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Key Points:

- We use ambient noise tomography at both onland and offshore stations to examine crustal structure across the eastern North American margin
- We observe three structural portions of the passive ENAM: continental crust, transitional crust, and oceanic crust
- The East Coast magnetic anomaly demarks the boundary between continental and oceanic crustal material

Supporting Information:

- Supporting Information S1

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Crustal structure across the eastern North American margin from ambient noise tomography

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Abstract Passive tectonic margins, like the eastern North American margin (ENAM), represent the meeting of oceanic and continental material where no active deformation is occurring. The recent ENAM Community Seismic Experiment provides an opportunity to examine the crustal structure across the ENAM owing to the simultaneous deployment of offshore and onshore seismic instrumentation. Using Rayleigh wave phase and group velocities derived from ambient noise data, we invert for shear velocity across the ENAM. We observe a region of transitional crustal thicknesses that connects the oceanic and continental crusts. Associated with the transitional crust is a localized positive gravitational anomaly. Farther east, the East Coast magnetic anomaly (ECMA) is located at the intersection of the transitional and oceanic crusts. We propose that underplating of dense magmatic material along the bottom of the transitional crust is responsible for the gravitational anomaly and that the ECMA demarks the location of initial oceanic crustal formation.

1. Introduction

The present-day eastern North American margin (ENAM) represents the boundary between the oceanic and continental portions of the North American plate. Passive margins, like ENAM, result from the rifting of continental material during the formation of new oceanic basins. Rifting that led to the formation of the ENAM began ~200 Ma with the breakup of the Pangaea supercontinent [Schlische, 2003; Thomas, 2006]. The ENAM has been interpreted as a strongly volcanic margin, characterized by a large volume of volcanism associated with rifting [e.g., White and McKenzie, 1989]. Volcanic margins are generally indicated by the presence of dipping reflectors within the transitional crust and crustal underplating of igneous materials, both of which have been observed along the ENAM [Klitgord et al., 1988; Tréhu et al., 1989; Holbrook et al., 1994]. The ENAM has been previously described as having three distinct crustal regions: continental crust, transitional crust, and oceanic crust [e.g., Klitgord et al., 1988]. Each region represents a different structural part of the margin that may have associated magnetic or gravitational anomalies.

Several magnetic anomalies are present along the ENAM, the most prominent one being the East Coast magnetic anomaly (ECMA; see Figures 1 and S1 in the supporting information). The ECMA is a magnetic high that runs parallel to the passive margin [Bankey et al., 2002]. The ECMA has often been interpreted as the initial site of spreading along the margin and has been used to denote the boundary between continental and oceanic regimes [Klitgord et al., 1988; Austin et al., 1990; Talwani et al., 1995]. Another prominent feature along the ENAM commonly interpreted to delineate the continent-ocean boundary is a positive gravitational anomaly (PGA) (Figure 1) seen in both the free-air and isotropic gravity fields [Behn and Lin, 2000; Bonvalot et al., 2012] (Figure S2). While geographically proximal, the ECMA and the PGA may not reflect the same structure.

In order to study the meeting of the oceanic and continental portions of the ENAM, we integrate seismic data sets that cross the margin from the North American continental interior onto the oceanic portion of the plate. Margin crossing data sets such as this are rare owing to the dearth of synchronous ocean bottom and onland seismic deployments. Studying both the onshore and offshore components of the ENAM simultaneously is only possible thanks to the recent ENAM Community Seismic Experiment (ENAM-CSE) that deployed 30 ocean bottom seismometers (OBS) and 3 coastal stations offshore of North Carolina contemporaneously with the arrival of the easternmost EarthScope Transportable Array stations. Combining onshore and offshore data sets results in a dense seismic array that spans the margin.

A recently developed tool for studying deep crustal structure is ambient noise tomography (ANT). ANT takes advantage of the ambient noise wavefield allowing us to extract Rayleigh wave group and phase velocity

Figure 1. (a) Map of the ENAM-CSE (red triangles) and other seismic stations (white triangles) used in this study. (b) Our shear velocity model at 25 km depth. The thick black line corresponds to the cross section shown in Figure 3, and the stars correspond to the velocities and sensitivities shown in Figure 2. The outer banks (OB), PGA, and ECMA are highlighted along the margin. The PGA is shown here by the >20 mGal contour in the free-air gravity field [Bonvalot et al., 2012]. The ECMA is defined by the >200 nT contour [Bankey et al., 2002]. The area outside of which we have any resolution is shaded in both panels.

dispersion measurements [e.g., Bensen et al., 2008; Porritt et al., 2011]. Since the ambient noise wavefield acts upon all contemporaneously active seismic stations, we can get exceptionally dense ray coverage over the ENAM study area (Figures 1, S3, and S4). Using group and phase velocities from periods of 8 s to 35 s, we are able to constrain shear velocity across the ENAM to the upper part of the lithospheric mantle. Here we present an ambient noise-derived shear velocity model spanning the ENAM using shoreline crossing seismic data.

2. Data and Methods

We use data from ENAM-CSE, EarthScope Transportable Array, and other local seismic stations operating from April 2014 to April 2015 to map regional differences in Rayleigh wave group and phase velocities. We follow the processing methods of Bensen et al. [2008]. Single-day, vertical component waveforms at each station are de-meaned, de-trended, tapered, corrected for station response, and whitened. We then apply a running-absolute-mean temporal normalization to remove signal from earthquake-generated sources. The single-day seismograms are then cross correlated between all available station pairs and stacked to form empirical Green's functions (EGFs) (Figure S5) (see Bensen et al. [2008] for further details).

We analyze the EGFs using traditional frequency-time analysis to measure Rayleigh wave phase and group velocity dispersions [e.g., Levshin et al., 1992]. This is performed at periods between 8 and 35 s in 1 s increments. We retain dispersion values from EGFs with a signal-to-noise ratio (SNR) greater than 10; SNR is defined as the peak signal amplitude relative to an average RMS of a leading and trailing window (see supporting information Text S1 for further details on SNR calculations). We further restrict the data set by rejecting dispersion values from interstation paths with path lengths less than two wavelengths. In the EGFs, we note the presence of a low-velocity wave traveling behind the Rayleigh wave (Figure S5). The trailing arrival is due to the water column above the ocean bottom stations [e.g., Cheng et al., 2015]. The asymmetrical nature of the water waves in the EGFs can be attributed to an asymmetrical distribution of noise sources due to the stations' location along the coast. The velocity of this water wave is so slow, ~ 1 km/s, that we do not observe significant interference between it and the Rayleigh wave signal. While this slow water wave likely has little-to-no impact on the Rayleigh phase and group velocity measurements, we nonetheless filter out signals with velocities less than 1.5 km/s in a further attempt to avoid contamination.

Figure 2. (top row) Group and phase velocities and (bottom row) phase velocity sensitivity kernels for the representative points in the (left column) continental and (right column) oceanic regions shown in Figure 1. The calculated group and phase velocity dispersion curves are shown in red. Sensitivity kernels are colored by period.

Using interstation phase and group velocities for each period, we invert for regional phase and group velocity maps on a two-dimensional grid with grid spacing of 0.1° by 0.1° following *Barmin et al.* [2001]. We then invert for shear velocity using the results of the regional group and phase velocities [*Herrmann, 2013*]. The shear velocity inversion is parameterized using 2 km thick layers in the upper 50 km, with increasing layer thicknesses below 50 km. Our initial model is set to a constant velocity of ~ 4.5 km/s in the upper 130 km (Figure S6) to avoid introducing low-velocity artifacts and AK135 [*Kennett et al., 1995*] in the rest of the mantle to transition zone depths. By extending the depth of our model far below the peak sensitivities of the group and phase velocities, we prevent the artificial mapping of deep velocity structures into the upper portions of our model. Finally, we correct our model for topography to remove effects of deployment depths of the OBS stations.

3. Results

The dichotomy between the oceanic and continental portions of the ENAM can be readily seen in dispersion curves from each region (Figure 2). Dispersion patterns throughout the oceanic region are consistent with previous studies where group and phase velocities follow similar trends [e.g., *Ewing and Press, 1952*;

Figure 3. Cross section along the profile shown in Figure 1. The ECMA and PGA are highlighted. The 4.2 km/s contour is shown in white.

Herrmann, 2013]. Beneath the continental region, group and phase velocities at short periods (<16 s) diverge before following similar trends at longer periods, which is also consistent with previous ANT observations in continental interiors [e.g., Moschetti *et al.*, 2010]. These different trends in phase and group velocities arise from compositional differences of continental and oceanic material. Differences between oceanic and continental regimes may also play a role in the model fits at the shortest periods since the continental crust generally exhibits greater lateral variability of shallow structures than the oceanic region. We are confident that the EGFs used in our inversion are sampling crustal structure and are not due to spurious errors from noncorrelated noise sources, interference from the water wave discussed above, or from earthquake-generated signals.

Shear velocity results across the ENAM can be seen in Figures 1 and 3. Throughout this study we define the crustal Moho interface by the 4.2 km/s velocity contour [e.g., Delph *et al.*, 2017]. The 4.2 km/s velocity contour is a good proxy for Moho depth as it represents the maximum potential shear velocity for typical crustal material [e.g., Christensen, 1996]. The use of a velocity contour as a proxy for Moho depth is necessary for ambient noise studies because Rayleigh wave group and phase velocity measurements are not sensitive to sharp discontinuity structure. At stations where receiver-function (RF) analyses have been performed, there is good agreement (typically within ~ 4 km) between the RF inferred and our 4.2 km/s inferred Moho depths [e.g., Abt *et al.*, 2010]. At the ENAM-CSE stations, no RF analyses have been performed, but our inferred Moho depths are in close agreement with those derived from the active source component of the project [e.g., Shuck and van Avendonk, 2016] and with previous active source studies near the ENAM-CSE [e.g., Klitgord *et al.*, 1988; Tréhu *et al.*, 1989; Holbrook *et al.*, 1994]. We argue, therefore, across our study region using the 4.2 km/s velocity contour as a Moho proxy is appropriate.

We observe variable Moho depths both within the oceanic region and across the margin between the oceanic and continental regimes. Offshore, we see ~ 20 km thick oceanic crust and a low-velocity (<2.2 km/s) anomaly at very shallow depths (<4 km) due to oceanic sediments. The slow oceanic crust sits upon high shear velocity (>4.6 km/s) lithospheric material. The continental crust is characterized by relatively uniform seismic velocities (~ 3.6 km/s) and ranges from ~ 35 to 40 km thick. A transitional region of crustal thickness between the oceanic and continental crusts extends over ~ 100 km. In the transitional region, there may be a larger volume of high-velocity lower crustal material than in the surrounding areas (Figure S7). This relationship is primarily based, however, on results from active-source studies in the region [Klitgord *et al.*, 1988; Tréhu *et al.*, 1989; Holbrook *et al.*, 1994]. Immediately adjacent to the transitional region, the oceanic crust thins from ~ 20 km to ~ 17 km.

We consider that our model is well resolved in our region of interest and of sensitivity, (see Figures S3, S4, and S8 for lateral resolution estimates), which extends from the near surface (~ 2 – 3 km) to depths of ~ 40 km. Laterally resolvable structures are on the order of ~ 75 km in the short-period part of the model and degrades to ~ 150 km at longer periods (Figure S8). Our peak vertical resolution lies between ~ 10 km and ~ 25 km depth. At these depths, relatively small (~ 4 km) structures are resolvable but

Figure 4. (a) Schematic and (b) interpretation of our proposed model for the ENAM. The continental crust extends to depths of ~40 km, while the oceanic crust is ~20 km thick. The two regimes are connected by a transitional crustal region that has been underplated by dense magmatic material causing the PGA. The ECMA (shown in white) is located at the boundary between the transitional and oceanic crusts. This boundary is associated with a slight local shallowing in the Moho.

region (at 25 km depth), slow crustal velocities dominate, while beneath the oceanic region, lithospheric velocities are observed. The velocity contrast originates from the rapid change in crustal thickness crossing the margin. This juxtaposition of crustal and lithospheric shear velocities highlights the nature of the margin. In cross section (Figure 3), all three structural portions of eastern North American margin are apparent; the thick (~40 km) continental crust, the relatively thin (~20 km) oceanic crust, and a region of transitional crust that connects the two (schematically shown in Figure 4). The transitional crustal region spans ~100 km and accommodates a change in Moho depth of ~20 km, with the majority of that change occurring over lateral distance of ~40 km. The transitional crust likely formed via rifting induced deformation of the continental crust prior to the initiation of oceanic crustal formation [e.g., Hopper and Buck, 1996; Skogseid et al., 2000; Huismans and Beaumont, 2008; Van Avendonk et al., 2009]. Once deformation related to spreading is localized at a spreading center, the formation of oceanic crust begins and the thinning of the transitional crust stops.

While the majority of the oceanic crust throughout our study region is ~20 km thick, there appears to be a slight thickening of the easternmost oceanic crust. This is, however, confined to the very edge of our model where we have poor ray coverage (Figures S3 and S4). The slight thickening is likely due to lacking ray coverage and not due to real crustal structure. Similar to results from previous studies from other portions of the margin [e.g., Klitgord et al., 1988; Tréhu et al., 1989; Holbrook et al., 1994], the oceanic crust in our study region is thicker than is observed for the majority of the oceanic crust in the Atlantic and globally [e.g., Laske et al., 2013; Van Avendonk et al., 2017]. Since the ENAM is a volcanic passive margin, initial rifting was likely characterized by voluminous magmatism [e.g., White and McKenzie, 1989] resulting in the thick oceanic crust we observe [e.g., Kelemen and Holbrook, 1995].

The boundary between transitional crustal depths and the oceanic crustal region is closely associated with the ECMA (Figure 3). We also observed a slight shallowing (~3 km) of the Moho relative to the oceanic crust coincident with this boundary. While the vertical resolution of our model makes interpretations of such small-scale structures tenuous, this region benefits from the densest ray coverage in our study area. We therefore interpret the localized thinning as a real structure, but we note that this should be revisited in future analyses. Due to the coincidence of the ECMA, the termination of transitional crustal depths, and the locally thinned crustal feature, we propose the ECMA and the crustal thinning are both associated with the initial spreading along the ENAM and the formation of the earliest oceanic crust [e.g., Hutchinson et al., 1983]. We suggest that the ECMA marks the boundary between oceanic and continental material.

Crustal thinning and the emergence of an active spreading center occurred along the ENAM with the continued divergent motion of the North American and African plates. Initially, the Late Triassic rifting along the

should be interpreted with caution. Below ~25 km depth, both vertical and lateral resolution decreases. We have little resolution in the upper ~2 km of our model because Rayleigh wave sensitivities at our shortest periods do not extend to the surface (Figures S3 and S4). While there is some streaking in our model, it generally runs orthogonally to the structures we observe. We are therefore confident of the resolvability of our observations.

4. Discussion

A sharp contrast in shear velocity structure between the oceanic and continental crustal portions of the margin can be seen in Figure 1. Beneath the continental

margin would have been accommodated by deformation and thinning of the continental crust creating the transitional crustal region [e.g., Hopper and Buck, 1996; Skogseid et al., 2000; Huismans and Beaumont, 2008; Van Avendonk et al., 2009]. In the Early Jurassic, an active spreading center, and thus the formation of new oceanic crust, is required to allow continued divergence and to arrest the deformation of the continental crust.

Previous studies have argued that the ECMA was emplaced along with the first oceanic crust during the initial spreading of the margin [e.g., Klitgord and Behrendt, 1979; Austin et al., 1990]. Our shear velocity model supports this notion as the ECMA is closely aligned with the boundary between the transitional and oceanic crustal regions. The slight shallowing in the Moho, as well as the tightening of the velocity contours in Figure S7, further supports the hypothesis that the ECMA demarks the initial site of spreading. Modern spreading centers are generally characterized by locally thinned crust [e.g., Hammond et al., 2011; Dunn et al., 2013; Jin et al., 2015], and numerical modeling of passive margin formation and evolution suggests that locally thin crust may be preserved at the ocean-continent boundary [e.g., Davis and Lavier, 2017].

The source of the PGA is likewise associated with the initial rifting of the margin via crustal underplating of dense magmatic material. Our results show that the PGA is closely associated with the bulk of the transitional crust (Figure 3). Active-source studies to the north and south of the ENAM-CSE [e.g., Klitgord et al., 1988; Tréhu et al., 1989; Holbrook et al., 1994], and to a lesser extent our V_s model (Figure S7), are characterized by high lower crustal velocities in the transitional crust. The high lower crustal velocities are less apparent in our V_s model than in the active-source P wave velocity models primarily due to the sensitivities of the different techniques. During the breakup of Pangaea and the initiation of rifting, underplating of dense magmatic material beneath the transitional crust occurred along the margin [e.g., Holbrook and Kelemen, 1993; Eldholm and Grue, 1994; Skogseid et al., 2000]. Underplating was likely the result of melt generation due to the divergent motion of the plates or due to the emplacement of central Atlantic magmatic province material coincidentally with the formation of the margin [e.g., White and McKenzie, 1989; Hames et al., 2000]. The presence of anomalously dense magmatic material beneath the transitional crust would lead to the observed PGA.

5. Conclusion

We present a shear velocity model spanning the eastern North American margin from ambient noise-derived Rayleigh wave group and phase velocities. There is stark transition in crustal thickness across the margin from ~20 km thick oceanic crust to ~40 km thick continental crust. This transition occurs over ~100 km lateral distance and coincides closely with the positive free-air and isostatic gravity anomalies that run along the margin. We suggest that the PGA is the result of magmatic underplating beneath the transitional crust. Immediately adjacent to the transitional crustal region, we observe a localized thinning of the oceanic crust from ~20 km to ~17 km that aligns directly with the ECMA. We propose that ECMA and the locally thin crustal structure are a consequence of the initial formation of oceanic crust along ENAM. The ECMA, therefore, highlights the boundary between continental and oceanic crustal material.

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