The relative roles of inheritance and long-term passive margin lithospheric evolution on the modern structure and tectonic activity in the southeastern United States

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ABSTRACT

We perform inversions for the shear-wave velocity structure of the southeastern United States (SEUS) using Rayleigh-wave phase and amplitude data from the broadband stations of the South Eastern Suture of the Appalachian Margin Experiment (SESAME) and EarthScope Transportable Array (TA). Our tomographic images of shear-wave velocities in the upper mantle beneath the SEUS provide new constraints on the evolution of mantle lithosphere, both from the inheritance of structures from repeated Wilson cycles and from processes that have occurred while at a passive margin setting. Our images also allow us to correlate these structures to evidence of Eocene to recent tectonism observed at the surface. We find evidence for both inherited structures and more recently evolved structures, both of which bear some correlation to observations of ongoing tectonism. Our results suggest that lithospheric mantle continues to evolve while in a passive margin setting and that even relatively “stable” continental mantle lithosphere is subject to episodes of delamination, foundering, and erosion due to processes that are still not well understood. Our results provide structural constraints on the types of processes that may be ongoing and on possible explanations for the numerous observations of comparatively recent tectonic activity occurring along this passive margin setting.

INTRODUCTION

The east coast of North America is an archetypal passive margin and, therefore, an obvious locale for the study of continental lithosphere that is neither actively evolving at a plate boundary nor part of a cratonic continental core. The southeastern United States (Fig. 1) in particular is an ideal location to study the temporal evolution of continental lithosphere. It is the product of repeated Wilson cycles, composed of multiple accreted terranes of different ages, subsequently exposed to flood basalt volcanism and failed rifting prior to the opening of the Atlantic Ocean and Gulf of Mexico. Here, we broadly define the southeastern United States as extending south from the latitude of the Mason-Dixon line (northern Maryland, ~40°N) and as far west as western Tennessee and Kentucky (~89°W).

The extent to which the modern lithospheric mantle beneath the southeastern United States (SEUS) is dominated by structures inherited from past episodes of convergence and rifting has long been the subject of research (e.g., Vauchez and Barruol, 1996; Cook and Vasudevan, 2006; Thomas, 2006). Surface-wave tomography is ideally suited for imaging lithospheric structure due to its sensitivity to velocity changes in the crust and uppermost mantle, especially as a function of depth. This allows us to look for evidence of possible inherited structures, such as the edge of the cratonic keel, dipping shear zones, or petrological discontinuities in the mantle preserved from past accretionary events, relict restite produced by the formation of flood basalts, or thinned lithosphere from past rifting events. The location and geometry of these structures can shed light on the details of the tectonic history of the region. Another key question is how these types of inherited structures relate to evidence for postrift instability and ongoing tectonism. Here, we present the results of our inversions of teleseismic Rayleigh-wave phase and amplitude data to image the shear-wave velocity structure of the uppermost mantle beneath the SEUS. We find evidence of both inherited structures and structures that do not obviously correspond to known accretionary, collisional, or rift geometries. These structures suggest that lithospheric evolution does not end with the transition to a passive margin and that all but the oldest continental interiors may not be as stable as previously assumed.

Tectonic History

The geologic history of our study area begins with the formation of Rodinia, the supercontinent that juxtaposed the Granite Rhyolite Province of the Laurerntian basement against cratons found in present-day South America, Africa,
Figure 1. Map of study area and surrounding region. Background colors show topography and bathymetry. Green circles show earthquakes from the U.S. Geological Survey earthquake catalog from 2007-2017. Purple circles show large events (M >4) sized according to the magnitude scale shown at the bottom left. Magenta outlines show Grenville rift basins from Whitmeyer and Karlstrom (2007). The brown line indicates the location of the Grenville Front (Whitmeyer and Karlstrom 2007). Red triangles show the locations of Eocene volcanics (Mazza et al., 2014, 2017). Yellow dashed line shows the location of the Blue Ridge Escarpment, and the green hexagon indicates the location of Mount Mitchell (highest peak east of the Mississippi). Colors in eastern South Carolina and North Carolina show exhumed sediment patterns associated with the Cape Fear Arch (Soller, 1988). Dark-gray shaded region shows the location of the South Georgia Rift Basin (McBride, 1991). Dashed line with red diamonds indicates the location of the Fall Line and associated river crossings and waterfalls. Dotted line shows the Suwannee Suture Zone (Higgins and Zietz, 1983). Blue line shows the approximate location of the Central Piedmont Shear Zone (Hatcher et al., 2007). Red line shows the location of the Brunswick Magnetic Anomaly (e.g., Daniels et al., 1983). Abbreviations: RR—Reelfoot Rift; RT—Rome Trough; CPS—Central Piedmont Shear Zone; SCSZ—South Carolina Seismic Zone; CVSZ—Central Virginia Seismic Zone; ETSZ—Eastern Tennessee Seismic Zone; SGR—South Georgia Rift Basin; GF—Grenville Front; SSZ—Suwannee Suture Zone.
and Antarctica during the Grenvillian Orogeny (1.35–1.1 Ga) (Thomas and Astini, 1996; Loewy et al., 2003; Tohver et al., 2004; Thomas, 2006; Whitmeyer and Karlstrom, 2007; Loewy et al., 2011). Deformation from this collision extended as far westward as the Grenville Front (Fig. 1). Subsequent rifting (760–530 Ma; Thomas, 2006; Whitmeyer and Karlstrom, 2007) resulted in the opening of the Iapetus Ocean. It also produced a number of failed rifts, including the Rome Trough and Reelfoot Rift (Fig. 1) (Whitmeyer and Karlstrom, 2007).

Subsequent to Iapetus rifting but prior to the formation of Pangaea, the southeastern Laurentian margin (current orientation) was subjected to a sequence of accretionary events resulting in the addition of the Inner Piedmont and Carolina terranes (also known as Carolina) (Hibbard, 2000; Miller et al., 2006; Anderson and Moecher, 2009; Hatcher et al., 2017). The exact number of accreted terranes and associated orogenic events is still debated (Hibbard et al., 2002; Hatcher et al., 2007; Hatcher, 2010; Hibbard et al., 2010), as are the locations of some of the terrane boundaries and vergence directions of associated subduction zones (Merschat and Hatcher, 2007; Anderson and Moecher, 2009). There is consensus on the location of the exposed suture between the Carolina and Inner Piedmont terranes along the Central Piedmont Suture (CPS) (Fig. 1). However, the subsequent Alleghanian Orogeny is believed to have formed an extensive thrust sheet, 5–10 km thick, across much of the SEUS that translated the uppermost crust tens to hundreds of kilometers inland (e.g., Hatcher, 1972; Cook and Vasudevan, 2006; Hopper et al., 2017), making it difficult to connect terrane boundaries at the surface, to basement terrane boundaries at depth.

The collision of Gondwana and Laurentia and the formation of Pangaea during the Alleghanian Orogeny at ca. 350–300 Ma produced the massive Appalachian orogen. The location of the suture is believed to lie offshore or beneath the easternmost continental margin along most of the eastern seaboard (McBride and Nelson, 1988; Kellner and Hatcher, 1999; Cook and Vasudevan, 2006), except in the south, where the accreted Gondwanan-affinity Suwanee and related terranes were left behind after rifting. The location of the Suwanee suture has often been placed along the Brunswick Magnetic Anomaly (BMA) (Fig. 1), which crosses southern Georgia and Alabama (e.g., Chown and Williams, 1983; Daniels et al., 1983; Nelson et al., 1985a, 1985b; McBride and Nelson, 1988; Mueller et al., 1994; Heatherington and Mueller, 1999; Heatherington and Mueller, 2003). However, recent work has revisited another hypothesis (e.g., Higgins and Zietz, 1983): that the Alleghanian margin between Laurentia and Gondwana lies farther north (e.g., Mueller et al., 2014; Boote and Knap, 2016; Hopper et al., 2017) and that the BMA represents an intra-Gondwanan suture (Boote and Knap, 2016). The degree and timing of convergence versus strike-slip motion on the transpressional Suwanee Suture Zone (SSZ) is also still debated (e.g., Mueller et al., 2014; Boote and Knap, 2016; Hopper et al., 2017).

The breakup of Pangaea began in the Middle to Late Triassic with the development of extensive rift basins across Georgia (the South Georgia Rift Basin) (Fig. 1) (e.g., Chown and Williams, 1983; McBride, 1991) and a series of smaller basins along the east coast (also known as Newark Super Group rifts) (e.g., Reinemund, 1955; Olsen et al., 1991; Schlische, 1993). This failed rifting episode culminated with the extensive flood basalt volcanism of the Central Atlantic Magmatic Province (CAMP) at ca. 190 Ma (e.g., Marzoli et al., 1999; McHone, 2000; Whalen et al., 2015). Rifting resumed in the earliest Jurassic, and the lithosphere of the SEUS transitioned from rift to drift by the mid-Jurassic (e.g., Withjack et al., 1998). Despite this transition to a passive margin setting, the eastern United States has continued to exhibit tectonic activity including deformation, uplift, seismicity, and volcanism.

**Evidence of Ongoing Tectonism**

**Blue Ridge Topography**

In the southeastern United States, the Blue Ridge Mountains are what remain of the Andean-scale orogen produced during the collision between Gondwana and Laurentia in the Pennsylvanian–Permian Alleghanian Orogeny. The persistence of both the total elevation of the Blue Ridge (which includes Mount Mitchell [2037 m], the highest point in the eastern United States) and the pronounced topography, in particular across the Blue Ridge Escarpment in North Carolina and southern Virginia, is surprising given models that predict the elimination of topographic relief within ~100 m.y. of the end of convergence (e.g., Ahnert, 1970; Tucker and Slingerland, 1994). A number of competing theories exist on the persistence of elevation and steep topography. These range from theories that require no postorogenic uplift and posit persistent topography due to dynamic equilibrium and lithologic variability (Hack, 1980; Baldwin et al., 2003; Matmon et al., 2003) to those that do require recent uplift (e.g., Hack, 1982), which may be due to various epeirogenic processes (e.g., Pazzaglia and Brandon, 1996; Fischer, 2002; Moucha et al., 2008; Flament et al., 2013; Linari et al., 2016) or climatic variability (e.g., Molnar, 2004).

One intriguing line of evidence comes from recent studies that have found migrating river knickpoints and relict topography consistent with hundreds of meters of relative base-level fall in the past 3.5–15 m.y. (Gallen et al., 2013; Miller et al., 2013; Prince and Spotila, 2013). The cause of these knickpoints has been variably attributed to a renewed uplift of the orogen (Gallen et al., 2013; Miller et al., 2013) or stream capture (e.g., Prince et al., 2010) resulting in the retreat of the Appalachian escarpment over time (e.g., Tucker and Slingerland, 1994; Spotila et al., 2004). Such unsteadiness in the postrifting landscape may be episodic, as indicated by thermochronology data that provide evidence for similar rejuvenation of topography in the Late Cretaceous (McKeon et al., 2014).

**Cape Fear Arch**

Another example of comparatively recent deformation in the southeastern United States is the uplift of the Cape Fear Arch, sometimes also referred to as the mid-Carolina Platform High (e.g., Riggs and Belknap, 1988) (Fig. 1). While
this arch has no topographic expression, it was first identified on the basis of river valley morphology in the late nineteenth century (e.g., Kerr, 1875). The arch is typically described either as an upwarping of basement topography as determined by active source seismic profiles and borehole data (e.g., Hersey et al., 1959; Bonini and Woollard, 1960; Soller, 1988), as patterns of exhumed late Cenozoic sedimentary layers that indicate pronounced erosion along the axis of the proposed ridge (e.g., Riggs and Belknap, 1988; Soller, 1988), and/or as the displacement of river channels in both North and South Carolina as the result of Pleistocene to modern uplift (e.g., Soller, 1988; Baldwin et al., 2006; Bartholomew and Rich, 2012). Uplift rates are weakly constrained but are conservatively estimated to have ranged between ~0.006 mm/yr from 2.75 to 1.75 Ma, increasing to ~0.04 mm/yr over the past 100,000 years (Cronin, 1981; Soller, 1988; Gardner, 1989 and references therein).

The axis of the Cape Fear Arch lies along the Cape Fear River, roughly parallel to the South Carolina–North Carolina border. The geometry of this uplift does not correlate with known terrane boundaries, gravity anomalies, or magnetic anomalies. There are few theories to explain the uplift of the Cape Fear Arch. Morgan (1983) attributed the arch to the passing of the Bermuda Hot Spot during the Paleocene, though this would not explain the ongoing deformation observed (Vogt, 1991). Others have postulated the existence of a Cape Fear fault (e.g., Zullo and Harris, 1979; Bartholomew and Rich, 2012), though direct evidence for discrete faulting is largely absent. The Cape Fear Arch also coincides with uplifted regions of the Orangeburg Scarp, a mid-Pliocene shoreline that extends from central North Carolina to northern Florida. Recent work has indicated that some of this uplift might be explained through dynamic topography associated with mantle upwelling (e.g., Rowley et al., 2013; Liu, 2015; Rovere et al., 2015), although this alone cannot explain the relatively short wavelength variations in uplift observed in the Carolinas.

**Eocene Volcanism**

The formation of the east coast passive margin and the transition from rift to drift was complete by 175 Ma (Withjack et al., 1998). However, two distinct pulses of volcanism have been identified in western Virginia and eastern West Virginia that postdate the removal of the SEUS from proximity to a plate boundary. The first pulse comprises Late Jurassic (ca 145 Ma) alkaline volcanics that have variably been described as adakites (Meyer and van Wijk, 2015) or phonolites (Mazza et al., 2017). The second pulse occurred 125 m.y. after the completion of rifting. These Eocene basalts (Fig. 1) have modeled melt equilibration conditions of 2.32 ± 0.31 GPa (~77 ± 9 km depth) and 1412 ± 25 °C (Southworth et al., 1993; Mazza et al., 2014, 2017) and represent the youngest known magmatism in the eastern United States. Theories on the cause of this volcanism include the presence of a basement fracture zone along the 38th parallel (e.g., Fullagar and Bottino, 1969; Dennison and Johnson, 1971), the passing of a hotspot track (Chu et al., 2013), and the delamination of eclogitized lower crust or mantle lithosphere (Mazza et al., 2014, 2017).

**Significant Earthquakes and Variable Seismicity Patterns**

Despite its passive margin setting, the SEUS has been the site of several moderate to large earthquakes over the past two centuries. The two most notable within our study area were the M = 7 1886 Charleston, South Carolina earthquake (e.g., Tarr et al., 1981; Talwani, 1982; Cramer and Boyd, 2014; Chapman et al., 2016) and the Mw = 5.8 2011, Virginia event (e.g., Wolin et al., 2012; McNamara et al., 2014) (Fig. 1). These large earthquakes are associated with localized regions of increased seismic activity. Seismicity in the immediate vicinity of the Charleston, South Carolina, event may represent a lingering aftershock sequence, given the locations and focal mechanisms of these events (Chapman et al., 2016). The Virginia earthquake occurred in an area that had previously been identified as a region of moderately increased seismicity known as the Central Virginia Seismic Zone (CVSZ) (e.g., Bollinger, 1973; Çoruh et al., 1988; Kim and Chapman, 2005). In addition to the seismicity clusters associated with these large events, other distinct regions of increased seismicity exist in the SEUS. The Eastern Tennessee Seismic Zone (ETSZ) comprises a concentrated cluster of seismic activity located parallel to the strike of the Appalachian Mountains in eastern Tennessee and westernmost North Carolina (e.g., Powell et al., 1994; Chapman et al., 1997; Powell and Thomas, 2016). Other more subtle seismic patterns are also visible along the east coast passive margin. In particular, there is a discernable decrease in the amount of seismic activity in the Piedmont and coastal plains of North Carolina and southern Virginia relative to areas to the north and south (Fig. 1). Part of this difference is due to increased seismicity in the region of the 1886 Charleston earthquake, but ongoing seismicity is also present over much of central and western South Carolina, parts of northern Georgia, and southwestern North Carolina. This region of increased seismicity is commonly referred to as the South Carolina Seismic Zone (SCSZ) (e.g., Bollinger, 1973; Tarr et al., 1981; Domoracki et al., 1998; Li et al., 2007) and has been incorporated into the U.S. Geological Survey (USGS) hazard maps (Petersen et al., 2008; Petersen et al., 2012). Finally, there is a notable “earthquake shadow” or gap in observed seismicity relative to surrounding areas (e.g., Bollinger and Gilbert, 1974) near the Eocene volcanics in western Virginia and in the central part of eastern West Virginia, just west of the Central Virginia Seismic Zone. A better understanding of lithospheric-scale structures across the SEUS will help to constrain the potential contributing factors associated with ongoing seismicity in passive margin settings.

**DATA**

We use data from the South Eastern Suture of the Appalachian Margin Experiment (SESAME) deployed across Georgia and into North Carolina, Tennessee, and Florida between July 2010 and May 2014 (Fischer et al., 2010) (Fig. 2). SESAME comprised three transects—two north-south–trending transects through the eastern and western portions of Georgia and Florida and one NW-SE transect that extended from Tennessee to near Augusta, Georgia.
The first seven stations for SESAME were installed in 2010. The western transect and portions of the diagonal transect were installed in May 2011, and the eastern transect was installed in May 2012. In addition to SESAME, we use data from the EarthScope Transportable Array (TA), which moved from west to east across the study area during the time of the SESAME deployment. We include all TA stations between 89°W and 75°W and from 28°N to 40°N. We also include stations from the temporary deployment Pre-Hydrofracking Regional Assessment of Central Carolina Seismicity (PHRACCS) (Wagner, 2012).

We analyze all events of magnitude greater than 6.2 that occurred at depths of less than 50 km at a distance of at least 25° from the center of our study area. The time period of our study extends from May 2011 (after the western transect of SESAME was installed) until June 2015. The last year (after the demobilization of SESAME) provides a unique opportunity to analyze the impact of SESAME on seismicity in the southeastern United States.
tion of SESAME) is included in order to have sufficient data at the northeasternmost portion of our study area where the TA was not installed until 2013. In total, we include data from 94 events in our inversions (Fig. 2 inset). Because of the temporary nature of these stations, different stations recorded various subsets of these events. In order to be included, a given station must have recorded at least ten events in that inversion. Most stations recorded >35 events over a wide range of back azimuths (Fig. 2). During the course of this study, we determined that the dense station spacing of the SESAME deployment, particularly over the three-year duration of the western transect, appeared to create suspicious anomalies in phase-velocity inversions (Fig. S1 in the Supplemental Material1) at periods >65 seconds that closely paralleled the station locations. We therefore err on the side of caution and remove SESAME data from phase-velocity inversions at periods >65 seconds.

Our data processing follows that in Forsyth and Li (2005) and Wagner et al. (2010). Each event is visually inspected at each station over a range of periods from 33 to 143 seconds. In order to inspect at a particular period, the data are first normalized to a standard station response file and then bandpass filtered around the central frequency using a 7–10-mHz-wide filter. We eliminate any data with signal to noise ratios of less than five, and we ensure a clear separation of fundamental modes and overtones before including a given event at a given frequency. Data are cut to include only the fundamental mode for each station for each event and a 50 second taper was applied. We then use a Fourier analysis to determine the amplitude and phase at the desired period for that station and event.

**METHODS**

**Phase-Velocity Inversions**

We follow the finite-frequency, two-plane-wave-tomography (TPWT) approach of Forsyth and Li (2005) and Yang and Forsyth (2006). This approach addresses scattering of the incoming plane wave due to structures outside of the study area by approximating the observed waveforms as the convolution of two plane waves with different back azimuths, amplitudes, and phases. Of particular importance in this study area was the need for an accurate starting velocity model. Typically, starting phase-velocity maps at these periods include a priori information on the crustal thickness in order to have a starting model closer to the final model. In the southeastern United States, particularly along the coastal plain in Georgia and northern Florida, the presence of thick sedimentary basins such as the South Georgia Rift Basin (e.g., Nelson et al., 1985a) produces complications in the determination of crustal thickness using automated P-s receiver functions. We use constraints on crustal thickness from P-s and S-p receiver functions (Parker et al., 2013; Parker et al., 2015; Parker et al., 2016; Hopper et al., 2017) and P-s wavefield migration (Hopper et al., 2016) at SESAME and TA stations to determine an accurate map of crustal thickness (Fig. 3A). Offshore crustal thicknesses were obtained from Crust 1.0 (Laske et al., 2013).
In addition, it was necessary to take into account the very thick and slow sedimentary basins that significantly affected the starting phase-velocity models, especially at shorter periods along the Gulf of Mexico and Mississippi Embayment. For this, we use the sedimentary thickness map of Crust 1.0 (Laske et al., 2013) (Fig. 3B). In order to determine the best average velocity for the sedimentary layer to calculate our starting phase velocities for the TPWT, we use the shorter-period Rayleigh waves from existing ambient noise tomography results that are more sensitive to these shallow crustal structures than our longer-period TPWT results. To do that, we calculate the predicted phase velocities for our starting model at the shorter periods determined by ambient wave tomography (10–25 s) using a range of different velocities for the sedimentary layer (1.6–2.4 km/s in 0.05 km/s increments). The starting shear-wave velocity model for a 40-km-thick crust with no sedimentary layer is shown in Figure 4. This model is then adjusted to the correct crustal thickness and to incorporate sedimentary layers as needed. Examples of this adjustment are also shown in Figure 4. The model layer thicknesses were kept consistent throughout the crust and mantle except adjacent to the base of the crust. At the Moho, the layer boundary closest to that Moho depth was moved to the Moho depth to allow for an abrupt discontinuity. The Vp/Vs ratio in the crust is held fixed at 1.726 and the mantle at 1.8 above 200 km depth, consistent with the Vp/Vs ratios in IASP91 in the crust and uppermost mantle, respectively (Kennett, 1991). While Rayleigh waves are somewhat sensitive to P-wave velocities at very shallow depths, our focus on mantle structures makes this assumption acceptable for this application. Velocities below 200 km are from IASP91 (Kennett, 1991). Predicted phase velocities for each sedimentary layer velocity at each point in map view were calculated using the method of Saito (1988). We calculate a root mean square (RMS) misfit between these calculated phase velocities and the phase velocities from USArray ambient noise tomography results (USANT13; Ekström, 2013). This process was repeated for each of the aforementioned sedimentary layer velocities. A sediment velocity of 2.1 km/s produced the lowest RMS misfits within the study area and was therefore used in the construction of the starting phase-velocity models, using the same layer parameterization described above.

For the TPWT inversion, we use a grid-node spacing of 0.5° and an a priori estimation of standard deviation of 0.25 km/s. We also invert for the effects of azimuthal anisotropy using 1° grid-node spacing. Outermost grid nodes are underdamped in order to absorb effects of velocity perturbations outside of the study area. Phase velocities are approximated by

\[ C(\omega, \theta) = B_2(\omega) + B_1(\omega) \cos(2\theta) + B_3(\omega) \sin(2\theta). \]

We omit higher-order 4θ terms (e.g., Smith and Dahlen, 1973). Peak-to-peak anisotropy is calculated as 2\(\pi\)\(\arctan(B_2/B_1)\)\(1/2\), and the direction of fastest velocity is calculated with 0.5\(\pi\)\(\arctan(B_2/B_1)\). Starting anisotropy at all grid nodes in our preferred model is set to zero. The regularization of anisotropy terms is achieved with damping. In our preferred model, this is set to 0.04. The effects of varying regularization and starting model on anisotropy terms are described in the resolution section below.

### Shear-Velocity Inversions

In order to obtain a 3D shear-wave velocity model, we use the aforementioned phase velocities to calculate predicted 1D shear-wave velocity profiles at each grid node in the phase-velocity inversion. In order to account for the effects of laterally variable crustal structure, we include the ambient noise results of Herrmann et al. (2016) for periods between 10–25 seconds. These results include the full EarthScope Transportable Array (TA) data set across our study area (Fig. S2 [footnote 1]). The method for the shear inversion follows that of Weeraratne et al. (2003) and Wagner et al. (2012a), using the forward calculation of Saito (1988). The layer parameterization and starting shear-wave velocities for this step are the same as that used to calculate the starting phase velocities in the previous step (Fig. 4). The regularization for the shear inversion is controlled by the standard deviations of each phase-velocity measurement provided by the TPWT inversions. For shorter periods from the ambient noise results, we use the standard deviation of the shortest TPWT period used (33 seconds) and multiply that value by 1.5 to place more weight on the results.
of the TPWT we performed. We only invert at those grid nodes defined by the TPWT inversions where at all periods the final standard deviation is <25 m/s and where the ambient noise tomography results of Herrmann et al. (2016) are defined by more than 10 ray paths. This eliminates most offshore grid nodes, as well as some of the coastal nodes where there are few crossing rays. Shear-wave velocity depth maps are smoothed between grid nodes (shown as black squares) using the Generic Mapping Tools triangulate function (Wessel and Smith, 1998). The shear-wave layer parameterization and lateral node locations can be seen in cross sections as square boxes as well.

## RESOLUTION

### Effect of Regularization and Parameterization on Phase Velocities

As with any inversion, the choice of regularization plays a significant role in the determination of velocity structures. For the phase-velocity inversions, the most relevant parameters are grid-node spacing and damping. The use of broader grid-node spacing limits the ability to image smaller-scale structures but reduces the need for damping, which can suppress the absolute velocities of areas that differ significantly from the starting model. By improving our starting velocity model with the inclusion of a priori information on sedimentary basins, we reduced the need for damping at a given grid-node spacing.

In order to test the effects of changes in grid-node spacing and damping on the phase-velocity results, we have included the results of test inversions using different grid-node spacing (0.33°) and different damping (0.01 and 0.4). These results are shown in Figures S3–S5 (footnote 1). In the case of the reduced grid-node spacing (Fig. S3), at shorter periods, the results are similar to those of our preferred model. However, at longer periods, it is clear that the results are underdamped for this grid-node spacing, and the results are unrealistically pixelated. At our preferred grid-node spacing, even the underdamped model looks comparatively stable, and very similar to our preferred model, at least over the well-resolved portions of our model. The over-damped model shows that the major features discussed in the text are required by the data even when the regularization is strongly suppressing the observed velocity deviations.

### Spatial Resolution of Phase Velocities

Resolution within the study area can be seen in checkerboard tests (Fig. 5), which require that a variety of check sizes are tested in order to demonstrate resolvability (Lévêque et al., 1993). We ran checkerboard tests with alternating positive and negative 5% velocity deviation anomalies that were 2, 3, and 4 grid nodes wide for all periods. Longer periods have more difficulty recovering the smaller anomalies, especially at the edges of the model space. Offshore, and in areas outside of the array of stations used, streaking is evident in all tests. However, anomalies are well recovered both in size and amplitude at all but the longest periods at the smallest anomaly sizes within the central portion of our study area.

### Effect of Starting Model and Regularization on Phase-Velocity Anisotropy

Given that the dominant anisotropy observed in our results is very small, we considered the possibility that our anisotropy parameters were over-damped in the inversion, or that the choice of a zero starting anisotropy biased the solution toward zero-magnitude anisotropy. To test this, we ran our inversions, first with greatly reduced damping on the anisotropy terms, and then again, with our preferred damping and a 5% E-W or N-S starting anisotropy across the entire study area. The results of these tests are shown in Figure 6 and Figures S6 and S7 (footnote 1). These figures show clearly the areas where changes in damping or starting anisotropy affect the results and those areas that are insensitive to these parameters. Of particular importance to our results is that when we include a starting 5% anisotropy, the inversion changes the magnitude of the fast direction to match that of those inversions where we had no anisotropy in the starting model. This, together with the test in which we reduced the damping parameter, suggests that the small magnitudes we observe in anisotropy are not the result of insensitivity of the inversion to anisotropy but are in fact required by the data.

### Vertical Resolution of Shear-Wave Velocity Structures

We calculate the RMS average misfit between the phase velocities used as data in the shear-velocity inversion and the phase velocities predicted for our preferred shear-wave velocity model over all periods at each point, and we plot these misfits in Figure S8 (footnote 1). To assess the ability of our inversion to recover the structures we observe, we perform recovery tests on the simplified uppermost mantle structures seen in two orogen-parallel cross sections. We calculate the predicted phase velocities for the input model (Figs. 7A and 7C) and then perform the shear-velocity inversions using the same starting model and same standard deviations assigned to the phase velocities at those grid nodes in our actual preferred inversion. The results of the recovery test are shown in Figures 7B and 7D. These recovery tests show a slight tendency to smear anomalies vertically. The magnitude of the velocity deviations is also not fully recovered. However, the basic shapes of the structures are clearly recognizable, as are the patterns of velocity deviations. We therefore limit our interpretation to those aspects of our results.

## RESULTS AND COMPARISON WITH PREVIOUS RESULTS

### Phase Velocities

The results of our phase-velocity inversions can be seen in Figure 8 and Figures S9 and S10 (footnote 1). The phase-velocity deviations in Figure 8 and Figure S9 reflect the difference between the starting phase-velocity maps (which account for crustal thickness variations and sedimentary basins) and the absolute velocities determined by the inversion.
Figure 5. Checkerboard tests: The left-hand column shows the input phase-velocity deviations. The top row is for a 2 × 2 grid-node deviation pattern. The middle row is for a 3 × 3 grid-node deviation pattern, and the bottom row shows a 4 × 4 grid-node deviation pattern. The middle column shows the recovered result for our 40-second phase-velocity inversion. The right-hand column shows our recovered result for our 125-seconds phase-velocity inversion.
These deviations therefore emphasize significant differences in crust and mantle structures not accounted for by the sedimentary basins and crustal thicknesses included in the starting velocity model. Azimuthal phase-velocity anisotropy is also plotted on these figures and is discussed below. In general, shorter periods (<77 s) are dominated by high phase velocities across most of our study area. Phase velocities are on the order of 3%–5% faster than the predicted starting phase velocities, with the highest velocities occurring to the northwest and moderately fast anomalies occurring closer to the coast. There are a number of distinct exceptions. The first is located in northern Virginia and eastern West Virginia. Phase velocities here are on the order of 2%–3% lower than the predicted starting phase velocities, putting this area in distinct contrast with surrounding areas at shorter periods. Another distinct anomaly is visible at 33 and 40 seconds (and to a lesser extent at 45 s), where there is a low-velocity region centered on the Cape Fear River beneath the Cape Fear Arch; both parallel the border between North Carolina and South Carolina near the coast. Low velocities are also observed at most periods in northern Florida and southeastern Georgia. Moderately lower phase velocities are also observed along and east of the Fall Line, and to a lesser extent in northern Georgia. Finally, we observe a distinct region of low velocities in central Kentucky at 33–46 seconds.

Starting with 77 seconds (Fig. 8), phase-velocity maps show broad regions of lower than predicted phase velocities over much of the eastern seaboard, extending west to the Appalachian orogen (at 77 seconds) and by the longest periods, across much of Tennessee, West Virginia, and Kentucky. One notable exception is in Alabama, where slightly higher than predicted velocities persist even to long periods.

Our results are broadly consistent with recently published continental-scale and regional-scale phase-velocity maps (Schmandt et al., 2015; Pollitz and Mooney, 2016; Shen and Ritzwoller, 2016; Zhao et al., 2017), although some of the smaller-scale structures in our phase-velocity maps are blurred or are difficult to see in other studies. For example, while most models show low velocities in the region of northern Virginia, most do not observe a decrease in velocities along the Cape Fear River or east of the Fall Line. This is likely simply due to a difference in resolution and grid-node spacing employed by these larger spatial-scale inversions.

**Azimuthal Anisotropy**

The results of the anisotropy terms of our phase-velocity inversions are shown in Figure 8 and Figure S9 (footnote 1). We plot only those measurements that are the most robust—those that are insensitive to starting model and regularization as described in the Resolution section below (Fig. 6). The direction of the bars at each node shows the fast direction, and the length of each bar indicates the percent anisotropy. Percent anisotropy is also shown in the diamond located at each node, color coded from black (no anisotropy) to red (2% anisotropy). Any node with anisotropy greater than 2% is shown with a white diamond. There are very few white diamonds, indicating that anisotropy recovered by our inversions is first and foremost small. We also
note that fast directions seem to follow few consistent patterns. To facilitate comparison with previous results, we have also plotted on Figure 8 and Figure S9 the boundaries of subregions defined by the teleseismic shear-wave splitting studies of Long et al. (2015) and Yang et al. (2017), which are broadly, though not entirely, consistent with one another. We recognize that XKS shear-wave splitting studies are integrated measurements from the core-mantle boundary to the surface, whereas our results are sensitive only to a specific range of depths for any given period. What follows is a description of previous determinations of anisotropy for each region (from west to east) and a comparison of these findings to our results over all periods and to new teleseismic shear-wave splitting measurements from the stations of the SESAME array (Fischer et al., 2015). While beyond the scope of this paper, future work might include a more detailed analysis of the relative contributions of anisotropy at different depths to the anisotropy observed here and through XKS splitting studies.

**Northwest Corner: Regions A3 and B2 (Yang) and Region E (Long)**

This region lies north and west of the highest elevations of the modern Appalachian orogen (Fig. 1), in an area that is dominantly west of the Grenville Front (Whitmeyer and Karlstrom, 2007). SKS splitting results indicate fast directions that are close to, if not identical to, absolute plate motion (APM) (Fig. 8; Figure 7. Shear-wave recovery test: Plots (A) and (C) show the input shear-wave velocity model used to calculate predicted phase velocities. Plots (B) and (D) show the respective recovered models, using the same standard deviations as were found in our actual inversions for cross sections D2 (A and B) and D1 (C and D). Topography is shown (in meters) above each cross section. See text for details.)
Figure 8. Isotropic phase-velocity deviations and anisotropic fast directions. Phase-velocity deviations are shown in colors, and anisotropy is shown by bars located at the anisotropy grid-node locations. The phase-velocity deviations shown here are the percent difference between the 2D starting phase velocities and the obtained absolute phase velocities shown in Figure S10 (footnote 1). The sensitivities of these phase velocities to shear velocities as a function of depth can be found in Figure 4. Diamonds at each anisotropy grid node are color coded according to the magnitude of the anisotropy observed. Black arrows offshore show absolute plate motion from Gripp and Gordon (2002). Also plotted are anisotropy region boundaries from Yang et al. (2017) (green lines and labels) and Long et al. (2015) (dotted black lines and black labels). Periods not shown here are included in Figure S9 (footnote 1).
Long et al., 2015; Yang et al., 2017). Uppermost mantle anisotropy determined from Pn arrivals suggest large variability in fast directions immediately below the Moho, ranging from due north-south in central and western Tennessee to NE-SW in southern Indiana (Buehler and Shearer, 2017). Our results are consistent with significant variability in fast directions, both laterally and as a function of period/depth. We do not observe a consistent trend at or close to APM. At shorter periods, we see some indication of north-south fast directions in western Tennessee and perhaps NE-SW fast directions in the northwesternmost corner of our resolved study area, consistent with the results of Buehler and Shearer (2017).

**Interior Appalachian Orogen: Regions C3 (Yang) and B (Long)**

SKS splitting measurements along the interior of the orogen from previous studies generally indicate fast directions that are parallel to both the strike of the orogen and APM (e.g., Fouch et al., 2000; Wagner et al., 2012b; Long et al., 2015; Yang et al., 2017), largely in agreement with SKS splitting fast directions obtained at SESAME stations in this region (Fischer et al., 2015). Pn fast directions are also generally oriented NE-SW except in northwestern Virginia and central and eastern West Virginia, where fast directions rotate to a more N-S orientation (Buehler and Shearer, 2017). Our results indicate orogen-parallel fast directions are more common to the southwest and western portions of this area, though this is not consistent across all frequencies. Longer periods generally show stronger orogen-parallel fast directions than shorter periods, though the anisotropy observed at the longest periods also tends to be very small (<1%).

**Outer Appalachian Orogen and Coastal Plain: Regions C2 (Yang) and C (Long)**

This area comprises the eastern portions of North and South Carolina, as well as the southeastern half of Virginia and portions of central Georgia. Long et al. (2015) and Wagner et al. (2012b) find dominantly null shear-wave splitting results across this region. In contrast, Yang et al. (2017) find in addition to these nulls a significant number of shear-wave splitting measurements that indicate a rotation from the orogen-parallel fast directions to the west to a more NNE-SSW–trending fast direction along the coast, particularly in North Carolina. Fischer et al. (2015) found similar NNE-SSW fast directions from their analysis of TA and permanent stations in the Carolinas. However, in central Georgia, SKS fast directions from the dense stations of the SESAME array manifest a strong back-azimuthal dependence that is consistent with N-S fast directions at shallower depths and APM-parallel fast directions at greater depths. Uppermost mantle anisotropy from Pn analyses suggests dominantly N-S fast directions at the easternmost margin of North Carolina and Virginia but a gradual rotation to NW-SE in central North Carolina, eastern Virginia, and most of South Carolina, reaching W-E azimuths in central Georgia (Buehler and Shearer, 2017). Our observations are dominated by a mix of N-S or NNE-SSW fast directions or near-null anisotropy across most of this area at all periods. In central Georgia, the N-S or NNE-SSW fast directions are consistent with the shallower layer of azimuthal anisotropy indicated by the SESAME SKS splitting, even at periods of 143 s, suggesting that the deeper layer of anisotropy inferred from SESAME data lies at greater depths than can be imaged with surface waves. In the eastern portion of Region C in central and eastern North Carolina, northeastern South Carolina, and southeastern Virginia, our results show variable anisotropy ranging from NNW-SSE to NNE-SSW anisotropy at shorter periods, consistent with the results of Yang et al. (2017) and Fischer et al. (2015). At longer periods in this region, anisotropy values are generally very small and are more consistent with the results of Long et al. (2015).

**Suwannee Terrane: Regions C4 (Yang) and D (Long)**

This area includes Florida and southernmost Georgia and Alabama. All previous studies indicate dominantly east-west fast directions, rotating somewhat to NE-SW along the Atlantic coast (Fischer et al., 2015; Long et al., 2015; Buehler and Shearer, 2017; Yang et al., 2017). We have limited resolution this far to the south for seismic anisotropy, but to the extent that we do, our observations are consistent with those of existing SKS shear-wave splitting studies.

**Shear-Wave Velocity Structure**

Our shear-wave velocity results are shown in map view in Figure 9 and in cross section in Figures 10–12. Given the heterogeneity of the crust across our study area, it was necessary to include the shorter-period, ambient-noise phase-velocity data in our shear-wave velocity inversions. However, the primary focus of this paper is on upper-mantle structure, as constrained by the earthquake-induced, phase-velocity maps that are the product of our research. As such we will focus our discussion on structures observed below the Moho.

**High-Velocity Layer in the Uppermost Mantle**

A high shear-wave velocity layer (4.6–4.7 km/s) extends across much of our study area between the Moho and 100–200 km depth. This layer is interrupted or altered only in very discrete regions, which are discussed below. The thickness of this layer is variable. To the northwest, velocities greater than 4.6 km/s persist to depths of at least 150 km, but in areas of the southeast, the base of this layer shallows to 75–90 km depth. Over most of the study region, this layer appears to thin uniformly from northwest to southeast (Fig. 11; cross sections X1–X4), showing only moderate variation parallel to the strike of the orogen (Fig. 10). However, in Alabama, the high-velocity layer is particularly thick (cross section X4), extending to at least 150 km depth.
Figure 9. Map view plots of shear-wave velocities. Colors show shear-wave velocity deviations, and contours show absolute shear-wave velocity in 0.1 km/s increments. X and Y axes correspond to longitude and latitude, respectively. Top left and bottom right plots show the locations of cross sections plotted in Figures 10–12. Black squares indicate the grid-node locations for the phase-velocity inversions at which shear-wave velocity inversions were performed.
Figure 10. North-south cross sections through shear-wave velocity model. Velocity deviations are shown in color, and absolute velocities are contoured in 0.1 km/s intervals. Topography is shown at the top as a black line (vertical scale is from 0 to 1000 m). Red and green dots show the locations of earthquakes. Red dots are projected to the zero-elevation line. Green dots are plotted at hypocentral depths. Earthquakes within 30 km of the transect are projected onto each cross section. X-axis is in degrees latitude. Labels show major tectonic features and locations of intersecting cross sections. Abbreviations: V—Eocene volcanoes; NY-AL—New York–Alabama Magnetic Lineament; CPS—Central Piedmont Shear Zone; CFA—Cape Fear Arch; GF—Greenville Front; SSZ—Suwannee Suture Zone; BMA—Brunswick Magnetic Anomaly; ETSZ—Eastern Tennessee Seismic Zone; FL—Fall Line; RR—Reelfoot Rift; RT—Rome Trough; SGR—South Georgia Rift Basin.
Figure 11. Orogen-perpendicular cross sections through shear-wave velocity model. Velocity deviations are shown in color, and absolute velocities are contoured in 0.1 km/s intervals. Topography is shown at the top as a black line (vertical scale is from 0 to 1000 m). Red and green dots show the locations of earthquakes. Red dots are projected to the zero-elevation line. Green dots are plotted at hypocentral depths. Earthquakes within 50 km of the transect are projected onto each cross section. X-axis shows distance along cross section in km. Labels show major tectonic features and locations of intersecting cross sections. Abbreviations: V—Eocene volcanoes; NY-AL—New York–Alabama Magnetic Lineament; CPS—Central Piedmont Shear Zone; CFA—Cape Fear Arch; GF—Grenville Front; SSZ—Suwannee Suture Zone; BMA—Brunswick Magnetic Anomaly; ETSZ—Eastern Tennessee Seismic Zone; FL—Fall Line; RR—Reelfoot Rift; RT—Rome Trough; SGR—South Georgia Rift Basin.
The presence of an uppermost-mantle, high-velocity layer has been seen in previous shear-wave velocity inversions of the SEUS (e.g., Yuan et al., 2014; Schmandt et al., 2015; Biryol et al., 2016; Pollitz and Mooney, 2016; Shen and Ritzwoller, 2016; Burdick et al., 2017; Savage et al., 2017). Previous surface-wave inversions (e.g., Yuan et al., 2014; Schmandt et al., 2015; Pollitz and Mooney, 2016; Shen and Ritzwoller, 2016; Savage et al., 2017) have shown very similar structures to those shown here, with some exceptions as discussed below. Teleseismic body-wave tomography studies (e.g., Biryol et al., 2016; Burdick et al., 2017), while less sensitive to depth variations in the uppermost mantle, do show evidence of high uppermost-mantle velocities with a particularly thick, high-velocity layer across Alabama, central Tennessee, and Kentucky.

**Virginia Anomaly**

The most prominent anomaly in map view is the "bulls-eye" located in western Virginia and eastern West Virginia. This feature is most strongly observed at only four grid nodes in map view (Fig. 9 and Fig. S10a [footnote 1]). The size of this anomaly in map view is therefore controlled, to some extent, by the grid-node spacing we employ in our phase-velocity inversion. However, in our test of finer grid-node spacing (Fig. S3a [footnote 1]), the phase-velocity maps do not show a significant change in the size or shape of this anomaly. It is unlikely that the anomaly is much larger than observed in our inversions, because the regularization employed would, if anything, tend to smear out an anomaly of this size; although it is possible that the anomaly is smaller. Our checkerboard tests (Fig. S5) indicate that an anomaly that is 2 × 2 grid nodes wide can be resolved at the shorter periods in question without significant smearing, giving us confidence in our results. In cross sections A, X1, and D1 (Figs. 10–12, respectively), velocities of <4.4 km/s extend from the base of the crust to ~90 km depth, underlain by somewhat less slow velocities persisting to at least 200 km depth. Our recovery tests (Fig. 7) indicate that a similar anomaly would be smeared downward and the velocity deviation reduced by the inversion. This suggests that the actual velocity structure is shallower and/or slower than the anomaly in our preferred model.

Almost all previous tomographic inversions that cover the eastern United States show some evidence of reduced seismic velocities in northern Virginia and easternmost West Virginia (e.g., Schmandt et al., 2015; Biryol et al., 2016; Pollitz and Mooney, 2016; Shen and Ritzwoller, 2016; Buehler and Shearer, 2017; Burdick et al., 2017; Savage et al., 2017). However, the regional model of Pollitz and Mooney (2016) does not show the low-velocity anomaly as shallow as it is recovered here. That study uses a somewhat different methodology than the one presented here. Another possible explanation for the observed difference is the use by the present study of a revised crustal thickness map and sediment thicknesses both for the phase-velocity inversion and the shear-wave velocity inversions. The presence of this feature at periods as short as 33 seconds gives us confidence in the shallow depth of our recovered anomaly.

Figure 12. Orogen-parallel cross sections through shear-wave velocity model. Velocity deviations are shown in color, and absolute velocities are contoured in 0.1 km/s intervals. Topography is shown at the top as a black line (vertical scale is from 0 to 1000 m). Red and green dots show the locations of earthquakes. Red dots are projected to the zero-elevation line. Green dots are plotted at hypocentral depths. Earthquakes within 30 km of the transect are projected onto each cross section. X-axis shows distance along cross section in km. Labels show major tectonic features and locations of intersecting cross sections. Abbreviations: V—Eocene volcanoes; CFA—Cape Fear Arch; SSZ—Suwannee Suture Zone; BMA—Brunswick Magnetic Anomaly; NETSZ—Northernmost Eastern Tennessee Seismic Zone; SGR—South Georgia Rift Basin.
**Rift Anomalies**

In a few discrete areas, the high-velocity layer that underlies the crust across much of our study area does not extend all of the way to the base of the crust. For example, along the Rome Trough and Reelfoot Rift in central and western Kentucky (cross sections C, D, X2, and X3; Figs. 10 and 11), the high-velocity layer is separated from the Moho by 10–20 km. For the Reelfoot Rift (cross sections D and X3; Figs. 10 and 11), the high-velocity layer appears to be deflected downward while maintaining constant thickness beneath the sub-Moho low-velocity layer. In contrast, the high-velocity layer beneath the low velocities that coincide with the Rome Trough (cross sections C and X2; Figs. 10 and 11) does not appear to be deflected downward at all. The reduced sub-Moho velocities can be seen clearly in map view at 60 km depth in central and western Kentucky (Fig. 9). These shallow low velocities are not seen in earlier continent-scale surface-wave inversions (e.g., Yuan et al., 2014; Schmandt et al., 2015; Shen and Ritzwoller, 2016), but they are observed in regional surface-wave inversions in the eastern United States (e.g., Chen et al., 2016; Pollitz and Mooney, 2016; Savage et al., 2017). However, the shallow low-velocity anomalies observed in this study contrast with the more dramatic and deeper low-velocity structures observed by Chen et al. (2016) beneath the southeastern extension of the Reelfoot Rift, which lies outside our study region. Similarly, the continent-wide Pn tomographic inversion of Buehler and Shearer (2017) shows reduced uppermost-mantle velocities along the Reelfoot Rift, but these do not extend as far to the east as those presented here.

**Cape Fear Arch Anomaly**

The moderately low-velocity anomaly at 60 and 75 km depths along the North Carolina–South Carolina border can be seen in map view and in cross sections A (Fig. 10), X2 (Fig. 11), and D2 (Fig. 12). Here, shallow low velocities appear above the high-velocity layer. The high-velocity layer is not thinned to accommodate this low-velocity feature but appears deflected downward around the subcrustal low velocities by 25–50 km, resulting in a greater maximum depth of this high-velocity layer beneath this low-velocity anomaly. This apparent downward deflection of the high-velocity layer in this area is significantly greater than that observed beneath the rifts to the northwest. To the southwest (cross section D2) and south (cross section A) of the arch, the downward-deflected, high-velocity layer appears to shallow and possibly thin as it approaches the South Georgia Rift Basin. Northeast of the arch (cross section D2), the high-velocity anomaly is located immediately below the crust but appears thicker than the high-velocity anomaly to the southwest of the arch. To the north and northwest of the arch (cross sections A and X2), the high-velocity layer is directly sub-Moho and transitions abruptly to its offset position beneath the arch to the south and southeast of the Fall Line. To our knowledge, this anomaly has not previously been discussed in the literature, although the regional surface-wave model of Pollitz and Mooney (2016) does show some indication of a downward deflection of the high-velocity layer beneath the Triassic basins and coastal plain.

**Northern Georgia Anomaly**

Another significant disruption of the high-velocity layer is observed across northern Georgia and western South Carolina (Fig. 9; 60–105 km depth). Here, the 4.6–4.8 km/s high-velocity layer is broken up by a region where velocity deviations are close to zero, and absolute shear-wave velocities are ~4.5 km/s. In some locations, this average velocity region extends as far north as central Tennessee. While not as pronounced as the low-velocity anomaly in northern Virginia, this comprises a significant disruption of the otherwise fairly uniform <4.6 km/s high shear-wave velocity layer observed in the uppermost mantle across most of our study area.

This is perhaps the most subtle of the anomalies presented here, but its proximity to the increased station density of the SESAME deployment along with our resolution tests give us confidence that this feature is well resolved. A similar feature is recovered by Shen and Ritzwoller (2016) at 70 km depth, and to a lesser extent by Savage et al. (2017) at 50 km depth. The location of this lower-velocity region also coincides with a reduction in Moho reflectivity found by Hopper et al. (2016) beneath the Blue Ridge Mountains and with decreased Pn velocities found by MacDougall et al. (2015) across the northernmost stations of the SESAME array.

**DISCUSSION**

One of the important questions this study seeks to address is the extent to which the modern mantle lithosphere in the SEUS is composed of inherited structures from earlier plate boundary processes and to what extent the existing mantle lithosphere reflects processes that have occurred since the region resumed a passive margin setting. We can also address the question of how these structures, whether inherited from plate boundaries or evolved in a passive margin setting, correlate with ongoing tectonism observed at the surface, such as seismicity and deformation. It is beyond the scope of this study to explore fully any possible causality between our seismic observations and the geological observations to which they correlate. We propose some possible explanations, and we hope this study will spur further investigations into the evolution of mantle lithosphere in passive margin settings.

A key facet of the shear-velocity structure is the marked variation in velocity in the depth range typically associated with the mantle lithosphere (Moho to 150 km). Multiple possible origins for these anomalies exist, including: (1) variations in lithospheric thickness; and (2) variations in the properties of intact mantle lithosphere, for example due to metasomatism and/or infiltration by partial melt. Depending on their location, such variations may be related to the transition from Proterozoic craton to Phanerozoic lithosphere, Phanerozoic tectonism (orogenesis and rifting), or more recent alteration of the lithosphere (e.g., delamination, other forms of thermal or mechanical lithospheric loss, or metasomatism and/or melt percolation).
In this section, we evaluate the plausibility of these alternative scenarios, drawing on complementary geological, geochemical, and geophysical constraints as available.

**Evidence for Structures Inherited from Past Plate Boundary Processes**

We find strong evidence for inherited structures in our shear-wave velocity model in a number of different regions. Figure 13 compares our shear-wave velocity model at 75 and 150 km depth to gravity anomalies, magnetic anomalies, and geologic observables at the surface. At 150 km depth, we see a strong correlation between the location of the Grenville Front (defined as the western margin of Grenville deformation) and the boundary between high velocities to the northwest and lower velocities to the southeast. We interpret these high velocities, along with the high-velocity layer that is dominant across much of our study area, as evidence of mantle lithosphere. Areas northwest of the Grenville Front represent those regions with the thickest mantle lithosphere, consistent with a continental core unaffected by plate boundary processes for >1 Ga. The only disruption to this region of thick mantle lithosphere is in those areas affected by the failed rifting of Rodinia—the Reelfoot Rift and Rome Trough. In these areas, we see evidence of structural changes to the mantle lithosphere in the form of reduced velocities in the uppermost mantle and, in some cases, a downward deflection of the lower boundary of the mantle lithosphere. These structures are in contrast to the larger velocity reductions observed in the southwestern Reelfoot Rift, where alteration of the deep lithosphere appears to have been more pronounced (Chen et al., 2018).

We do not, however, see consistent evidence for a change in mantle lithospheric structure across the New York–Alabama (NY-AL) Magnetic Lineament, the Central Piedmont Shear Zone, the Suwannee Suture Zone, or the Brunswick Magnetic Anomaly. These proposed terrane boundaries might be expected to show evidence of different mantle lithosphere corresponding to the differing provenances of the accreted terranes. While some cross sections do show changes near these boundaries, these changes are not consistent and do not appear to follow the terrane boundaries in map view. In the case of the NY-AL Magnetic Lineament and the Central Piedmont Shear Zone, this is consistent with earlier studies that propose that Grenville basement rocks underlie most of the SEUS due to the formation of an Alleghanian thrust sheet (e.g., Cook and Vasudevan, 2006; Duff and Kellogg, 2017).

There may be some evidence of alteration of mantle lithosphere associated with the location of the Suwannee Suture Zone (e.g., Mueller et al., 2014; Boote and Knapp, 2016; Hopper et al., 2017). To the north-northwest of the Suwannee Suture Zone in northern Georgia and South Carolina, we observe average shear-wave velocities (~4.5 km/s) in lieu of the high-velocity layer (>4.6 km/s) that is dominant across most of our study area. One possibility is that the average velocities observed north of the suture represent the alteration of mantle lithospheric material as the result of the collision between Laurentia and Gondwana and the accretion of the Suwannee terrane. Seismic velocities could have been reduced by the hydration of the mantle lithosphere above

Figure 13. Comparison of shear-wave velocity deviations at 75 and 150 km depth and geologic structures discussed in text. Acronyms and symbols are the same as in earlier figures. (A) Velocity deviations at 75 km depth. (B) Velocity deviations at 150 km depth. Velocity deviations are shaded according to topography to make the location of the modern Appalachian orogen visible. Lavender circles show earthquake locations. Abbreviations: BMA—Brunswick Magnetic Anomaly; GF—Grenville Front; NY-AL—New York–Alabama; RR—Reelfoot Rift; RT—Rome Trough; SGR—South Georgia Rift Basin.
a northwestern-oriented subducting slab (present orientation) (e.g., Whalen et al., 2015). However, this subduction polarity is opposite to the dip of the structure in the crust (e.g., Hatcher, 1972; Cook and Vasudevan, 2006; Hopper et al., 2017). Moreover, subsequent CAMP flood basalts produced extensive dikeing across the southeastern half of this average-velocity region (e.g., Hames et al., 2000 and references therein). Such extensive volcanism would most likely have dehydrated the mantle lithosphere, resulting in faster seismic velocities across those portions of this feature, which we do not observe.

Another possible explanation is that the mantle lithosphere has been gradually removed by a lithospheric thinning event in this zone. Biryol et al. (2016) found evidence for a high-velocity anomaly dipping to the east in the upper mantle beneath our study area. The shallowest portion of their anomaly extends from Alabama to Kentucky, consistent with deeper high velocities observed in our model, both west of the Grenville Front and across much of Alabama. Biryol et al. (2016) argue that this feature is most plausibly explained by downwelling mantle lithosphere drawn from lithosphere farther to the east, which allows the dip to be explained by the westward absolute plate motion of the overriding plate. However, the exact source of this removed mantle lithosphere was difficult for Biryol et al. (2016) to discern due to the reduced resolution of regional teleseismic body-wave tomographic images above 100 km depth. We propose that one possible source location for the foundered lithospheric material may be from this area of average seismic velocities across northern Georgia and South Carolina. However, the absence of significantly slower seismic velocities suggests that the removed material was not simply replaced by hot asthenospheric mantle, as is observed elsewhere. More work is needed to understand better the genesis of this velocity change in the uppermost mantle and its relationship to other upper-mantle seismic velocity structures and/or tectonic history.

Our results are generally consistent with the presence of Grenville lithosphere beneath most of the SEUS north of the Suwannee Suture Zone. This mantle lithosphere is, however, not homogeneous. We see a general thinning of the mantle lithosphere toward the Atlantic coast, consistent with lithospheric thinning due to rifting and continental breakup. There is perhaps some correlation between the location of the South Georgia Rift Basin (SGR) and a sub-Moho low-velocity region similar to what is observed at the Reeffoot Rift and Rome Trough (cross sections B–D; Fig. 10), but this is not consistent (cross sections X4 and D2; Figs. 11 and 12, respectively). However, a number of structures (discussed below) cannot easily be associated directly with inherited Grenville lithosphere or subsequent rifting and likely reflect structures that evolved in the mantle lithosphere while located in a passive margin setting.

Evidence for Structures Associated with Observed Passive Margin Tectonism

In several regions, our observed mantle lithospheric structures correlate closely with passive margin tectonic activity observed at the surface. The clearest evidence of this is the very localized low-velocity anomaly that directly underlies the Eocene volcanism in western Virginia and eastern West Virginia. It is unlikely that such a small-scale structure, unassociated with any other known geologic boundary or feature, would have always had a near absence of mantle lithosphere while surrounded by otherwise relatively uniform and unremarkable mantle lithosphere, leading us to believe this material was lost. Our results do not indicate how this material was lost. However, we would argue there is strong suggestion that the material was lost during the Eocene. In addition to the strong spatial correlation in map view, thermobarometric modeling indicates that the overlying Eocene basalts last equilibrated at pressures consistent with depths only ~30 km below the Moho (Mazza et al., 2014). This is consistent with our observation that high-velocity mantle lithosphere appears largely absent below the crust in this region. We note that the size of the anomaly is on the order of 100 x 150 km or less in map view. The high-velocity layer surrounding this low-velocity anomaly suggests that the thickness of the mantle lithosphere removed is on the order of 50–70 km. This adjacent high-velocity layer is also indistinguishable in velocity and thickness from the high-velocity layer seen across large regions of our study area, suggesting that the mantle lithosphere surrounding this feature was not thinned or altered by whatever process created this low-velocity anomaly. The exception is due east of the anomaly, where velocities are less fast than those observed elsewhere (4.5–4.6 km/s). These velocities are consistent with other velocities observed east of the Fall Line and may therefore be due to lithospheric thinning along the continental margin rather than to processes attributable to the formation of this low-velocity anomaly. While seismic tomography cannot determine how this low-velocity anomaly formed, possible explanations include thermomechanical erosion of the continental lithosphere (e.g., Ranalli et al., 2007; Foley, 2008 and references therein), possibly assisted by decompression melting due to upwelling induced by variations in lithospheric thickness (e.g., Till et al., 2010) or a very localized delamination of mantle lithosphere due to the removal of eclogitized lower crust as proposed by Mazza et al. (2014, 2017). A further question remains about how such a low velocity can be sustained over a period of 48 m.y. There is no evidence that water was introduced to form these melts, making hydration an unlikely explanation for these low velocities. More work is needed to constrain how such an abrupt lithospheric structure could form and how it might persist for a period of ~48 m.y.

Another example of a correlation between observed mantle lithospheric structure and postrift tectonism lies along the Cape Fear Arch. The location of exhumed Upper Cretaceous sediments adjacent to Upper Oligocene sediments along the coast corresponds closely to an apparent delamination of the mantle lithosphere in our tomographic images. The high-velocity layer is offset from the Moho by up to 50 km and is replaced by moderately lower velocities. The location of this anomaly also corresponds to the local maximum uplift of the Orangeburg Scarp located along the Fall Line (e.g., Rovere et al., 2015). The Orangeburg Scarp, as a Middle Pliocene shoreline, was presumably developed at uniform elevation at sea level; thus, differential uplift along this scarp provides constraints on deformation across the Cape Fear Arch during the past ~3 m.y.
A number of recent studies have attempted to model this deformation with dynamic topography caused by a combination of glacial isostatic adjustment and large-scale mantle flow patterns identified by tomographic imaging of the upper and lower mantle beneath the United States (e.g., Mouca et al., 2008; Spasovčić et al., 2008; Rowley et al., 2013; Liu, 2014). Results across these models differ, but none are able to replicate the short wavelength of the observed uplift across the Cape Fear Arch and Orangeburg Scarp. Recently, Mouca and Ruetenik (2017) added flexure due to sediment loading to their dynamic topography models and were able to successfully model the uplift along the Orangeburg Scarp. It is difficult to know without further modeling studies what the predicted effect of the downward deflection and apparent delamination of the mantle lithosphere beneath the Cape Fear Arch might be on surface elevations. It is also impossible to know the temporal evolution of this structure from the seismic tomography alone. More work is needed to incorporate the effects of these localized lithospheric structures into our understanding of the development of this uplift.

**Ongoing Tectonism Not Clearly Correlated with Observed Mantle Lithospheric Structure**

**Blue Ridge Escarpment**

The steepness of the topography observed across the southern Appalachians, particularly across the Blue Ridge Escarpment, has been explained by a number of different mechanisms ranging from rift flank retreat (e.g., Tucker and Slingerland, 1994; Spotila et al., 2004), to flexural bending associated with continental erosion and sediment deposition (e.g., Pazzaglia and Gardner, 1994; Liu, 2014), to isostatic response to the delamination of the lower crust and mantle lithosphere (e.g., Wagner et al., 2012c). However, we do not find any remarkable structures beneath the topography of the southern Appalachians. Indeed, the steepest portions of the orogen along the Blue Ridge Escarpment are uniformly underlain by the high-velocity layer we are interpreting as mantle lithosphere (cross sections B and X2, Figs. 10 and 11, respectively).

The only exception is in northwestern Georgia, where the average velocities observed north of the Suwannee Suture Zone extend beneath portions of the southernmost Appalachian Mountains (34°5′S–35°S, cross section C; Fig. 11; 800–650 km, cross section D1; Fig. 12). We do not, however, observe a change in elevation associated with the northern edge of this anomaly.

These observations suggest that any observed uplift or disequilibrium landscapes in the southern Appalachians are not caused by isostatic responses to changes in mantle lithospheric structure. This suggests that if indeed there is an uplift and/or rejuvenation of the southern Appalachians, this is more likely due to dynamic topography effects (e.g., Liu, 2014, 2015). However, it is also possible that observations of disequilibrium landscapes are due to stream capture and base-level change rather than a net uplift of this ancient orogen (e.g., Prince et al., 2010).

**Seismicity and Mantle Lithospheric Structure**

The ongoing seismicity in the southeastern United States, as with intraplate seismicity in general, is not well understood. A variety of contributing factors to the development of intraplate seismicity have been proposed, including: inherited zones of weakness, possibly associated with oceanic fracture zones (e.g., Sykes, 1978); thin or weak mantle lithosphere or abrupt changes in lithospheric thickness or strength (Liu and Zoback, 1997; Assumpção et al., 2004; Mooney et al., 2012); the presence of terrane boundaries or sutures (Babuška et al., 2007); the presence of increased fluid pore pressure either from fluctuations in meteoric water or from increased mantle CO₂ emissions (e.g., Brauer et al., 2003; Costain, 2008); intersecting fault zones (Taiwani, 1988); the presence of failed rifts, especially from the most recent episode of orogenesis (e.g., Chapman and Beale, 2010; Bartholomew and van Arsdale, 2012); and faults favorably aligned with the regional stress field (Zoback, 1992; Bartholomew and van Arsdale, 2012).

While this study does not have the ability to test all of these hypotheses, it does allow us to look for correlations between the locations of increased seismicity and mantle lithospheric structure. The largest cluster of seismicity within our study area is the Eastern Tennessee Seismic Zone (ETSZ). This northeast-trending band of seismicity is located just northwest of the highest topography of the southern Appalachians. Our results indicate that for the most part, this cluster is located above the high-velocity layer we are interpreting as mantle lithosphere, just to the northeast of where this high-velocity layer ends and is replaced by the average velocity structure in northern Georgia described earlier (cross sections C, X3, and D1; Figs. 10–12, respectively, and in map view; Fig. 13A). This observation is consistent with Mooney et al. (2012), who observed an increase in intraplate seismicity along lateral gradients in lithospheric thickness globally. However, in this case, the seismicity would be almost exclusively located within the region of thicker and presumably stronger mantle lithosphere.

A similar pattern is observed with the Central Virginia Seismic Zone (CVSZ). This cluster, associated with the Mw = 5.8 2011 Mineral, Virginia earthquake, is located just east of the low-velocity anomaly that underlies the Eocene volcanoes described earlier. That low-velocity anomaly is, in turn, the location of the seismic shadow first described by Bollinger and Gilbert (1974). We also do not observe any significant seismicity along the Cape Fear Arch anomaly.

While a full assessment of the implications of our model on the causes of intraplate seismicity is beyond the scope of this paper, we hope that these first-order observations might encourage future work on this important topic.

**CONCLUSIONS**

1. We see evidence of tectonic inheritance in the greater thickness of the mantle lithosphere to the west of the Grenville Front compared to that of the mantle lithosphere east of the Grenville Front and in the overall continuity of the mantle lithosphere beneath our study area, consistent
with previous work suggesting the underthrusting of Grenville basement beneath most of the Piedmont and coastal plains (e.g., Cook and Vasudevan, 2006; Duff and Kellogg, 2017). We see some evidence of structures that may be due to episodes of failed rifting, both in the early Paleozoic and in the Mesozoic.

(2) We also see a number of structures that appear strongly correlated with surface observations of Eocene to recent tectonism. These include evidence for the wholesale removal of a small patch of mantle lithosphere beneath the Eocene volcanics of western Virginia and eastern West Virginia and the apparent delamination of the mantle lithosphere from beneath the Cape Fear Arch in North and South Carolina.

(3) We do not see a strong correlation between the high elevations of the Appalachian Mountains and mantle lithospheric structure. This is in contrast to earlier work using a very limited number of stations that suggested the uplift of the Blue Ridge Escarpment may be due to an earlier episode of mantle lithospheric delamination (Wagner et al., 2012c).

(4) We also do not see a consistent correlation between patterns of seismicity and mantle lithospheric structures. While not all regions of increased seismicity are located above homogeneous mantle lithosphere, it is difficult to draw any direct links between the observed structures and earthquake hazards. More work is needed to understand how mantle lithospheric structures might affect seismic activity in a passive margin setting.

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