silver, 2009; Savage, 1999] (Figure 9B). Nulls are produced either from an S-wave propagating through a region of isotropy or else the initial polarization parallels the fast or slow directions of the anisotropic medium [Long and Silver, 2009]. Although nulls do not constrain $\delta t$, and $\phi$ reflects either the fast or slow axis, they are potentially useful for SWS analysis [Wüstefeld and Bokelmann, 2007]. Nulls can reflect both the geometry and strength of anisotropy because $\phi$ corresponds to the initial polarization in the presence of isotropy, which is often the backazimuth of XKS waves [Wüstefeld and Bokelmann, 2007]. Like splitting measurements, nulls were rated according to criteria created by Wüstefeld and Bokelmann [2007]. Good nulls required a delay time ratio ($\rho = \frac{\delta t_{RC}}{\delta t_{SC} \text{ or } EV}$) of $0 < \rho < 0.2$ and $\Delta \phi$ around $45^\circ$ ($37^\circ \leq \Delta \phi \leq 53^\circ$) while near-nulls needed $0 < \rho < 0.3$ and $32^\circ \leq \Delta \phi \leq 58^\circ$. Measurements that did not meet these criteria were deemed poor quality.

RESULTS

I. Simple Waveform Analysis

A total of 175 moderately to well constrained (good and fair) splits and 164 well defined nulls were measured. The majority of splits are SKS (121) followed by 44 SKKS splits (almost exclusively recorded at the LaBarge array) and 10 PKS measurements. A minimum of one split was found at 78 stations. To better understand observed anisotropy
at different length scales, the results below are analyzed over short distances (LaBarge stations) and broader, more regional distances (non-LaBarge stations).

A. LaBarge Passive Source Seismic Array

The LaBarge Array yielded 44 good and fair splitting measurements and 44 nulls (Figure 10; Appendix I). Strong anisotropy is observed, as shown by an average $\delta t$ of $\sim 1$ second. The estimated $\phi$ direction aligns with the absolute plate motion (APM) of North America ($249^\circ \pm 10.7^\circ$) [APM in HS3-NUVEL 1A; Gripp and Gordon, 2002] and agrees with previous SWS studies of the Yellowstone wake (i.e. the eSRP) by Schutt et al. [1998], Schutt and Humphreys [2001], and Waite et al. [2005] (Figure 5). These authors concluded that the North American plate warms as it passes over the Yellowstone hotspot, allowing it to reorient the asthenosphere via plate-induced shearing.

Only one SKKS event (2009-042) exhibited shear-wave splitting at LaBarge, making it difficult to rule out complex models of anisotropy such as dipping fast axes or multiple layers. The signal-to-noise-ratio (SNR) of this event is moderate to low for most stations. Nonetheless, the $\phi$ direction is remarkably coherent across the array, even among poor results, although this is to be expected, because station spacing is dense. There is, however, a notable discrepancy between splits derived from the SC method and those measured by the RC and EV techniques. This and other inconsistencies were more easily identified by dividing the dense array into three segments: arm 1 (stations L01-L22), arm 2 (stations L23-L41), and arm 3 (stations L42-L55).

For stations along arms 1 and 2, splitting parameters calculated by the RC and EV methods largely agree with one another (Figures 11 and 12). The $\delta t$ values along arms 1
and 2 gently increase moving east, but \( \phi \) stays approximately constant across strike. Individual splits, as well as mean \( \text{RC}_\phi \) (73.8°) and mean \( \text{EV}_\phi \) (74.1°), are parallel to the APM direction of North America. Although \( \text{RC}_\phi \) is roughly constant west to east, \( \text{EV}_\phi \) decreases from \( \sim 89° \) to less than 60° at station L20 before gradually increasing to \( \sim 75° \) at the most eastern stations in arm 2. Individual and average \( \text{SC}_\phi \) (57.6°) tends to be smaller than directions calculated by the two other techniques, instead aligning better with the Yellowstone hotspot track [241.0° ± 23.8°; Gripp and Gordon, 2002]. \( \text{SC}_\phi \) estimates increase by 20°, from \( \sim 45° \) to \( \sim 65° \), moving east along arms 1 and 2. \( \text{SC}_{\delta t} \) is less consistent than results produced by the other methods, with values \( \sim 0.2-0.6 \) seconds greater than the highly coherent \( \text{RC}_\phi \) and \( \text{EV}_\phi \). It is, however, intriguing that \( \text{SC}_{\delta t} \) scatter abruptly disappears at station L27. This is also where \( \text{RC}_{\delta t} \) and \( \text{EV}_{\delta t} \) begin converging with \( \text{SC}_{\delta t} \). Measurements made at arm 3 show the same \( \phi \) and \( \delta t \) patterns as arm 2 (Figures 11 and 12 respectively). It is important to note, however, that SC measurements are within error of the other two techniques even though SC results are considerably more scattered.

Measurements made at the A-stations were generally noisier than the dense-array results (Figure 13). Three of the eight A-stations (A02, A07, and A08) yielded a usable split, the same SKKS event observed at the L-stations (Appendix I). Splitting parameters correspond with measurements taken at the most eastern dense-array stations; although, \( \text{SC}_{\delta t} \) and \( \text{RC}_{\delta t} \) at stations A02 and A08 have greater coherency than \( \text{EV}_{\delta t} \). Unfortunately, this event did not produce a useful phase at any of the regional stations.
B. Regional Individual Splitting

At the regional scale, splitting measurements lack the coherency observed at the LaBarge array (Figures 14 and 15A; Appendix II). Our results from stations within or near the eSRP (west of the Wyoming Range) have a mostly uniform NE-SW trend of φ direction (parallel to APM direction) and therefore agree well with previous splits reported by Shcutt et al. [1998] and Waite et al. [2005] (Figures 14 and 15A, Appendix II). A notable deviation from the APM direction is at station L17A, where a PKS arrival exhibits φ_{ALL} (average of the three SWS methods) at an extremely high angle to the APM direction (φ_{ALL} = 8.0°). Upon closer inspection, the PKS split at station L17A is only moderately constrained (fair split) and is the only usable split yielded by this station. Stations near the eSRP report a strong null direction (~60°) subparallel to the APM direction when null events are plotted against modulo-180° (the backazimuth subtracted by 180°) (Figures 15B). A second group of nulls have backazimuths (modulo-180°) that are subparallel to the inferred slow direction (approximately normal to φ). This consistency between null backazimuths and the fast or slow splitting direction is evidence of simple anisotropy and is further supported by the absence of backazimuthally dependent splitting parameters shown in Figure 16.

Station results between the Wind River Range and Sevier fold and thrust belt (Wyoming Range), including measurements made at LPSSE, indicate that the φ and null direction patterns observed near the eSRP continues to the east until about longitude -109.5°W (approximate position of the Wind River thrust). Although this suggests simple anisotropy, stations in this area report many nulls and few split waveforms, most of
which are noisy. TA station L18A, for instance, yielded only one split, but eight nulls. Similarly, TA stations J18A (three splits, six null measurements), K17A (two splits, seven nulls), and L19A (six nulls and no splits) yielded low split to null ratios. One split, one null, and several events that could not be classified as a null or split were observed at LPSEE, which is centered between the aforementioned TA stations. Immediately west of longitude -109.5°W (Wind River Range) stations PD31 and BW06 measured five nulls and seven nulls respectively. Waite et al. [2005] also reported only finding null results at station BW06 during their deployment between June 2000 and June 2001. Furthermore, stations PD31 and BW06 are located between the A-stations and Y22, all of which reported no usable results (or only one well constrained split in the case of stations A02, A07, and A08). Additional stations within the basin that yielded no results were installed south of 42° N, with the exception of TA station L19A, which measured six nulls. Some of these southern stations were only deployed for a few months, but half of them are TA stations and were deployed for an appropriate time needed to conduct a SWS investigation. Furthermore, some φ measurements vary drastically from APM direction, specifically at TA stations L18A (SKKS split, mean $\varphi_{ALL} = 86.4^\circ$), J18A (SKS split, mean $\varphi_{ALL} = 106.2^\circ$), and M19A (PKS split, mean $\varphi_{ALL} = 9.0^\circ$). Nonetheless, most station results within the Green River Basin indicate φ parallels the APM direction; therefore, measurements in the basin and those observed at stations closer to the eSRP are grouped together.

Instead of clustering around the APM direction, splits observed east of the Wind River thrust (roughly longitude -109.5°W) have a significantly greater φ range.
Measurements generally strike E-W to NW-SE, which agree with the most northern CD-ROM φ estimates reported by Fox and Sheehan [2005], and nulls come from directions that align or are approximately perpendicular to the wider swath of φ angles. There is also a greater array of backazimuths sampled at eastern stations, but the backazimuthal coverage is not robust enough at individual stations to identify complicated anisotropy. Nonetheless, Figure 17 illustrates that plotting all measured φ directions against modulo-180° (backazimuth - 180°) reveal a possible pattern of π/2 periodicity, which is evidence of either two or more anisotropic layers or a single layer of smoothly varying media [Silver and Savage, 1994; Özalaybey and Savage, 1994; Rümpker and Silver, 1998]. Confidently arguing either case is difficult, however, because some backazimuths are poorly represented, primarily between ~139° and ~179°, and δt does not exhibit π/2 periodicity as there is minimal variation between backazimuths. However, this evidence of complicated anisotropy, as well as consistently shallow φ angles, imply the existence of at least two major splitting domains: a NE-SW splitting domain (stations results west of the Wind River Range) and an E-W splitting domain (station splits east of the Wind River Range).

C. Regional Station-Average Splitting

Obtaining the average splitting parameter of individual stations is a useful approach to utilizing temporary stations with few well-constrained results [Waite et al., 2005]. However, averaging nearly vertically travelling waves is only appropriate for waves with small angles of incidence (<15°), like XKS arrivals [Waite et al., 2005]. Secondly, this approach must only be used in areas of simple anisotropy since dipping
fast axes and multiple layers of anisotropy result in backazimuthally-dependent splitting parameters [e.g. Schutt et al., 1998; Waite et al., 2005]. This can result in unreliable and unrealistic station-average splits. Station-average splitting results for all stations used in our study are shown in Figure 18 and are also plotted against station longitude to simplify interpretations in Figure 19. A list of the station-average splitting parameters can be found in Appendix III.

The estimated δt_{average} for most stations in the array fall between 0.5 and 1.5 seconds. A sizable number of stations with large δt_{average} (≥1.2 seconds) are observed just south of Yellowstone (DCID1, FXWY, I16A, I18A, IMW, MOOW). Waite et al. [2005] similarly reported a station-average δt of over 2 seconds within the Yellowstone caldera. Station-average φ generally rotates and increases eastward over the array. Like average splits reported by Schutt et al. [1998] and Waite et al. [2005], the φ_{average} of stations proximal to the Yellowstone wake cluster around the APM direction and the trend of the Yellowstone hotspot track. Station averages located within or near the Wind River Basin (e.g. K19A, K20A, J21A, S101, Y35) tend to deviate from this trend, instead having E-W and NW-SE orientations. However, our individual splitting measurements indicated possible complications due to backazimuthal dependence of φ measurements, which make station averages east of the Wind River Range less reliable than those obtained further west. Stations L20A, N00, N01, and N02 are located over a NW dipping high-velocity anomaly found by Dueker et al. [2001]. The anomaly, interpreted as an ancient slab [Yuam and Dueker, 2005], is reported to have plunging anisotropy [Fox and Sheehan, 2005], making splitting averages from these four instruments less reliable as
well. To first order, it can still be concluded that our station-average results, like the individual measurements, demonstrate two separate splitting domains.

II. Frequency Analysis

The sensitivity of a teleseismic split to anisotropic structure is highly dependent on frequency [e.g. Favier and Chevrot, 2003; Long et al., 2008]. Analyzing splitting results for frequency dependence can better improve constraints on the possible source depth and thickness of anisotropy [Long, 2010; Wirth and Long, 2010]. Only the highest quality splits ($\text{SNR}_{\text{RC}} \geq 14.5$ with a clear, strong XKS pulse) were used to characterize frequency-dependent splitting. Selected measurements were subjected to three discrete frequency bands: the low-frequency band (L-band) at 10 to ~33 seconds, the intermediate-frequency band (I-band) at ~3 to 10 seconds, and the high-frequency band (H-band) at 1 to ~3 seconds. Discrepancies in splitting results between bands indicate frequency-dependent splits, which occur when the surface errors of splits measured at different frequencies do not overlap (Figure 20). Long [2010], however, has cautioned that it is difficult to identify frequency-dependent SWS when only using XKS arrivals because XKS energy is often dominated by periods greater than 8 seconds. For this reason, local S phases have been used in many studies to characterize frequency-dependent SWS, as they tend to contain energy at higher frequencies [e.g. Wirth and Long, 2010; Marson-Pidgeon and Savage, 1997; Fouch and Fisher, 1998].

Well-defined results were nevertheless yielded from eight XKS events, six of which exhibited energy in either the L- and I-bands or H- and I-bands. Stations L39 and K19A are the only instruments that produced results in all three frequency bands. Results
of this analysis are shown in Appendix IV and an example of a consistent and discrepant split are shown in Figures 20A and 20B, respectively. Frequency dependence is observed at all seven station-event pairs that displayed energy in both the I- and L-bands. The I-band produced $\delta t \sim 0.7$ seconds larger than those measured in the L-band. Only two splitting discrepancies were found out of the five station-event pairs that recorded well-constrained results in both the H- and I- bands. Each discrepancy revealed $\delta t$ estimates larger in the I-band than in the H-band at its respective station. However, $\phi$ was not affected by frequency content for the seven station-event pairs exhibiting frequency dependence. Figure 21 shows that significant splitting energy is present in the I-band; in fact, every selected event, across both splitting domains, exhibited energy at intermediate frequencies. The L-band, in contrast, resolved good or fair splits exclusively in the NE-SW splitting domain. Splitting energy in this bandwidth was only observed at stations I16A and IMW, which are immediately adjacent to the eSRP and Yellowstone mantle plume. In contrast, all stations displaying splitting energy at high frequencies (stations J21A, K19A, and K20A) were located in the E-W splitting domain. Most null measurements were yielded in the L-band for both splitting domains (83% of NE-SW domain nulls and 67% of E-W domain nulls), while 17% of NE-SW domain nulls and 33% of E-W domain nulls were observed in the H-band and none were found in the I-band for either splitting domain.

In summary, XKS arrivals in our study area are frequency-dependent. More specifically, $\delta t$ values, but not $\phi$ results, are affected by frequency content. Low-frequency splitting is only observed in the NE-SW splitting domain. The E-W splitting
domain, in contrast, exclusively yields high-frequency splitting. The majority of splitting energy occurs at intermediate frequencies, but most nulls are found at low frequencies.

**III. Ray-Piercing-Point Analysis**

As previously discussed, XKS raypaths have steep angles of incidence, which provide excellent lateral resolution but poor vertical resolution. This implies that anisotropy can occur anywhere on the receiver side of the raypath. A new and intriguing technique by Gao et al. [2010] may improve depth resolution and, by extension, constraints on the depth of anisotropy. Based on the work of Alsina and Sneider [1995], Gao et al. [2010] treated XKS waves as tubes centered near the geometric raypath. These tubes are influenced by the material property of the medium and their diameter is dependent upon the first Fresnel zone, which is most sensitive to anisotropic structure [Alsina and Sneider, 1995]. Sample area (tube diameter) varies with period, traveltime (epicentral distance and focal depth), and backazimuth because of the nonvertical geometry of most XKS raypaths [Alsina and Sneider, 1995; Gao et al., 2010]. Knowing the dimensions of the Fresnel zone improved their estimation of source depth. Their method demonstrates that measurements placed at ray-piercing points corresponding to the “true” depth of anisotropy will have high spatial coherency because two or more events sample the same subsurface area [Gao et al., 2010] (Figure 22). Obviously, this suggests that splitting estimates placed at incorrect depths have low spatial coherency. Thus, Gao et al. [2010] formulated an equation to estimate source depth:

\[
F_r = \frac{1}{N} \sum_{i=1}^{N} \left( \frac{1}{M_i - 1} \left( w_\phi \sqrt{\sum_{j=1}^{M_i} (\phi_{ij} - \bar{\phi}_i)^2} + w_\alpha \sqrt{\sum_{j=1}^{M_i} (\alpha_{ij} - \bar{\alpha}_i)^2} \right) \right)
\]
where $N$ is the number of blocks, $M_i$ is the number of measurements for the $i^{th}$ block, $\phi_{ij}$ and $\delta t_{ij}$ are the $j^{th}$ fast direction and delay time estimate in the $i^{th}$ block, $\bar{\phi}_i$ and $\bar{\delta t}_i$ are the averages of all measurements in block $i$, and $w_\phi$ and $w_\delta t$ are the weighing factor for their respective splitting parameter. Because the maximum possible value of $\phi$ in the NE-SW splitting domain is 180° and maximum observed $\delta t$ is ~1.8 seconds, we use $w_\phi = 1/180^\circ$ and $w_\delta t \approx 1/1.8$ seconds (~0.56 seconds). Likewise, because the maximum possible value of $\phi$ in the E-W splitting domain is 180° and maximum observed $\delta t$ is ~1.4 seconds, we use $w_\phi = 1/180^\circ$ and $w_\delta t \approx 1/1.4$ seconds (~0.71 seconds). Using these weighing factors allows us to combine the misfits of the splitting parameters. The resulting spatial-variation factor of the measurements, $F_v$, is dimensionless and varies with depth.

The frequency analysis and simple waveform analysis results suggest a sharp transition in $\phi$ direction from NE-SW to E-W across the Wind River Range, which is also reflected by ray-piercing point plots down to 400 km, approximately the 410 km seismic discontinuity (i.e. the maximum depth of dislocation creep in the upper mantle). Therefore, the variation factor, and by extension depth to anisotropy, was determined based on dividing the data into the two aforementioned splitting domains (Figures 23 and 24). The E-W splitting domain is roughly located within the Wind River Basin, bounded to the west by the Wind River thrust fault, which may be a reactivated structure related to subduction during the late Archean [Chamberlain et al., 2003], and bounded to the south by the Oregon Trail structure - Geochronologic Front (OTS-GF) [Chamberlain et al., 2003], another late Archean structure, which additionally serves as a border to prevent the influence of dipping anisotropy from the Cheyenne slab on variation-factor
calculations. The northern border of the E-W splitting domain similarly ensures that
temperature results are not affected by the Yellowstone hotspot plume. The area to the
west and north of the E-W splitting domain makes up the NE-SW splitting domain.
Splitting parameters within the box located at the bottom right-hand corner of Figures 23
and 24 are not used for the variation factor calculations due to evidence of dipping
anisotropy, which leads to piercing-point location dependence of the splitting
measurements.

The minimum splitting variation determined for the E-W splitting domain clearly
indicates that the main source of anisotropy is centered within the lithosphere at a depth
of 100-130 km (Figure 23A). At this depth, mean φ ≈ 88°, consistent with the E-W
orientation observed in the simple waveform and frequency analyses, and mean δt ≈ 1
second (Figure 23B). However, the corresponding depth to minimum variation factor for
the NE-SW splitting domain is less constrained, estimating a central source of anisotropy
between 110 and 210 km depth, with a mean φ of ~ 67°, nearly parallel to the North
American APM, and like the E-W splitting domain, a mean δt of ~ 1 second (Figure 24).

DISCUSSION

I. LaBarge Waveform Statistical Analysis

We find it unusual that only one usable split was found during the six month
deployment of the LaBarge Seismic Array. Possible reasons for the low number of results
include: stations were deployed during a period of unfavorable events for splitting
analysis, random noise occurred in the waveforms, or the subsurface geology
complicated the XKS signal.
As previously discussed, the SC method produces the smallest $\varphi$ (parallel to the Yellowstone hotspot track) but largest $\delta t$ of the three SWS techniques, albeit within error of the other two techniques (Figure 11). EV and RC splitting parameters are nearly identical and their $\varphi$ estimates align with the APM direction. These large differences between splitting techniques may reveal the robustness of each method. For example, the standard deviation (S.D.) of measurements indicates the degree of coherency achieved by a splitting technique. The average and S.D. of each SWS method is listed in Appendix V.

The S.D. values for $\delta t$ are virtually identical between the three splitting techniques, thus measurement coherency is better indicated by $\varphi$ direction. Overall, the RC method is the most reliable method utilized for our analysis because it has the lowest S.D. of the three SWS techniques (4.5°); therefore, it has the highest $\varphi$ coherency between individual stations. SC$_\varphi$, conversely, has the largest S.D. (11.9°), thus the least robust measurements. However, the disagreement between splitting parameters estimated via the SC method and measurements made with the EV and RC techniques seems to demonstrate that complex geology is located beneath LPSSE. Subsurface complexities and noise from the nearby oil fields may also contribute to the low number of splitting results measured at the dense array and A-stations. Nearby stations also reported few or no split waveforms, but many null measurements, including BW06, PD31, Y22, and the more distant TA stations K17A, K18A, L18A, and L19A.

II. Evidence of transition from Simple to Complicated Anisotropy

The remarkably uniform nature of $\varphi$ direction measured adjacent to and within the eSRP has been interpreted as relatively shallow asthenospheric flow [e.g. Schutt et al.
Splits measured by stations near the eSRP for our study generally agree with this interpretation. Both the consistency of φ directions and the approximate alignment of null backazimuths with the fast and slow seismic directions also suggest a simple flow model involving only a single layer of anisotropy. Most stations in the Green River Basin, including LPSSE, have splits that align with the APM direction (NE-SW orientation), and therefore also suggest that observed splitting is largely produced by mantle flow in the asthenosphere, albeit with some pronounced complexities as indicated by the low ratio of splits to nulls (Figures 15). In other words, we observe simple anisotropy as documented at stations within or proximal to the eSRP, but we find greater complexity moving east across the region. Complexities are greatest at stations east of the Wind River Range. Fox and Sheehan [2005], for example, report that dipping anisotropy exists south of the Wind River Basin and is produced by a dipping feature interpreted as the Cheyenne slab by Dueker et al. [2001]. Waite et al. [2005] similarly report that splits in the Wind River and Bighorn Basins could be caused by a NW-dipping high-velocity anomaly related to the Cheyenne slab. However, Yuan and Dueker [2005] and Waite et al. [2006] argue that the anomaly represents the down-welling of cold, lithospheric mantle to accommodate hotter, new material that has risen up through the plume. Still, large-scale mantle-convection models by Smith et al. [2009] indicate that the high-velocity feature may instead be a remnant of the tectosphere, thus consistent with the interpretation of Waite et al. [2005]. Our findings in the Wind River Basin and preliminary results from Erslev et al. [2009] in the Bighorn Basin show that φ is generally E-W and does not align in the dip direction of
the high-velocity structure, implying that observed splitting does not result from a plunging axis of symmetry.

**III. Frequency-Dependent Splitting**

Frequency-dependent splitting has been documented at multiple sites worldwide including Japan [Long and van der Hilst, 2006; Wirth and Long, 2010; Huang et al., 2011], the Mariana subduction zone [Fouch and Fischer, 1998], New Zealand [Marson-Pidgeon and Savage, 1997], and the Gulf of California [Long, 2010]. According to ray theory, the diameter of the first Fresnel zone, the region most sensitive to anisotropic structure, increases with decreasing frequency (i.e. increasing depth) [e.g. Alsina and Sneider, 1995; Favier and Chevrot, 2003]. Therefore, higher frequencies are sensitive to small-scale heterogeneities while lower frequencies generally smooth out these features [Wirth and Long, 2010 Saltzer et al., 2000]. Rümpker et al. [1999] demonstrated that SWS estimates are sensitive to filtering and that frequency-dependent splitting is evidence of depth-dependent anisotropy. Long et al. [2010] similarly reports that vertically and laterally changing anisotropy has the potential to produce frequency-dependent splitting parameters as a consequence of finite-frequency effects on teleseismic waves. Waveform modeling studies reveal that in the presence of vertically varying anisotropy, splitting measurements are biased toward shallow fabrics at higher frequencies [e.g. Rümpker and Silver, 1998; Saltzer et al., 2000]. In other words, near-surface features located in the lithosphere and crust (at the highest frequencies) are more likely to affect short-period energy while longer period waves reflect deeper structures in the asthenosphere [Wirth and Long, 2010; Long, 2010]. If anisotropic heterogeneities are
strong enough, multiple scattering will reduce energy on the transverse-component
seismogram, thus decreasing $\delta t$, which can result in null measurements despite the
presence of significant anisotropy [Saltzer et al., 2000]. For these reasons, interpretations
of observed anisotropy should be made with frequency dependence in mind.

From previous studies (cited above), we conclude that the L-band detects
asthenospheric anisotropy, the H-band lithospheric anisotropy, and the I-band may reflect
either of these layers depending on the depth of the lithosphere-asthenosphere boundary
(LAB). Lithospheric thickness was obtained from a receiver-function study of the
western United States, which provides the LAB contours shown in Figures 23 and 24
[Miller and Levander, 2011]. Our own results indicate that frequency-dependent splitting
occurs beneath Wyoming, which supports our findings that more complex anisotropy
generally exists east of the eSRP (Figure 21). In particular, $\delta t$ is affected by frequency
content while $\varphi$ is independent of frequency (Appendix IV), thus suggesting that the
intensity of birefringence may vary laterally, vertically, or both.

Low-frequency splitting energy detected at stations I16A and IMW (near
Yellowstone) is therefore most likely produced by asthenospheric flow. This is consistent
with earlier investigations [e.g. Schutt et al., 1998; Waite et al., 2005], which argued that
splitting observed within or adjacent to the Yellowstone wake results from relatively
shallow mantle flow in the asthenosphere. In contrast, the high-frequency splitting
observed exclusively at stations J21A, K19A, and K20A (east of the Wind River Range)
is likely from a lithospheric source, which is further supported by the presence of a
deeper LAB (>130 km) roughly east of the Wind River Range. The lack of discrepancies
between H- and I-band results in the E-W splitting domain implies that anisotropy is coherent with depths in the lithosphere, which points to a single anisotropic layer. In addition, the large δt values yielded by the H- and I-bands, as opposed to the substantial null energy in the L-band, implies that the main source of SWS in the E-W domain is a single layer of anisotropy that exists within the lithosphere and extends down to the shallow asthenosphere depending on LAB depth. The deep asthenosphere, however, appears to be weakly anisotropic west of the eSRP, as supported by the high percentage of null energy in the L-band. This contrast in splitting intensity between the lithosphere and deep asthenosphere may explain the π/2 periodicity characteristic of vertically varying anisotropy at stations in the E-W splitting domain, as well as the low yield of splitting results measured at stations within the Green River Basin.

IV. Source of Anisotropy beneath Wyoming

The trade-off between the source thickness of anisotropy and the intensity of birefringence can be used to approximate the depth of anisotropy and is calculated with the formula

\[ h = \frac{\delta t \times v_s}{\gamma} \]

where \( h \) is the thickness of the anisotropic source, \( v_s \) is the average upper mantle shear-wave velocity, and \( \gamma \) is the strength of anisotropy (%). Assuming an average upper mantle \( v_s \) of 4.5 km/s and \( \gamma \) of 4%, we can determine the percent contribution of the crust, lithosphere, and asthenosphere to observed splitting by centering the calculated anisotropic thickness \( h \) (~113 km) on the minimum splitting variation depth (from our
ray-piercing-point analysis) and then, using crustal and LAB thickness values, determine the amount of anisotropy contained within each of the uppermost layers of the Earth. For instance, because crustal thickness is no greater than 55 km within our study area [Miller and Levander, 2011], anisotropy is of sufficient thickness to penetrate the crust at only five stations (stations K20A, K21A, Y32, Y35, and Y46), all of which are located in the E-W splitting domain where the optimal depth to anisotropy is estimated to be at an average depth of 115 km. Coincidentally, all stations within this domain show that ~63-87% (mean of ~70% and S.D. of 7.19%) of the anisotropic source is within the lithosphere. This agrees well with the high-frequency (shallow) splitting results measured at stations J21A, K19A, and K20A in our frequency analysis. Because φ does not align with Laramide structures, any mantle deformation produced by the Laramide Orogeny is probably limited. Thus, splitting parameters from the E-W splitting domain are most likely the result of late Archean tectonic fabric related to several temporally related episodes of crustal growth and shortening along the Sweetwater subprovince [Frost et al., 2000; Chamberlain et al., 2003; Frost et al., 2006] (Figure 25). However, φ measurements made at stations near the Tetons and northern Green River Basin do not parallel the lithospheric fabric observed east of the Wind Rivers.

Instead, splitting parameters from the Green River Basin and near the eSRP align with the North American APM and Yellowstone hotspot track, which suggests a deeper source of anisotropy. However, the spatial variation calculations are not as robust as estimates made in the E-W splitting domain (Figure 23). Although anisotropy may be as deep as 210 km, the variation factor plot shows that the optimal depth to anisotropy in the
NE-SW splitting domain may be as shallow as the E-W splitting domain. The simple waveform and frequency dependence analyses, nevertheless, indicate that anisotropy is more likely to be deep west of the Wind River Mountains. We therefore argue that anisotropy measured in the NE-SW splitting domain is mostly asthenospheric and reflects simple shear produced by mantle flow or drag-induced strain of a stationary mantle. Any fossil fabrics that existed in the lithosphere within or immediately proximal to the eSRP were likely reworked by passage over the Yellowstone hotspot as previously concluded [e.g. Schutt et al., 1998; Waite et al., 2005], which explains why φ directions suddenly rotate from E-W to NE-SW across the Wind River Range. Although deformation from the Yellowstone plume overprinted older lithospheric fabric, some complexities are still observed in the Green River Basin as noted by the lack of splits and abundance of nulls at stations within the basin.

A possible explanation for these complexities is that the Green River Basin represents a transition in anisotropic structure and contribution between the asthenospheric dominated SWS of the eSRP and the lithospheric-induced splitting of the Wyoming Craton as suggested by the sharp contrast in φ across the Wind River Range. West of the range, the Archean fabric of the Sweetwater subprovince is probably of sufficient strength and thickness to dominate observed patterns of mantle anisotropy. There is also evidence that the mantle beneath the Wind River Basin is particularly capable of resisting tectonism. As previously discussed, a lack of range-parallel splitting within the basin suggests that Laramide deformation was limited in the underlying mantle. Additionally, Chamberlain et al. [2003] report that Proterozoic extension (ca.
2.1-2.0 Ga) centered in the SAT did not continue north of the Oregon Trail structure (c.a. 2.62 Ga) or the southern extent of the 2.9-2.75 Ga magmatic province, which is the core of the BBMZ (Figure 1), as a result of Archean sutures or the lithospheric contrasts across these structures (based on geophysical profiles by Gorman et al., 2002). West of the Wind River Range, the Archean fabric likely dissipates and becomes subordinate to asthenospheric anisotropy as indicated by the alignment of $\phi$ with the APM direction. The lack of high-frequency-splitting measurements in the Green River Basin also points to a weakening of anisotropic strength in the lithosphere. Nevertheless, the splitting strength of the lithosphere is intense enough to induce strong vertical heterogeneity. In other words, the SWS techniques are attempting to resolve the shallow, weakened lithospheric signature and the deeper mantle flow of the asthenosphere. The combination of differing anisotropic structures within the same region, but at different depths, produces scatter, which reduces the tangential-component seismograms and ultimately results in null-like measurements [Saltzer et al., 2000] such as the ones observed in the Green River Basin.

**CONCLUSIONS**

Each of the three shear-wave-splitting analyses performed at the 121 stations utilized for our study indicate at least two domains of anisotropic structure beneath western Wyoming. SWS measurements observed west of the Wind River Range are highly coherent and align with the absolute plate motion of North America, as well as the Yellowstone hotspot track, suggesting a simple layer of asthenosphere-induced anisotropy caused by SW-directed mantle flow. This is further supported by the
dominance of splitting energy at longer periods (i.e. deeper depth) at western stations. Although the optimal depth to anisotropy possibly coincides with the estimated depth of the E-W splitting domain, the preponderance of our evidence suggests that anisotropy near the eSRP and the Green River Basin is largely within the asthenosphere. Fast directions east of the Wind River Range are more complicated, possibly revealing vertically changing anisotropy as shown by the π/2 periodicity of estimated fast directions, and are predominantly oriented E-W, thus implying a non-asthenospheric source of anisotropy. Additional evidence of this interpretation includes the observation of high-frequency-splitting energy at stations within the Wind River Basin and the shallow mean depth to anisotropy estimated at 115 km. The lack of range-parallel splitting demonstrates that the majority of Laramide tectonism within the region was limited to the crust. Instead, fast directions parallel the late Archean fabric of the Sweetwater subprovince, therefore, providing evidence that the subcrustal lithosphere of the Wyoming Craton was left largely intact after flat-slab subduction of the Farallon plate. Although splitting direction sharply changes across the Wind River Range, delay times remain constant at ~1 second and suggest that anisotropic thickness for both splitting domains is ~113 km. On another note, the few splitting results and notable number of nulls reported at stations in the Green River Basin may be the result of a transition from asthenosphere-induced splitting at the Eastern Snake River Plain to lithosphere-dominated anisotropy within the Sweetwater subprovince. A westward decrease in the anisotropic strength of the lithosphere is our preferred cause of this
transition and may be supported by our observations of delay-time dependence with depth, which is indicative of vertically and/or laterally changing anisotropy.
ACKNOWLEDGEMENTS

The author would first like to thank his advisors Susan Beck and George Zandt for accepting me into their research group (GSAT) and providing me their guidance and support. I also thank Megan L. Anderson of Colorado College for the insightful discussions on frequency and ray-piercing-point analyses, as well as for providing us the matlab code, written by John Spence Hornbuckle III and Jeff Rahl of Washington and Lee University, used to plot the spatial variation factor of our data. Funding for this research was provided by ExxonMobil. Some figures were created with Generic Mapping Tools (GMT) software [Wessel and Smith, 1995].

I would also like to give many thanks to C. Berk Biryol who was immensely helpful in improving my understanding of shear-wave splitting analysis and computer coding. I thank Christine R. Gans for all her help with coding and editing, for being a good friend, and for serving as a role model for what a successful graduate student should be. And finally I wish to thank all members (current and former) of GSAT, all of my mentors at outside institutions, and my family and friends for their love and support.
Figure 1: (A) Map of the Wyoming Craton and the Proterozoic Trans-Hudson and Colorado (Cheyenne Belt) Orogenies modified from Sims et al. [2001]. The yellow dotted line bounds the magmatic core of the Beartooth-Bighorn Magmatic Zone (BBMZ), and the black dotted line marks the Oregon Trail structure (OTS_GF), a late Archean suture [Chamberlain et al., 2003]. Irregular shapes outlined in black represent exposures of Precambrian basement rocks. (B) Tectonic map of Laramide structures and trends after Brewer et al. [1980]. Laramide arches are based on Neely and Erslev [2009].
Figure 2: (A) Modified cartoon illustrating the anisotropic character of olivine [Blackman, 2007]. (B) Schematic cut-view of the Earth showing the raypath of $SK(K)S$ waves and the two parameters ($\phi$, $\delta t$) determined by shear wave splitting (SWS) analysis.
Figure 3: Schematic showing the different layers within the Earth. Names highlighted in red denote confirmed observations of SWS.
Figure 4: An illustration depicting Shape Preferred Orientation (SPO) versus Lattice Preferred Orientation (LPO) in the upper 400 km of the Earth; i.e. the dominant contributing portion to observed SWS. The four circumscribed numbers each indicates one of the four layers that constitute the Earth’s upper 400 km. Layer 1 is the upper crust (up to 10-15 km depth) where SPO of cracks bipolarizes shear phases. Layer 2 is the lower crust, which generates SWS via pure shear deformation or the SPO of structures, such as shear zones. Layers 3 and 4 are the upper mantle, which creates seismic anisotropy via olivine-induced LPO. However, SWS in layer 3 is produced via “frozen” LPO in the mantle lithosphere and layer 4 is caused by modern-day mantle flow in the asthenosphere. Black triangles represent stations and the red bars plotted on top of them are hypothetical splitting measurements. For this example, the splits demonstrate that observed anisotropy results from multiple layers of anisotropy, as reflected by the measurement discrepancies between stations (i.e. φ and δt vary with backazimuth).
Figure 5: Collection of Western U.S. splitting parameters: solid circles represent individual splits; station averages are open circles [Waite et al., 2005]. Blue splits represent results reported by Waite et al. [2005] and red splits are from other studies. Within the yellow box is our regional study, and the blue box contains LPSSE. Take note of the remarkably coherent splitting in the western portion of our regional study and the inconsistent measurements southeast of LPSSE in Colorado.
Figure 6: Location map of the 63 seismometers that LPSSE. (A) The 55 L-stations (dense array) separated into Arms 1, 2, and 3. (B) The eight A-stations. The black box encompasses the LaBarge dense array.
Figure 7: Regional map showing all utilized seismometers. Green diamonds are regional (public) stations; red diamonds are LPSSE stations.
Figure 8: Rose diagram showing the distribution of events used for our study.
Figure 9: Examples of good results recorded at station K19A. (A) A good splitting event from April 19, 2009. The top panels show the diagnostic plots of the RC method made in SplitLab. From left to right they include the corrected fast (dashed, blue) and slow (solid, red) components, the corrected radial (dashed, blue) and transverse (solid, red) components, and the corrected (solid) and uncorrected (dashed) particle motion diagrams. The bottom panel illustrates the uncorrected radial and transverse seismograms. (B) A good null event from April 4, 2009. Diagnostic plots are as in part (A). The yellow box highlights the time window selected for analyzing the SKS arrival.
Figure 10: Result map of the L-stations, with $\varphi$ indicated by the bars aligned with the seismic fast directions and $\delta t$ is scaled to the length of the stick. The large arrow represents the absolute plate motion direction. Note, only one usable split was detected during the six month deployment (October 2008 - June 2009).
Figure 11: Fast direction ($\phi$) plotted as a function of station longitude for arms 1 and 2 (A) and arms 1 and 3 (B). Note the difference in $\phi$ estimates between the SC and RC/EV methods. The dark line indicates the direction of APM.
Figure 12: Same as Figure 11 but for delay time. Note the difference in $\delta t$ values between the SC and RC/EV methods. The dark line indicates the average $\delta t$ for the L-stations.
Figure 13: Result map of the A-stations and L-stations, with $\phi$ and $\delta t$ plotted as in Figure 10. The large arrow parallels the direction of APM. Note, only one usable split was detected during the six month deployment (October 2008 - June 2009).
Figure 14: Individual splitting parameters for regional and LaBarge stations, with $\varphi$ and $\delta t$ plotted as in Figures 10 and 13. The large arrow is aligned with the direction of APM.
Figure 15: (A) Fast direction ($\phi$) plotted against station longitude for all events. The APM direction, within error, is indicated by the orange bar between 60° and 80°. Note that the majority of fast axes reported by western stations align with the APM direction. (B) Null backazimuth (mod-180°) is plotted as a function of station longitude. The agreement of null backazimuths with the fast or slow splitting direction indicates simple anisotropy.
Figure 16: Backazimuthal variations of SC splitting parameters for events measured at stations within the NE-SW splitting domain. Splitting parameters are plotted as a function of modulo-180° in order to increase backazimuthal coverage and identify evidence of complex anisotropy. (A) Fast direction (φ) plotted against modulo-180°. (B) Same as (A) but for delay time (δt). Both (A) and (B) indicate the existence of simple anisotropy.
Figure 17: Same as Figure 16 but for events measured at stations within the E-W splitting domain. (A) Note the $\pi/2$ periodicity, shaded orange, which may be indicative of vertically varying anisotropy. (B) Unlike $\phi$, there is little systematic variation in $\delta t$ with mod-$180^{\circ}$. 
Figure 18: Station averages of splitting measurements with $\varphi$ and $\delta t$ plotted as in Figures 10, 13, and 14. The large arrow represents the direction of APM. Note that station average splitting is likely appropriate only for stations near the Eastern Snake River Plain (eSRP) and within the Green River Basin.
Figure 19: Station averages of the splitting parameters: (A) fast direction and (B) delay time. Violet diamonds represent the LaBarge dense array. The APM direction, within error, is indicated by the orange bar between 60° and 80° in (A).
Figure 20: Examples of frequency dependence for two SKS waveforms shown on SC diagnostic plots. The left panel displays the uncorrected radial (dashed, blue) and transverse (solid, red) seismograms, the center panel the corrected (solid, red) and uncorrected (dashed, blue) particle motion diagrams, and the right panel the error plot. (A) An example SKS arrival (event 2008.130.21:51 at station IMW) exhibiting no splitting discrepancies (i.e. the error surfaces of the two measurements overlap). The L-band yields a $\phi$ of $52 < 62 < 72^\circ$ and $\delta t$ of $1.1 < 1.2 < 1.4$ seconds. The I-band produces similar, but more robust, splitting parameters of $68 < 74 < 78^\circ$ and $1.2 < 1.5 < 1.7$ seconds. (B) An example SKS phase with frequency-dependence (event 2008.197.03:26 at station K19A). The L-band measures a null measurement. The I-band yields a good split with a $\phi$ of $60 < 66 < 74^\circ$ and $\delta t$ of $0.6 < 0.7 < 0.8$ seconds. Note, the surface errors for the two bands do not overlap.
Figure 21: Chart displaying the percentage of splitting energy and null energy exhibited in each discreet frequency band for the highest quality splitting measurements. (A) Results from the NE-SW splitting domain. (B) Results from the E-W splitting domain.
Figure 22: A cartoon illustrating two ray paths and their corresponding ray-piercing points. A piercing point is the position on the Earth’s surface directly above a seismic wave at a particular depth. The “true” depth of anisotropy can be estimated from the minimum spatial variation in splitting measurements as a function of depth.
Figure 23: Figure shows estimated depth to anisotropy in the E-W splitting domain (100-130 km depth). The shaded region indicates splits not used as a result of dipping anisotropy. (A) Spatial variation factor of splitting measurements as a function of assumed depth of anisotropy. (B) Splits plotted on a LAB contour map [adapted from Miller and Levander, 2011] at estimated anisotropic depth.
Figure 24: Same as Figure 23 but for the NE-SW splitting domain. (A) The estimated depth to anisotropy is less constrained than the variation factor plot in Figure 23A, which may suggest that the two splitting domains are at the same depth. (B) The estimated depth to anisotropy is placed at an arbitrary deep depth (170 km), because the simple waveform and frequency dependence analyses indicate asthenospheric dominated anisotropy.
Figure 25: Map illustrating the simplified anisotropic structure of our study area. Red bars represent asthenospheric induced SWS. Black bars represent lithospheric-dominated SWS. The Yellowstone mantle plume (red shape) is imaged at 180 km depth, and the high-velocity anomaly (blue shape) crossing the Green River, Wind River, and Bighorn Basins is imaged at 90 km depth [Waite et al., 2006]. The brown shape represents the approximate extent of the Cheyenne slab beneath Wyoming [Yuan and Dueker, 2005; Dueker et al., 2001]. The green dashed line outlines the eSRP and the blue solid line traces the rim of the Yellowstone caldera (YS). Archean boundaries are indicated by purple dashed lines dividing the Wyoming Craton into five subprovinces (four of which are within the map): the Montana medasedimentary province (MMP); the Beartooth-Bighorn magmatic zone (BBMZ); the Sweetwater subprovince, and the Southern accreted terranes (SAT). The Oregon Trail Structure (OTS-GF) is a late Archean suture that separates the Sweetwater subprovince and SAT.
Appendix I. Summary of splitting results from the LaBarge Array.

Explanation of abbreviations: Date = dd/mm/yy; J-Day = Julian Calendar day; Station = station name; Sta. LAT = station latitude in degrees; Sta. LON = station longitude in degrees; Phase = phase of event; BAZ = backazimuth in degrees; Mod180 = backazimuth (modulo-180°); \( \varphi(\text{RC, SC, EV}) \) = fast direction calculated by the rotation correlation, minimum energy, and eigenvalue methods respectively; \( \delta(\text{RC, SC, EV}) \) = delay time estimated by the rotation correlation, minimum energy, and eigenvalue methods respectively; Quality = ranking; Null? = is the event a splitting or null measurement?

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Appendix II. List of individual splitting results from regional stations.

Explanation of abbreviations: Date = dd/mm/yy; J-Day = Julian Calendar day; Station= station name; Sta. LAT = station latitude in degrees; Sta. LON = station longitude in degrees; Phase = phase of event; BAZ = backazimuth in degrees; Mod180 = backazimuth (modulo-180°); ϕ(RC, SC, EV) = fast direction calculated by the rotation correlation, minimum energy, and eigenvalue methods respectively; δ(RC, SC, EV) = delay time estimated by the rotation correlation, minimum energy, and eigenvalue methods respectively; Quality = ranking; Null? = is the event a splitting or null measurement?

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Appendix III. Station averages for regional stations. Note, that averaged results are most appropriate for western regional stations (AHID, DCID1, FXWY, I16A, LI6A, L17A, LOHW, MOOW, RR12, Y18, and Y19).

Explanation of abbreviations: Station= station name; Sta. LAT = station latitude in degrees; Sta. LON = station longitude in degrees; Mean $\varphi$(RC, SC, EV) = station average fast direction calculated by the rotation correlation, minimum energy, and eigenvalue methods respectively; Mean $\delta$t(RC, SC, EV) = station average delay time estimated by the rotation correlation, minimum energy, and eigenvalue methods respectively.

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Appendix IV. Summary of frequency analysis results. Splitting estimates from 16 high quality station-event pairs. Each arrival was subject to filtering at low (10-33.3s), intermediate (3.3-10 s), and high (1-3.3 s) frequencies. The final column indicates whether a measurement discrepancy was found for a particular station-event pair.

Explanation of abbreviations: Station= station name; Date = dd/mm/yy; J-Day = Julian Calendar day; BAZ = backazimuth in degrees; $\varphi$, H-band = high frequency fast direction; $\delta t$, H-band = high frequency delay time; $\varphi$, I-band = intermediate frequency fast direction; $\delta t$, I-band = Intermediate frequency delay time; $\varphi$, L-band = low frequency fast direction; $\delta t$, L-band = low frequency delay time; Discrepant? = Are values between frequency bands within error?

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<th>J-day</th>
<th>BAZ</th>
<th>$\varphi^*$, H-band</th>
<th>$\delta t^*$ (s), H-band</th>
<th>$\varphi^*$, I-band</th>
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Appendix V. Statistical Analysis of the LaBarge Dense Array.

Table A calculations of fast direction ($\varphi$) in degrees. Table B contains calculations of delay time ($\delta t$) in seconds.

Explanation of abbreviations: Segments = L01-L22 is Arm 1, L23-L41 is Arm 2, and L42-L55 is Arm 3, and All stations includes all dense array stations; Avg.(RC, SC, EV) = average splitting parameter for an arm of the array, which is calculated by the rotation correlation, minimum energy, and eigenvalue methods respectively; S.D.(RC, SC, EV) = the standard deviation of a splitting parameter for an arm of the array, which is calculated by the rotation correlation, minimum energy, and eigenvalue methods respectively; All methods = average splitting parameter for each arm.

### Table A

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### Table B

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REFERENCES


Humphreys, E., Relation of flat subduction to magmatism and deformation in Western United States, Geological Society of America, 204, 85-98, 2009.


Miller, M.S., and A. Levander, Lithospheric Structure beneath the Western US Using USArray Data. IRIS Image Gallery, 2011.


