Shear Wave Splitting Beneath Eastern North American Continent: Evidence for a Multi-layered and Laterally Variable Anisotropic Structure

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Key Points:

- We made consistent shear wave splitting measurements from 24 earthquakes along a 1300-km-long transect in eastern North America
- Fast polarizations vary systematically with backazimuth, while delay times increase from the center of the continent to the coast
- We interpret the anisotropic structure to be multi-layered, and the lateral change in delay time marks the edge of the craton at depth
Abstract

Eastern North America records a tectonic history of over 3Ga in duration. Much of this record is preserved within the lithosphere, and may be unraveled by detailed studies of its interior structure. Past episodes of tectonic activity likely left their imprints in the form of anisotropy-forming rock fabric presently preserved in lithosphere of the continent. We perform shear wave splitting measurements using observations of core refracted waves collected from a ~1300 km long array extending from James Bay in Quebec to the Fundy basin in Maine, with lateral spacing of 10 ~ 100 km between instruments. Close spacing of instruments helps us associate anisotropic properties with geological boundaries. We find that the fast polarizations concentrate between N60°E and N90°E with an average of N80°E and change systematically with backazimuth. In addition, we observe a lateral increase in delay time from 0.56 ± 0.25s at the NW end of the array to 0.90 ± 0.41s at the SE end. The location of lateral change in delay time does not match geological boundaries on the surface but seems to match the geophysical boundary at depth of 160 km. We interpret this boundary in splitting values to be the edge of cratonic lithosphere at depth. Our observations suggest that the anisotropic structure beneath our study area is complex and possibly both multi-layered and laterally variable.

1 Introduction

Eastern North America contains some of the thickest lithosphere on Earth within the cratonic core of the continent, as well as the record of the two most recent Wilson cycles corresponding to the assembly and breakup of the supercontinents Rodinia and Pangea. Thus, it is an ideal place to study the formation and evolution of continents through geological time.

Seismic anisotropy, a directional dependence of seismic velocity, is a property that can be used as a proxy for deformation in the deep interior of the Earth, and thus can provide constraints on the processes of continental formation and evolution. Olivine is the most abundant mineral in the upper mantle, and it is intrinsically anisotropic (Christensen, 1984; Kumazawa & Anderson, 1969; Ribe, 1989). The orientation of its crystals will align with the strain direction when it is deformed (Zhang & Karato, 1995). Both lithosphere and asthenosphere can contribute to seismic anisotropy in the upper mantle (Silver, 1996; Vinnik et al., 1989). Thus, by developing constraints on mantle anisotropy, we can infer both the past
deformation reflected by rock fabric frozen into the lithosphere, as well as the present deformation due to shearing of the asthenosphere by plate motion or mantle flow (Forte et al., 2010; Long & Becker, 2010), or a combination of both.

Our study area (Figure 1(a)) includes all major tectonic units and boundaries of eastern North America. The Superior Province is an Archean craton which has been stable since 2.8 – 2.5 Ga (Card, 1990; Percival, 2007). Tectonic subprovinces in the Superior craton are trending nearly east to west and are surrounded by early Proterozoic orogens (Calvert & Ludden, 1999; Hoffman, 1988). The boundary between the Archean and the Proterozoic terranes, the Grenville Front was formed during the assembly of the supercontinent Rodinia between ~1.19 and 0.98 Ga (Hynes & Rivers, 2010; Rivers, 1997; Whitmeyer & Karlstrom, 2007). The Grenville Province is the oldest post-Archean terrane. It consists of rocks with high degree of metamorphism and includes both old Archean material and relatively younger volcanic arc material (Rivers, 2015). The Appalachians were formed by a sequence of orogenies: Taconic, Acadian-Neoacadian, and Alleghanian (Hatcher, 2010) that together resulted in the assembly of the supercontinent Pangea. Multiple tectonic boundaries can be recognized within the Appalachian Orogen, most of which separate terranes with distinct tectonic histories, such as Meguma, Gander, and Avalon (e.g. Hibbard et al., 2007). The Appalachian Front marks the boundary between the Appalachian Orogen and the Grenville Province on the surface. However, Grenville-aged rocks likely extend east of the Appalachian Front at depth, so the surface geology and deep crustal lithology of Appalachian Orogen likely do not match (Hynes & Rivers, 2010).

Although numerous previous studies have explored seismic anisotropy in the same or neighboring regions using shear wave splitting (Barruol et al., 1997; Darbyshire et al., 2015; Eaton et al., 2004; Fouch et al., 2000; Gilligan et al., 2016; Levin et al., 1999; Long et al., 2016) and surface wave inversions (Darbyshire & Lebedev, 2009; Yuan & Romanowicz, 2010), studies of anisotropy that cover the entire region (Darbyshire et al., 2015; Yuan & Romanowicz, 2010) typically lack the lateral sampling necessary to associate anisotropic structure with elements of the regional tectonic structures. In other cases, a number of studies that did have good lateral sampling (Long et al., 2016; Yang et al., 2017) were limited to parts of the region. This makes comparison between different tectonic units difficult.
In this study, we perform shear wave splitting measurements using observations of core-refracted shear waves on a 1300 km long array crossing the eastern part of the North American continent, from James Bay to the Fundy basin (Figure 1(a)). Inter-station spacing between 10 and 100 km along this continent-scale array allows us to relate lateral variations in observed properties to specific tectonic boundaries and geological units. Having operated our instruments for 2-3 years, we have enough observations to explore likely vertical variations in anisotropic structure.

2 Methods and data

Shear wave splitting (Long & Silver, 2009; Silver & Chan, 1991; Vinnik et al., 1989) is one of the most widely adopted methods for the study of seismic anisotropy in the Earth. When a shear wave propagates through an anisotropic medium, it will split into two orthogonally polarized components travelling at different velocities. As they propagate, a delay time between the fast and the slow components will accumulate, in proportion to the length of their path, as well as the strength of anisotropy of the medium. Under the subcontinental upper mantle conditions, the fast polarization is expected to align with the direction of deformation (Long & Silver, 2009; Park & Levin, 2002). Core-refracted phases including SKS, SKKS, SKIKS and PKIKS (called XKS hereafter) are commonly used in this method to avoid source-side contamination of the signal (Savage, 1999). In addition, since teleseismic waves have nearly vertical travel paths beneath the receiver, shear wave splitting provides very good lateral resolution. By analyzing the estimated splitting parameters, we are able to examine lateral changes in anisotropic properties and make comparisons between anisotropic structures at depth and tectonic divisions evident on the surface.

We adopt the SplitLab software and choose Rotation-Correlation (called RC hereafter) method (Figure 2) to estimate the shear wave splitting parameters (fast polarization and delay time). This method searches for the coordinate system where two components of the split shear wave are most similar by rotating horizontal seismograms into a sequence of test coordinate systems and comparing their pulse shapes via cross-correlation (Wüstefeld & Bokelmann, 2007). To assess the reliability of the splitting parameters we also estimate them using a Minimum Transverse Energy (called SC hereafter) method (Silver & Chan, 1991), which seeks to minimize the power of the horizontal (SH) component in the observed shear wave. We established a set of criteria concerning the splitting parameters and the quality of
measurements by taking into consideration measurements from both methods. A detailed description of criteria of analysis in this study (Text S1) along with examples of measurements of different qualities (Figure S1) is provided in the supplement.

In this study, we examined seismograms recorded at 64 seismic stations, including both permanent and temporary observatories (Canadian National Seismograph Network; POLARIS stations in Quebec (http://ds.iris.edu/mda/PO); New England Seismic Network; USArray Transportable Array; temporary network deployed in the framework of the EarthScope project). Our starting data set included 662 station-event records (some records contain more than one XKS phase) from 24 different earthquakes from 2012 to 2015 (Mw > 6.8) at distances 90° to 150° away from our region (Figure 1(b)). A table listing event information is in the Supplement (Table S1). Of these, XKS phases from 13 events (yellow in Figure 1(b)) were observed over the entire 1300 km length of our array, while the remaining 11 events yielded XKS phases for parts of the array only.

Visual inspection for the clarity of signals, low SNR traces, gaps in records or no data being recorded at all reduced the dataset to 900 records of individual XKS phases that were subsequently analyzed. To estimate splitting parameters using SplitLab, we choose time windows for the target XKS phases manually. Bandpass filters were applied depending on the signal to noise ratio (SNR) of the raw data. In this study, the lower corner of the filter was mostly 0.01 Hz or 0.02 Hz, and the upper corners varied from as low as 0.05 Hz to no filter at all. Measurements were given qualities ranked by the analyst as “good”, “fair” or “poor” on the basis of criteria such as the SNR, and the stability of measurements when filter settings or time windows were changed. For interpretation in this paper, we included the measurements of all qualities. We verified that “fair” and “poor” measurements do not show systematic differences from “good” measurements (Figure S2 and S3)). In the dataset consisting of 900 records, 639 yielded observations of splitting (Figure 2(i)) and 261 were designated as NULLs (Figure 2(ii)).

Theoretically, NULL measurements can be assigned when 1) we have an observation of a rectilinear particle motion without correction for anisotropy; 2) both RC and SC methods yield nearly zero delay times; or 3) there is no energy on transverse component before correction for anisotropy. In this study, since we combine measurements made using two methods, we adopt one more criterion in addition to the three mentioned above. According to
the synthetic test by Wüstefeld and Bokelmann (2007), characteristic differences between RC and SC methods can be identified when the backazimuths are near the true fast polarization. In that case, RC method tends to yield fast polarizations with ±45° from true fast polarization, and the delay times are close to 0. However, SC method yields a large scatter of fast polarization values, and delay values close to the maximum of the search range (3 s in our study). Levin et al. (2007) documented a similar disparity of measurement results for synthetic seismograms simulated in models with very small amount of anisotropy. Thus, in this study NULL measurements can also be identified when measurements using the two different methods yield very different answers. When applied to real data, this criterion should be considered together with three other criteria so that NULL measurements can be distinguished from those where the splitting parameters can be measured but the quality is poor.

3 Results
3.1 Shear wave splitting values

We find that evidence of splitting in XKS records can be observed at all stations along the entire length of our array. Figure 3(a) shows all shear wave splitting measurements obtained from 24 earthquakes. Fast polarizations measured along the array are similar, though not completely the same, and generally fall into a range from N50°E to N120°E. In Figure 3(c), we plot the unweighted mean fast polarization at each station. We treat the fast polarizations as scalars varying from N0°E to N180°E and calculate the arithmetic average of all fast polarizations estimated at each station using RC method. This procedure excludes NULL measurements from the average. Given the uneven distribution of sources with backazimuth (mainly West, North and East), an unweighted average of all values at a station makes implicit assumptions that directional variability is not systematic, that there is only one set of true splitting parameters, and that the scatter in the values reflects noise in the measurements. Averaged fast polarizations do not vary significantly from station to station (Figure 3(c)), however systematic differences may be seen between sections of the linear array. Stations in the NW half of the array have mean fast polarizations generally close to E-W. This is most clear at the stations in the Superior Province. For stations in the SE half of the array, fast polarizations are close to N80°E.
At each station, fast polarizations measured from different earthquakes are not consistent with each other, but vary from event to event. To better show that fast polarizations at each station will change according to backazimuths of incoming rays, we plot splitting patterns for individual earthquakes in Figures 4(a)-(d). Splitting patterns for all 24 earthquakes can be found in Figure S2. We select 4 XKS phases that come from 4 earthquakes of different directions and that were observed at most of the stations along the array. In Figure 4(a), event 2013.321 has a backazimuth of N167°E, and the fast polarizations at stations along the transect are ~N125°E. In Figure 4(b), event 2014.202 has a backazimuth of N269°E. Fast polarizations measured from this event are not very consistent: most stations yield fast polarizations N30°E~N70°E with a few others yield ~N-45°E. Figure 4(c) shows measurements from event 2015.150 with a backazimuth of N330°E. Most fast polarizations are ~N100°E. In Figure 4(d), fast polarizations measured from event 2014.043 with a backazimuth of N21°E are ~N65°E. We thus find a systematic change in fast polarizations with backazimuth. Specifically, shear waves arriving from the south and the NW show similar splitting patterns, while arrivals from west and NE produce significantly different ones.

A significant change in the values of the splitting delay times can be observed along the array. We find a lateral increase in delay times from the NW end of the array to the SE end. This can be seen in the individual splitting values measured from all events (Figures 3(a)-(b)). In Figure 3(a), we find that delay times measured at the stations in the NW part of the array are generally close to ~0.5 s. Moving toward to the SE along the line, delay times start to increase to ~1 s, then decrease to ~0.5 s in a relatively narrow area in the Grenville Province north of the Appalachian Front. Stations in the SE section of the array tend to have larger delay times of over 1 s, except for the circled area in Figure 3(b) where delay times are less than 1 s but still larger than those measured in the NW half of the array. It is also noticeable that the delay times measured at stations close to the coast are much larger than along the rest of the array, reaching values of 1.5 s.

To examine the lateral change in delay times in detail, we average the measurements at individual stations, as shown in Figure 3(c), and estimate their standard deviations (Figure 5). As was done previously with fast polarizations, we average all delays irrespective of the direction of wave propagation, making the implicit assumption of a single uniform source of splitting. However, in Figures 3(a)-(b) and Figure S3, we notice that delay times at many sites
vary according to backazimuth. Moreover, for certain stations we only have a few measurements. And as a result, relatively large error bars are not surprising. While this limits our ability to tell how statistically different delay times are at individual stations, it is not contradictory to our general observation that there is a lateral increase in both the mean delay value and the size of the error bar from the NW end of the array to the SE end (Figure 5). We only use averaged delay values in the following discussion.

In Figure 5, delay times measured at stations between 0 ~ 600 km along the transect (corresponding to the NW half of the array) are generally consistent and have an average of ~0.5 s. From ~600 km to ~790 km (corresponding to the circled area in Figure 3(a)), delay times increase southward, reaching ~1 s at 700 km along the transect, and then decrease to ~0.5 s at 790 km. From ~790 km southward (corresponding to the SE section of the array), delay times increase once again. In this part, delay times separate into two trends. One trend of the delay times is ~1 s to ~1.2 s, whereas the other trend of delay times is ~0.5 s to ~0.8 s. Delay times measured in the circled area in Figure 3(b) are marked with the cyan diamonds in Figure 5 and they correspond to the trend of delay times varying from 0.5 s to 0.8 s. Here we divide the splitting values based both on delay times measured from individual events before averaging, and the mean values after unweighted averaging. The exact location where this lateral change happens is hard to identify. In Figure 3(a) we identify a region where the delay times appear to vary from site to site, not forming a clear trend. We present statistics of delay times in these three sections in detail in the Discussion section.

We can also observe the lateral change in delay times in measurements from individual events. For each event observed over the entire length of the array, the pattern of delay times is not exactly the same as the averaged pattern, however the lateral increase in delay times is still very obvious. In Figures 6(a)-(b), the increase in delay times from NW to the SE can be seen very clearly. The nature of change in delay values with distance along the array is not always the same, and appears to be a function of event backazimuth. For example, delay times measured from event 2014.103 (Figure 6(a)) are ~1.0 s from array position 500 km southward, then are scattered near the Appalachian Front, and subsequently grow to nearly 2 s at the Atlantic coast. Event 2015.132 (Figure 6(b)) yields splitting values that increase steadily from ~0.5 s to 1 s at the Appalachian Front, and then continue to increase to ~1.5 s near the coast.
3.2 NULL measurements

Apart from the splitting signals, NULL measurements can also be observed over the entire length of the array (Figure 7(a)). Stations in the NW half of the array tend to have more NULL measurements per station than those at the other end. We had 261 NULL measurements in total, which includes 127 NULL measurements at 18 stations in the NW half of the array and the rest 134 NULL measurements at 50 stations in the SE half of the array. If we calculate the average of NULL measurements per station, we have ~7 NULL measurements per station in the NW, and less than 3 NULL measurements per station in the SE. This result is very well illustrated by a comparison between stations QM78 (in the NW half of array) and QM34 (in the SE half of array) shown in Figures 7(b)-(c).

We also look at the NULL measurements in individual events. In Figures 4(a)-(b), we can observe NULL measurements at most stations along the transect. Those earthquakes are from nearly south and west of our study area. We can also observe NULL measurements at stations from earthquakes that come from the north of our study area (Figures 4(c)-(d)), but those observed NULL measurements are either limited to a small region of the study area, or the number of NULL measurements are smaller compared to what we have from the earthquakes coming from the south or the west.

3.3 Comparison with previous studies

We compare our observations with previous studies in the neighboring areas and make a station to station comparison when possible (Table S2). Our measurements do match the general statistics of those from previous studies (Figure 3(c)). Surveys of shear wave splitting in the southern part of our region by Long et al. (2016) measure an average fast polarization of N77°E and Yang et al. (2017) shows mostly E-W fast polarizations, with some local variations. A study by Darbyshire et al. (2015) covers approximately the same area as our study does, and reports a range of fast polarizations varying from ENE-WSW to ESE-WNW. Delay times reported in the Superior Province by Darbyshire et al. (2015) are generally less than 1 s, similar to the range we report in this study, while results for the Appalachian Orogen by Long et al. (2016) and Yang et al. (2017) are closer to 1-1.5 s on average, once again in line with our findings.
A station-to-station comparison with previous studies of splitting parameters measured from the same individual events is documented in Table S2. We first compare our measurements with those from Darbyshire et al. (2015). The averaged fast polarizations at sites measured in our study and that of Darbyshire et al. (2015) are generally within \( \sim 10^\circ \). Our averages are closer to E-W. For the only event measured by both studies (2013.134) at stations LATQ and MATQ we find close matches in observed splitting values. It is worth mentioning that splitting averages at stations DMCQ and A64 included in both studies are very similar even though we use two completely different datasets without any event overlap. We do not find systematic differences in fast polarizations when comparing our results with Long et al. (2016) and Yang et al. (2017). We note that nearly all of our averaged delay times are systematically smaller than those of the other two studies. For event 2014.103, sites G64A and H66A (SKKS phase in our study) yield very close measurements between this study and Long et al. (2016) (fast polarizations: less than 3° different; delay times: less than 0.02 s different) whereas G65A yields very different measurement especially a much larger delay time in their study of 2 s. Compared with Yang et al. (2017), we have very close matches from event 2013.271 at stations F61A, G63A and G64A (fast polarizations: less than 5° different; delay times: less than 0.1 s different), but a discrepancy at station H66A (fast polarizations: \( \sim 30^\circ \) different; delay times: \( \sim 0.3 \) s different). For event 2015.132, we have close matches in fast polarizations at stations E61A, D63A, G63A and G65A (less than 5°), but the differences in delay times are around 0.1 s \( \sim 0.2 \) s. For other events, whether there are close matches or large discrepancies depends on specific stations. For instance, for event 2015.150, D61A shows large discrepancies in both splitting parameters (fast polarizations: \( \sim 25^\circ \) different; delay times: 0.4 s different) whereas G63A shows close matches (fast polarizations: \( \sim 7^\circ \) different; delay times: 0.1 s different).

There are several reasons for discrepancies between our study and previous studies. First, we include datasets in different time frames. Second, there is a difference in the selection of time windows, filters, and how the analysts decided whether a measurement is a NULL. Third, choices of teleseismic phases to measure are different. For instance, we include four types of XKS phases whereas Long et al. (2016) includes only the SKS phase. Even within the same event, different phases have different ray paths and thus sample different parts of Earth. Finally, different methods of measurements can lead to different results. For instance, in Long et al. (2016), only measurements for which RC and SC methods yielded close results were retained and averaged, while we use the data measured by RC method, and utilize SC
method for quality assessment and for declaring NULLs. Both Darbyshire et al. (2015) and Yang et al. (2017) adopt SC method. Wüstefeld & Bokelmann (2007) show that RC method that we have adopted tends to yield relatively smaller absolute values of delays in cases where the noise level is high. We should also note that at stations with only a few measurements, the averages are easily influenced by extreme values and thus give quite different averaged splitting values.

4 Discussion

4.1 Comparison with absolute plate motion (APM)

Observations of shear wave splitting reflect the cumulative effects of anisotropic structure along the ray path, combining effects of the lithospheric mantle, the asthenosphere and the lowermost mantle. We first compare the fast polarizations with the mantle flow patterns in the asthenosphere. While the details of the mantle flow beneath a continent may be complex (e.g., Forte et al., 2007), we can compare fast polarizations of split shear waves with the absolute plate motion (APM), which reflects the motion of the North American plate relative to the deeper part of the upper mantle. Figure 8(a) shows a histogram of fast polarizations along the entire array computed on the basis of all observations of splitting. NULL measurements are not included in this calculation.

Figure 8(a) shows that while fast polarizations measured in this study fall into a very wide range, from N0°E to N160°E, there is a single well-defined peak between N60°E to N90°E. The average value of all fast polarizations measured is N80°E. According to the NUVEL1A-HS3 model the APM in our area is N249°E, and varies by less than 10° along the array (Gripp & Gordon, 2002).

Based on the similarity between the average fast polarization and the APM in the HS3 reference frame, we conclude that the shearing of the asthenosphere is a major contributor to the seismic anisotropy in our study area.

4.2 Possible thickness of the anisotropic layer

To test this inference further, we also estimate the possible thickness of the anisotropic layer responsible for the observed splitting signal. Based on Helffrich (1995), the thickness of a
single homogeneous horizontal anisotropic layer can be estimated as \( L \approx \delta t \times V_s / dV_s \), where \( \delta t \) is delay time, \( V_s \) is the shear wave velocity, and \( dV_s \) is the percentage of velocity change due to anisotropy. For estimation, we use a shear wave velocity of 4.5 km/s and an average anisotropy strength of 4\% (Savage, 1999), which are the representative numbers for the parameters in the subcontinental upper mantle. Average delay times for the NW and SE ends of the array are 0.56 \( \pm \) 0.25\,s and 0.90 \( \pm \) 0.41\,s, respectively. The middle section of the array has an average delay value of 0.79 \( \pm \) 0.31\,s. Figures 8(b)-(d) show histograms of delays within each section and specify the extent of each section. Correspondingly, the thickness of a single anisotropic layer will be 63 \( \pm \) 28 km in the NW section, 89 \( \pm \) 35 km in the middle section and 101 \( \pm \) 46 km in the SE section.

The change in the vertical extent of the anisotropic layer from the NW to the SE is consistent with constraints on the vertical extent of the lithosphere. Under the central part of the North American craton it extends to the depths of 200 – 250 km (Gung et al., 2003; Jaupart et al., 1998; Romanowicz, 2009; Rudnick et al., 1998), while it is about 90 – 110 km at the eastern North America continental margin (Abt et al., 2010; Rychert et al., 2005; Rychert et al., 2007).

Thus, we find that first-order lateral change in the strength of the splitting signal may be explained by the laterally variable vertical extent of the asthenospheric mantle deformed by the motion of the North American plate.

4.3 Evidence for more than one layer of anisotropy

Fast polarization measurements forming a clear peak between N60°E to N90°E in the histogram shown in Figure 8(a) make up only 56\% of all observations. While noise in the data likely impacts the values, the width of the distribution and the fact that nearly half of the measurements fall outside the main peak suggest that there should be additional contributions of anisotropy from another source besides the mantle flow in the asthenosphere.

In Figure 4, we document systematic changes of fast polarizations according to the backazimuths of the incoming rays. We can also observe such changes with backazimuths in data from events that were observed at subsets of our array. Figures 7(b)-(c) and Figure S3 illustrate values of fast polarization changes with backazimuth at individual sites. Directional
dependence of splitting parameters is an expected consequence of multi-layered anisotropic structure (e.g., Silver & Savage, 1994), which means besides the anisotropic contribution from the asthenosphere, there has to be another contribution, possibly from the fossil fabrics in the lithosphere.

Apart from splitting measurements of individual events, NULL measurements also provide the evidence of a complicated anisotropic structure since our observation of NULL measurements is contradictory to the pattern predicted by a simple one-layered anisotropic model. We find NULL measurements from many directions (Figures 4 and 7), while in case of a single layer of anisotropy we expect them to concentrate at two orthogonal directions (cf. Savage, 1999), coincident with either the fast polarization or the slow polarization.

Presence of multiple (up to three) layers of seismic anisotropy has been previously proposed for this region. Yuan and Romanowicz (2010) used seismic tomography combining surface waves and SKS splitting data to argue for multiple layers of anisotropic material within the North American lithosphere. In particular, their model suggests that there are two different anisotropic layers beneath the North America craton, including the region where our array was deployed. Levin et al. (1999) analyzed shear wave splitting values in the Appalachians and built a two-layered anisotropic model by matching the observed values and the predicted ones generated by synthetic seismograms. A subsequent study by Yuan and Levin (2014) confirmed the presence of these layers using two decades of XKS observations at sites near the Atlantic coast. Other studies of shear wave splitting results in neighboring areas (e.g. Darbyshire et al., 2015; Long et al., 2016) also interpret the corresponding areas to have more than one layer of anisotropy. Thus, combining the results from previous studies, the contribution of anisotropy from the past deformation processes preserved in the lithosphere cannot be neglected.

4.4 Possible contribution of anisotropy from the lowermost mantle

Theoretically, anisotropy measured from shear wave splitting integrates the contribution starting from the lowermost mantle to the upper mantle. In addition to the frozen fabric in the lithosphere and the mantle flow in the asthenosphere, the anisotropic contribution from the D” layer cannot be neglected. Since SKS and SKKS phases sample different portions of the lower mantle and similar portions of the upper mantle, the discrepancies in splitting values
between these two phases measured at the same station from the same event can be interpreted as evidence for anisotropy in the D” layer (Lynner & Long, 2014). In our study, we pick three events (Table S3) to check the anisotropy from the lowermost mantle. For event 2014.043, we cannot identify any discrepancy between SKS and SKKS pairs. However, for event 2014.103, we find similar fast polarizations between measurements made from SKS and SKKS phases, and SKKS phase yields larger delay times from 0.2 s to 0.4 s except at QM15 and F61A. For event 2014.305, SKS and SKKS phases have similar fast polarizations but SKKS generally yields smaller delay times than SKS phase except at QM38. Thus, we conclude that while a contribution from the D” layer is possible, we do not see a clear evidence for it in the data set we have analyzed.

4.5 A laterally variable anisotropic structure

Out of 24 events analyzed in this study, 13 produced XKS phases observed over the entire 1300 km length of our array (Figure 1(b)). Measurements made on the same phase exclude likely complications from different paths taken through the Earth. Lateral changes in anisotropic parameters measured form the same phase have to reflect variations in earth structure beneath our region. Comparing average measurements from 13 events to those obtained using a full dataset we confirm that the splitting values (delay values and fast directions) are very similar. The fast polarizations are N60°E to N90°E, NULL measurements are more common at the NW end of the array and an increase in delay time from the NW end to the SE end of the array is clearly manifested. Our ability to see the same behavior in individual continuously observed phases (Figure 4) and in averaged values (Figure 5) adds confidence to the lateral variations we report below.

4.5.1 Variation in delay time

Histograms of delay times for three sections of the array (Figures 8(b)-(d)) document significant scatter of values, especially in the SE section (790~1300 km). This scatter can also be seen in station mean values shown in Figure 5. This relatively scattered pattern at the SE end of array is not consistent with a notion of a single source of anisotropy at depth being the smoothly flowing upper mantle material. Considering the complicated tectonic history of the Appalachians it is likely that rock fabric frozen into the lithospheric mantle varies between distinct terranes composing the orogen. It is interesting to note that there are noticeable
along-strike changes in delay values in the Appalachians (Figure 3(b)). These changes in delay values over relatively short distances provide additional support for the presence of anisotropy in both the lithosphere and the asthenosphere. If the vertical extent of the anisotropic layer in the asthenosphere beneath the Appalachians is at the higher end of our estimate (over 100 km), average delays smaller than 1 s will imply a partial cancellation of its signature. The opposite scenario is also possible, with an asthenospheric contribution being amplified locally so that the average delays significantly exceed 1 s.

On the other hand, within the Superior and Grenville Provinces delay histograms show well-defined single peaks, and station averages are more uniform. Small delay values (<0.6 s) are especially common in the NW of the array, over the Archean craton.

4.5.2 Delay time comparison with geological settings on the surface

We compare changes in delay times along the array with the geological boundaries that can be observed on the surface. We find the smallest average delay times in the Superior Province. Delay times remain consistently small (less than 0.6 s) across the Grenville Front and through most of the Grenville Province. Delay times over 1 s appear in the Appalachians. It is also very interesting to see that though we cannot identify the exact location where the lateral change in delay time takes place, it is clear that this lateral change in delay time does not correspond to any of the geological boundaries on the surface. In Figure 3(a), we circle the region where we identify a lateral change, and in Figure 5 we mark this section on the transect. While it is close to the Appalachian Front, it clearly is not coincident with this major tectonic boundary. The change in delay values takes place over a zone ~200 km wide (600 km to 790 km along the array) within the Grenville Province, to the northwest of the area affected by the Appalachian orogeny.

The fact that the change in the size of the splitting delay, from ~0.6 s on average to ~0.9 s on average (Figures 5 and 8) takes place over a distance of 200 km or less, is in general agreement with the rapid lateral decrease of the thickness of the continental lithosphere towards the eastern coast of North America (Artemieva, 2006; van der Lee & Nolet, 1997; Yuan et al., 2014).
4.6 Interpretation of the possible edge of the craton at depth

The change in delay times between 600 and 790 km along our array does not correspond to major tectonic boundaries on the surface. More generally, none of the tectonic boundaries seem to coincide with a significant change in shear wave splitting. We thus seek possible links with continental lithosphere structure at depth. In Figure 9, we compare shear wave splitting measurements with the distribution of shear wave velocity and anisotropic properties. We show values for the depth of 160 km which is within the lithosphere beneath the craton, but in the asthenosphere under the Appalachians. We plot velocity values from Yuan et al. (2014) and anisotropy values from Yuan et al. (2011).

Sites at the NW end of this array, which have smaller delay times, correspond to a relatively higher velocity area (Vs > 4.7 km/s, or 4% faster than the global model IASP91, (Kennett, 1991)), with very weak anisotropy. Conversely, sites at the SE end, which have larger delay times, correspond to a relatively lower velocity area (Vs is 4.5-4.6 km/s, or within 2% of IASP91) with stronger anisotropy. As Figure 9 shows, the place where we find the change in delay time corresponds to the changes in both the shear wave velocity and the azimuthal anisotropy. The tomography model of Yuan et al., (2011) has a lateral resolution of ~500 km. Therefore, even though we can see a transition from higher velocity to lower velocity, it is hard for us to locate where the transition happens. However, because the measurements from shear wave splitting provide very good lateral resolution, they put a better constraint on the change of properties at depth. Since the lateral change in delay times agrees with both the 4.65 km/s contour of shear wave velocity and 0.25% contour of azimuthal anisotropy, we interpret this boundary to be the edge of cratonic lithosphere at the depth of 160 km.

5 Summary

In this paper, we present shear wave splitting measurements of core-refracted shear waves on a 1300 km long array crossing the eastern part of the North American continent from James Bay to the Fundy Basin. We compare the shear wave splitting values with the absolute plate motion direction, tectonic boundaries on the surface and geophysical boundaries at depth.

We find splitting signals at all stations of this array, with predominant fast polarizations falling between N60°E and N90°E. The close similarity between this dominant value and the
direction of the absolute plate motion suggests that the deformation of the asthenosphere is the primary source of the signal we detect.

At each station, the polarizations are similar within each observed event, but are different from event to event, and a systematic change of fast polarizations can be observed at all stations along the array. This suggests the possibility of a structure with more than one layer of anisotropy beneath our study area, in agreement with previous studies. Delay times are relatively consistent at each individual station, and increase from ~0.5 s in the Superior Province to ~1 s in the Appalachian Province. The change takes place in the Grenville Province near the Appalachian Front.

We observe a smaller delay time over a much thicker lithosphere. This finding may imply an absence of anisotropy in the old cratonic lithosphere, or alternatively an efficient cancellation of contributions from it and the underlying asthenosphere. We favor the first choice as we do not find any examples of strong splitting at our stations on the craton. In the presence of two layers with near-orthogonal anisotropy orientations we would expect to detect strong splitting from events that arrive along the symmetry axis of one of them.

The lateral change in delay times is located approximately 100 km northwest of St. Lawrence River, and does not correspond to any major geologic structures at the surface. Rather, it appears to match the boundary where the shear wave velocity and the strength of azimuth anisotropy change at the depth of 160 km, which can be interpreted as the edge of cratonic lithosphere at that depth.

Splitting results in our study area rule out the possibility of a single layer of anisotropy and suggests the anisotropic structure beneath the eastern North America to be both multi-layered and laterally variable.

Acknowledgments

This work was supported by the NSF Earthscope grant EAR-1147831 and the graduate fellowship for the first author from the Rutgers School of Graduate Studies. Aresty Undergraduate Research Assistantship for Yiran Li was instrumental for her participation in the work. Data can be accessed at the Data Management Center (DMC) of the Incorporated Research Institutions for Seismology (IRIS) and Portable Observatories for Lithospheric
Analysis and Research Investigating Seismicity (POLARIS). Figures are drafted using GMT (Wessel & Smith, 1991). We thank two anonymous reviewers for their comments and suggestions.

References


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Figure 1. (a) Map of our study area in the eastern North America: black lines with teeth show the Grenville Front (GF) and the Appalachian Front (AF), brown lines show terranes of the Appalachian Orogen. Seismic stations used in this paper are dark red. JB: James Bay; FB: Fundy Basin; SLR: St. Lawrence River. (b) Locations of 24 earthquakes (Table S1) from 2012 to 2015 with magnitudes over 6.8 used in this paper (yellow:13 earthquakes with phases observed by the entire array, blues: additional 11 earthquakes; see text for details). This map is centered on our study area.
Figure 2. Examples of (i) a splitting measurement (at station LATQ) and (ii) a NULL measurement (at station QM76) exported from SplitLab. Plots (a)-(f) are similar in examples (i) and (ii). (a) ENZ components of the seismogram with predicted SKS (red solid line) and Sdiff (blue solid line) phases and the time window (time duration between ‘Start’ and ‘End’ as shown in the plot) marked with two black dashed lines. (b) RTZ components of the seismogram (same convention is adopted as ENZ components). In this figure, positive radial component is pointing away from the source whereas the positive radial component in Splitlab is pointing toward the source. Same conventions are adopted both in splitting and NULL examples. (c) Fast (blue solid line) and slow (red solid line) components after

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correction of delay time. (RC method is adopted in this example.) (d) Radial (blue solid line) and transverse (red solid line) components after correction of strength of anisotropy. (e) Particle motions before (blue solid line) and after (red solid line) correction of anisotropy. Black line here stands for the direction of backazimuth. (f) Map of correlation coefficient.
Figure 3. Splitting values measured from 24 earthquakes at 64 stations along the transect (red solid line). Orientations of the blue sticks show the fast polarizations and lengths of the blue sticks are proportional to measured delay times. Orange circles stand for stations that are used in this study. (a) Splitting values of all the stations. (b) An amplified plot of splitting values of the stations in the SE half of the array. Same convention is adopted for tectonic settings as in Figure 1. (c) unweighted mean splitting measurements in this study (blue sticks) plotted on the background of the mean splitting values from the observations of published studies (orange sticks). Data from previous studies acquired using a global shear-wave splitting database (IRIS DMC, 2012).
Figure 4. Splitting values and NULL measurements along the transect for selected events. Dates (Julian Days) of different events are in the corresponding figure titles. Meanings of different sticks are the same as those in Figure 3. Green sticks stand for NULL measurements. Their directions are aligned with the backazimuths plotted starting from the stations, and lengths are chosen to be 0.5 s. Red arrows stand for ray propagation directions from different earthquakes.
Figure 5. Unweighted mean fast polarizations and delay times plotted from NW end of the array to the SE end of the array according to the distances away from the 79W, 52N. The blue circles: mean fast polarizations; red diamonds: mean delay times. Cyan diamonds identify sites circled in Figure 3(b). In both figures, error bars are one standard deviation in each direction. Vertical grey lines, represent tectonic boundaries: GF: Grenville Front, AF: Appalachian Front. Red rectangle marks the circled area in Figure 3(a).
Figure 6. (a) and (b) are splitting values and NULL measurements for selected events. (Conventions of (a) and (b) are the same as similar plots in Figures 3-5). (c) and (d): splitting values of fast polarization and delay time measured from two selected events. Red lines in both upper panels stand for the polarizations aligned with the backazimuths. Grey lines in both upper panels stand for the averaged fast polarizations at all events. Black triangles stand for NULL measurements. For other symbols, blue circles, red diamonds and the vertical grey lines in the lower panels follow the same convention as Figure 5.
Figure 7. Examples of NULL measurements. (a) NULL measurements from 24 earthquakes at 64 stations along the transect (red solid line). Green sticks are aligned with backazimuths of the earthquakes (same convention as in Figure 4). Orange circles stand for stations that are used in this study. (b) Stereoplot of all splitting values and NULL measurements at QM78. (c) Stereoplot of all splitting values and NULL measurements at QM34. Stereoplots show splitting values as a function of backazimuth (positive clockwise from north) and incidence angle (positive outward from center, grid lines from 0 to 18° every 3°) of their respective rays. Red sticks stand for the splitting measurements using RC method and black circles stand for NULL measurements. Orientations of the red sticks stand for the corresponding fast polarizations and their lengths are proportional to delay times. The black stick on the right of each plot stands for 1 s delay time.
Figure 8. Histograms of all the fast polarizations and delay times in three sections of the array. For convenience, we subtract 180° from the true APM direction and plot N69°E onto the histogram instead. (a) Histogram of all fast polarizations in the data set. The red line is the average of all fast polarizations; the black line is an average of absolute plate motion directions estimated at all stations using the HS3-NUVEL1A model and HS3 reference frame. (b) Histogram of delay times at the NW end of the array (stations from 0 km to 600 km); (c) Histogram of delay times at stations between 600 km and 790 km; (d) Histogram of delay times at the SE end of the array (stations from 790 ~ 1300 km). Both red lines mark the average delay times in the corresponding section of the array. Ranges of the transect distances used in three histograms are marked at the upper right corner of each figure.
Figure 9. Comparison of shear wave splitting values with contours (red lines) of (a) shear wave velocity (unit is km/s) from Yuan et al. (2014) and (b) azimuthal anisotropy amplitude (unit is %) at the depth of 160 km from Yuan et al. (2011). Orientations of blue sticks which are centered at the stations represent averaged fast polarizations, while the lengths are proportional to the corresponding averaged delay times. In (a), only contours of 4.7 km/s and 4.65 km/s are shown. In (b), only contours of 0.15% and 0.25% are shown. Velocity decreases from the NW to the SE, and the strength of azimuthal anisotropy increases from the NW to the SE.