Anisotropic Zonation in the Lithosphere of Central North America: Influence of a Strong Cratonic Lithosphere on the Mid-Continent Rift

O. Ola^a, A.W. Frederiksen^{a,*}, T. Bollmann^b, S. van der Lee^b, F. Darbyshire^c, E. Wolin^b, J. Revenaugh^d, C. Stein^e, S. Stein^b, M. Wysession^f

^a*Department of Geological Sciences, University of Manitoba, Winnipeg, Manitoba, R3T 2N2, Canada*

^b*Department of Earth and Planetary Sciences, Northwestern University, 2145 Sheridan Road, Evanston IL, 60208-3130, USA*

^c*Centre de recherche GEOTOP, Université du Québec à Montréal, CP 8888, succ. Centre-Ville, Montréal, Québec, Canada, H3C 3P8*

^d*Department of Earth Sciences, University of Minnesota, 310 Pillsbury Drive SE, Minneapolis, MN 55455-0231, USA*

^e*Department of Earth and Environmental Sciences, University of Illinois at Chicago, 845 W. Taylor St., Chicago, IL 60607-7059, USA*

^f*Department of Earth and Planetary Sciences, Washington University in St. Louis, One Brookings Drive, St. Louis, MO 63130-4899, USA*

Abstract

We present shear-wave splitting analyses of SKS and SKKS waves recorded at sixteen Superior Province Rifting Earthscope Experiment (SPREE) seismic stations on the north shore of Lake Superior, as well as fifteen selected Earthscope Transportable Array instruments south of the lake. These instruments bracket the Mid-Continent Rift (MCR) and sample the Superior, Penokean, Yavapai and Mazatzal tectonic provinces. The data set can be explained by a single layer of anisotropic fabric, which we interpret to be dominated by a lithospheric contribution. The fast S polarization directions are consistently ENE-WSW, but the split time varies greatly across the study

Preprint submitted to Tectonophysics *May 27, 2016*

*Corresponding author.

Email addresses: kunleola15@yahoo.com (O. Ola),
Andrew.Frederiksen@umanitoba.ca (A.W. Frederiksen),
trevor@earth.northwestern.edu (T. Bollmann), suzan@earth.northwestern.edu (S. van der Lee), darbyshire.fiona_ann@uqam.ca (F. Darbyshire),
emilyw@earth.northwestern.edu (E. Wolin), justinr@umn.edu (J. Revenaugh),
cstein@uic.edu (C. Stein), seth@earth.northwestern.edu (S. Stein),
michael@wucore.wustl.edu (M. Wysession)

29 area, showing strong anisotropy (up to 1.48 s) in the western Superior, mod-
30 erate anisotropy in the eastern Superior, and moderate to low anisotropy in
31 the terranes south of Lake Superior. We locate two localized zones of very
32 low split time (less than 0.6 s) adjacent to the MCR: one in the Nipigon
33 Embayment, an MCR-related magmatic feature immediately north of Lake
34 Superior, and the other adjacent to the eastern end of the lake, at the south-
35 ern end of the Kapuskasing Structural Zone (KSZ). Both low-splitting zones
36 are adjacent to sharp bends in the MCR axis. We interpret these two zones,
37 along with a low-velocity linear feature imaged by a previous tomographic
38 study beneath Minnesota and the Dakotas, as failed lithospheric branches of
39 the MCR. Given that all three of these branches failed to propagate into the
40 Superior Province lithosphere, we propose that the sharp bend of the MCR
41 through Lake Superior is a consequence of the high mechanical strength of
42 the Superior lithosphere ca. 1.1 Ga.

43 *Keywords:* Mid-Continent Rift, Superior Province, shear-wave splitting,
44 lithosphere, anisotropy, Nipigon Embayment.

45 1. Introduction

46 Rifting a continent necessarily involves both the crust and the entire litho-
47 sphere. The mechanical strength of the continental lithosphere plays an im-
48 portant role in this process (Gueydan et al., 2008; Huismans and Beaumont,
49 2011), as the presence or absence of a strong lithosphere is a major control
50 on the geometry and deformation mechanisms of the evolving rift. In addi-
51 tion, the mechanical fabric of the lithosphere may influence the directionality
52 of the rifting process (Tommasi and Vauchez, 2000). Rifting processes are,

53 to some extent, recorded in the lithospheric fabric beneath active (Bastow
54 et al., 2010) as well as long-stable (Vauchez et al., 2000) rift zones, though
55 strain localization in active rifts implies that broad anisotropic features will
56 primarily record the early stages of rift development.

57 The Mesoproterozoic Mid-Continent Rift (MCR), in central North Amer-
58 ica, abuts on the Archean Superior Province (SP), the largest Archean craton
59 in existence. The MCR cross-cuts the Proterozoic Penokean, Yavapai and
60 Mazatzal orogens with both its eastern and western arms (Fig. 1), but avoids
61 penetrating deep into the SP, instead bending sharply through Lake Supe-
62 rior. The MCR was recently instrumented with broadband seismographs as
63 part of the Superior Province Rifting Earthscope Experiment (SPREE; Stein
64 et al., 2011; Wolin et al., 2015), yielding the first detailed seismic constraints
65 on the lithosphere of the MCR/Superior contact. In this study, we present
66 the first observations of upper-mantle anisotropy made using this data set.
67 We measure the S polarization anisotropy of the upper mantle using SKS
68 splitting methods, control for possible non-lithospheric sources of splitting
69 effects, interpret the measured splitting in terms of variations in lithospheric
70 fabric, and examine the relationship between the MCR and the SP litho-
71 sphere. We suggest that rifting did not extend further to the north owing to
72 the strong SP lithosphere, though MCR magmatism may have propagated
73 into the Superior lithosphere in several places.

74 2. Tectonic and geophysical background

75 The Canadian Shield, the Precambrian core of North America, is an amal-
76 gam of Archean and Proterozoic tectonic blocks and orogens. The largest of

77 the Archean blocks is the Superior Province, which stabilized ca. 2.6 Ga via
 78 accretion of a series of older terranes (Card, 1990; Calvert and Ludden, 1999;
 79 Percival et al., 2006). In the western Superior, these terranes form narrow
 80 belts with a consistent E-W alignment; sutures between these belts have been
 81 found to traverse the Moho in LITHOPROBE seismic sections (White et al.,
 82 2003), indicating that tectonic accretion had a role in the formation of the
 83 Superior lithosphere. The lithosphere beneath the Superior Province is thick
 84 and seismically fast (Darbyshire et al., 2007; Frederiksen et al., 2007, 2013a)
 85 as well as strongly anisotropic (Darbyshire and Lebedev, 2009; Frederiksen et
 86 al., 2013b; Ferré et al., 2014), possibly as a result of accretionary processes.
 87 The lithosphere beneath the eastern Superior is seismically slower and con-
 88 tains an anomaly attributed to the Great Meteor hotspot track (Rondenay et
 89 al., 2000; Eaton and Frederiksen, 2007; Frederiksen et al., 2007). The eastern
 90 Superior was affected by uplift along the ca. 1.9 Ga Kapuskasing Structural
 91 Zone (KSZ; Percival and West, 1994).

92 The Superior Province is surrounded by Proterozoic orogens (Fig. 1). The
 93 oldest of these are the roughly contemporaneous Trans-Hudson and Penokean
 94 orogens, which accreted to the west and south of the Superior, respectively,
 95 ca. 1.8 Ga (Whitmeyer and Karlstrom, 2007). The Yavapai and Mazatzal
 96 orogens accreted further juvenile crust ca. 1.7 and 1.6 Ga, respectively,
 97 followed by extensive plutonism (Whitmeyer and Karlstrom, 2007; Amato et
 98 al., 2008). Further accretion continued southward with the Granite-Rhyolite
 99 Province ca. 1.55-1.35 Ga, which extends beyond our study area (Whitmeyer
 100 and Karlstrom, 2007). The last and largest of these orogens is the Grenville
 101 Orogen, which accreted to the east of the Superior in stages from 1.3 to 1.0

102 Ga as part of a major continent-continent collision (Davidson, 1998).

103 While Grenvillian orogenesis was in progress, a major magmatic feature
104 cross-cut the preexisting Penokean, Yavapai and Mazatzal provinces: the
105 Mid-Continent Rift (MCR). The MCR is a ca. 3000 km long, arcuate rift
106 structure that curves through Lake Superior, with arms extending southwest
107 and southeast (Van Schmus and Hinze, 1985; Ojakangas et al., 2001); rifting
108 along the MCR may have been related to the opening of an ocean between
109 Amazonia and Laurentia ca. 1.1 Ga (Stein et al., 2014). The rift contains
110 large volumes of basaltic magma, generating a significant gravity anomaly
111 (see, e.g., Merino et al., 2013); the high volume and geochemistry of the
112 basalts suggest hotspot participation in the rifting process (Hutchinson et
113 al., 1990; White, 1997; Hollings et al., 2012, 2014) and the MCR has been
114 described as a hybrid of a rift and a large igneous province (Stein et al., 2015).
115 A late compressional stage of the MCR’s development may have reactivated
116 structures related to the KSZ (Manson and Halls, 1997).

117 The Nipigon Embayment (NE; Fig. 1) is a magmatic feature north of
118 Lake Superior, adjacent to the most sharply-curved section of the MCR.
119 Its mafic and ultramafic rocks are contemporaneous with the early stages
120 of the MCR (Hollings et al., 2007), but are predominantly emplaced in the
121 form of sills rather than dykes. The dominance of sills is suggestive of a
122 non-extensional tectonic regime (Hart and MacDonald, 2007), though sills
123 are not in themselves incompatible with extensional processes. The NE has
124 been found to overlie anomalous mantle in a number of studies (Ferguson et
125 al., 2005; Frederiksen et al., 2007, 2013a).

126 Limited geophysical constraints are available on the lithosphere of the

127 MCR. The MCR crust was examined by the Great Lakes International Mul-
128 tidisciplinary Program on Crustal Evolution (GLIMPCE), which included a
129 number of marine seismic reflection surveys performed within Lake Superior
130 (Cannon et al., 1989). These surveys revealed varying asymmetry along the
131 rift, and evidence for significant crustal thinning during rifting, followed by a
132 late-stage compressional event (Mariano and Hinze, 1994; Samson and West,
133 1994; Sexton and Henson, 1994). The western arm of the MCR was exam-
134 ined using ambient noise and surface-wave tomography by Shen et al. (2013),
135 who found thickened crust along the MCR and an intermittent low-velocity
136 feature in the lithospheric mantle beneath the rift axis. The teleseismic P-
137 wave model of Frederiksen et al. (2013a) also showed a low-velocity anomaly
138 at lithospheric depth (50-250 km) beneath part of the western arm, but its
139 resolution of the MCR is limited. The lithospheric expression of the MCR
140 at regional scales has not been well imaged by published studies, nor has
141 the relationship between the MCR and the lithospheric anomaly beneath the
142 NE.

143 **3. Data and processing**

144 Deployment of the Earthscope Transportable Array (TA) reached Min-
145 nesota in 2010 and Wisconsin in 2011, occupying the south shore of Lake
146 Superior from mid-2011 through mid-2013 with instruments spaced approxi-
147 mately 70 km apart. To coincide with this deployment, 83 broadband Earth-
148 scope FlexArray instruments were deployed in the Superior Province Rifting
149 Earthscope Experiment (SPREE; Stein et al., 2011; Wolin et al., 2015). The
150 SPREE deployment consisted of dense lines of instruments along and across

151 the axis of the southwest arm of the MCR, in Minnesota and Wisconsin,
 152 along with a sparser deployment of stations north of Lake Superior, in On-
 153 tario, at a spacing comparable to the TA (Fig. 2). In this study, we examine
 154 data from the sixteen Canadian SPREE stations as well as fifteen selected TA
 155 stations south of Lake Superior, thus building a data set that straddles the
 156 meeting point of the eastern and western arms of the MCR. Eight of the TA
 157 stations (C39A, C40A, D37A, D41A, E38A, E39A, E43A, and E44A) were
 158 previously analyzed in Frederiksen et al. (2013b) using the same approach
 159 as is used here; the results in this study are based on a larger data set and
 160 should be considered more robust.

161 In an anisotropic layer, an incoming S wave will excite one or both of
 162 two possible shear-like (quasi-S) wave modes with different velocities; if the
 163 anisotropy is weak, the two quasi-S modes will have approximately orthogo-
 164 nal polarizations. We use teleseismic ray paths (SKS and SKKS) that have
 165 a radial plane polarization in the absence of anisotropy, and are near-vertical
 166 in the upper mantle; thus, we are able to observe the effect of anisotropy
 167 on a vertically-propagating wave of varying polarization, as SKS waves from
 168 earthquakes in different regions will arrive along different azimuths and so
 169 with different directions of polarization. Analysis of the SKS or SKKS pulse
 170 yields the polarization azimuth of the faster (qS1) mode, often referred to
 171 in the literature as the "fast direction" or "fast axis" (though we use the
 172 term "fast polarization direction" in this paper), as well as the time sepa-
 173 ration between the two quasi-S modes, which is known as the "split time".
 174 If the incident SKS wave travels near-vertically, the fast S polarization di-
 175 rection obtained by splitting analysis may be attributed to the projection of

176 material fabric onto the horizontal plane, while the split time represents a
177 combination of the thickness and strength of this projected fabric.

178 We obtained data for events of magnitude ≥ 6 at distances correspond-
179 ing to angles from 90° to 130° from a point at the centre of the array
180 (47.79°N , 87.70°W). In this distance range, the SKS pulse is expected to
181 be well-separated from other body-wave arrivals, and so is suitable for split-
182 ting analysis. Of the 196 events considered, 128 events exhibited usable (i.e.,
183 of quality ≥ 3 ; see below for details) SKS or SKKS pulses at at least one
184 station. For each trace, the data were filtered in a frequency band of 0.02-
185 0.2 Hz, windows were manually chosen around the expected SKS and SKKS
186 arrival times, and a Hanning taper was applied to the ends of the SKS and
187 SKKS windows.

188 The earthquakes for which usable SKS or SKKS pulses were obtained
189 (Fig. 3a) are concentrated in back azimuths ranging from WSW clockwise
190 to NNE, with a few scattered events from the south. This gives nearly
191 continuous coverage over almost half of the possible back-azimuth range. If
192 we assume that the incident SKS and SKKS waves travel nearly vertically in
193 the upper mantle, then their splitting should depend only on the polarization
194 plane of the incident wave, which will be the same for two events 180° apart in
195 back azimuth. If we consider only the polarization directions (back azimuths
196 modulo 180°) of these events, we have nearly complete coverage (Fig. 3b),
197 with all but one 10° swath containing at least one event. In the case of layered
198 anisotropy with horizontal fast axes, the anisotropic response will have 180°
199 back-azimuthal symmetry and will depend only on the polarization direction,
200 even for a non-vertical incidence angle; if a plunging anisotropic symmetry

axis or a dipping interface are present, the symmetry will be broken. Given that our data are largely restricted to a single hemisphere, and that we are considering SKS and SKKS arrivals with very steep incidence angles, we will be unable to detect deviations from layered structure and horizontal axis orientations. However, our very complete polarization-direction coverage will allow us to detect the effect of multiple anisotropic layers, should they be present.

A crude estimate of the anisotropic influence on these data may be obtained by examining the average ratio of transverse to radial energy in the SKS or SKKS pulse. In an isotropically layered Earth, the transverse component of a core-refracted wave should consist entirely of noise. If anisotropic material is present on the receiver side of the ray path, there will be coherent energy on the transverse component, unless the earthquake is aligned with one of the anisotropic symmetry axes; if we average the ratio of transverse to radial energy over a range of event azimuths, we expect that stations with stronger anisotropy will exhibit a higher ratio of transverse energy. We averaged the transverse/radial (T/R) energy ratio for all acceptable SKS and SKKS time windows at each station (Fig. 4). The map is spatially coherent over long distances, indicating that the T/R ratio is measuring large-scale structure rather than localized effects at individual sites. The high ratios north of Lake Superior indicate that SKS waves deviate strongly from radial polarization at these stations, indicating a strong anisotropic influence.

Shear-wave splitting analysis for individual events was carried out using the eigenvalue minimization approach of Silver and Chan (1991), in which a grid search is performed over a range of split time (δt) and fast S polarization

226 azimuth (ϕ) values to find the values that, when applied as a correction, min-
 227 imize the second eigenvalue of the covariance matrix between the corrected
 228 traces, and so recover the most linear initial particle motion (Fig. 5). As
 229 an additional check, we also minimized the energy on the corrected trans-
 230 verse component, which should yield approximately the same result. Based
 231 on plots similar to Fig. 5, we assigned a quality to each SKS and SKKS
 232 pulse on a subjective scale from 0 through 5, based on the apparent noise
 233 level on the input traces, the degree of linearization of the particle motion,
 234 the degree of minimization of the transverse energy, and the correspondence
 235 between the eigenvalue and transverse-energy solutions. Only traces with
 236 quality levels of 3 or more were retained for further analysis; examples of
 237 arrivals with qualities of 3, 4 and 5 are provided as supplementary figures.

238 The single-event measurements at each station show considerable scatter
 239 (Fig. 6), particularly in the recovered split time. This is a problem inherent
 240 to shear-wave splitting analysis: single events are more sensitive to fast S
 241 polarization direction than split time, and often a considerable portion of
 242 the $\phi, \delta t$ surface returns low values of misfit (Fig. 5). Particularly strong
 243 ambiguities arise when the incident-wave polarization is close to the polar-
 244 ization of the fast or slow quasi-S wave, in which case no splitting is observed,
 245 the split time is not constrained, and there is a 90° ambiguity in fast S po-
 246 larization direction. As noted by Wolfe and Silver (1998), the error surface
 247 (i.e., the value of the second eigenvalue of the covariance matrix, calculated
 248 over a grid of $\phi, \delta t$ values) is a more robust observable than the actual split-
 249 ting parameters, so averaging error surfaces over multiple events is a safer
 250 approach than averaging together single-event splitting measurements if a

single-layered model is sufficient to explain the data.

The single-event measurements from our data set (Fig. 6) are too scattered to indicate whether the splitting parameters vary systematically with back azimuth, which would be an indicator of complex or multilayered anisotropic structure. To check for back-azimuthal variations, we stacked the error surfaces for events falling in polarization swaths at each station (Fig. 7), with the polarization direction taken as the back azimuth modulo 180° (i.e., the remainder of the back azimuth $\div 180^\circ$). The swaths used are the same as the ones used in the histogram in Fig. 3b. The swath-stacked error surfaces for the example station vary significantly by direction, with some directions (e.g., 140° - 150°) showing a null-like pattern with 90° directional ambiguity and no split-time resolution, and others (e.g., 90° - 100°) constraining the split time while having limited directional resolution. The final set of splitting parameters (white dots on all panels) fall in low-misfit regions in all of the swaths, indicating that a one-layer model is compatible with the entire data set, though the differences between some swaths (e.g., 90° - 100° and 130° - 140°) suggest that more complex structures may be present. The sample station is typical of the data set, in that none of the stations examined unambiguously required multiple anisotropic layers to explain the observed error surfaces. Therefore, we proceeded with a one-layer analysis at all stations.

Final measurement of splitting parameters was done using a directionally-balanced variant (Frederiksen et al., 2006, 2007, 2013b) of the error-surface stacking method of Wolfe and Silver (1998). The error surfaces were stacked twice, first by forming directional swaths as described above, and then by stacking the swath stacks with equal weight. This last procedure evens out

276 the directional coverage (within the limitations of the data set) and so yields
 277 results that are not dominated by the most seismically-active directions. The
 278 minimum quality threshold for inclusion in the stack was taken to be either
 279 3 or 4 depending on the number of available events at the station and the
 280 appearance of the stacked error surface. The final stack of all swaths (Fig.
 281 8) yielded an estimate of the splitting parameters at the station, as well as
 282 an error bar (Figure 9b) obtained from the error surface using the Fischer
 283 F-test (Silver and Chan, 1991); the obtained error bar treats the composite
 284 error surface as though it were obtained from a single trace, and is therefore
 285 a pessimistic estimate. At station D46A, the measured split time of 0.28 s is
 286 less than the error bar of ± 0.33 s, indicating a null measurement (anisotropy
 287 is not necessary to explain the data). Stations SC07 and K42A are near-null
 288 cases where the error bar comes within 0.05 s of the split time, so their fast
 289 S polarization directions should be interpreted with caution.

290 **4. Results**

291 The final splitting parameters are given in Table 1 and plotted as direc-
 292 tional arrows in Fig. 9a. The map also includes results from a number of
 293 other studies in the area, divided into measurements done using the same
 294 methodology as our study (Frederiksen et al., 2006, 2007, 2013b) and other
 295 published measurements (Silver and Kaneshima, 1993; Barruol et al., 1997;
 296 Kay et al., 1999; Rondenay et al., 2000; Eaton et al., 2004; Ferré et al., 2014).
 297 The study of Yang et al. (2014) is omitted from this map due to a difference
 298 in methodology – unlike the other splitting measurements shown here, their
 299 measurements are based on averaging of single-event splitting measurements

rather than the stacking of error surfaces. As noted by Kong et al. (2015), splitting results obtained by averaging splitting parameters rather than stacking error surfaces tend to produce somewhat higher averaged split times from the same data sets; we have therefore excluded the Yang et al. (2014) measurements from our quantitative analysis, though their spatial pattern is in keeping with the other studies.

Of the 31 stations we examined, all but one of the fast polarization directions lie within the northeast quadrant, ranging from 36° to 107° (the one exception) with an average of 69° and a standard deviation of 14° . The exception is D46A, which, as noted above, is a null measurement whose fast polarization direction may not be meaningful. A contour plot of the fast S polarization directions is shown in Fig. 10. North of Lake Superior, the fast polarization azimuth is consistently close to 70° (ENE-WSW); immediately south of the lake, there is more variability, with some stations having fast polarization directions closer to 45° (NE-SW), while ENE-WSW directions resume further south. Looking at variations over a broader area (including previous studies), we can see that the fast polarization direction rotates to NE-SW between 44 and 49°N along the western edge of the map area, and to E-W in the SE corner of the map.

The split time (Fig. 11) shows considerable variation over the study area, averaging 0.62 s with a standard deviation of 0.26 s. SPREE stations north of Lake Superior and west of ca. 89°W exhibit split times of 0.8 s or greater, including our strongest observed split (1.48 s at SC02). East of 89°W , stations north of the lake exhibit moderate to low split times, with split times less than 0.6 s concentrated in two clusters along the lakeshore:

one centered at 49°N, 88°W, at the western edge of the Nipigon Embayment,
and one at 47°N, 84°W at the eastern end of Lake Superior. South of the lake,
split times are also low, and decrease southward into the Mazatzal Orogen.

5. Discussion

5.1. Depth of anisotropic variations

Shear-wave splitting of teleseismic phases is diagnostic of anisotropic fabric, but provides no direct constraint on the depth of the anisotropy. When core-refracted phases are used, as is done here, the splitting effect is physically limited to the receiver-side path above the core-mantle boundary, but in principle, the anisotropy calculated from a split SKS or SKKS pulse may be present at any depth between the receiver and the core-mantle boundary (CMB).

Whole-mantle tomographic models that include anisotropy (e.g., Panning and Romanowicz, 2006; Auer et al., 2014) typically assume radial anisotropy (i.e. anisotropy with a vertical symmetry axis), in contrast to the azimuthal anisotropy (anisotropy with a horizontal symmetry axis) that shear-wave splitting is able to detect. These models generally show that the strongest anisotropy present is in the upper mantle, but that anisotropy in D'' is also strong. Targeted body-wave studies (e.g., Garnero et al., 2004; Long, 2009; He and Long, 2011; Nowacki et al., 2010) detected significant azimuthal anisotropy in portions of D'', the strongest fabric being associated with regions of inferred downwelling (Pacific subduction zones) and upwelling (large low-shear-velocity provinces; Garnero and McNamara, 2008). As SKS ray paths necessarily pass through the D'' layer, it is likely that our data set

contains some degree of contamination from the base of the mantle, and possible that the cumulative effect of weak mid-mantle anisotropy may also affect our data. Our use of eigenvalue minimization for splitting analysis, which maximizes the linearity of the incident wave’s polarization, is robust in the presence of deviations from SV polarization, and in any case we have not observed any systematic deviations of this nature. Therefore, it is only deep-mantle anisotropy capable of splitting the SKS wave that must be considered.

To address this, we examined splitting parameters for individual events averaged over all stations for which the event was recorded with acceptable quality. This is a somewhat ad hoc approach that cannot completely isolate deep-mantle effects. However, the averaging should enhance the effect of deep-Earth contributions, as the ray paths for an event will be closer together at the CMB than in the lithosphere. The resulting maps (Fig. 12) show coherent spatial variations in the fast S polarization direction (upper panel), with nearby events generally exhibiting similar fast polarizations. The split time (lower panel), which is more difficult to measure robustly from single events, shows no obvious coherence. We take the coherent fast-polarization clusters to be evidence of at least some deep-mantle influence on our data set.

The relatively short spatial wavelength of variations in the event-averaged fast S polarization direction (Fig. 12, upper) suggests that the deep-mantle contribution to our split measurements is not systematic over large distances, and so may be suppressed by directional averaging. The general consistency observed in swath-stacked misfit surfaces (see e.g., Fig. 7) suggests that

374 deep-mantle contributions are already being averaged away within individ-
375 ual polarization swaths. Furthermore, our two-stage stacking approach, in
376 which the swath stacks are themselves stacked with equal weight to form
377 composite error surfaces for each station, should suppress any remaining di-
378 rectional variation. We will therefore interpret our station-averaged results
379 under the assumption that they represent only upper-mantle anisotropy. We
380 further adopt the commonly-made assumption that upper-mantle anisotropy
381 is dominantly due to the preferential alignment of olivine crystals (see e.g.
382 Nicolas and Christensen, 1987; Silver, 1996), and so that our measurements
383 reflect fabric above the 410 km discontinuity, below which the olivine phase
384 is absent.

385 The remaining possible depth ranges for the anisotropy we observe are the
386 asthenosphere (representing active deformation), the lithosphere (represent-
387 ing frozen deformation), and the crust. The crustal contribution to shear-
388 wave splitting may be evaluated based on existing constraints on crustal
389 structure. In particular, the velocity structure of the western Superior Province
390 north of Lake Superior was examined using two perpendicular refraction lines
391 (Musacchio et al., 2004) as a component of the LITHOPROBE Western Su-
392 perior transect. The refraction survey located a ≈ 10 km-thick lower crustal
393 layer with P velocities of 7.5 km/s and 6.9 km/s in perpendicular directions,
394 representing 8.3% P anisotropy if the fast quasi-P axis is parallel to the north-
395 south line (i.e., perpendicular to the locally E-W geologic strike of Superior
396 subprovinces); assuming the S velocity has the same symmetry axis (which
397 will be the case for simple anisotropic symmetry models) and a comparable
398 percentage of anisotropy, this layer would generate a split of ≈ 0.2 s between

399 the fast and the slow wave. Even if strongly anisotropic, thick layers like this
400 were a common feature in the Superior crust, their contributions would be
401 insufficient to account for more than a small part of the observed splitting.
402 Ferré et al. (2014) also concluded that the crustal contribution to SKS split-
403 ting is weak in the southwest Superior, based on modelling of the seismic
404 effects of observed metamorphic foliation.

405 The question of asthenospheric versus lithospheric contributions is more
406 difficult to answer. The vast majority of the fast polarization directions we
407 observe are parallel to the direction of absolute plate motion calculated from
408 model HS3-NUVEL-1A (Fig. 9, green arrow; Gripp and Gordon, 1990) as
409 well as to the general tectonic fabric of the western Superior Province; the
410 absolute plate motion direction in this area is consistent between different
411 plate-motion models, and is fairly uniform over the study area. The split
412 times, by contrast, vary significantly over short length scales (Fig. 11). For
413 a ray of approximately 11,000 km in length (typical for a teleseismic SKS
414 phase) recorded at 0.2 Hz, the Fresnel zone at 250 km depth will be ≈ 106
415 km in diameter, indicating that stations less than this distance apart will
416 be sampling overlapping volumes within the asthenosphere. Given that the
417 split times we observe vary rapidly over short distances (SC04 and SC07 are
418 120 km apart and have split times of 0.90 and 0.25, respectively; SC05 and
419 C40A are 46 km apart and have split times of 0.85 and 0.55), we conclude
420 that, though there may be some asthenospheric contribution to the regional
421 anisotropy, the spatial variations that we see are the result of variations
422 within the lithosphere. It is worth bearing in mind, however, that strong
423 topography on the lithosphere-asthenosphere boundary can modify the as-

424 thenospheric flow pattern and cause local flow to be enhanced by channeling
425 effects, a process which can enhance shear-wave splitting (Fouch et al., 2000);
426 given that surface-wave models of the area (e.g., Darbyshire et al., 2007; Yuan
427 and Romanowicz, 2010) indicate a consistently thick lithosphere, we will in-
428 terpret our results largely in terms of lateral variations in lithospheric fabric.

429 Large-scale surface-wave models of North America indicate that mantle
430 anisotropy in the mid-continent is multi-layered (Darbyshire and Lebedev,
431 2009; Yuan and Romanowicz, 2010), with a lithospheric fabric that changes
432 across a mid-lithospheric discontinuity. Though our observations do not re-
433 quire multiple layers to explain the observed SKS/SKKS arrivals (see e.g.
434 Fig. 7), we cannot rule this out, given the lack of depth resolution in tele-
435 seismic shear-wave splitting analysis. Our horizontal resolution, by contrast,
436 is vastly superior to that of these surface-wave studies, so can make a much
437 more detailed interpretation of lateral changes in fabric. Future studies com-
438 bining SKS, surface-wave, and receiver-function observations will be required
439 to completely constrain the three-dimensional pattern of anisotropy in central
440 North America.

441 5.2. *Relationship to lithospheric velocity structure*

442 The Superior Province has been the subject of several tomographic studies
443 (Sol et al., 2002; Frederiksen et al., 2007; Darbyshire and Lebedev, 2009;
444 Frederiksen et al., 2013a), which detected significant lateral variations in
445 lithospheric velocity. As teleseismic tomography, being based on near-vertical
446 rays, has similar lateral resolution characteristics to SKS splitting, we will
447 examine the relationship between our results and the most recently-published
448 teleseismic P-velocity model (Frederiksen et al., 2013a). Fig. 13 shows the

split-time contours from Fig. 11 overlain on two depth slices through the velocity model. Although there is no simple relationship between split time and seismic velocity, there are a number of interesting spatial relationships between the pattern of split times and the pattern of velocities.

The most evident relationship is that the strongest splits are associated with a large region of elevated velocities in the northwest of the map. This feature is termed the Western Superior Mantle Anomaly (WSMA) by Frederiksen et al. (2013a,b): a region of high lithospheric velocity in tomographic images and strong, consistent ENE-WSW fabric inferred from SKS measurements, bounded by sharp gradients in both velocity and split time. Our new measurements sharpen the eastern edge of the WSMA significantly, particularly near the Nipigon Embayment, and confirm that the transition between the WSMA and the more moderate fabric in the eastern Superior is sharp rather than gradational.

The Frederiksen et al. (2013a) model contains two low-velocity anomalies in the eastern Superior: a large feature interpreted to correspond to the northwestern limit of the Great Meteor hotspot track, and a smaller feature corresponding to the Nipigon Embayment (NE). Our new splitting measurements show that, while the Great Meteor feature corresponds to moderate split times typical of the eastern Superior, the NE feature corresponds fairly closely to a zone of very low split times. The low splits are displaced slightly eastward of the NE, which may be a consequence of the dominance of ray paths from the west and north (Fig. 3); for a source-receiver distance of 100° , the SKS pierce point at 250 km depth (around the base of the lithosphere) will be displaced 0.46° , or 51 km, toward the source. The apparent shift of

474 the low-splitting contour lines is larger than this; however, given that those
475 contour lines are constrained by a small number of stations, it is possible
476 that the apparent shift is largely a contouring artifact; denser measurements
477 in and around the NE would be required in order to resolve this issue. With
478 this caveat kept in mind, and given previous magnetotelluric observations of
479 anomalous phase at lithospheric depth in the NE (Ferguson et al., 2005), we
480 now have three lines of geophysical evidence indicating that the embayment
481 is underlain by lithosphere significantly different from that of the surrounding
482 Superior Province.

483 We also detected a similar zone of very low splits immediately east of Lake
484 Superior. The velocity model does not contain a corresponding low velocity
485 feature; note, however, that the ray coverage of this zone was quite poor (the
486 region in question is greyed out due to lack of sampling in the 150 km depth
487 slice, Fig. 13). A similar sampling issue is also at play along the axis of
488 the MCR, given the lack of instrumentation within Lake Superior itself. The
489 relationship between seismic velocity and fabric along the rift axis should
490 become clearer once the SPREE data are incorporated into tomographic
491 models.

492 *5.3. Mantle domains north of Lake Superior*

493 As noted in the previous section, our major new observation is the pres-
494 ence of two localized zones of minimal shear-wave splitting along the edge of
495 the MCR, on the northern and eastern shores of Lake Superior (Fig. 11), one
496 of which coincides with a known mantle velocity anomaly beneath the NE
497 (Fig. 13). The near-zero split times in these two zones are similar to those
498 previously detected beneath the Minnesota River Valley Terrane (Frederiksen

et al., 2013b), though more localized, and are slightly lower than the values detected along the MCR axis, given our limited set of measurements within the MCR itself. The fact that the low-splitting zones are both adjacent to the MCR suggests some causal relationship.

The first question is whether these low-splitting zones actually represent an absence of coherent fabric, versus an interference effect of more complex layering (as suggested by Ferré et al., 2014, for a similar low-split region in southern Minnesota). We can address this by comparing the split time measurements in Fig. 11 to the transverse/radial (T/R) energy ratios in Fig. 4. We generated synthetic back azimuth-averaged T/R ratios and split times for a two-layer model whose layers exhibit split times of 0.4 and 0.6 s, respectively, with varying angles between the fast S polarization directions of the two layers. The results (Fig. 14a) indicate that the required degree of cancellation only occurs if the two layers' fast polarization directions deviate by less than 10° from perfect 90° opposition, while the T/R ratio is more sensitive to misalignment. Thus, for moderately misaligned layers, we would expect the T/R ratio to be low in proportion to the split time. In Fig. 14b, the T/R ratio and the split time are shown to be closely correlated, with no obviously low T/R values; we conclude that our observations do not require a contribution from multiple-layer interference. The stations at which we see very low splits also correspond to low ($< 7\%$) energy ratios, indicating that the low-splitting zones are zones where very little energy is rotated out of the radial plane by any means, including 3-D velocity variations or spatially-varying anisotropy with a horizontal axis; anisotropy with a vertical axis of symmetry cannot be ruled out by SKS splitting data.

524 The northern low-splitting zone is the easiest to interpret, given that
525 it corresponds fairly closely to the Nipigon Embayment, as well as to a
526 low-velocity anomaly and a magnetotelluric phase anomaly at lithospheric
527 depths. The NE was a locus of extensive magmatism in the Proterozoic,
528 roughly contemporaneous with the MCR (Hart and MacDonald, 2007). The
529 predominance of sills over dykes suggests that the NE was not extensional
530 at the time of emplacement, though north-trending extensional structures
531 in the NE predate MCR magmatic activity by ca. 200 Ma. The lack of a
532 gravity anomaly and the relatively small change in heat flow associated with
533 the NE indicate that the overall volume of intrusives in the crust is small
534 (Perry et al., 2004); however, the very low split times that we observe in the
535 NE, the negative P-velocity anomaly (Frederiksen et al., 2007, 2013a), and
536 the magnetotelluric anomaly at lithospheric depths (Ferguson et al., 2005)
537 all indicate that the NE overlies a significantly modified lithosphere.

538 It is difficult to explain the loss of lithospheric fabric in the NE by purely
539 deformational processes, particularly given the lack of evidence for extension.
540 Given the evidence for mantle plume involvement in MCR magmatism
541 (Nicholson and Shirey, 1990; Hutchinson et al., 1990; Hollings et al., 2012),
542 we propose that the NE lithospheric anomaly represents thermal/chemical
543 modification by a locus of plume impingement on the lithosphere, located
544 somewhat off-axis from the associated rifting. This displacement of the rifting
545 may indicate that the western Superior lithosphere was unusually resistant
546 to deformation at the time (as previously suggested by Frederiksen et
547 al., 2007, on other grounds).

548 The similar zone of weak splitting on the eastern shore of Lake Superior

549 lacks an associated magmatic feature, though it is very similar in size and
550 split time to that underlying the NE. The eastern weak-splitting zone does
551 straddle the southern end of the KSZ (Fig. 10), suggesting a relationship;
552 given the likely reactivation of KSZ structures by the MCR (Manson and
553 Halls, 1997), we propose the possibility that MCR-related melt or fluids
554 followed a KSZ-related zone of lithospheric weakness for ca. 150 km, but
555 failed to develop into an additional rift branch. The lack of a velocity anomaly
556 beneath the low-splitting region may indicate that the infiltrating material
557 was sufficient to reset the lithospheric fabric, but insufficient in volume to
558 greatly affect the bulk composition, perhaps as a result of being more distal
559 to the magma source than the NE.

560 The tomographic model of Frederiksen et al. (2013a) detected a linear
561 low-velocity feature in the lithosphere beneath western Minnesota and the
562 Dakotas, for which one suggested interpretation was a failed branch of the
563 MCR. Unlike the features detected by this study, the linear low-velocity
564 zone does not correspond to a zone of near-zero splitting (Frederiksen et
565 al., 2013b). The Minnesota/Dakotas feature, the NE feature, and the KSZ
566 feature all connect to the MCR at points at which the rift axis bends sharply
567 (Fig. 13, lower panel), where a triple junction would be expected. Given the
568 association between bends in the MCR and the features we have interpreted
569 as failed branches, the possibility that additional cryptic failed branches exist
570 at other sharp angles in the MCR axis would merit further investigation.

571 A major implication of our interpretation is that the Superior lithosphere
572 controlled the trajectory of the MCR. We interpret three features (the KSZ,
573 the NE and the failed branch) to represent failure of rifting to propagate

574 into the Superior Province. If the Superior lithosphere was particularly re-
 575 sistant to being rifted, then the path of least resistance for the rift axis would
 576 run along the Superior margins, as in fact it does (Fig. 1). Frederiksen et
 577 al. (2007) previously argued that the lithosphere of the Western Superior is
 578 unusually strong, while the eastern Superior is weaker, based on the appar-
 579 ent deflection of the Great Meteor hotspot track by lithospheric deformation
 580 (Eaton and Frederiksen, 2007) and the lack of deflection of the Nipigon Em-
 581 bayment feature. If the entire Superior lithosphere was a barrier to rifting ca.
 582 1.1 Ga, then the weakening of the eastern Superior must have occurred at a
 583 later date, and may have been related to the Great Meteor hotspot itself.

584 **6. Conclusions**

585 We have obtained shear-wave splits from teleseismic SKS and SKKS
 586 phases recorded at sixteen newly-deployed stations in the Superior Province
 587 north of Lake Superior, on the edge of the Mid-Continent Rift. This data set
 588 is supplemented by fifteen Earthscope Transportable Array stations south of
 589 the lake. Fast S polarization directions are consistently ENE-WSW to NE-
 590 SW, averaging 69° , while the split time varies strongly, ranging from 0.25
 591 to 1.48 s. Our data indicate that the lithosphere north of Lake Superior
 592 contains two highly localized domains of weak anisotropy, located adjacent
 593 to the Mid-Continent Rift axis as well as to the strongly anisotropic west-
 594 ern Superior Province. One closely corresponds to a known mantle anomaly
 595 beneath the Nipigon Embayment, a magmatic feature whose relationship to
 596 the MCR is not completely understood. The other lies immediately east of
 597 Lake Superior and is aligned with the southern extremity of the Kapuskasing

598 Structural Zone. We interpret these zones to represent resetting of lithospheric
599 fabric by MCR-related activity; along with an additional low-velocity feature
600 previously detected beneath Minnesota and North and South Dakota, we in-
601 terpret three offshoots of the MCR extending into the Superior lithosphere,
602 all of which failed to generate crustal rifting. Our interpretation suggests
603 that the lithosphere of the Superior Province was a barrier to rift propaga-
604 tion, and may have been indirectly responsible for the arcuate shape of the
605 MCR.

606 7. Acknowledgments

607 Instrumentation for the Canadian portion of the SPREE project was pro-
608 vided by Earthscope and the IRIS PASSCAL program, and deployed by the
609 kind permission of Ontario Parks and individual landowners. Algoma Univer-
610 sity and Lakehead University generously provided work space during deploy-
611 ment. Instrument deployers included Ghassan Aleqabi, Miguel Merino, Mulu
612 Serzu, Allen Wywrot, and Taras Zaporozan. The SPREE field project was
613 funded by NSF grant EAR-0952345; additional NSF grant PIs were Donna
614 Jurdy and Doug Wiens. Additional field work and support for Oyekunle Ola
615 were provided by NSERC grant RGPIN/261640-2008. Figures were prepared
616 using Generic Map Tools (GMT); absolute plate motion was calculated using
617 an online calculator provided by the University of Tokyo.

618 References

619 Amato, J.M., Boullion, A.O., Serna, A.M., Sanders, A.E., Farmer, G.L.,
620 Gherels, G.E., Wooden, J.L., 2008. Evolution of the Mazatzal province

621 and the timing of the Mazatzal orogeny: Insights from U-Pb geochronology
622 and geochemistry of igneous and metasedimentary rocks in southern New
623 Mexico. *Geol. Soc. Am. Bull.* 120, 328–346, doi:10.1130/B26200.1.

624 Auer, L., Boschi, L., Becker, T.W., 2014. Savani: A variable resolution whole-
625 mantle model of anisotropic shear velocity variations based on multiple
626 data sets. *J. Geophys. Res.* 119, 3006–3034, doi:10.1002/(ISSN)2169-9356.

627 Barruol, G., Helffrich, G., Vauchez, A., 1997. Shear wave splitting around
628 the northern Atlantic: frozen Pangaeon lithospheric anisotropy? *Tectono-*
629 *physics* 279, 135–148.

630 Bastow, I.D., Pilidou, S., Kendall, J.-M., Stuart, G.W., 2010. Melt-
631 induced seismic anisotropy and magma assisted rifting in Ethiopia: Ev-
632 idence from surface waves. *Geochem. Geophys. Geosyst.* 11, Q0AB05,
633 doi:10.1029/2010GC003036.

634 Bickford, M.E., Wooden, J.L., Bauer, R.L., 2006. SHRIMP study of zircons
635 from Early Archean rocks in the Minnesota River Valley: Implications for
636 the tectonic history of the Superior Province. *Geol. Soc Am. Bull.* 118,
637 94–108. doi:10.1130/B25741.1

638 Calvert, A., Ludden, J., 1999. Archean continental assembly in the south-
639 eastern Superior Province of Canada. *Tectonics* 18, 412–429.

640 Cannon, W.F., Green, A.G., Hutchinson, D.R., Lee, M., Milkereit, B.,
641 Behrendt, J.C., Halls, H.C., Green, J.C., Dickas, A.B., Morey, G.B., Sut-
642cliffe, R., Spencer, C., 1989. The North American Midcontinent Rift be-

643 neath Lake Superior from Glimpce seismic reflection profiling. *Tectonics*
644 8, 305–332, doi:10.1029/TC008i002p00305.

645 Card, K., 1990. A Review of the Superior Province of the Canadian Shield,
646 a Product of Archean Accretion. *Precambrian Research* 48, 9–156.

647 Darbyshire, F.A., Lebedev, S., 2009. Rayleigh wave phase-velocity het-
648 erogeneity and multilayered azimuthal anisotropy of the Superior
649 Craton, Ontario. *Geophys. J. Int.* 176, 215–234, doi:10.1111/j.1365-
650 246X.2008.03982.x.

651 Darbyshire, F.A., Eaton, D.W., Frederiksen, A., Ertolahti, L., 2007. New
652 insights into the lithosphere beneath the Superior Province from Rayleigh
653 wave dispersion and receiver function analysis. *Geophys. J. Int.* 169, 1043–
654 1068, doi:10.1111/j.1365-246X.2006.03259.x.

655 Davidson, A., 1998. An overview of Grenville Province geology, Canadian
656 Shield, in: *Geology of the Precambrian Superior and Grenville Provinces*
657 and Precambrian Fossils in North America, *Geology of North America*, vol.
658 C-1, edited by S. B. Lucas and M. R. St-Onge, pp. 205–270, Geological
659 Survey of Canada, Ottawa.

660 Eaton, D.W., Frederiksen, A., 2007. Seismic evidence for convection-
661 driven motion of the North American plate. *Nature* 446, 428–431,
662 doi:10.1038/nature05675.

663 Eaton, D.W., Frederiksen, A., Miong, S.K., 2004. Shear-wave splitting obser-
664 vations in the lower Great Lakes region: Evidence for regional anisotropic

domains and keel-modified asthenospheric flow. *Geophys. Res. Lett.* 31,
L07610, doi:10.1029/2004GL019438.

Ferguson, I., Craven, J., Kurtz, R., Boerner, D., Bailey, R., Wu, X., Orellana,
M., Spratt, J.E., Wennberg, G., Norton, A., 2005. Geoelectric response of
Archean lithosphere in the western Superior Province, central Canada.
Phys. Earth Planet. Int. 150, 123–143, doi:10.1016/j.pepi.2004.08.025.

Ferré, E.C., Gébelin, A., Conder, J.A., Christensen, N., Wood, J.D.,
Teyssier, C., 2014. Seismic anisotropy of the Archean crust in the Min-
nesota River Valley, Superior Province. *Geophys. Res. Lett.* 41, 1514–1522,
doi:10.1002/2014GL060504.

Fouch, M., Fischer, K., Parmentier, E., Wysession, M., Clarke, T., 2000.
Shear wave splitting, continental keels, and patterns of mantle flow. *J.*
Geophys. Res. 105, 6255–6275.

Frederiksen, A.W., I.J. Ferguson, D. Eaton, S.-K. Miong and Gowan, E.,
2006. Mantle Fabric at Multiple Scales Across an Archean-Proterozoic
Boundary, Eastern Ontario, Canada. *Phys. Earth Plan. Int.* 158, 240–263.

Frederiksen, A., Miong, S.K., Darbyshire, F.A., Eaton, D.W., Rondenay, S.,
Sol, S., 2007. Lithospheric variations across the Superior Province, Ontario,
Canada: Evidence from tomography and shear wave splitting. *J. Geophys.*
Res. 112, B07318, doi:10.1029/2006JB004861,2007.

Frederiksen, A., Bollmann, T., Darbyshire, F., Lee, S., 2013a. Modification
of Continental lithosphere by tectonic processes: a tomographic image of
central North America. *J. Geophys. Res.* 118, 1–16, doi:10.1002/jgrb.50060.

688 Frederiksen, A., Deniset, I., Ola, O., Toni, D., 2013b. Lithospheric fabric vari-
689 ations in central North America: Influence of rifting and Archean tectonic
690 styles. *Geophys. Res. Lett.* 40, 1–5, doi:10.1002/grl.50879.

691 Garnero, E.J., Maupin, V., Lay, T., Fouch, M.J., 2004. Variable azimuthal
692 anisotropy in Earth’s lowermost mantle. *Science* 306, 259–261.

693 Garnero, E.J., McNamara, A.K., 2008. Structure and Dynamics of Earth’s
694 Lower Mantle. *Science* 320, 626–628, doi:10.1126/science.1148028.

695 Gripp, A.E., Gordon, R.G., 1990. Current plate velocities relative to the
696 hotspots incorporating the NUVEL-1 global plate motion model. *Geophys.*
697 *Res. Lett.* 17, 1109–1112.

698 Gueydan, F., Morency, C., Brun, J.-P., 2008. Continental rifting as
699 a function of lithosphere mantle strength. *Tectonophysics* 460, 83–93,
700 doi:10.1016/j.tecto.2008.08.012.

701 Hart, T.R., MacDonald, C.A., 2007. Proterozoic and Archean geology of the
702 Nipigon Embayment: implications for emplacement of the Mesoproterozoic
703 Nipigon diabase sills and mafic to ultramafic intrusions. *Can. J. Earth Sci.*
704 44, 1021–1040, doi:10.1139/EO7-026.

705 He, X., Long, M.D., 2011. Lowermost mantle anisotropy beneath the
706 northwestern Pacific: Evidence from PcS, ScS, SKS, and SKKS phases.
707 *Geochem. Geophys. Geosyst.*, 12, Q12012, doi:10.1029/2011GC003779.

708 Hollings, P., Hart, T., Richardson, A., MacDonald, C.A., 2007. Geochem-
709 istry of the Mesoproterozoic intrusive rocks of the Nipigon Embayment,

710 northwestern Ontario: evaluating the earliest phases of rift development.
711 Can. J. Earth. Sci. 44, 1087–1110, doi:10.1139/E06-127.

712 Hollings, P., Smyk, M., Cousens, B., 2012. The radiogenic isotope character-
713 istics of dikes and sills associated with the Mesoproterozoic Midcontinent
714 Rift near Thunder Bay, Ontario, Canada. Precambrian Research 214-215,
715 269–279, doi:10.1016/j.precamres.2011.11.006.

716 Hollings, P., Smyk, M., Heaman, L.M., Halls, H., 2010. The geochemistry,
717 geochronology and paleomagnetism of dikes and sills associated with the
718 Mesoproterozoic Midcontinent Rift near Thunder Bay, Ontario, Canada.
719 Precambrian Research 183, 553–571, doi:10.1016/j.precamres.2010.01.012.

720 Huisman, R., Beaumont, C., 2011. Depth-dependent extension, two-stage
721 breakup and cratonic underplating at rifted margins. Nature 473, 74–78,
722 doi:10.1038/nature09988.

723 Hutchinson, D.R., White, R.S., Cannon, W.F., Schulz, K.J., 1990. Ke-
724 weenaw hot spot: Geophysical evidence for a 1.1 Ga mantle plume be-
725 neath the Midcontinent Rift System. J. Geophys. Res. 95, 10869–10884,
726 doi:10.1029/JB095iB07p10869.

727 Kay, I., Sol, S., Kendall, J., Thomson, C., White, D.J., Asudeh, I., Roberts,
728 B., Francis, D., 1999. Shear wave splitting observations in the Archean
729 craton of Western Superior. Geophys. Res. Lett. 26, 2669–2672.

730 Kong, F., Gao, S.S., Liu, K.H., 2015. Applicability of the Multiple-Event
731 Stacking Technique for Shear-Wave Splitting Analysis. Bull. Seis. Soc. Am.
732 105, 3156–3166, doi:10.1785/0120150078.

- 733 Long, M.D., 2009. Complex anisotropy in D'' beneath the eastern Pacific
734 from SKSSKKS splitting discrepancies. *Earth Plan. Sci. Lett.* 283, 181–
735 189, doi:10.1016/j.epsl.2009.04.019.
- 736 Manson, M.L., Halls, H.C., 1997. Proterozoic reactivation of the southern
737 Superior Province and its role in the evolution of the Midcontinent rift.
738 *Can. J. Earth Sci.*, 34, 562–575, doi:10.1139/e17-045.
- 739 Mariano, J., Hinze, W., 1994. Structural interpretation of the Midcontinent
740 Rift in eastern Lake Superior from seismic reflection and potential-field
741 studies. *Can. J. Earth Sci.* 31, 619–628.
- 742 Merino, M., Keller, G.R., Stein, S., Stein, C., 2013. Variations in Mid-
743 Continent Rift magma volumes consistent with microplate evolution. *Geo-
744 phys. Res. Lett.* 40, 1513–1516, doi:10.1002/grl.50295.
- 745 Musacchio, G., White, D.J., Asudeh, I., Thomson, C., 2004. Lithospheric
746 structure and composition of the Archean western Superior Province from
747 seismic refraction/wide-angle reflection and gravity modeling. *J. Geophys.
748 Res* 109, B03304, doi:10.1029/2003JB002427.
- 749 Nicolas, A., Christensen, N.I., 1987. Formation of anisotropy in upper
750 mantle peridotites—a review. In *Composition, Structure and Dynam-
751 ics of the Lithosphere-Asthenosphere System*, K. Fuchs and C. Froide-
752 vaux, eds., American Geophysical Union, Washington, D.C., USA,
753 doi:10.1029/GD016p0111.
- 754 Nicholson, S.W., Shirey, S.B., 1990. Midcontinent rift volcanism in the Lake

755 Superior Region: Sr, Nd, and Pb isotopic evidence for a mantle plume
756 origin. *J. Geophys. Res.* 95, 10851–10868, doi:10.1029/JB095iB07p10851.

757 North American Magnetic Anomaly Group, 2002. Digital data grids for the
758 magnetic anomaly map of North America. U.S. Geological Survey Open
759 File Report 02–414, U.S. Department of the Interior, Washington, D.C.,
760 USA.

761 Nowacki, A., Wookey, J., Kendall, J.-M., 2010. Deformation of the
762 lowermost mantle from seismic anisotropy. *Nature* 467, 1091–1094,
763 doi:10.1038/nature09507.

764 Ojakangas, R., Morey, G., Green, J., 2001. The Mesoproterozoic Midcon-
765 tinent Rift System, Lake Superior Region, USA. *Sediment. Geol.*, 141,
766 421–442.

767 Panning, M., Romanowicz, B., 2006. A three-dimensional radially anisotropic
768 model of shear velocity in the whole mantle. *Geophys. J. Int.* 167, 361–379,
769 doi:10.1111/j.1365-246X.2006.03100.x.

770 Percival, J., West, G., 1994. The Kapuskasing Uplift – a Geological and
771 Geophysical Synthesis. *Can. J. Earth Sci.*, 31, 1256–1286.

772 Percival, J.A., Sanborn-Barrie, M., Skulski, T., Stott, G.M., Helmstaedt,
773 H., White, D.J., 2006. Tectonic evolution of the western Superior Province
774 from NATMAP and Lithoprobe studies. *Can. J. Earth. Sci.* 43, 1085–1117,
775 doi:10.1139/e06-062.

776 Perry, H., Jaupart, C., Mareschal, J.C., 2004. Heat flow in the Nipigon arm of

777 the Keweenawan rift, northwestern Ontario, Canada. *Geophys. Res. Lett.*
778 31, L15607, doi:10.1029/2004GL020159.

779 Rondenay, S., Bostock, M.G., Hearn, T., White, D.J., Wu, H., Senechal, G.,
780 Ji, S., Mareschal, M., 2000. Teleseismic studies of the lithosphere below
781 the Abitibi-Grenville Lithoprobe transect. *Can. J. Earth Sci.* 37, 415–426.

782 Shen, W., Ritzwoller, M.H., Schulte-Pelkum, V., 2013. Crustal and upper-
783 most mantle structure in the central U.S. encompassing the Midcontinent
784 Rift. *Journal of Geophysical Research* 118, doi:10.1002/jgrb.50321.

785 Samson, C., West, G., 1994. Detailed basin structure and tectonic evolution
786 of the Midcontinent Rift System in eastern Lake Superior from reprocessing
787 of GLIMPCE deep reflection seismic data. *Can. J. Earth Sci.* 31, 629–639.

788 Sexton, J., Henson, H., Jr, 1994. Interpretation of seismic reflection and
789 gravity profile data in western Lake Superior. *Can. J. Earth Sci.* 31, 652–
790 660.

791 Silver, P. G., 1996. Seismic anisotropy beneath the continents: Probing the
792 depths of geology, *Annu. Rev. Earth Planet. Sci.*, 24, 385–432.

793 Silver, P. G., Chan, W.W., 1991. Shear-wave splitting and subcontinental
794 mantle deformation, *J. Geophys. Res.*, 96, 16,429–16,454.

795 Silver, P., Kaneshima, S., 1993. Constraints on Mantle Anisotropy Beneath
796 Precambrian North America From a Transportable Teleseismic Experi-
797 ment. *Geophys. Res. Lett.* 20, 1127–1130.

798 Sol, S., Thomson, C., Kendall, J., White, D.J., VanDecar, J., Asudeh, I.,
799 2002. Seismic tomographic images of the cratonic upper mantle beneath
800 the Western Superior Province of the Canadian Shield – a remnant Archean
801 slab? *Phys. Earth. Planet. Int.* 134, 53–69.

802 Stott, G.M., 2011. A revised terrane subdivision of the Superior Province
803 in Ontario. Ontario Geological Survey, Miscellaneous Release–Data 278,
804 ISBN 978-1-4435-6399-4.

805 Stein, S., van der Lee, S., Jurdy, D., Stein, C., Wiens, D., Wyses-
806 sion, M., Revenaugh, J., Frederiksen, A., Darbyshire, F., Bollmann, T.,
807 Lodewyk, J., Wolin, E., Merino, M., Tekverk, K., 2011. Learning from
808 failure: The SPREE Mid-Continent Rift Experiment. *GSA Today* 21, 5–7.
809 doi:10.1130/G120A.1.

810 Stein, C.A., Stein, S., Merino, M., Keller, G.R., 2014. Was the Mid-Continent
811 Rift part of a successful seafloor-spreading episode? *Geophys. Res. Lett.*
812 41, doi:10.1002/2013GL059176.

813 Stein, C.A., Kley, J., Stein, S., Hindle, D., Keller, G.R., 2015. North Amer-
814 ica’s Midcontinent Rift: When rift met LIP. *Geosphere* 11, 1607–1616,
815 doi:10.1130/GES01183.1.

816 Tanner, J.G. and Members of the Committee for the Gravity Anomaly Map
817 of North America, 1988. Gravity anomaly map of North America. *The*
818 *Leading Edge* 7, 15–18.

819 Tommasi, A., Vauchez, A., 2001. Continental rifting parallel to ancient
820 collisional belts: an effect of the mechanical anisotropy of the litho-

821 spheric mantle. *Earth Planet. Sci. Lett.* 185, 199–210, doi:10.1016/S0012-
822 821X(00)00350-2.

823 Van Schmus, W., Hinze, W., 1985. The midcontinent rift system. *Ann. Rev.*
824 *Earth Planet. Sci.* 13, 345–383.

825 Vauchez, A., Tommasi, A., Barruol, G., Maumus, J., 2000. Upper man-
826 tle deformation and seismic anisotropy in continental rifts. *Physics and*
827 *Chemistry of the Earth, Part A: Solid Earth and Geodesy* 25, 111–117.

828 White, D.J., Musacchio, G., Helmstaedt, H.H., Harrap, R.M., Thurston,
829 P.C., van der Velden, A., Hall, K., 2003. Images of a lower-crustal oceanic
830 slab: Direct evidence for tectonic accretion in the Archean western Supe-
831 rior province. *Geol.* 31, 997–1000, doi:10.1130/G20014.1.

832 White, R., 1997. Mantle temperature and lithospheric thinning beneath the
833 Midcontinent rift system: Evidence from magmatism and subsidence. *Can.*
834 *J. Earth Sci.* 34, 464–475.

835 Whitmeyer, S.J. and Karlstrom, K.E., 2007. Tectonic model for the
836 Proterozoic growth of North America. *Geosphere* 3, 220–259, doi:
837 10.1130/GES00055.1.

838 Wolfe, C.J., Silver, P.G., 1998. Seismic anisotropy of oceanic upper mantle:
839 Shear wave splitting methodologies and observations. *J. Geophys. Res* 103,
840 749–771.

841 Wolin, E., Van der Lee, S., Bollmann, T.A., Wysession, M.E., Stein, S.,
842 Wiens, D.A., Darbyshire, F.A., Frederiksen, A.W., Revenaugh, J., 2015.

- 843 Seasonal and diurnal variations in long-period noise at SPREE stations:
844 the influence of soil characteristics on shallow stations' performance. Sub-
845 mitted to Bull. Seis. Soc. Am.
- 846 Yang, B.B., Gao, S.S., Liu, K.H., Elsheikh, A.A., Lemnifi, A.A., Refayee,
847 H.A., Yu, Y., 2014. Seismic anisotropy and mantle flow beneath the north-
848 ern Great Plains of North America. J Geophys. Res. 119, 1971–1985,
849 doi:10.1002/2013JB010561.
- 850 Yuan, H., Romanowicz, B., 2010. Lithospheric layering in the North Ameri-
851 can craton. Nature 466, 1063–1068, doi:10.1038/nature09332.

Figure 1: Geologic setting of this study, overlain on a map of magnetic anomalies (North American Magnetic Anomaly Group, 2002). MCR: Mid-Continent Rift, MRVT: Minnesota River Valley Terrane, NE: Nipigon Embayment, KSZ: Kapuskasing Structural Zone, GMHST: Great Meteor Hotspot Track. Solid black lines are tectonic province boundaries from Whitmeyer and Karlstrom (2007); dashed lines are boundaries of interest within the Superior Province (MRVT boundary from Bickford et al. (2006); NE boundary from the National Atlas of Canada, <http://atlas.gc.ca/site/english/maps/geology.html>). Hotspot track is from Eaton and Frederiksen (2007). Shaded regions are clastic (lighter) and volcanic (darker) rocks associated with the MCR, from Ojakangas et al. (2001). Inset shows location of study within North America (box).

Figure 2: Seismic instrumentation in the study area, overlain on Bouguer gravity (Tanner et al., 1988). Black symbols show sites used in this study. Inverted triangles: Superior Province Rifting Earthscope Experiment (SPREE) instruments; upright triangles: Earthscope Transportable Array instruments; circles: Canadian National Seismograph Network instruments.

Figure 3: a) Locations of events used in the shear-wave splitting analysis and judged to be of quality 3 (out of 5) or greater. Larger circles represent higher-quality events; the star indicates the approximate centre of the study area. b) Histogram of the events in (a), binned by the associated polarization direction.

Figure 4: Average ratio of transverse to radial energy at each station, for all events of quality ≥ 3 . This ratio is a direct measure of the effect of anisotropy on the SKS and SKKS traces.

Figure 5: Splitting measurement for a sample high-quality SKS pulse (quality level 5) recorded at station SC05. The event occurred on 2011/12/14 at 05:04:59 in Papua New Guinea. Top two panels show unprocessed data (band-pass filtered between 0.02 and 0.2 Hz): radial and transverse traces (left) and particle motion (right). Middle two panels show the grid search over split time and fast polarization direction: second-eigenvalue misfit surface (left, darker values are lower misfit) and best-fit values with 95% confidence contour from F-test (both eigenvalue and transverse-energy results are shown). Bottom two panels show the result of correcting the traces using the eigenvalue solution: recovered radial and transverse traces (left) and recovered particle motion (right).

Figure 6: Recovered single-event splitting parameters for all events of quality ≥ 3 recorded at station SC05, plotted against the incident polarization angle (the back-azimuth modulo 180°). Black circles are SKS measurements, grey circles are SKKS measurements; circle size indicates measurement quality (from 3 through 5). Horizontal dashed lines are the final fast S polarization direction and split time values obtained at SC05; vertical dashed lines indicate the expected null directions given the obtained fast polarization.

Figure 7: Eigenvalue misfit surfaces for all events of quality ≥ 3 recorded at station SC05, stacked in 10° incident-polarization swaths. The white circle is the composite solution for SC05. Note how it is consistent with all of the swath misfit surfaces.

Figure 8: Top: final misfit-surface stack for events of quality ≥ 3 at station SC05, formed by stacking all of the swath stacks in Fig. 7 with equal weight. Bottom left: best-fit solution and 95% error contour, compared to the individual-event solutions (dots). Bottom right: scaled misfit of each individual trace to the best-fit solution.

Figure 9: a) Map of splitting measurements in the study area. Arrow orientation is fast S polarization direction, and arrow length is proportional to split time. Red arrows are the results of this study; black arrows are previously published studies using the same methodology (Frederiksen et al., 2006, 2007, 2013b), and grey arrows are other published studies (Silver and Kaneshima, 1993; Barruol et al., 1997; Kay et al., 1999; Rondenay et al., 2000; Eaton et al., 2004; Ferré et al., 2014). White circles indicate split times > 1 s. Blue lines are tectonic boundaries, as in Fig. 1, with the addition of Superior subprovince boundaries within Ontario (thin dashed lines; Stott, 2011). The green arrow indicates the direction of absolute plate motion, from model HS3-NUVEL-1A (Gripp and Gordon, 1990). b) Error bars on splitting measurements. The width of the wedge indicates the range of angles included in the error bar; the length of the black wedge represents the minimum possible split time, while the red wedge represents the maximum.

Figure 10: Contour map of fast S polarization directions (in degrees) across the study area. Large circles: this study; medium circles: published studies with the same methodology; small circles: other studies. Grey lines are tectonic boundaries (see Fig. 1).

Figure 11: Contour map of split times (in seconds) across the study area. Large circles: this study; medium circles: published studies with the same methodology; small circles: other studies. Grey lines are tectonic boundaries (see Fig. 1).

Figure 12: Fast polarization directions (top panel) and split times (bottom panel) for individual events, averaged over all stations. The fast polarization directions show coherent spatial variations indicating a deep-mantle influence.

Figure 13: Contours of split time in seconds (black) overlain on tomographic P velocities at 150 km depth (top) and 250 km depth (bottom) from the teleseismic model of Frederiksen et al. (2013a); black dots are locations of stations used to obtain the contours. Light grey regions indicate that the tomographic model lacks ray coverage. Dark grey lines are tectonic boundaries (see Fig. 1). The 250 km slice is overlain with interpretation (see text for details).

Figure 14: a) Synthetic directionally-averaged split times and transverse/radial (T/R) energy ratios for a two-layer model in which the upper layer has a split time of 0.4 s and the lower layer has a split time of 0.6 s, for a range of angles between the two layers' fast polarization directions. b) Split times plotted against T/R ratios for real data.

Table 1: Final splitting measurements at all stations. Stations with codes starting with “SC” are SPREE stations, while the others are TA stations. ϕ is the fast S polarization direction and δt is the split time. The last three columns indicate the number of SKS and SKKS traces contributing to the final result, and the quality threshold used.

Station	Lat.	Lon.	ϕ ($^\circ$)	δt (s)	SKS	SKKS	Quality
C39A	47.817	-90.129	66 ± 12	0.55 ± 0.18	13	17	4
C40A	47.915	-89.151	58 ± 10	0.55 ± 0.20	16	8	4
D37A	47.160	-92.430	40 ± 19	0.58 ± 0.25	31	21	4
D41A	47.061	-88.566	36 ± 19	0.33 ± 0.20	14	11	4
D46A	46.890	-84.040	107 ± 41	0.28 ± 0.33	9	5	3
E38A	46.606	-91.554	58 ± 16	0.78 ± 0.30	23	10	4
E39A	46.378	-90.556	50 ± 10	0.60 ± 0.20	21	10	4
E43A	46.376	-86.995	61 ± 8	0.88 ± 0.25	21	15	4
E44A	46.620	-85.921	71 ± 11	0.58 ± 0.25	20	13	3
H41A	44.616	-89.653	65 ± 12	0.53 ± 0.23	18	10	4
H43A	44.470	-87.770	79 ± 12	0.63 ± 0.33	19	11	3
I40A	43.892	-90.618	86 ± 16	0.48 ± 0.25	20	10	4
K42A	42.779	-89.346	86 ± 33	0.25 ± 0.20	15	10	3
L41A	42.075	-90.498	66 ± 16	0.38 ± 0.23	21	8	4
L44A	42.178	-87.912	68 ± 17	0.45 ± 0.28	18	11	3
SC01	49.250	-90.568	74 ± 11	0.85 ± 0.33	18	16	3
SC02	49.895	-91.141	66 ± 8	1.48 ± 0.35	22	8	4
SC03	50.254	-89.094	68 ± 19	0.78 ± 0.38	23	11	3
SC04	49.624	-89.675	75 ± 10	0.90 ± 0.25	16	13	4
SC05	48.280	-89.443	64 ± 10	0.85 ± 0.28	19	9	4
SC06	48.905	-88.446	67 ± 31	0.43 ± 0.35	8	7	3
SC07	49.651	-88.088	67 ± 28	0.25 ± 0.20	14	8	4
SC08	48.888	-87.357	73 ± 17	0.45 ± 0.23	15	12	3
SC09	49.740	-86.755	80 ± 12	0.75 ± 0.30	14	11	3
SC10	49.753	-83.817	74 ± 13	0.93 ± 0.28	12	13	3
SC11	49.084	-85.856	89 ± 11	0.73 ± 0.25	20	12	3
SC12	49.189	-84.763	61 ± 12	0.83 ± 0.25	18	9	4
SC13	48.613	-85.258	70 ± 8	0.90 ± 0.23	10	9	4
SC14	47.723	-84.814	81 ± 23	0.38 ± 0.25	16	13	4
SC15	47.861	-83.354	62 ± 25	0.40 ± 0.33	22	17	3
SC16	47.305	-84.588	80 ± 17	0.40 ± 0.23	20	10	4