Mantle structure and flow across the continent-ocean transition of the eastern North American margin: anisotropic *S*-wave tomography

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Key Points:

- We conducted mantle-scale velocity-anisotropy tomography across the continent-ocean transition of eastern North America.
- The results capture layers of anisotropy preserved from collision and extension, as well as produced from modern mantle flow.
- The imaged lithospheric and asthenospheric structure supports edge-driven convection and margin parallel asthenospheric flow.

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Abstract

Little has been seismically imaged through the lithosphere and mantle at rifted margins across the continent-ocean transition. A 2014-2015 community seismic experiment deployed broadband seismic instruments across the shoreline of the eastern North American rifted margin. Previous shear-wave splitting along the margin shows several perplexing patterns of anisotropy, and by proxy, mantle flow. Neither margin parallel offshore fast azimuths nor null splitting on the continental coast obviously accord with absolute plate motion, paleo-spreading, or rift-induced anisotropy. Splitting measurements, however, offer no depth constraints on anisotropy. Additionally, mantle structure has not yet been imaged in detail across the continent-ocean transition. We used teleseismic *S*, *SKS*, *SKKS*, and *PKS* splitting and differential travel times recorded on ocean-bottom seismometers, regional seismic networks, and EarthScope Transportable Array stations to conduct joint isotropic/anisotropic tomography across the margin. The velocity model reveals a transition from fast, thick, continental keel to low velocity, thinned lithosphere eastward. Imaged short wavelength velocity anomalies can be explained by edge-driven convection. We also find layered anisotropy. The anisotropic fast polarization is parallel to the margin within the asthenosphere. This suggests margin parallel flow beneath the plate. The lower oceanic lithosphere preserves paleo-spreading-parallel anisotropy, while the continental lithosphere has complex anisotropy reflecting several Wilson cycles. These results demonstrate the complex and active nature of a margin which is traditionally considered tectonically inactive.

Plain Language Summary

North America was once connected to Africa, but the continents rifted apart and are now separated by the Atlantic Ocean. The nature of rifting on land has been thoroughly studied. However, it is much more difficult to study the offshore region where the thinned continent pinches out and the tectonic plate transitions to sea-floor produced after continental breakup. Using a new dataset from ocean-bottom seismic stations, we construct a 3-D image of seismic wavespeeds, which are diagnostic of rock type and temperature. We also image seismic anisotropy, which is the directional dependence of seismic velocity. Anisotropy is often used as a proxy for the direction of stretching or mantle flow. We find wavespeed anomalies diagnostic of convective cells driven by a step in the thickness of the lithosphere. The anisotropy models suggests that, in this region, the mantle beneath the plates is currently flowing along the margin. Within the tectonic plates, the mantle preserves anisotropy developed during cycles of rifting and collision. These seismic wavespeed and anisotropy models demonstrate the complex and active nature of a continental margin that is traditionally considered tectonically inactive.

1 Introduction

Continental rifting (*e.g.,* McKenzie, 1978; Wernicke, 1985) and sea-floor spreading (*e.g.,* Hess, 1962; Vine & Mathews, 1963) are fundamental tectonic processes. The transition from continental rifting to the production of seafloor and thus continental drifting, however, remains unclear (*e.g.,* Kendall et al., 2005; Shillington et al., 2006; Van Avendonk et al., 2006; Crosby et al., 2008; Begg et al., 2009; Huismans & Beaumont, 2011; Yuan et al., 2017; Larsen et al., 2018). Seismic resolution across the rift-drift transition, particularly in the mantle, is extremely limited due to the sparsity of broadband ocean-bottom seismometers (OBSs) offshore at rifted margins.

The eastern North American passive margin (ENAM) is an excellent location to study the tectonics of the rifted continent-ocean transition (COT). ENAM is a mature passive margin resulting from the rifting of Pangaea at ∼230-200 Ma (Withjack et al., 2012). There has been relatively little deformation at ENAM since the transition from rifting to continental drifting in the Jurassic (Schlische, 2003; Withjack & Schlische, 2005).

Structures associated with rifting along the margin are thus likely unperturbed and can offer insights into rifting processes.

ENAM is a natural laboratory for studying rifting processes (Worthington et al., 2021). It was selected as a primary site for a Geodynamic Processes at Rifting and Subducting Margins (GeoPRISMS) community seismic experiment (CSE) (Lynner et al., 2020). Thirty broadband OBSs were deployed in 2014-15, while the Transportable Array (TA) was in the eastern US (Figure 1). The TA provided excellent on-land broadband seismic coverage in the eastern US throughout 2012-2015, supporting interrogation of the continent. Combined, the ENAM-CSE and TA provide dense, co-temporal seismic data coverage crossing the COT. The ENAM-CSE constitutes one of the only rifted-margin crossing broadband OBS datasets, which is capable of interrogating a rifted COT in the mantle. This dataset has already been utilized for crustal-scale ambient noise tomography (Lynner & Porritt, 2017; Li & Gao, 2019), shear-wave splitting analyses (Lynner & Bodmer, 2017), multi-channel reflection imaging (Bécel et al., 2020), and crustal to uppermost-mantle tomography based on wide-angle seismic data (Shuck et al., 2019). However, no margin-spanning body-wave velocity or 3-D anisotropy models have been developed that illuminate the offshore mantle with the coverage offered by these OBSs, leaving mantle structure and flow across the COT largely unknown.

1.1 The ENAM continent-ocean transition

During continental breakup, the lithosphere thins (*e.g.,* Wernicke, 1985; Ziegler & Cloetingh, 2004; Crosby et al., 2008; Huismans & Beaumont, 2011). However, it is unknown whether there is a transition in plate thickness, wavespeed, or other properties across the ENAM COT. It is also unknown to what extent the oceanic lithosphere has thermally grown in with time. It is not known whether the lithosphere-asthenosphere boundary bears a signature of the COT (*e.g.,* Yuan et al., 2017) because the lithospheric and mantle structure at the COT has not been imaged in detail.

Several studies have used dense on-land seismic coverage to image the ENAM continental crust and mantle (*e.g.,* Forte et al., 2007; Van Der Lee et al., 2008; Bedle & van der Lee, 2009; Schmandt & Lin, 2014; Biryol et al., 2016; Pollitz & Mooney, 2016; Savage et al., 2017; Golos et al., 2018; Wagner et al., 2018; Savage, 2021). These informed the continent's lithospheric structure, which contains a mid-lithospheric discontinuity in the continental interior, thins toward the ocean, and is highly thinned at the Harrisonburg anomaly (HA) in Virginia (*e.g.,* Abt et al., 2010; Byrnes et al., 2019; Savage, 2021). However, studies to date have focused primarily on continental structures and have not incorporated the new OBS data.

Utilizing ENAM-CSE OBSs, recent Rayleigh wave ambient noise phase velocity tomography has shown crustal thinning across the margin and a correlation between the East Coast Magnetic Anomaly (ECMA) and a region of thinned crust (Lynner & Porritt, 2017). Full-waveform ambient-noise tomography reinforced these results (Li & Gao, 2019). The presence of the ECMA at the edge of the margin suggests that it is correlated with the first oceanic material emplaced after rifting. Active source results show that the crust is thin (down to about 6-8 km) and highly faulted between the ECMA and the Blake Spur magnetic anomaly (BSMA) (Shuck et al., 2019; Bécel et al., 2020), which is approximately 100-200 km east of the ECMA. The localized, thin crust suggests that a $\sim$ 150 km swath of crust between the magnetic anomalies is proto-oceanic and formed during ultra-slow spreading. The crust thickens to about 8.5-10 km and attains a smoother topography at the BSMA. This may imply that full sea-floor spreading did not initiate until the emplacement of BSMA ($\sim$ 170 Ma).

Ambient noise surface waves and long-offset refraction data are primarily sensitive to structure in the crust. The relationship of the lithospheric mantle wavespeed structure and crystalline fabric to crustal structure, magnetic anomalies, and stages of rift-

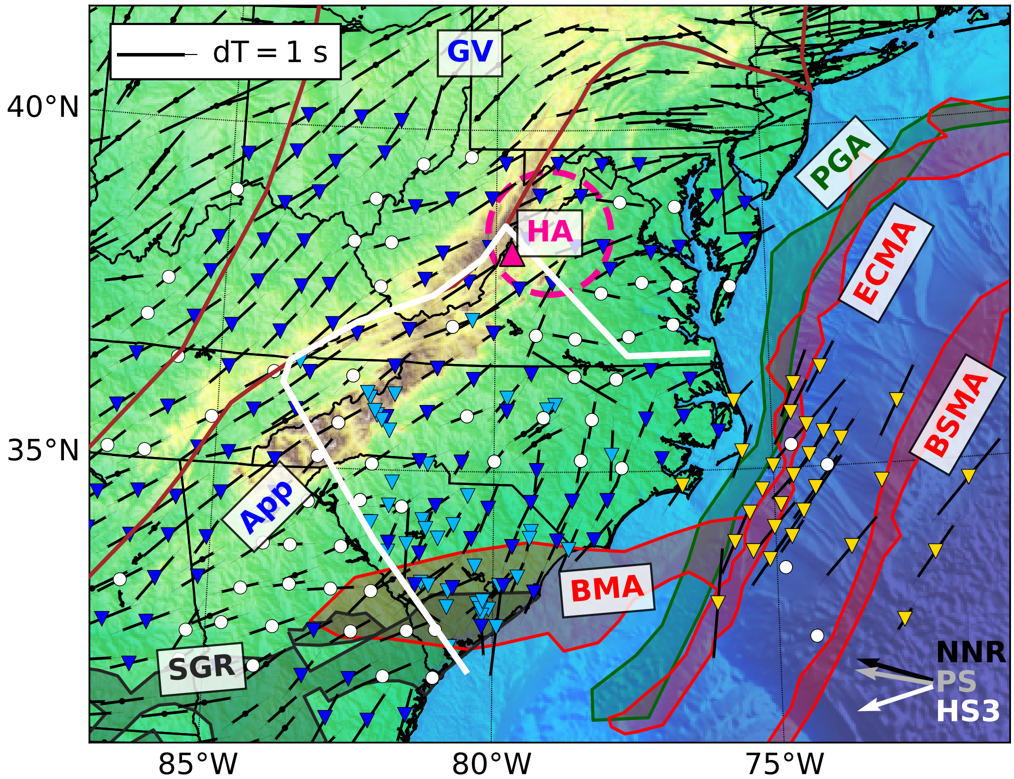

Figure 1. Map of the study area, showing stations used in our inversion (inverted triangles and white circles) and previous splitting measurements (Long et al., 2016; Lynner & Bodmer, 2017; Yang et al., 2017) that we averaged at each station. Splitting measurements are black lines centered at stations. Stations with dominantly null splitting are white circles. These null splitting stations had quality null arrivals and no split arrivals in Lynner and Bodmer (2017) and Long et al. (2016), but may still have split arrivals in Yang et al. (2017). For non-null splitting stations, the OBSs and land stations that were deployed as part of the ENAM-CSE are yellow triangles, the TA is dark blue triangles, and other networks are light blue triangles. The white line shows the boundary between stations where we measured splitting and stations where we only measured differential travel times. We also show splitting averages outside our seismometer array, where stations are indicated as small black dots. The Appalachian (App) and Grenville (GV) province boundaries are indicated by brown lines. The low velocity Harrisonburg anomaly (HA) is indicated with a pink, dashed circle (HA location is from our results; see also Biryol et al., 2016; Pollitz & Mooney, 2016; Savage, 2021). The pink triangle is the approximate location of Eocene volcanics (Mazza et al., 2014). Arrows at bottom right show approximate directions of Paleo-spreading (PS) (Becker et al., 2014) and plate-motion in no-net-rotation (NNR; DeMets et al., 2010) and hot-spot (HS3; Gripp & Gordon, 2002) reference frames. We highlighted the Positive Gravity anomaly (PGA) based on Sandwell and Smith (2009). We highlighted magnetic anomalies (BMA, ECMA, BSMA) based on Maus et al. (2009). South Georgia Rift basin (SGR) modified from Akintunde et al. (2014) and Chowns and Williams (1983). OBSs: ocean-bottom seismometers. ENAM-CSE: Eastern North American margin community seismic experiment. TA: Transportable Array. HA: Harrisonburg anomaly. SGR: South Georgia Rift basin. App: Appalachian. GV: Grenville province. BMA: Brunswick Magnetic Anomaly. PGA: Positive Gravity Anomaly. ECMA: East Coast Magnetic Anomaly. BSMA: Blake Spur Magnetic Anomaly. PS: Paleo-spreading direction. NNR: No-net-rotation reference frame. HS3: Hot-spot reference frame.

ing remains unknown. Here, we image the structure of the lithosphere at the COT. We investigate how the structures are associated with the transition from continental break up to continental spreading and how the lithosphere has subsequently been modified.

1.2 Mantle flow

Several important geodynamic phenomena have been proposed at ENAM (*e.g.,* Long et al., 2010). Fouch et al. (2000) showed that shear-wave splitting within the continent is consistent with a model where mantle flow is redirected by a deep continental lithospheric keel to flow perpendicular to the margin. Low shear velocity near ENAM could indicate volatile abundance and upwelling material associated with subducted Farallon slab (Van Der Lee et al., 2008). The strong horizontal temperature gradient in the mantle near the cratonic edge (the "keel") can induce edge-driven convection (EDC) (*e.g.,* King & Ritsema, 2000; Ramsay & Pysklywec, 2011; Savage et al., 2017). EDC has been invoked to explain specific seismic velocity features at ENAM (*e.g.,* Savage et al., 2017). Conversely, the slow velocity features may indicate lithoshperic delamination and asthenospheric upwelling, which could also account for enigmatic Eocene volcanism (Figure 1; *e.g.* Mazza et al., 2014; Biryol et al., 2016). Margin-parallel shear-wave splitting results offshore have been interpreted as reflecting large scale density-driven flow (Lynner & Bodmer, 2017).

Observational constraints on mantle flow at ENAM, particularly beneath the ocean, are limited (*e.g.,* Yuan et al., 2011; Lynner & Bodmer, 2017; Yang et al., 2017). Seismic anisotropy is a crucial observational constraint on mantle flow. Our joint velocity/anisotropy tomography model, with sensitivity extending through the asthenosphere, is poised to address mantle flow near the COT of ENAM.

1.3 Anisotropy and shear-wave splitting

Seismic anisotropy measurements can offer insight into patterns of mantle deformation (*e.g.,* Silver, 1996; Long & Becker, 2010; Skemer & Hansen, 2016). Deformation via dislocation creep produces a crystallographic preferred orientation (CPO) due to heterogeneity in the strength of internal slip systems (Karato & Wu, 1993; Maupin & Park, 2007; Karato et al., 2008). Olivine CPO produced in this way is one of the dominant anisotropic signatures associated with mantle flow. Other phenomena can also result in seismic anisotropy, including aligned fractures or melt pockets (*e.g.,* Vauchez et al., 2000; Kendall et al., 2005).

Shear-wave splitting is a common method used to examine seismic anisotropy (Silver & Chan, 1991). Solutions to the Christoffel equation generally give three wave speeds and particle motion polarizations corresponding to a P wave and two quasi-S waves (Maupin & Park, 2007). Because the quasi-S waves travel at different velocities in an anisotropic medium, a time delay between them can accrue. Using the polarization and time delay between the quasi-S waves, the strength and orientation of anisotropy can be inferred.

There is a first-order question across the ENAM whether CPO fabrics and anisotropy are dominated by recent processes or record deformation associated with continental breakup. If associated with recent processes, splitting may align with absolute plate motion or paleo-spreading in the ocean (*e.g.,* Silver, 1996; Long et al., 2010; Becker et al., 2014). If related to past deformational events, splitting may align with tectonic boundaries (*e.g.,* Silver, 1996; Long et al., 2010). *SK(K)S* phase splitting across the ENAM exhibits a complex pattern of anisotropy that does not fit with either simple explanation. A region of dominantly null and very weak splitting on the continent (Figure 1; Wagner et al., 2012; Long et al., 2016) might be caused by roughly isotropic material, vertical mantle flow, or depth varying anisotropy that effectively cancels. Further, splitting at the OBSs reveals margin-parallel fast-axes (Figure 1; Lynner & Bodmer, 2017). This is neither consistent with paleo-spreading parallel frozen-in anisotropy in the lithosphere nor absolute-plate-motion-parallel anisotropy in the sheared asthenosphere (*e.g.* Becker et al., 2014). Lynner and Bodmer (2017) proposed the splitting is a consequence of modern margin parallel mantle flow. However, splitting of *S(K)KS* phases provides few constraints on the depth of anisotropy in the upper mantle, making it difficult to interpret what geodynamic processes are occurring (*e.g.,* Long et al., 2016; Lynner & Bodmer, 2017; Yang et al., 2017).

In this paper, we provide the first high resolution constraints on 3-D anisotropy and velocity heterogeneity across the COT of the rifted ENAM with sensitivity through the lithosphere-asthenosphere system. We obtained improved anisotropic depth resolution by combining *S*-phase splitting (*e.g.,* Hammond & Toomey, 2003; Boyd et al., 2004) with *S(K)KS* and *PKS* phases in a tomographic method (Eilon et al., 2016). We used a 1500-km-wide seismic array of broadband stations to produce a model that extends from the Appalachians to $\sim$ 300 km offshore (Figure 1). We interrogate anisotropy that developed during previous tectonic events, and utilize anisotropy to inform modern asthenospheric flow. We utilize isotropic velocity heterogeneity to interrogate the structure of the mantle, which further informs geodynamic processes, as well as to understand the lithospheric-scale structure of the rift and transition from continent to ocean.

2 Methods

We applied a joint velocity and anisotropy tomography method which uses differential travel times and splitting times of *S*, *SK(K)S*, and *PKS* phases to simultaneously solve for 3-D seismic velocity (synonymous with wavespeed) and azimuthal anisotropy (Eilon et al., 2016). Inversions which assume no anisotropy can produce significantly biased velocity models (Bezada et al., 2016). In addition to providing new depth constraints on anisotropic structure, an important strength of this approach is that it addresses the trade-off between anisotropic and isotropic controls on travel times (Eilon et al., 2016).

We used broadband data from the ENAM-CSE (up to 1.5 years of data from 30 OBSs and 3 land seismometers), the Transportable Array (TA) (which was present in the eastern U.S. from 2011-2015), and several regional networks for a total of 245 stations (Figure 1). Our earthquake selection criteria (Section S1.1) leave 2326 earthquakes which we evaluated from January 2003 to May 2020.

We first measured shear wave splitting times jointly with differential travel times (Section 2.1). We used an augmented multi-channel cross-correlation (MCCC) approach (Eilon et al., 2016). Splitting times were measured as either margin-parallel fast (positive) or margin-perpendicular fast (negative) (Figure 2). Splitting times constrain anisotropy. We jointly measured differential travel times, relative to the arrival times predicted by the IASP91 1-D velocity model (Kennett & Engdahl, 1991), between all stations. The primary role of the differential travel times is to constrain isotropic velocity, although they also inform anisotropy.

Splitting and differential travel times were the input for the tomographic method (Section 2.2; Figure 3, 4, and 5). We decompose velocities into margin parallel ($V_{\|}$) and perpendicular ($V_{\perp}$) components (Figure 4). We jointly invert these velocities using our splitting and differential travel times (Eilon et al., 2016). Isotropic velocity is simply $\left(V_{\|} + V_{\perp}\right)/2$. Velocity is reported as percent deviation from each layer's average. Differential travel times cannot constrain absolute velocity; by construction, velocity deviations within each layer have an average of zero. Azimuthal anisotropy strength is simply $\left(V_{\|} - V_{\perp}\right) / \left(V_{\|} + V_{\perp}\right)$, so positive values indicate margin-parallel fast and negative values indicate margin-perpendicular fast (Figure 4).

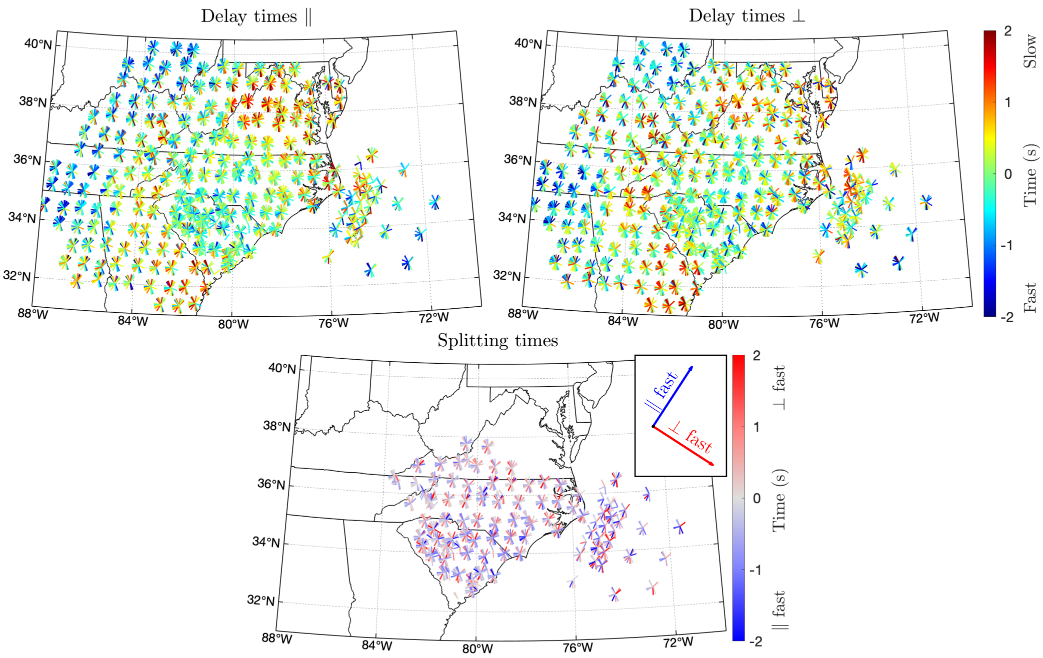

Figure 2. Inverted differential travel times for teleseismic *S*, *S(K)KS*, and *PKS* phases polarized parallel and perpendicular to the symmetry axis, $\delta T_{\|}$ and $\delta T_{\perp}$, as well as the measured splitting times, dT_{splt}. Each measurement is plotted as a line that points toward the earthquake. Note that smaller array bounds are used for measuring splitting to avoid anisotropy with non-constant geometries.

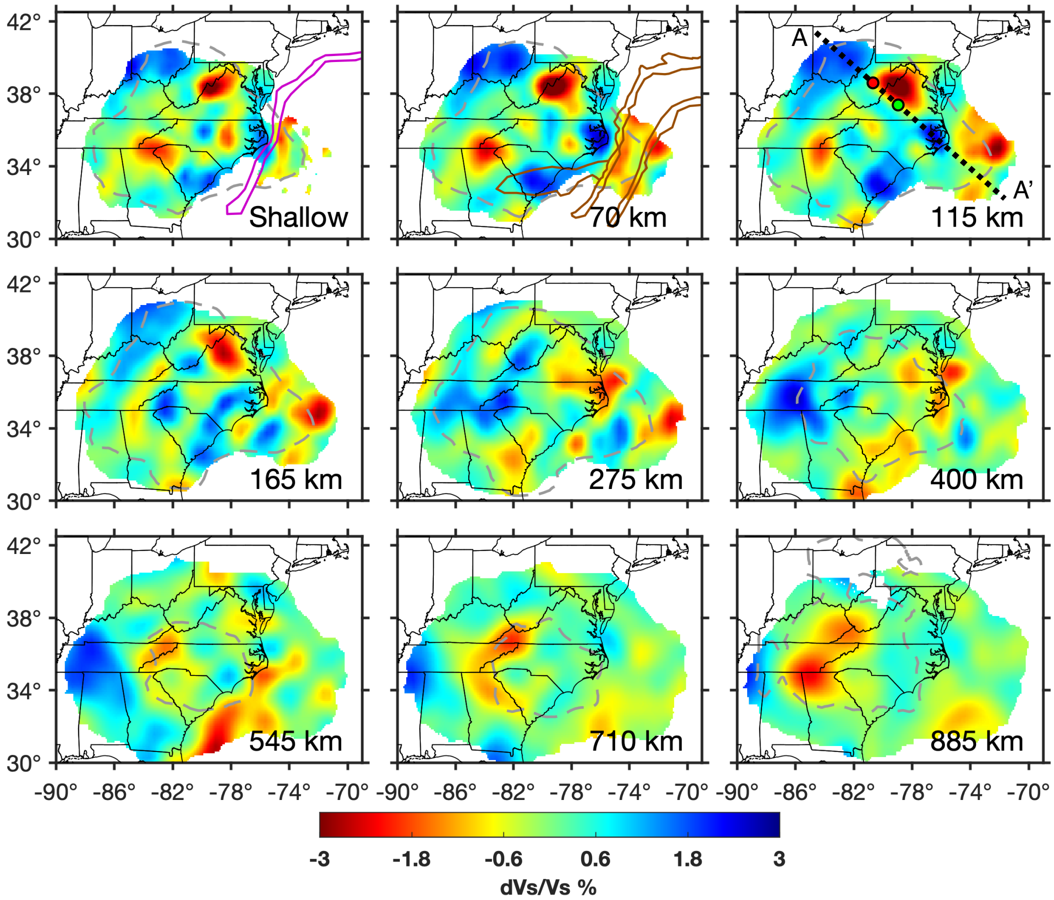

Figure 3. Final isotropic shear velocity model. Each layer shows percent deviation from the layer's average velocity. "Shallow" is our shallowest layer. It contains structure which is too shallow to be vertically resolved and is essentially averaged above ∼ 70 km. Positive Gravity Anomaly (PGA), drawn based on Sandwell and Smith (2009), shown on shallow depth slice as magenta line. Magnetic anomalies, drawn based on Maus et al. (2009), shown on 70 km slice as brown lines. From west to east, they are the BMA, ECMA, and BSMA. Models are only plotted where hit quality exceeds 0.7. The dashed gray contour shows where semblance (a measurement of the similarity between synthetic input and output checkerboard models) exceeds 0.8. Red and green dots on the 115 km slice border the high topography region of the Appalachians. PGA: Positive Gravity Anomaly. BMA: Brunswick Magnetic Anomaly. ECMA: East Coast Magnetic Anomaly. BSMA: Blake Spur Magnetic Anomaly.

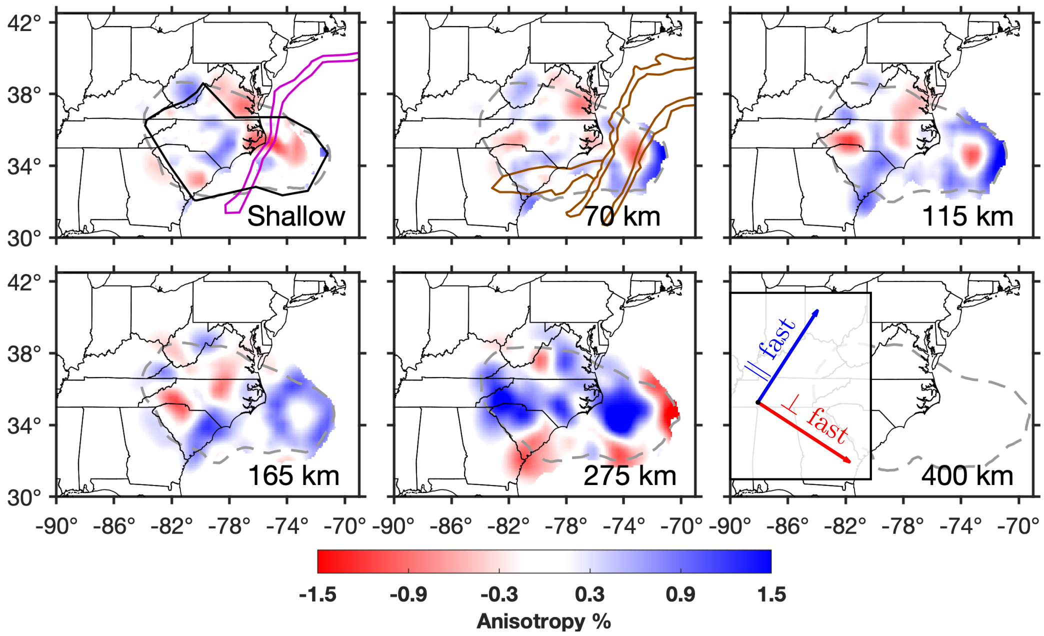

Figure 4. Final anisotropy model. See Figure 3 for description of shallow layer, potential field contours, and masking based on hit quality and semblance. Blue indicates margin-parallel-fast anisotropy and red indicates margin-perpendicular-fast anisotropy. Anisotropy is assumed to be 0 beneath 300 km. The black contour on the "shallow" slice is the region within which we measured splitting times.

2.1 Anisotropic multi-channel cross-correlation

To jointly measure splitting and differential travel times, we used the augmented multi-channel cross-correlation (MCCC) method of Eilon et al. (2016). This augmented method builds on traditional MCCC (Section S1; VanDecar & Crosson, 1990). We assume a constant horizontal hexagonal symmetry axis with orientation $\phi =$ N33°E. The fixed symmetry axis is not equivalent to assuming a fixed fast splitting orientation. Rather, we assume that structure will be organized according to the tectonic geometry: either margin-parallel or margin-perpendicular fast. This geometry is seen in the OBS splitting measurements (which have an average fast polarization of N33°E: Lynner & Bodmer, 2017) as well as splitting measurements in the Appalachians (Figure 1; Long et al., 2016). The assumed anisotropy can produce split quasi-S waves polarized approximately parallel and perpendicular to ϕ. These have differential travel times $\delta T_{\parallel}$ and $\delta T_{\perp}$, respectively. The difference between these times at a station is the splitting time dT_{splt}. By incorporating both quasi-S waves into MCCC, we simultaneously measured all three delay times. In detail, the relative amplitude of both quasi-S waves, and thus the feasibility of measuring splitting, depends on a wave's particle motion polarization (Section S1.2). For some earthquakes, we can only measure $\delta T_{\parallel}$ or $\delta T_{\perp}$.

We require a constant assumed symmetry axis direction for the tomography method to work. We evaluated whether the preponderance of SKS splits at any station is concordant with the assumed symmetry. This is quantified as previous splitting fast polarizations being on average within $\sim 25°$ of parallel or perpendicular to the assumed symmetry axis (Figure S1). We only measured splitting times and carry out anisotropic tomography in this region. Outside this region, we measured differential travel times only and conducted only isotropic tomography, fixing anisotropy to zero (Figure 2 and 4). Fi-

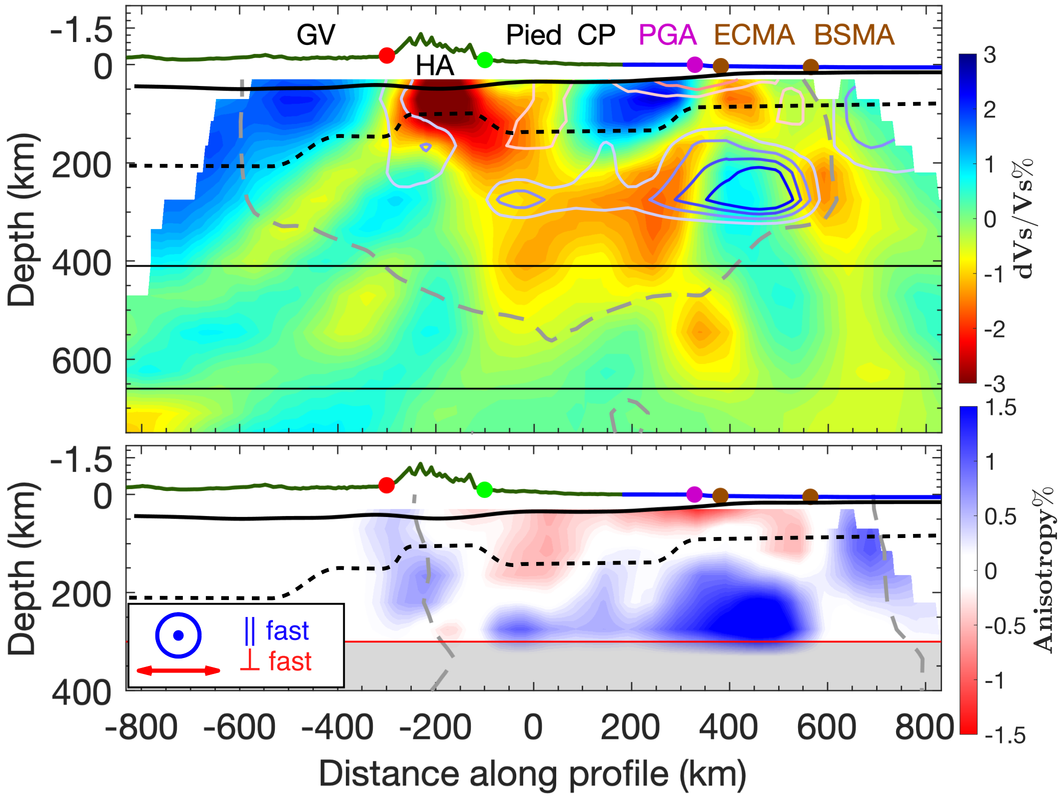

Figure 5. Cross-sections of the velocity (top) and anisotropy (bottom) models. Contours of anisotropy are shown on top of the velocity figure at $\pm 0.4\%, 0.75\%, 1.1\%$, and 1.45%. Elevation lines are blue in the ocean and green on land. The red and green dots on the elevation line correspond to the same dots in Figure 3. Mantle transition zone lines are shown at 410 km and 660 km. The approximate dislocation creep regime base is indicated as a red line at 300 km depth. We assumed no anisotropy beneath 300 km, which is greyed out. The Moho depth (Shen & Ritzwoller, 2016; Shuck et al., 2019) is shown as a solid black line. The approximate, interpreted lithosphere-asthenosphere boundary is shown as a black dashed line. Cross-sections run from northwest to southeast (A-A' in Figure 3). Only portions of the model where the hit quality is greater than 0.7 is shown. The dashed gray contour shows where semblance (a measurement of the similarity between synthetic input and output checkerboard models) exceeds 0.8. GV: Grenville province. HA: Harrisonburg anomaly. Pied: Piedmont. CP: Coastal plains. PGA: Positive Gravity Anomaly. ECMA: East Coast Magnetic Anomaly. BSMA: Blake Spur Magnetic Anomaly.

nally, we down-weighted splitting measurements made where the measured symmetry axis deviates from our assumed symmetry axis. We weighted measurements by $\sqrt{|(45^\circ - \Delta\phi)|/45^\circ}$ where $\Delta\phi$ is the difference between our assumed symmetry axis and the mean fast polarization of literature measurements within the proximity of a station ("splitting misorientation" in Figure S1).

Our method also assumes that symmetry axes are horizontal, and it is not sensitive to radial anisotropy. Any radial anisotropy will be mapped into our results, but only weakly. This effect is well within the uncertainty of the data (Eilon et al., 2016). Variations in anisotropy orientation with depth could be further problematic. A lack of strong back-azimuthal variability in splitting measurements offers some support that layered anisotropic fabrics are either parallel or perpendicular to each other where we measure splitting times (Yang et al., 2017). Our method is optimal for such layering. We verified that the complexity of splitting at ENAM is consistent with our simplified anisotropic orientation, supporting that the first order azimuthal anisotropy can be captured by our inversion.

The supplementary information (Section S1) describes quality control and data processing steps used when measuring splitting and travel times, as well as how we incorporated multiple MCCC datasets for different sub-regions (Section S1.1), along with splitting measurements from the literature (Long et al., 2016; Lynner & Bodmer, 2017; Yang et al., 2017).

2.2 Tomography

We jointly inverted the velocity and anisotropy models using the splitting and differential travel times. It is not feasible to independently resolve the number of parameters required to describe even relatively simple hexagonal anisotropic elasticity. This is largely due to inherent non-linearities and the limits of the data. Instead, we followed the methodology of Eilon et al. (2016) in applying several key assumptions to parsimoniously parameterize the anisotropic elastic tensor (Section S2). This approach reduces the required parameters at each model node to two: the velocities of shear waves traveling vertically with particle motion polarities parallel ($V_{\parallel}$) and perpendicular ($V_{\perp}$) to the anisotropic symmetry axis. These parameters easily translate to isotropic velocity and anisotropy. Based on simplified formulae for Vsh and Vsv (Section S2), we then calculated the differential travel times ($\delta T_{\parallel}$, $\delta T_{\perp}$), and splitting times (dT_{splt}).

We applied *a priori* crustal corrections to account for the influence of known crustal heterogeneity on differential travel times (Section S3; Figure S2; *e.g.,* Sandoval et al., 2004). We additionally solved for event and station static terms for splitting and differential travel times (Section S3). The static terms account for heterogeneity outside the model. We accounted for finite frequency effects using a first Fresnel zone paraxial approximation (Section S4), which is updated from Schmandt and Humphreys (2010) and Eilon et al. (2015).

The discretization and ray geometry influence where the model is reliably recovered. The model space extends from 30 km to 1080 km depth, with inter-node vertical spacing that increases linearly with depth from 40 km to 100 km. Above the depth where rays cross ($\sim$ 70 km), the inversion cannot accurately constrain the depth of heterogeneity. The delay times are still sensitive to structure in this depth range, and structure here will be mapped into station static terms or the shallow portion of the model. We display the 30 km "shallow structure" layer (Figure 3 and 4) because, although structure here is formally not vertically resolved, this layer still illuminates important lateral heterogeneity. The horizontal span of the model is between latitudes 25°N and 46°N and longitudes 96°W and 63°W. This includes a buffer region beyond the seismic array on all sides, required for well-behaved tomography. We do not interpret or display structure in this buffer region. Horizontal node spacing within the seismometer array is 40 km.

We addressed the mixed-determined tomographic inverse problem using smoothed, damped, least squares (*e.g.,* Menke, 2012). This is equivalent to imposing *a priori* assumptions of relatively simple structure and relatively modest perturbations in velocity and anisotropy. We used "L-tests" to determine the appropriate regularization parameters (Section S5; Figure S3).

Anisotropy in the upper mantle is dominantly controlled by the CPO of olivine. This conventionally develops through deformation within the dislocation creep regime. However, with increasing depth, the dominant deformation mechanism of olivine in the mantle transitions from dislocation to diffusion creep (Karato & Wu, 1993). In 1-D Earth reference models, anisotropy (Vsh/Vsv) tends toward zero by about 300 km depth (Chang et al., 2015). This is much shallower than the base of our model. To prevent erroneous mapping of anisotropy to depths where CPO, and hence anisotropic fabrics, is unlikely, our preferred models assume zero anisotropy beneath 300 km depth. For completeness, we also present models where this assumption is relaxed (Figure S4 and S5). As expected, in this case, anisotropy extends deeper than the anticipated dislocation creep regime. However, we find that upper mantle features that we interpret remain.

We evaluated how simplifications regarding wave polarization in the forward model (Section S2) might bias the data fit. We conducted a synthetic splitting test with a more complete parameterization. We propagated Gaussian pulses through our anisotropic model along a given ray path, solving the full Christoffel equation in each layer to find quasi-S wave velocities and polarizations. These calculations utilized the back-azimuths and ray parameters from the actual data. In each layer, we sequentially apply splitting to the wavelet. At the top of the model, we measured splitting parameters on the final synthetic waveform using transverse energy minimization, mimicking the processing of real splitting data (Silver & Chan, 1991). Figure S6 shows one example resulting transverse energy surface. The resulting synthetic fast polarizations closely match observed splitting fast polarizations from the literature over the vast majority of the region (Figure 6). The only notable exception is in the north of our model, where isolated synthetic split fast polarizations approach orthogonal to their literature counterparts. This mismatch is primarily at stations where Long et al. (2016) measured only nulls. Other Splitting measurements in the literature become highly variable here, for example rotating from margin parallel to perpendicular across the MAGIC array just north of our study region (Aragon et al., 2017). Our synthetic splitting delay times are small here ($<\sim 0.3$ s), which is consistent with previous null measurements at those stations, and we place low emphasis on the polarization of the almost null synthetic splits.

3 Results

3.1 Delay times and shear wave splitting

We used MCCC to measure splitting and differential travel times. The well-aligned and linearized waveforms after undoing the effects of splitting indicate success of the approach (Figure S7). Weighted variance reduction for differential travel times ($\delta T_{\parallel}$ and $\delta T_{\perp}$) is 66.4%, and for splitting times (dT_{splt}) is 74.3%. On average, particle motion ellipticity for splitting-corrected shear waves is 51% the original ellipticity. We calculated ellipticity as the ratio of eigenvalues in the particle motion covariance matrix (*e.g.,* Silver & Chan, 1991). Of the 2326 earthquakes, we applied MCCC to 742, yielding $48,428$ delay and splitting measurements (Figures 2 and S8). The remaining earthquakes were rejected based on the quality control criteria (*e.g.*, poor signal-to-noise ratio or poorly aligned waveforms after applying MCCC: Section S1.1).

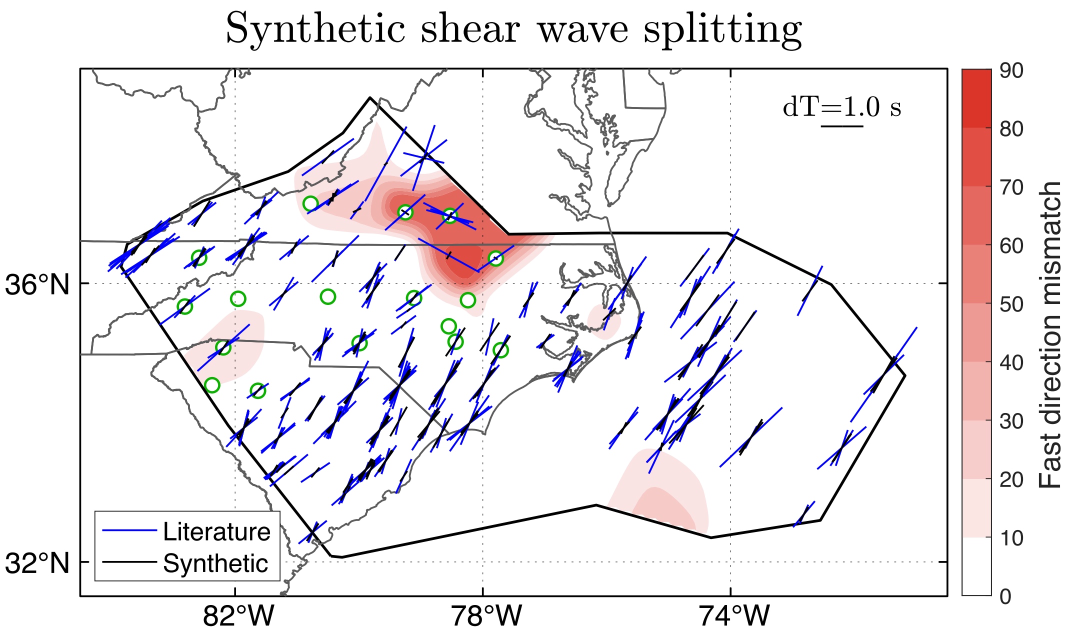

Figure 6. Results of synthetic splitting tests applied to rays for which splitting has been measured in the literature (Long et al., 2016; Lynner & Bodmer, 2017; Yang et al., 2017). We calculated these by first synthetically splitting waveforms and then measuring splitting using transverse energy minimization. This synthetic splitting method is not dependent on the simplifying assumptions regarding waveform polarization from Section 2.2 and S2. The orientations of black and blue lines indicate fast polarizations, and line lengths indicate splitting delay times. Background color indicates the angular misorientation between our synthetic splits and splitting measurements from the literature, which is interpolated to 2-D using a Gaussian filter. Green/white circles indicate stations identified as null splitting stations in Long et al. (2016). We did not apply this analysis where we did not measure splitting, outside the thick black line.

The differential travel times show up to ~ 1.5 s fast arrivals in the Grenville orogenic belt and up to ~ 1.3 s slow arrivals at the Harrisonburg Anomaly (HA) near the Virginia/West Virginia border (Figure 2 and S8). The splitting times indicate dominantly margin-parallel-fast splitting offshore. They also show dominantly margin-parallel-fast anisotropy near North/South Carolina within a region encompassed by small splitting times, consistent with Long et al. (2016). Different from most previous literature, we measured margin-perpendicular splitting over parts of the coastal plains (*e.g.,* Long et al., 2016; Yang et al., 2017). However, these measurements are consistent with the transition from margin parallel to perpendicular splitting moving oceanward across the MAGIC seismometer array north of our study region (Aragon et al., 2017). Splitting measurements can often be resolved only if the delay times are fairly large (*e.g.*, at least 0.5 s splitting using periods longer than 8 s in Long et al., 2016). MCCC improves precision over single station measurements (VanDecar & Crosson, 1990), allowing us to identify splitting trends where splitting times are small.

3.2 Resolution

We conducted synthetic checkerboard resolution tests with approximately 2° sized checkers (Figures S9 and S10). Beneath 300 km, structure is more smeared and reduced in amplitude, but is still recovered. The edges of the plotted model are notably less well resolved than the center by ~ 400 km depth. While velocity heterogeneity is partially resolved beneath 400 km onshore, Figure S9 suggests we should not interpret > 400 km deep velocity anomalies offshore. Anisotropy checkers are well resolved above 300 km.

We calculated semblance, comparing the original and inverted checkerboard models within 260 km radius from each model node (Zelt, 1998). We found semblance > 0.8 corresponds to checkers that we consider well resolved (Figure S9 and S10). We do not interpret, and suggest caution, regarding features where semblance is lower. Hit quality assesses the number of and back-azimuthal distribution of rays throughout model. We do not plot the model where hit quality < 0.7 (this value is roughly consistent with semblance < 0.8 in the shallow part of the model).

We conducted spike tests to further assess resolution (Section S6; *e.g.,* Rickers et al., 2013; Rawlinson & Spakman, 2016). These approximate the impulse response of the inversion (Menke, 2012). We used a spherical spike centered at 165 km depth with 85 km radius (Figure S11 and S12). The output structures had vertical smearing over approximately 200 km (Figure S11 and S12). Velocity and anisotropy amplitude was reduced by $\sim 40\%$. Smearing and reduced amplitude are normal consequences of damping and smoothing. The spike test demonstrates independence of anisotropy and velocity. For the velocity spike test with 0% anisotropy input, the maximum inverted anisotropy magnitude was 0.15%. The anisotropy spike test showed maximum inverted dV of 0.06%. The basic input structure is clearly represented in the output structures, despite some normal distortions.

Squeezing tests indicate the depth to which the data require structure (Section S7; Figure S13). They suggest that velocity heterogeneity is required to at least 660 km depth, which is the limit of our interpretations. Squeezing tests also suggests that noise and velocity-anisotropy trade-off are erroneously mapped toward to base of the anisotropy model. This corroborates our decision to enforce zero anisotropy beneath 300 km depth, the approximate dislocation creep regime depth (Chang et al., 2015).

3.2.1 Synthetic resolution tests for specific features

To understand how specific features of interest might be imaged in our models, we conducted synthetic input-output tests using structures which match different regional predictions (Figure S14 and S15). We included features within our final models which we seek to interpret. Key input velocity features included a fast, thick, Precambrian lithosphere in the northwest of the model and the slow Harrisonburg Anomaly (HA) (Figure S14). The main features we interpret, above about 400 km, are all well resolved. The HA was recovered with smearing over approximately 100 km. High velocity anomalies at the continent-ocean transition (COT) were recovered well. Two deep anomalies within our models, a high velocity anomaly centered near 500 km depth and a low velocity anomaly centered near 800 km, were recovered in shape but with only about 30% of their original amplitude. This suggests caution for interpreting mantle transition zone and deeper mantle features, which are not our focus.

The anisotropy input models included two scenarios (Figure S15). For the offshore region, we tested 1.5% paleo-spreading parallel (margin-perpendicular) fast frozen-in lithospheric anisotropy overlying an equal magnitude margin parallel mantle-flow induced anisotropy. In the continent, we tested a lithospheric layer overlying an asthenospheric layer. This anisotropy could cause previously observed null splitting (Long et al., 2016). Both layers were recovered offshore with approximately 50% amplitude loss and a lateral limit to good lithospheric layer recovery about 200 km from the continent. The continental layers were recovered with similar amplitude loss but with better shape preservation. These tests are strong evidence that first order anisotropic mantle structure, including depth variations, should be faithfully imaged by our models.

3.3 Tomography results: isotropic velocity models

The shear velocity models can be seen in Figure 3, 5, S16, and 3-D models can be viewed interactively using the supplementary linked file `brunsvik-tomog.html`. A prominent fast velocity structure is observed furthest into the continent above about $200-300$ km depth (extending to a maximum depth of ~ 400 km). This structure is as much as 2% fast compared to any layer's average. Within the +1% velocity isosurface of this feature, the mean wavespeed is +1.5%. The shallower ($<\sim$200 km depth) portion of this is the cold, thick, continental interior lithosphere (*cf.* Savage, 2021). However, this feature is near the edge of our array and only its basic structure is clear (Figure S9). The high velocity lithosphere shallows toward the ocean until it meets the most prominent slow-velocity feature in our model, in Virginia ($\sim 38.5°$N, 79°W). This is the previously imaged low-velocity Harrisonburg anomaly (HA) (*e.g.,* Shen & Ritzwoller, 2016; Savage et al., 2017; Wagner et al., 2018). This feature dips oceanward from the surface. It is up to $\sim 5\%$ slow at 70 km depth. The average velocity within the -1% slow isosurface of the HA is -1.9%. Oceanward of the HA, from ~ -100 to 350 km horizontally in Figure 5, we observe a low velocity anomaly just above the 410 km transition zone. This feature appears to connect to the HA.

Several features in the model correspond with magnetic and gravity anomalies. A high velocity feature (up to $\sim 2\%$ fast at 70 km depth) in southern Georgia closely follows the trend of the Brunswick Magnetic Anomaly (BMA) and South Georgia Rift (SGR) (Figure 1 and 3). Beneath the OBSs, upper mantle velocity tends to be slower than on the continent. The offshore 70 km layer is 0-2% slow compared to the whole layer average. A low velocity band above ~ 100 km closely follows the trend of the East Coast Magnetic Anomaly (ECMA) and Positive Gravity Anomaly (PGA). This is in the better resolved portion of the offshore region, though resolution of such a fine structure is suspect given our recovery tests (Figure S9). We also note a low velocity feature, near the edge of the array and thus likely poorly resolved, that correlates with the Blake Spur Magnetic Anomaly (BSMA) (Figure 3). With caution regarding reduced ray coverage offshore, increased delay/splitting noise, and synthetic test results (Figure S9), we focus our interpretation on only the dominant trends offshore. Some oceanic structures may be artifacts at the edge of our seismometer array. For instance, the nearly 3% slow anomaly at 72°W, 35°N, and 165 km depth is likely an artifact (Figure 3 and S16).

We also observe anomalies deep in the mantle. Checkerboard tests (Figure S9) suggest not to interpret the low velocity anomalies at 545 km depth offshore of Georgia and Florida. These features are outside the semblance > 0.8 contour (Figure 3 and S9). The $< 3\%$ fast velocity anomaly that is strongest beneath Tennessee near ~ 400 km depth has been previously imaged (*e.g.,* Schmandt & Lin, 2014; Biryol et al., 2016). We do not interpret the strong anomalies beneath about 660 km, which are less well resolved (Figure S9) and may be a result of using steeply incident *SK(K)S-PKS* rays. Nevertheless, other body wave tomography models similarly show strong anomalies at such depths here (*e.g.,* Schmandt & Lin, 2014; Golos et al., 2018; Wang et al., 2019).

3.4 Tomography results: anisotropy models

The anisotropy models can be seen in Figure 4, 5, S16, and interactively in the supplemental linked file `brunsvik-tomog.html`. As a key result, we observe two layers of anisotropy, both onshore and offshore (Figure 5). Deeper than $\sim 100-150$ km offshore, approximately within the asthenosphere, anisotropy is dominantly margin parallel (generally $> 1\%$ fast). In the offshore lower lithosphere, anisotropy is generally margin-perpendicular/paleo-spreading parallel, up to about 0.8% fast. Our results do not place depth constraint on upper lithospheric anisotropy, and are instead primarily sensitive to the lower lithosphere. The cross-section in Figure 5 runs through the center of the OBSs to give the most reliable sense of offshore anisotropy. However, lithospheric/asthenospheric layering becomes increasingly inconsistent away from the cross-section, where hit quality and resolution decrease (Figure 4). We suggest the model is strong evidence for lithosphere-asthenosphere anisotropic layering.

On the continent, our model shows margin-parallel-fast anisotropy in the asthenosphere ($> 1\%$ fast at 275 km depth) (Figure 4 and 5). This is consistent with dominantly margin parallel splitting from previous work (*e.g.,* Yang et al., 2017). At 165 km depth and above, the model shows some margin-perpendicular anisotropy up to almost 1% fast in the Piedmont and coastal plain (North/South Carolina and Virginia). This is the same region where Long et al. (2016) observed dominantly null splitting. This shallower anisotropy is complex, pocketed with ~ 100 km wavelength features. Margin-perpendicular-fast anisotropy is consistent with the *Pn* analysis of Buehler and Shearer (2017), which indicates margin-perpendicular-fast anisotropy just beneath the Moho in the coastal plain. This is also consistent with anisotropic surface-wave phase velocities in the low topography region east of the Appalachians, which rotate from margin parallel for periods longer than about 77 s to margin perpendicular for periods between about 77 s and 40 s (Wagner et al., 2018). The continental lithosphere has complicated anisotropy, while asthenospheric anisotropy is dominantly margin parallel.

4 Discussion

Our shear velocity and anisotropy models inform hypotheses of rift and drift dynamics, as well as interpretations of present day structures and processes. We first discuss the velocity and anisotropy structures associated with rifting. Second, we discuss the transition from rifting to drifting. Third, we discuss processes and structures which likely occurred and developed during and after the formation of the passive margin. Finally, we discuss the complex relationship between strain and anisotropy. Our observations and interpretations are summarized in Figure 7.

4.1 Rift structure

Our velocity models show thick Precambrian lithosphere in the northwest, which thins toward the margin (Figure 3 and 5). At the HA, the lithosphere is greatly thinned. However, the precise depth of the lithosphere-asthenosphere boundary is not easily es-

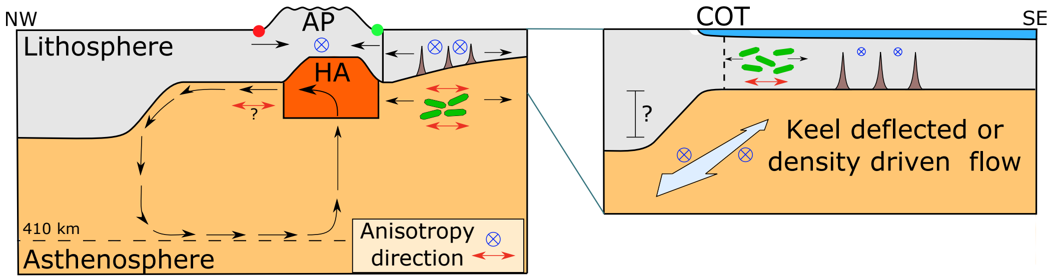

Figure 7. Schematic illustration of the structure and kinematics of ENAM. Red arrows and blue crosses indicate the orientations of anisotropy. Black arrows indicate previous or current extension, compression, or mantle flow. Brown dikes are illustrated in the lithosphere. Olivine crystals are illustrated with CPO. Red and green dots correspond to the same regions as in Figure 3 and 5. CPO: crystallographic-preferred orientation. AP: Appalachian plains. HA: Harrisonburg Anomaly. COT: continent-ocean transition.

tablished using body-wave tomography, which has less sensitivity to variation in velocity with depth than, for example, receiver functions (*e.g.,* Yuan et al., 2017; Liu & Gao, 2018).

Near the South Georgia Rift (SGR), we observe a shallow, high velocity feature, about 2% fast ($\sim 81°$W, 33°N, 70 km depth) (Figure 3). This distinctly follows the Brunswick positive magnetic anomaly (BMA) and more subtly follows positive gravity anomalies. Wide-angle seismic results suggest that across the failed Georgia rift, high velocity material intruded the lower crust as part of the Central Atlantic Magmatic Province (CAMP) (Marzen et al., 2020). The imaged fast velocity feature is potentially crustal CAMP underplating, with recovery smeared vertically through the lithosphere according to synthetic tests (Figure S9 and S14). This fast feature extends from the SGR to the continent-ocean transition (COT), and arguably, to some extent, delineates the COT. The connection of the BMA- and SGR-correlated fast velocity anomaly to more fast velocity anomalies delineating the COT suggests that these fast features are of the same cause. Possibly, igneous material from the initial phase of rifting, now solidified, eventually localized from the failed SGR to the COT as part of continental break-up. Alternatively, the fast features near the BMA might be related to an Alleghanian suture (*e.g.,* Lizarralde et al., 1994; Hopper et al., 2016). This fast feature was not clearly resolved in recent mantle tomography models, possibly due to the exclusion of margin-crossing data (*e.g.,* Biryol et al., 2016; Wagner et al., 2018; Savage, 2021).

Our models show laterally complex anisotropy on the continent, above ~ 200 km depth (Figure 4 and 5). Many rifts exhibit extension-perpendicular (rift-parallel) anisotropy (*e.g.,* Vauchez et al., 1998, 2000; Kendall et al., 2005; Eilon et al., 2014). This is in part due to shape-preferred orientation structures such as dike intrusions and melt lenses. However, if extensional strain and consequent CPO dominates anisotropy, then extension parallel splits should be seen (Tommasi et al., 1999). This is indeed observed in some well-developed rifts (Eilon et al., 2014, 2016) and at mid-ocean ridges (Wolfe & Solomon, 1998). We see a mix of margin parallel anisotropy (*i.e.*, approximately parallel to the Appalachians) and margin perpendicular anisotropy (*i.e.* approximately parallel to spreading) near the Piedmont and coastal plains (Figure 4 and 5). This is consistent with Aragon et al. (2017). Extension-induced CPO, now frozen in, may explain the lithospheric extension-parallel anisotropy (*e.g.,* Tommasi et al., 1999). Fossil-melt shape-preferred orientation (SPO) may explain some margin parallel anisotropy. However, igneous SPO at rifts is attributed to the strong velocity contrast between melt and host material (*e.g.,* Kendall et al., 2005). This velocity contrast is strongly reduced once melt solidifies. Igneous SPO is a poor candidate for explaining present-day anisotropy, except perhaps near the HA, where melt may be present.

The Appalachians also exhibit convergence-induced anisotropy that is frozen-in (*e.g.,* Long et al., 2016). Splitting is dominantly margin/orogen parallel at the west border of our anisotropic model, where rift deformation is less prominent (Figure 4). The competing influence of convergence, extension, and possibly igneous SPO can produce the complex lithospheric anisotropy we imaged (Figure 4). Such complexity is further expected from laterally heterogeneous volcanism (*e.g.,* Greene et al., 2020) and extension (*e.g.,* Withjack & Schlische, 2005).

Orthogonal and effectively cancelling anisotropic layers is one proposed explanation for null splitting in portions of the continental coast (Wagner et al., 2012; Long et al., 2016). We imaged variations in anisotropy with depth, and in particular, the transition to margin parallel in the asthenosphere (Figure 5). Depending on the local magnitudes of anisotropic layers, this is capable of causing null splitting. However, our results cannot rule out the contributions of vertical mantle flow to null splitting measurements because we assumed a horizontal symmetry axis.

4.2 Rift-drift transition

4.2.1 Lithospheric structure

Recent work suggests that proto-oceanic crust was emplaced during the transition from rifting to drifting between the East Coast magnetic anomaly (ECMA) and Blake Spur magnetic anomaly (BSMA) (Shuck et al., 2019; Bécel et al., 2020). We observe a shallow, $\sim$ 1.5% slow velocity anomaly above $\sim$ 100 km at the ocean which parallels the ECMA ($\sim$ 74°W, 35°N) (Figure 3). No such feature or trend manifests in the anisotropy model. The gradient in velocity between fast values inboard of the ECMA and slow values seaward of this lineament is suggestive of a relatively rapid contrast in lithospheric thickness (Figure 3, 5, and 7). This is consistent with localized crustal thinning observed by Lynner and Porritt (2017) and Li and Gao (2019). We are hesitant to over-interpret small-scale velocity anomalies beneath the ocean in our models. Synthetic tests indicate that resolution here is relatively poor (Figure S9 and S14). This slow anomaly may reflect persistent continental crust remnant from the rifting process, but that assertion requires further investigation.

There is limited detailed imaging of passive margins at a lithospheric scale to compare our results. In NW Namibia, receiver functions indicate a lithospheric feature that thins from 120 km to 80 km at the COT. Thinned lithosphere at rifted COTs may be a common theme (Figure 7). Because the private African dataset is the only other broadband OBS dataset to cross a rifted passive margin, our results are some of the first detailed mantle-scale seismic models of a rifted COT.

4.2.2 Preserved extensional fabric in transitional-oceanic lithosphere

Globally, anisotropic fast directions within the oceanic lithosphere tend to align with paleo-spreading (*e.g.,* Wolfe & Solomon, 1998; Becker et al., 2014). This supports the notion that upper mantle anisotropy develops parallel to plate motion and is then frozen within the lithosphere. Although the nearby Cretaceous Atlantic lithosphere shows spreading-parallel fast lithospheric anisotropy (Gaherty et al., 2004), splitting measurements offshore at ENAM are instead margin-parallel-fast (Lynner & Bodmer, 2017). This could corroborate many other studies showing mismatch of anisotropy to paleo-spreading and absolute-plate-motion (*e.g.,* Dunn et al., 2005; Takeo et al., 2016; Eilon & Forsyth, 2020).

Synthetic tests (Figure S15) suggest that simple layering is resolved in our models, albeit with reduced amplitudes. We image margin parallel anisotropy in the asthenosphere rather than lithosphere ($> 1\%$ anisotropy beneath ~ 100 km). The asthenospheric anisotropy produces the surprising offshore splitting measurements (Figure 5). Margin parallel flow can explain the asthenospheric anisotropy (Section 4.3).

We imaged approximately paleo-spreading-parallel anisotropy in the offshore lower lithosphere (Figure 4 and 5). This supports that the oceanic lithosphere preserves crystallographic-preferred orientation (CPO) of olivine which developed parallel to mid-ocean ridge spreading (*e.g.,* Becker et al., 2014; Russell et al., 2019). We find this layering persists in our model, independent of regularization scheme or whether we assume no anisotropy beneath 300 km (Figure S4 and S5). Synthetic tests (Figure S10 and S15) demonstrate that this result is robust, though we only expect to resolve heterogeneity beneath ~ 70 km. Above this, it is possible that anisotropy rotates to be paleo-spreading perpendicular (*e.g.,* Shuck & Van Avendonk, 2016).

Offshore ENAM, there are only global seismic models for comparison in the mantle. Although global models are highly variable, they tend to show instead margin perpendicular anisotropy in the asthenosphere and have little consensus in the lithosphere (see compilation of Schaeffer et al., 2016). Our models capture the offshore anisotropic layers ultimately absent in global models, demonstrating the importance of utilizing broadband OBSs to accurately characterize the oceans.

4.3 Active mantle processes at the passive margin

The causes of low velocity anomalies at ENAM, in particular the prominent Harrisonburg anomaly (HA), are subject to debate (*e.g.,* Chu et al., 2013; Mazza et al., 2014). High temperature and possibly partial melt may cause the HA. Savage (2021) estimated up to 2% melt based on the magnitude of their inverted Vs anomaly, and our velocity anomaly is of similar magnitude (about 5% slow). An abrupt increase in attenuation at the HA (Byrnes et al., 2019), high conductivity (Evans et al., 2019), and coincidence with 48 Ma volcanics (Figure 1; Mazza et al., 2014) also suggests the presence of partial melt. The HA is associated with a dynamic topography anomaly, which likely resulted from buoyant mantle (Ramsay & Pysklywec, 2011; Rowley et al., 2013). Receiver functions indicate thinned lithosphere (Evans et al., 2019). Our models add to a preponderance of evidence that there is a present-day mantle upwelling that significantly perturbs the lithosphere beneath Harrisonburg, VA.

Edge-driven convection (EDC) is density-driven flow that is excited by strong lateral gradients in temperature at the edge of cold, continental lithosphere (*e.g.,* King & Ritsema, 2000; Shahnas & Pysklywec, 2004; King, 2007). This process may be important at ENAM (*e.g.,* Ramsay & Pysklywec, 2011; Menke et al., 2016). Some have conjectured that the HA represents the low wavespeed, low density upwelling limb of EDC along the margin (*e.g.,* Savage et al., 2017; Byrnes et al., 2019). Our models are consistent with this theory. Despite relative tectonic quiescence in this region, the HA is the slowest feature in the models, suggesting active processes must maintain a velocity contrast. The presence of well established, high velocity lithosphere beginning ~ 400 km northwest from this feature is consistent with a cold, thick lithospheric edge where the downwelling limbs of convective cells could originate (Figure 5). Unfortunately, anisotropic coverage is limited where we interpret EDC. We are cautious to evaluate EDC mantle flow using the anisotropy model. However, we do not detect strong azimuthal anisotropy at the HA (Figure 5). This is (non-uniquely) consistent with EDC upwelling (*e.g.,* Long et al., 2010). These structures match the geodynamic setting for EDC (*e.g.,* King & Anderson, 1998).

EDC upwellings can occur in laterally isolated cells (Ramsay & Pysklywec, 2011). Our model shows similar, lower amplitude anomalies elsewhere along the margin (Figure 3). The presence of a low velocity anomaly just above the 410 km mantle transition zone, southeast of the HA (Figure 5), may further be associated with EDC. This could result from a convection cell or upwelling feature between the COT and the HA. Some 3-D EDC models might predict similar features (Kaislaniemi & Van Hunen, 2014). Modelling of the analogous African margin (Kaislaniemi & Van Hunen, 2014) also suggests a margin-parallel component of flow is possible with EDC. This could explain some disconnect between the expected margin perpendicular convective flow and anisotropy. The EDC-like low-velocity anomalies imaged here are in addition to the low velocity Georgia anomaly (only peripherally imaged here: Biryol et al., 2016), a low velocity mantle anomaly in Texas (Pollitz & Mooney, 2016), and the Northern Appalachian Anomaly (Menke et al., 2016). EDC is an attractive theory for explaining a variety of discontinuous, short-wavelength, upper mantle velocity features imaged here and elsewhere without invoking multiple processes (Menke et al., 2016).

There are other possible causes of the HA. Although fertile mantle with reduced Mg# can decrease Vs, a reasonable Mg# likely only contributes -1% dVs (Pollitz & Mooney, 2016). Volatiles, possibly originating from the subducted Farallon slab (Van Der Lee et al., 2008), could also reduce velocity. However, velocity reduction is likely less than 3% (Pollitz & Mooney, 2016; Savage, 2021), and the anomaly is at least 5% slow in our models. Plume presence (Chu et al., 2013) may not be supported. We see no low velocity plume track connected to the HA in our results and others (Pollitz & Mooney, 2016), and melting temperatures were too cold (Mazza et al., 2014).

Another frequently invoked geodynamic process beneath ENAM is delamination of the lithosphere. In this scenario, the HA results from asthenospheric return flow (*e.g.,* Mazza et al., 2014; Biryol et al., 2016; Byrnes et al., 2019). Previous delamination may have carved the lithospheric gap and promoted EDC (*e.g.,* Byrnes et al., 2019). Similarly, if a plume did erode the lithosphere, the modified lithospheric topography may have promoted asthenospheric inflow or EDC (Tao et al., 2021).

The increase in lithospheric thickness moving into the craton (the keel) might have an important influence on mantle flow. The keel can redirect horizontally flowing mantle around the continent, producing keel-parallel flow. Some splitting trends in the continent have been attributed to this phenomenon (*e.g.,* Fouch et al., 2000; Yang et al., 2017). Our model shows margin parallel asthenospheric anisotropy well within the continent (Figure 5), consistent with keel-deflected flow. Similarly, we note that a step in the thickness of the lithosphere at the COT could promote margin parallel flow (*e.g.,* Wang & Becker, 2019).

4.3.1 Offshore asthenospheric flow and anisotropy

Density-driven flow may also contribute to margin parallel anisotropy offshore ENAM (Lynner & Bodmer, 2017). Globally, asthenospheric anisotropy beneath the oceans tends to align with plate motion, with maximum match at ~ 200 km depth (Becker et al., 2014). Limited data makes this trend difficult to assess at rifted continent-ocean transitions. In the asthenosphere, our model can resolve the first order anisotropic structure (Figure S15). The model shows instead anisotropy perpendicular to current plate motion and paleo-spreading (Figure 1 and 5).

Plate motion has only a partial control on asthenospheric shear, and inclusion of density-driven flow is needed to explain anisotropy in much of the oceanic asthenosphere (Becker et al., 2014). Density driven flow could help explain oceanic margin-parallel anisotropy seen in the deeper layers of our models (Lynner & Bodmer, 2017). The two layer lithosphere-asthenosphere mantle flow model of Wang and Becker (2019) predicted margin-perpendicular splitting offshore. By adding 3-D flow driven by density anomalies, splitting becomes more margin parallel (Wang & Becker, 2019). Some other density-driven mantle flow models also show approximately margin parallel flow here (*e.g.,* Rowley et al., 2013).

4.4 Relationship between anisotropic fabric and strain

The fast polarization of splitting is usually assumed to indicate modern mantle deformation (*e.g.,* Zhang & Karato, 1995). However, recent experiments have revealed several complexities in mantle CPO development (*e.g.,* Skemer & Hansen, 2016) that require consideration of time-integrated strain patterns (Kaminski & Ribe, 2002). For example, static annealing can modify otherwise steady CPO through time (Boneh et al., 2017). Our model does show some paleo-spreading perpendicular anisotropy in the offshore lithosphere (Figure 4), albeit where resolution is reduced (Figure S9). Since this is nearly 200 ma lithosphere, static annealing may partially account for reoriented CPO.

If CPO is already present, then overprinting fabrics to reflect changed asthenospheric flow can require substantial strain, sometimes up to several hundred percent (Skemer et al., 2012; Boneh & Skemer, 2014; Boneh et al., 2015). For small strain, CPO may be in a transient state and not reflect modern asthenospheric flow in a simple way. CPO may similarly be in a transient state if asthenospheric flow orientation changes over small spatial and temporal scales (Kaminski & Ribe, 2002; Skemer et al., 2012). Anisotropy may not clearly reflect asthenospheric flow in convective systems spanning short distances, or where mantle flow changes through time, such as EDC (*e.g.* Kaislaniemi & Van Hunen, 2014). In the asthenosphere, our anisotropy model shows some heterogeneity at wavelengths down to ~ 100 km (Figure 4). We speculate that flow at this scale might have produced transient state anisotropy with fast orientations not clearly reflecting modern mantle flow. Northwest of the HA, within the asthenosphere, we predict margin-perpendicular EDC to produce margin-perpendicular anisotropy. However, we observe complicated, yet more dominantly margin parallel, anisotropy. This may be a result of transient-state CPO. In contrast, for larger-scale margin parallel asthenospheric flow, particularly beneath the ocean, CPO should reach steady state and produce margin parallel anisotropy. This matches the more strongly margin-parallel asthenospheric anisotropy offshore (Figures 3 and 4).

5 Conclusion

We present S-wave tomography models from a passive broadband dataset spanning the continent-ocean transition of the eastern North American rifted margin (ENAM). Our inversion technique places depth constraints on isotropic and anisotropic structures. It also resolves trade-offs present in single-parameter inversions by simultaneous fitting of travel time and shear wave splitting data. The resultant models provide the first high-resolution images of seismic velocity and azimuthal anisotropy to sub-lithospheric depth across the margin.

Offshore, we find that the rifted continental to oceanic lower lithosphere preserves extension-parallel anisotropy. Onshore, complex lithospheric anisotropy likely reflects the competing effects of extension and convergence over several Wilson cycles. In the asthenosphere, margin parallel anisotropy dominates. This may reflect mantle flow due to density gradients or pressure gradients associated with a step in lithospheric thickness. Isotropic velocities within the continent show the thick, high-velocity continental keel inboard of the Appalachians and the low-velocity Harrisonburg Anomaly associated with Eocene volcanics. This latter feature, together with other small-wavelength velocity anomalies, are consistent with edge-driven convection and other active mantle flow processes at the passive margin.

These results, made possible by an unusual amphibious broadband dataset, demonstrate the dynamic and complex nature of mantle processes at the rifted continent-ocean transition. This study, together with other products of the ENAM-CSE, reinforces the importance of shoreline-crossing instrumentation.

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