Crustal structure beneath the East Coast Magnetic Anomaly from seismic refraction tomography

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Abstract

Syn-rift igneous addition is necessary for successful continental breakup. Past investigations of passive margins have focused on strike perpendicular structure, but potential field anomalies indicate that significant crustal variations may be present. Data from 21 ocean bottom seismometers was acquired as part of the Eastern North American Margin Community Seismic Experiment and was used for tomographic inversion to create 2D velocity models of the margin that are representative of crustal structure. Crustal thickness varies along-strike from ~20 km to ~24 km and a high velocity (Vp > 7 km/s) layer is present at the base of the crust above the Moho. The high velocity layer is interpreted as magmatic addition to the margin and has a significantly variable thickness along both velocity models. This suggests that magmatic addition to the margin is more variable on smaller scales than previously thought. Additionally, a region where elevated lower crustal velocities are missing is coincident to the Northern Fracture Zone and may be evidence of long-lived segmentation.
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1. Introduction

The eastern North American margin (ENAM) is a passive margin formed by the breakup of Pangea ~200 million years ago (Figure 1). Passive margins are a remnant of previous rifting, and the margins record rifting processes through time in their crustal structure. This thesis seeks to image crustal structure beneath ENAM along-strike using a recently acquired community seismic dataset to understand how igneous addition to the margin varies along-strike. In many continental rift environments, extensional tectonic forces are not strong enough to break apart continental lithosphere, but the rift can succeed with the assistance of igneous addition (Bott, 1991; Bialas et al., 2010).

Igneous addition to an actively rifting Pangea enabled the rift-to-seafloor-spreading transition leading to the modern Mid Atlantic Ridge. Passive margin crustal structure has traditionally been studied with long, strike perpendicular seismic lines but imaging of this nature does not provide a full picture of igneous addition because it does not capture smaller scale variability along-strike.

Previous seismic observations of ENAM have imaged igneous addition along the margin in the form of seaward dipping reflectors (SDRs) and high velocity lower crust (HVLC), therefore classifying it as magma-rich (volcanic) as opposed to magma-poor (nonvolcanic) (Tréhu et al., 1989; Sheridan et al., 1993; Austin Jr et al., 1990; Holbrook et al., 1994). The transition from regular continental crust to regular oceanic crust generally occurs over several hundred kilometers while the transition from rifted continental crust to early oceanic crust (igneous crust) occurs over ~100 km (Holbrook et al., 1994, 1992; Lynner and Porritt, 2017; Shuck et al., 2019). These surveys also
interpolated between strike perpendicular profiles to understand margin parallel structure, but smaller scale variability is likely as evidenced by variations of coastal geophysical anomalies. There have been no prior seismic surveys that have specifically imaged the along-strike structure of ENAM directly and therefore a data gap exists.

The prominent East Coast Magnetic Anomaly (ECMA) is a positive magnetic signal thought to be the result of volcanic addition to the margin. The ECMA has significant variability along-strike suggesting that there is more variation in crustal structure than strike-perpendicular profiles have constrained (Holbrook and Kelemen, 1993; Behn and Lin, 2000). Potential field modeling along the entire margin suggests broad changes in along-strike structure near the ECMA (Behn and Lin, 2000; Greene et al., 2017; Greene and Tominaga, 2018). ENAM can be divided into three regions of different seafloor spreading regimes and each region in turn has a unique source geometry for the ECMA (Greene et al., 2017; Greene and Tominaga, 2018) (Figure 1). Crustal structure models along-strike can provide geophysical ground truth to potential field interpretations; in addition, potential field models sensitive to a portion of the margin’s igneous addition (i.e. the ECMA source being the SDR packages) can be correlated with coincident seismic profiles to build a margin-wide understanding of igneous addition.

The goal of this work is to image the along-strike crustal structure of ENAM at a finer scale than previous surveys in order to better understand how magmatic addition varies along-strike. Inverted P-wave velocity models provide an image of depth vs. velocity variation with an along-strike resolution that has yet to be seen for this margin.
These velocity models illuminate along-strike variability of igneous addition to ENAM and give insight into the relationship between HVLC and the SDR package. Further, large scale velocity changes may indicate early evidence of margin segmentation that may extend to the current Mid-Atlantic Ridge. My results can be used to better understand how igneous addition varies along-strike, to build a relationship between intrusive and extrusive igneous activity, to groundtruth margin wide potential field modeling, and to look for evidence of long-lived segmentation at the Mid-Atlantic Ridge.

I use a recently acquired ocean bottom seismometer (OBS) dataset from the ENAM Community Seismic Experiment (ENAM-CSE) (Lynner et al., 2019) to seismically image the subsurface with refraction travel time tomography. Two along-strike velocity profiles, created using data from OBS line 4A and 4B, show regions of elevated lower crustal velocities (>7.2 km/s) that appear to be variable in thickness and discontinuous in some locations. Additionally, the Moho is imaged at depths of 23 – 26 km along-strike and shows some significant depth changes in regions where elevated lower crustal velocities are missing. The heterogeneous structure of the HVLC indicates that the igneous addition is not uniform along the margin and the overall distribution varies significantly. The HVLC has an apparent relationship with an interpreted magnetic source body: thicker regions of HVLC are co-located with thicker regions of the modelled magnetic source body and the thinnest section of magnetic source overlies no HVLC. The along-strike models of crustal structure may groundtruth potential field modeling of the margin as a whole. One previously extrapolated fracture zone is approximately
coincident to the discontinuous section of HVLC and may represent long lived segmentation of the Mid-Atlantic Ridge starting with the ECMA.

2. Tectonic Setting

2.1 The Breakup of Pangea

Pangea is Earth’s most recent supercontinent that was fully assembled by ~300 Ma and that began an early phase of rifting in the late Triassic (as early as 230 Ma), which did not succeed. A later phase in the early Jurassic (~190 Ma) successfully completed the rift-to-drift transition. On smaller scales, the breakup of Pangea and generation of new oceanic crust was diachronous along-strike with rifting starting earlier and occurring for a shorter time period in the southeastern US and occurring later and taking longer in the northeastern US and Canada (Withjack et al., 1998).

Large igneous provinces have frequently been found contemporaneously with the breakup of large continental areas (Coffin and Eldholm, 1992). The Central Atlantic Magmatic Province (CAMP) erupted 201 million years ago between Triassic and Jurassic rift phases, and can be found in locations across multiple continents (Withjack et al., 1998; Blackburn et al., 2013). CAMP and the breakup of Pangea are nearly contemporaneous, but it is unknown how they may be related or how CAMP may relate to igneous margin wide igneous addition.

2.2 ENAM

ENAM encompasses the North American coastal region from Nova Scotia, Canada to the Floridian peninsula, totaling a distance of approximately 2,500 km (Figure 1). The average trend of the margin is 37 degrees but there is significant variation of the
strike on smaller scales. ENAM contains several prominent coastal geophysical anomalies that are important for the understanding of the margin as a whole (Figure 2). The ECMA is a positive magnetic anomaly that follows the coastline for the length of ENAM and is variable along-strike with discrete peaks and troughs in amplitude. The ECMA has been interpreted by some as the site of initial seafloor spreading along the margin (Austin Jr et al., 1990; Talwani et al., 1995). The Blake Spur Magnetic Anomaly (BSMA) is found seaward of the ECMA and has been hypothesized as the site of a ridge jump following successful rifting (Greene et al., 2017) (Figure 2). Recent work from the ENAM CSE suggests that the BSMA is the site of initial seafloor spreading and that the ECMA is the start of proto-oceanic crust created by diffuse upwelling (Shuck et al., 2019). The coastal Positive Gravity Anomaly (PGA) is a free air gravity anomaly and has been interpreted to be primarily driven by the sharp relief of the continental slope and possibly a denser subsurface body. The PGA shows second order variability along-strike which may be more related to subsurface density differences (Behn and Lin, 2000; Lynner and Porritt, 2017).

Igneous addition during rifting is comprised of both magmatically intruded (HVLC) and volcanically erupted (SDRs) rocks with a basaltic composition. SDRs are formed as successive, subaerial volcanic flows and appear as seismically bright reflectors which forms wedges that thicken downdip (Buck, 2017). The HVLC is thought to result from mafic magmatic emplacement into the lower crust at or above the Moho. Partial decompression melting of the mantle, possibly with elevated mantle temperatures,
would generate the melt that collects near the base of the crust and can migrate upwards towards the eruption location (Van Avendonk et al., 2017; Shuck et al., 2019).

Fracture zones and transforms are a characteristic feature of mid ocean ridge segmentation. Mid Atlantic Ridge fracture zones can be traced eastward towards ENAM based on offset of seafloor magnetic anomalies, including the Kane and Northern fracture zones which are major offsets (Tucholke and Schouten, 1988). Behn and Lin (2000) and Klitgord and Schouten (1986) traced offsets (Kane, Northern and several minor) to M series magnetic chron M25, extrapolated the remaining distance to the BSMA based on plate motion reconstructions, and linearly extended them from the BSMA to the ECMA (Figure 3). Fracture zones cannot be traced farther than M25 due to the Atlantic Jurassic Quiet Zone (AJQZ), which exists between M25 and the US East Coast, where no seafloor magnetic chron can be defined or interpreted. The AJQZ has been subdivided into two further zones, the Inner Magnetic Quiet Zone (IMQZ), between the BSMA and the ECMA, and the Outer Magnetic Quiet Zone (OMQZ), between M25 and the BSMA (Greene et al., 2017; Klitgord and Grow, 1980). Offsets have been interpreted within the IJQZ that may relate to fracture zones found seaward of M25 (Greene et al., 2017) (Figure 3).

The presence of salt along ENAM was identified by early seismic surveys of the Carolina Trough south of Cape Hatteras; salt diapirs are extensive in the Carolina trough but are absent from the remaining Cape Hatteras region (Dillon et al., 1982; Tréhu et al., 1989) (Figure 1). Salt diapirs and layers are formed as an amalgamation of various salt minerals (e.g. Halite, Gypsum, Anhydrite etc.) with different physical properties. On
average, a salt diapir will have a density of ~2300 kg/m$^3$ and a seismic velocity of ~5.5 km/s (Jones and Davison, 2014). The salt layer for this portion of ENAM is located near the crystalline basement and diapirs rise through the sediments to near the surface (Tréhu et al., 1989).

2.3 Previous Geophysical Investigation

Previous large-source seismic surveys of in the Cape Hatteras region focused on long profiles perpendicular to the margin’s strike (Figure 2). All of the legacy surveys imaged the continent to ocean transition over ~100 km, a series of SDRs, and a region of HVLC. The EDGE 801 seismic line was acquired offshore of the Virginia-North Carolina state line (Holbrook et al., 1992; Sheridan et al., 1993). This survey found several distinct intracrustal reflections thought to be from the volcanic wedge (SDR package) and a northwest dipping reflector interpreted as the top of the underplated layer. The BA-6 seismic line is the from the southernmost survey in the Carolina Trough, acquired near the South Carolina – North Carolina boundary (Austin Jr et al., 1990; Holbrook et al., 1994). BA-6 imaged crustal structure across the Carolina trough through the BMA and the ECMA; the study found abundant seaward dipping reflectors and successfully modeled the BMA and ECMA using magnetized rocks above 20 km (their assumed Curie depth with no remnant magnetization below). The LASE study was the first large seismic profile in the Baltimore Canyon region acquired using expanding shot profiles. The LASE study did not penetrate to Moho depths and therefore could only provide constraint on more shallow structure (LASE Study Group, 1986). The USGS32 seismic line is located ~50 km north of BA-6 off of Cape Fear, North Carolina; USGS 32 provided the first
conventional deep crustal image along ENAM and acknowledged the presence of HVLC (Tréhu et al., 1989). The recent results of the strike perpendicular ENAM CSE lines found similar structure on either side of Cape Hatteras compared to the legacy data. The Moho beneath the ECMA has a reasonably consistent depth of ~21 km both north and south of Cape Hatteras, and the HVLC beneath the ECMA varies from ~6 km thick to the south and ~4 km thick to the north (Shuck and Van Avendonk, 2017; Shuck et al., 2019). The only other along-strike images of ENAM come from ENAM-CSE line 3, acquired within the BSMA, which found minimal variation in oceanic crustal structure for ~350 km along-strike (Shuck & Van Avendonk, 2017; Shuck et al., 2019). The strike perpendicular data and the far offshore along-strike image only sample a small portion of the area beneath the ECMA and provide an along-strike view of regular oceanic crust. Along-strike images sampling the crust directly beneath the ECMA are still necessary to understand igneous addition to the margin.

3. Data

3.1 ENAM Community Seismic Experiment

The OBS data for this study comes from the ENAM-CSE, centered on Cape Hatteras, North Carolina, in the middle of a data gap between previous seismic surveys (Figure 2). It was acquired in stages throughout 2014 and 2015 as a combination onshore-offshore survey with coincident OBSs and multichannel seismic (MCS) lines. The offshore active source component of the experiment, in 2014, deployed 47 short period OBSs from the R/V Endeavor and acquired ~4,816 km of MCS lines. The MCS and wide-angle seismic source was the R/V Marcus Langseth using its 6600-cu. in. tuned
airgun array. Complete acquisition of the offshore data was performed in 30 days from September 10 – October 10, 2014; the experiment required coordination between both vessels to deploy and retrieve OBSs while acquisition occurred.

For this study, we have focused on OBS lines 4A and 4B which follow the margin along-strike (Figure 3). Lines 4A and 4B intersect near the cusp of Cape Hatteras which also corresponds to a change in orientation of the margin itself. Previous interpretations of along-strike margin structure and variability have been made between widely spaced strike-perpendicular profiles, but lines 4A and 4B provide opportunity to image structures directly along-strike at much finer scales than previously possible.

3.2 OBS Data

Lines 4A and 4B follow the margin along-strike for ~375 km within the peak of the ECMA (Figure 3). Line 4A trends 26.97 degrees (N-NE) and is ~280 km long with 14 OBSs deployed during acquisition for an average spacing of ~15 km; Line 4B trends 6.85 degrees (N) and is ~230 km long with 9 OBSs deployed during acquisition for an average spacing of ~15 km (Figure 3). Acquisition parameters varied throughout the survey for MCS and OBS lines, but for both lines 4A and 4B the average shot spacing was ~225 meters; line 4A recorded 1228 airgun shots while line 4B recorded 998 airgun shots. The 23-total deployed OBSs recorded arrivals on four channels (vertical, two horizontals, and hydrophone). Arrivals were analyzed on both the vertical and hydrophone channel as this study’s primary focus is P-wave velocity structure and P-wave arrivals appear the best on those channels.
The signal-to-noise ratio for lines 4A and 4B is generally high thanks to the Langseth’s well-tuned source, but OBS gathers can be split into several groups based on the range and complexity of observed arrivals. OBS gathers from A401, A402, A403, and A406 have visible but complex arrivals within 40 km of the station and a reappearance of clear, high amplitude arrivals past 120 km (Figure 4a). OBS gathers from A404, A405, A407, and A408 have complex arrivals from the south, only visible within ~40 km offset, but have high quality arrivals on the north sides for the entire distance range (Figure 5a). OBS gathers from A408, A410, A412, A414, A415, A416, A418, B409, B411, B413, B415, B417, B420, B421, and B422 have high quality arrivals visible for the full range of offsets with minimal data gaps (Figures 6a-8a).

Previous studies of ENAM identified multiple sedimentary phases (Psed) in their wide-angle data along with a crustal refraction (Pg), an igneous basement reflection (PgP), a mantle refraction (Pn), and a Moho reflection (PmP). I initially picked the first arrivals and then added secondary phases. Two distinct sedimentary phases (sub vertical at a reduction velocity of 6 km/s) can be identified across most OBS gathers from both lines but a reflection between the phases cannot be easily identified. The basement refraction phase (Pg) constitutes arrivals faster than the sedimentary phase (>6 km/s apparent velocity) and sub-horizontal at a reduction velocity of 8 km/s. The mantle refraction (Pn) is visible at offsets farther than ~120 km on every OBS gather with some very far offsets arrivals visible (> +/- ~200 km). The Moho reflection (PmP) is picked between ~60 km and ~80 km on a majority of gathers while the basement reflection (PgP) is picked between ~10 km and ~30 km.
Every data gather was filtered with a Butterworth bandpass using frequencies of 4 Hz and 15 Hz for the edges of the pass range. Background noise is significantly reduced following the filter but still is prevalent on some gathers. In addition, previous shot noise occurs on every gather but any further processing to remove it was not performed because it obscures only a small section of arrivals. Multiphase picking was performed from the highest reduction velocity and then moving to the lowest, first with 8 km/s and then with 6 km/s. Any arrivals with faster than an apparent velocity faster or equal to 8 km/s was picked as Pn. Any arrivals with an apparent velocity faster than 6 km/s but slower than 8 km/s were picked as Pg. Any arrivals with an apparent velocity slower than 6km/s was picked as Psed. The igneous basement reflection is difficult to identify on a majority of the data gathers, but there are enough picks available to confidently use the arrivals for a sediment – igneous basement – mantle model.

Low-quality OBS gathers, with low signal to noise ratios and low amplitude regions, are located on the southern end of line 4A and several low-quality gathers exhibit a unique arrival with an apparent velocity of ~ 6km/s that arrives before the first sedimentary phase (Figure 4). Previous studies of ENAM noted a series of along-strike salt diapirs within the ECMA that are sourced from near the basement, reach near the seafloor in some locations, and appear approximately coincident to line 4A (Figure 1). Coincident MCS shipboard stacks from the ENAM CSE image two salt diapirs on the southern end of line 4A which would be in the immediate vicinity of OBS A403 and A404. Salt diapirs can have P-wave velocities in the 5 – 6 km/s range and would significantly distort the geometry of raypaths traveling through the diapir.
Uncertainties were assigned to all picked arrivals based on the source-receiver offset. The lowest uncertainty assigned was 50 ms for arrivals with less than 25 km of offset, an uncertainty of 75 ms was assigned to arrivals with an offset between 25 and 50 km, an uncertainty of 100 ms was assigned to arrivals with an offset between 50 and 120 km, and an uncertainty of 125 ms was assigned to arrivals with offsets farther than 120 km. These uncertainties are similar to many large wide angle experiments with similar survey parameters (Van Avendonk et al., 2004; Accardo et al., 2018; Shuck et al., 2019) In addition, all picks made on OBSs A401, A402, A403, and A404 were given an additional 25 ms of uncertainty to help overcome the inversion artifacts that arise due to the salt diapirs. The average prescribed uncertainty for line 4A is 95 ms, the average prescribed uncertainty for line 4B is 91 ms. Table 1 details the number of picks for each phase on each OBS and shows the average uncertainties respectively.

3.3 Potential Field Data

Several other geophysical datasets offer tools for interpreting the modeled velocity structure of ENAM. EMAG2v3, a global high-resolution magnetic grid compiled from satellite, ship, and airborne measurements with a 2 arc-minute resolution, was obtained from the National Oceanic and Atmospheric Administration (NOAA) data center (Meyer et al., 2017). A global 1 arc-minute gravity grid was obtained from University of California, San Diego (Sandwell et al., 2014). A global high-resolution bathymetry grid, ETOPO1, was also obtained from NOAA with a 1 arc-minute resolution (Amante, 2009) (Figure 2, 3a, 12a).
In addition to the global datasets, several regional potential field models were obtained including the isostatic gravity anomaly, the reduced to pole (RTP) magnetic anomaly, and the total magnetic source thickness for the ECMA (Behn and Lin, 2000; Greene and Tominaga, 2018) (Figure 2b-d). These anomalies provide more correlation potential to the subsurface structure imaged by Lines 4A and 4B because the anomalies are more directly related to structures directly below them. The isostatic gravity anomaly assumes constant density structure within each earth layer and removes the effect of boundary topography between layers. The modeling used density contrasts of 1300 kg/m$^3$ for the water-sediment boundary, 400 kg/m$^3$ for the sediment-crust boundary and 600 kg/m$^3$ for the crust-mantle boundary. Calculating the isostatic anomaly here subtracted the boundary effects and considered local Airy isostasy for crustal thickness (Behn and Lin, 2000). Remaining variability of the isostatic anomaly should be the result of different density materials and different boundary topography. The RTP magnetic anomaly is calculated by assuming all magnetization is vertical only, which pushes the major amplitudes directly above their source body (Behn and Lin, 2000). The ECMA modeling results calculated a total magnetic source wedge thickness for the length of ENAM (Greene and Tominaga, 2018).

4. Methods

I used traveltime tomography to invert the picked arrival times for each seismic shot to resolve the subsurface velocity structure. Tomography was performed using the VMTomo software package following the methods of (Van Avendonk et al., 2004; Roland et al., 2012). VMTomo uses an iterative raytracing and inversion process which
seeks to minimize traveltime residuals between picked and calculated arrivals. Each iteration reduces the residuals until the desired level of accuracy is attained.

Several 1D starting models were tested in order to minimize potential input bias; 1D models were taken from the intersection of Lines 4A and 4B with lines 1 and 2 from the ENAM CSE (Shuck et al., 2019) and from the equivalent strike perpendicular distance from the EDGE/LASE tomographic profiles (LASE Study Group, 1986; Holbrook et al., 1992, 1994) (Figure 9). In addition, both of these models were smoothed to remove smaller variations and lower crustal velocities between 7.1 and 7.9 km/s to assure that any high lower crustal velocity was not artificially introduced (Figure 9). The chosen 1D models were hung from the seafloor depth and interpolated across the 2D model space to create a set of tomographic starting models. Initial boundaries were prescribed to the model space at depths of 11 km and 25 km for the basement and moho respectively.

The water column, from the sea surface (0 km depth) to the seafloor, has a constant velocity of 1.5 km/s. We chose a grid size (1 km x .2 km) small enough to resolve subsurface features we believe our tomography can resolve but large enough to have a reasonable number of rays per grid.

The raytracing component of VMTomo creates a raypath using the shortest path method and ray bending for each shot-receiver combination and calculates a traveltime through the subsurface based on the current velocity model (Van Avendonk et al., 2004) (Figure 12). The code allows for the user to prescribe a grid size to raytrace through separate from the grid size of the velocity model; for both lines 4A and 4B we raytraced the model with a grid size of .2 km.
The inversion component of VMTomo compares the traveltime of every calculated raypath to the equivalent picked travel time and attempts to minimize the residual by updating the subsurface velocities and structures. The code calculates the current Chi² error metric from the residuals and attempts to create a subsurface structure that lowers the Chi² to a set level (Figure 4c – 8c). The user can prescribe several parameters including damping, flattening, and smoothing that influence the inversion result based on the previous iteration and the model’s first and second derivatives (Van Avendonk et al., 2004; Roland et al., 2012).

The Chi² error metric is a representation of model misfit and is calculated as shown in equation 1. Chi² is prescribed in the inversion step as the designated error level for an inversion; methodologies vary in regard to the error reduction per iteration and for these models a combination of reductions was used. Chi² was initially reduced by one half (50%) per iteration until it equaled ~3 and then Chi² was reduced by ~25% until it reached ~1. When starting Chi² values are large (> 50), reducing the misfit by 50% allows for faster convergence without the potential for model artifacts to be introduced. As Chi² lowers towards a more acceptable range, the value per iteration reduction is lowered to 25% to help prevent the emergence of artifacts. Chi² values of 1 are considered to be the ideal solution as there is no misfit between the model and picked traveltimes, however a range of Chi² between 1.4 and 1 is considered to be acceptable (Van Avendonk et al., 2004).

\[
\chi^2 = \sum_{n=1}^{N} \frac{(t_{n,calc} - t_{n,pick})^2}{\sigma_n^2}
\]

Equation 1
Traditional tomography inverts all input phases simultaneously but in some situations an adaptation to this process can improve the final inversion result. Early model iterations on Line 4A resolved some of the shallow salt structure but simultaneously introduced physically unlikely streaking and velocity inversions into the crust and mantle. To counter these features, a layer stripping approach was used in which the layers are isolated in the tomography process and run separately from shallow to deep until their individual $\chi^2$ are within the desired tolerances. Through this isolation, the shallow salt structure can be preserved when inverting deeper arrivals and allow for more realistic features in the crust and mantle. Line 4B does not have the prevalence of salt but in order to keep the process consistent layer-stripping was also applied.

5. Results

5.1 Line 4A Velocity Model

The velocity model for line 4A covers ~280 km of distance along-strike and reaches depths of ~30 km (Figure 10a). Following 17 iterations using 10236 picks (2000 Psed picks, 3655 Pg picks, 976 PgP picks, 1344 PmP picks, and 2261 Pn picks), the model has a $\chi^2$ error of 1.4 and an overall rms uncertainty of 100 ms. Ray coverage for line 4A is excellent from ~125 km to ~250 km along-strike but limited due to the anomalously high shallow velocities from ~25 km to ~100 km (Figure 11a). The bathymetry of line 4A is relatively consistent with a smooth seafloor on the kilometer scale and minimal depth variation along-strike. Within my modeled sediment layer, extending from the seafloor
to the basement reflector, velocities range from \( \sim 2 \text{ km/s} \) to \( \sim 5.8 \text{ km/s} \). Velocities at the seafloor are \( \sim 1.8 \text{ km/s} \) near each OBS and \( \sim 2.5 \text{ km/s} \) elsewhere, but there is an anomalous region between OBS A402 and A404 where velocities near the seafloor approach \( \sim 4-5 \text{ km/s} \). From the seafloor to the basement reflector, velocities increase with a gradient of \( \sim 0.49 \text{ s}^{-1} \). Within my modeled igneous basement layer, from the basement reflector to the Moho, velocities increase from \( \sim 6 \text{ km/s} \) to \( \sim 7.6 \text{ km/s} \) with a gradient of \( 0.133 \text{ s}^{-1} \).

Velocities in the sedimentary layer appear smooth with minimal depth variations along strike. Some visible variation in shallower velocities is due to ray streaking, an artifact generated during the inversion, that we attempted to minimize as much as possible. Velocities in the igneous basement layer have significantly more variation along-strike with large depth changes. The 7.5 km/s contour is more consistent than slower velocities above with less along-strike depth variation. The Moho reflector has an average depth of 25 km but is deepest (\( \sim 26 \text{ km} \)) on the southern end of the line and shallowest (\( \sim 24 \text{ km} \)) on the northern end. There is minimal depth variation along-strike with some long wavelength changes of \( \pm 1 \text{ km} \). Within the mantle, the velocity structure is close to 8 km/s with a region between \( \sim 100 \text{ km} \) and \( \sim 200 \text{ km} \) that has a velocity closer to \( \sim 7.9 \text{ km/s} \).

5.2 Line 4B Velocity Model

The velocity model for line 4B covers \( \sim 230 \text{ km} \) of distance along-strike and reaches depths of \( \sim 30 \text{ km} \) (Figure 10b). Following 12 iterations using 8629 picks (2140 Psed picks, 2241 Pg picks, 1268 PgP picks, 1355 PmP picks, and 1625 Pn picks), the
model has a Chi² error of 1.37 and an overall rms uncertainty of 92 ms. Ray coverage for line 4B is excellent for the entire extent along-strike; raypaths throughout the entire model provide good crossing paths and turning depths (Figure 11b). The bathymetry of line 4B is relatively consistent with a smooth seafloor on the kilometer scale but shallows from ~3 km on the south end of the line to ~2 km on the north end of the line. Within our modeled sedimentary layer, extending from the seafloor to the basement reflector, velocities range from ~2 km/s to ~6 km/s similar to line 4A and from the basement reflector to the Moho velocities range from ~6 km/s to ~7.6 km/s. Velocities at the seafloor are ~2.3 km/s with small regions beneath OBS B409, B420, and B421 having velocities of ~1.9 km/s. From the seafloor to the igneous basement reflector, velocities increase with a gradient of ~0.373 s⁻¹ to ~6 km/s. From the basement reflector to the Moho, velocities increase from ~6 km/s to ~7.6 km/s with a gradient of ~1 s⁻¹.

Velocities within the sedimentary layer appear smooth with minimal depth variation or waviness. There is less noticeable ray streaking compared to line 4A, likely due to denser ray coverage. Velocities within the igneous basement layer have more variation and contour waviness compared to the shallower velocity structure. The 7.5 km/s contour shallows from ~21 km depth to ~18 km depth between ~40 km and ~90 km along-strike and deepens significantly from ~19 km to ~23 km between ~140 km and ~160 km along-strike. Between ~100 km and ~140 km, there is a large discontinuity where velocity contours greater than 7 km/s sharply deepen and terminate into the Moho. The Moho reflector has an average depth of 24.5 km but exhibits significant variation in the center of line 4B coincident with the HVLC disappearance. On the
southern end of the line, the Moho is consistent at a depth of 24 km until ~ 100 km along-strike where it deepens to ~26.5 km over 25 km of distance. The Moho stays at ~26.5 km depth until ~ 125 km along-strike where it shallows to ~ 24.5 km depth. The mantle is primarily uniform in velocity near 8 km/s but directly below the Moho deepening the velocity decreases to ~ 7.9 km/s.

5.3 Combined Velocity Structure

The majority of lines 4A and 4B are located within the peak of the ECMA but the northernmost end of line 4A and the southernmost end of line 4B (northward and southward of the 4A-4B intersection point respectively) extend outside of the anomaly (Figure 3). Examining the structure of line 4A to the intersection, and the structure of line 4B past the intersection creates a truer along-strike view of crustal structure beneath the ECMA (Figure 12). Velocity models for line 4A and 4B were inverted independently of each other but are in good agreement at their intersection point with velocity contours and the Moho depth closely matching.

Legacy strike perpendicular studies have interpreted a maximum sediment velocity in the Carolina Trough to be ~ 6 km/s while crossing tomography from the ENAM CSE has a sediment velocity near ~5.5 km/s. Legacy studies and tomography from the ENAM CSE also interpret a minimum velocity of the SDR package (the top of the igneous basement) as ~6.3 km/s. I consider the velocity structure of less than 5.5 – 6 km/s to represent the sedimentary portion of the imaged crustal structure and any portion of the crust faster than ~6.3 km/s but above the Moho to be the igneous basement. Primarily, the sediment section’s consistent velocity structure indicates
uniform deposition on the kilometer scale. The shallow high velocity region appears coincident to the MCS imaged salt diapirs, but the model’s resolution is not sufficient to fully constrain diapir geometry or true velocity structure.

The deeper section of the crust has velocities from \(~6.3\) km/s to \(~7.6\) km/s, indicating that volcanic SDRs and magmatic HVLC are both present in this region. South of the 4A-4B intersection, the 7.2 km/s contours has an average depth of and the 7.5 km/s contour has an average depth of 24 km. North of the 4A-4B intersection, the 7.2 km/s contour has an average depth of 18 km and the 7.5 km/s contour has an average depth of 21 km. In agreement with legacy surveys and other velocity modeling from the ENAM CSE, I consider the 7.2 km/s contour to represent the top of HVLC, meaning that the HVLC thickness south of Cape Hatteras is \(~5\) km and the thickness north of Cape Hatteras is \(~7\) km.

**6. Discussion**

**6.1 Magmatic Addition to the Lower Crust**

Legacy strike perpendicular studies imaged magmatic addition to the crust as HVLC, with profiles spaced several hundred kilometers, beneath the ECMA with apparent long wavelength variations along-strike. The along-strike velocity models of lines 4A and 4B show more variation in thickness than strike perpendicular seismic studies could image. The along-strike profiles image HVLC that is on average 5 km thick south of Cape Hatteras and on average 7 km thick north of Cape Hatteras; in addition, the HVLC on line 4B is discontinuous over a span of \(~40\) km (Figure 12). Legacy strike perpendicular models show HVLC thickness beneath the ECMA varies from \(~11\) km
south of Cape Hatteras to ~15 km near Cape Hatteras and ~10 km north of Cape Hatteras (Holbrook and Kelemen, 1993; Holbrook et al., 1994; Austin Jr et al., 1990; Tréhu et al., 1989). Strike perpendicular models from the ENAM CSE have HVLC that is ~6 km south of Cape Hatteras and ~8 km thick north of Cape Hatteras, which agrees with broad observations along-strike. Line 4A has HVLC with variable thickness from 3 km to 7 km and appears more chaotic while line 4B has a more uniform thickness along-strike that varies by +/- ~1 km. The intersection point of line 4A and 4B, coincident to Cape Hatteras, appears to serve as an important boundary where crustal structure is different on either side.

An explanation for the thickness variation of HVLC could be different mantle potential temperatures on either side of the 4A-4B intersection. A higher mantle potential temperature would generate more melt from the mantle and lead to thicker HVLC while a lower temperature would generate less melt and lead to thinner HVLC. At mid ocean ridges, a mantle potential temperature anomaly of 1 °C leads to igneous crust that is ~50 – 70 meters thicker (Van Avendonk et al., 2017). Petrologic modeling of Line 3 found that a 25 °C difference (a 1.25 – 1.75 km thickness difference) from south to north could explain the slight crustal thickness variation along-strike beneath the BSMA (Shuck et al., 2019). The total igneous thickness, however, is consistent on either side of Cape Hatteras which would indicate that mantle potential temperature may not have had an effect on magmatic addition to the margin (Figure 12).

Strike perpendicular studies used the cross-sectional area of HVLC to estimate a volume of magmatic addition to the margin as a whole. The along-strike models of lines
4A and 4B only represent a small portion of the total volume of magmatic addition, but it is possible to compare the cross-sectional area imaged by 4A and 4B to the equivalent area determined by strike perpendicular models. Along-strike, the top of HVLC on line 4B-4B combined extends for ~300 km and multiplying the thickness by this distance gives an area of 1420 km². Taking the HVLC thickness from BA-6 to the south and EDGE 801 to the north and multiplying by the same along-strike distance gives an area of 3600 km². Using the strike perpendicular thickness of lines 1 and 2 from the ENAM CSE and the same distance gives an area of 1650 km². The along-strike profile of 4A and 4B leads to a ~60% decrease in HVLC cross-sectional area compared to estimates from the legacy datasets and a ~14% decrease compared to line 1 and 2. The calculated cross-sectional areas could be used to calculate a rough volume of HVLC beneath the ECMA. The HVLC imaged by legacy datasets and lines 1 and 2 have an approximately consistent thickness, but this calculation leads to a similar estimated decrease as the areal calculation. The estimated decrease indicates that along-strike imaging of ENAM can significantly refine estimates of magmatic addition to the margin and even strike perpendicular results from the same experiment would overestimate the volume of magmatic addition.

Strike perpendicular velocities models imaged HVLC with maximum velocities of ~7.5 km/s that was present at most of the lower crust beneath the ECMA (Holbrook and Kelemen, 1993). Velocity models from lines 1 and 2 image HVLC with velocities >= 7.5 km/s but it is not laterally continuous across the HVLC region (Shuck et al., 2019). At the intersection points of line 1 with line 4B and line 2 with line 4A, HVLC with velocities >= 7.5 km/s are present along-strike but not present perpendicular to strike. It is possible
that the strike perpendicular models do not sample the lower crust beneath the ECMA well enough to resolve the 7.5 km/s structure, but it is present in other locations. An alternative explanation of this could be anisotropy of the HVLC which would exhibit faster velocities in an along-strike orientation and slower velocities in a strike perpendicular orientation. HVLC is a mafic gabbro, with increased MgO, similar to an olivine gabbro (White and McKenzie, 1989). Oceanic upper mantle anisotropy has been documented at mid ocean ridges where the Pn velocity is faster parallel to the spreading direction and slower perpendicular to it; anisotropy of layer 3 in oceanic crust, also an olivine gabbro, has also been documented in an opposite orientation where Pg through layer 3 is slower parallel to the fossil spreading direction and faster perpendicular to it (Christensen and Mooney, 1995). It is possible that the velocity reduction at the line intersection points due to anisotropy is enough to make the velocity structure mismatch, but further investigation would be needed to verify this phenomenon beneath the ECMA.

6.2 Connections between HVLC and SDRs

Margin wide 3D potential field modeling has estimated a magnetic source body responsible for the ECMA (Greene and Tominaga, 2018); the magnetic source body has been interpreted prior as the SDR package (Austin Jr et al., 1990; Holbrook et al., 1992, 1994). Overlaying the source body onto the velocity models from line 4A and 4B shows the source relation to the subsurface velocity structure (Figure 12). The magnetic source top approximately follows the 5.5 – 6 km/s velocity contours, while the source bottom approximately follows the 6.5 – 7.2 km/s velocity contours. Legacy margin perpendicular
modeling found the SDR package to have a velocity range from ~6 – 6.9 km/s and the HVLC to have a velocity range from ~7.1 – 7.5 km/s (Holbrook and Kelemen, 1993). Modeling of ENAM CSE lines 1 and 2 imaged igneous basement velocities ranging from ~5.6 km/s to 7.5 km/s (Shuck et al., 2019). The SDR package is hypothesized to be a series of volcanoclastic flows that have been imaged as reflectors and have an assumed velocity range from ~ 6 – 6.9 km/s. The velocity structure within the magnetic source is close to the interpreted SDR package for the margin, indicating that the previously interpreted SDR package and the magnetic source are located in the same depth range.

My velocity models of lines 4A and 4B lack a defined boundary between the upper crust and the HVLC, but the magnetic source body can be used as a proxy of the SDR package thickness for correlation with the HVLC.

The HVLC was formed as a magmatic intrusion and the SDR package was deposited extrusively. The HVLC-SDR system can be thought of as a magma chamber and the associated lava flows sourced from the magma chamber. The velocity models from lines 4A and 4B show a physical relationship between the HVLC and SDR thicknesses; south of Cape Hatteras, on line 4A, the SDR package has a semi-consistent thickness along-strike above HVLC which also has a semi consistent thickness while north of Cape Hatteras, on line 4B, the SDR package is the thinnest above the section of lower crust with no HVLC but then thickens above the return of HVLC (Figure 12). The HVLC is inherently related to the SDR package as if there is no HVLC present, then there is minimal melt available to form the SDRs.
The velocity profile of lines 4A and 4B along-strike images a significant change in crustal structure on either side of the 4A-4B intersection. The cusp of Cape Hatteras, coincident to the 4A-4B intersection, marks a change in orientation of the margin as well as the boundary between the Carolina (south of Cape Hatteras) and Mid-Atlantic (north of Cape Hatteras) early spreading regime regions (Greene et al., 2017) (Figure 1). The Carolina region has an estimated half spreading rate in the OMQZ of 25 mm/yr while the Mid-Atlantic region has an estimated half spreading rate of 18 mm/yr (Greene et al., 2017). Spreading rate determinations landward of the BSMA, within the IMQZ, are less constrained but estimated half-rates range from ~4 – 13 mm/yr (Lundin et al., 2018). The regional difference in OMQZ spreading rates may have been consistent through the IMQZ up to the formation of the ECMA. A faster extension rate to the south of Cape Hatteras would lead to enhanced melt extraction from the HVLC and therefore a thinner remnant magma chamber. Slower extension rates north of Cape Hatteras would inhibit melt extraction from the HVLC and lead to a thicker remnant magma chamber.

The 3D potential field modeling for the whole margin also interpreted three broad magnetic source regions that follow the previously interpreted regions with different early spreading rate regimes (Greene et al., 2017; Greene and Tominaga, 2018) (Figure 1). South of Cape Hatteras, the ECMA source body is interpreted to be broad and thin while north of Cape Hatteras the source body is narrower and thicker. This boundary is apparent in the along-strike view of crustal structure where the magnetic source body has a consistent thickness and depth south of Cape Hatteras and is thicker and deeper north of Cape Hatteras (Figure 12). This boundary also appears to separate
different lower crustal character as well; south of Cape Hatteras the HVLC is thinner and has a consistent thickness, but north of Cape Hatteras the HVLC is thicker and discontinuous (Figure 12). A broad and thin section of SDRs south of Cape Hatteras would indicate that volcanism was widely distributed across the margin for that region similarly the HVLC should be broadly distributed. The more narrow and thick section of SDRs north of Cape Hatteras indicate that volcanism was more focused for that region and would imply that the lower crust was more focused as well. These scenarios match what is observed on either side of Cape Hatteras, with thinner but more distributed HVLC to the south and thicker HVLC to the north.

6.3 Early Margin Segmentation

Fracture zones and transform faults are a characteristic feature of mid ocean ridges that create spreading center segmentation by separating ridge segments and forming both geophysical and geochemical boundaries (Hooft et al., 2000). It is unknown if mid ocean ridge segmentation starts after seafloor spreading initiation or if it is an inherent process due to structures present during continental rifting. My tomography results show that the region north of Cape Hatteras on line 4B, with discontinuous HVLC and vertically oriented lower velocity contours, is located within ~10 km of the Northern fracture zone’s extrapolated location (Figure 12). The region with no HVLC is coincident to a ~2 km deepening of the Moho and is directly correlated to a sharp decrease in the unfiltered Isostatic Gravity Anomaly (Figure 12). The fracture zone location also intersects line 4A at ~245 km along-strike and line 3 at ~190 km along-strike, but this location on line 4A is not within our resolvable area at depth.
(Figure 10), and there is no structural evidence present on line 3 (Shuck et al., 2019). This region has a velocity structure similar to the starting model for line 4B but testing of the gap by using a starting model with HVLC present also led to a similar structure. If this lower velocity region is caused by presence of the Northern fracture zone, then it has been a feature of the Mid-Atlantic Ridge since the formation of the ECMA at ~190 Ma.

Tomography results across the Gofar and Quebrada transform faults at the East Pacific Rise show two different velocity signatures, one with reduced velocities and vertically oriented contours reaching the Moho and the other with slightly reduced velocities over a broader region but still reaching deep into oceanic layer 3 (Roland et al., 2012). At another segment of the East Pacific Rise, the Clipperton transform shows an in-between case with more drastic velocity reductions but with minimal vertically oriented features (Van Avendonk et al., 1998, 2001). The Gofar-Quebrada study also crossed several young fractures zones, which exhibit some component of velocity reduction, albeit less than at the transform fault itself, that did not reach the Moho but did extend into layer 3 (Roland et al., 2012). Velocity reductions have been attributed to small fractures formed along the fault plane which increase the rock porosity and therefore decrease seismic velocity (Van Avendonk et al., 1998, 2001; Roland et al., 2012). Roland et al. (2012), hypothesized two cases to explain the velocity reduction at fossil fracture zones. The first has the fracture rocks in layer 2 and layer 3 healing with time, over ~ 1 myr, removing the velocity reduction in all but the shallowest crust. The second would require a weaker velocity reduction that never reached significant depths and the reduction has been locked in since the crust was moved off axis. In the second
case, the Gofar transform would be an example of a strong velocity reduction and the Quebrada and Clipperton transforms would be examples of weak reduction.

The velocity reduction on line 4B near the Northern fracture zone is similar to a strong reduction as at the Gofar transform, but the crust beneath the ECMA is significantly older. If the velocity reduction is indeed due to the Northern fracture zone, then two possibilities exist for the reduction lasting over such a length of time. The first possibility is that the velocity reduction was unable to heal in a manner similar to the fossil fracture zones of the East Pacific Rise. The thicker igneous crust from the ECMA to the BSMA has been interpreted as proto-oceanic crust created by diffuse upwelling, not focused spreading like at a normal mid ocean ridge. The crustal formation may inhibit the same type of healing possibly observed near the Gofar transform. The second possibility is that the velocity reduction was locked in at the time of formation, similar to the alternative hypothesis of Roland et al. (2012), regardless of the specific crustal formation process and has remained since.

The lack of velocity reduction on line 3 indicates that either healing did occur within the proto oceanic crust between the ECMA and the BSMA and that the crust beneath the ECMA is unique, or that the discontinuous HVLC on line 4B may not be evidence of a fracture zone and instead is a preexisting structural boundary that inhibited magmatic intrusion and emplacement beneath the ECMA. Further MCS imaging of the crustal structure beneath the ECMA is needed to look for further evidence of the fossil fracture zone on line 4B and to reconcile the different scenarios of discontinuous HVLC beneath the ECMA.
7. Conclusions

The tomographic results from this study provide the first high resolution along-strike views of crustal structure at the eastern North American Margin; my models show that the magmatic addition to ENAM is more variable on smaller scales than previously known. The heterogeneous structure of the HVLC imaged on lines 4A and 4B indicate that igneous addition is not uniform along the margin and the overall distribution varies significantly. Imaged HVLC on lines 4A and 4B show smaller scale thickness variability and a different thickness on either side of Cape Hatteras which corresponds to a boundary of different magnetic source and seafloor spreading rate regimes. The modeled magnetic source body, interpreted as SDRs and confirmed by the velocity structure within, has a relationship with the lower crust where thinner HVLC corresponds to thinner SDRs and thicker HVLC corresponds to thicker SDRs. The region just north of Cape Hatteras with no HVLC may be evidence that the Northern fracture zone has existed since the ECMA formation and deformation at the fracture zone has not healed, but further structural imaging with the coincident MCS data is needed for verifications.
8. References


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other passive margins: AAPG bulletin, v. 82, p. 817–835.
Figure 1. Map of the eastern North American margin, modified from Withjack et al., (1998) showing locations of Jurassic dikes, the East Coast Magnetic Anomaly, SDRs, and Jurassic basalt flows. Identified salt diapirs are shown as white circles (Tréhu et al., 1989). Early spreading rate regime regions are identified, and boundaries are shown as dashed lines (Greene et al., 2018). The ENAM CSE survey area is shown by the black box with its center near Cape Hatteras, North Carolina starred. The Carolina Trough (CT) and Baltimore Canyon Trough (BCT) are shaded and noted.
Figure 2. ENAM CSE offshore survey design with geophysical features of the region. Multichannel seismic reflection lines are shown by black solid lines, short period ocean bottom seismometer deployments are shown by yellow circles, and legacy seismic experiment lines EDGE-801, BA-6, USGS 32 and LASE are shown as black labeled lines. Fracture zone extrapolations of Klitgord and Schouten (1986) are shown as black dashed lines and interpreted inner magnetic quiet zone offsets from Greene et al. (2017) are shown as green lines. The background map is from EMAG2v3 (Meyer et al., 2017) with the East Coast Magnetic Anomaly (ECMA) and the Brunswick Magnetic Anomaly (BMA) labeled.
Figure 3. ENAM CSE lines 04A and 04B shown over maps of the Cape Hatteras, NC region. (a) Bathymetry map from ETOPO1 (Sandwell et al., 2014). (b) Isostatic gravity anomaly calculated for length of ENAM (Behn and Lin, 2000). (c) Calculated magnetic source wedge thickness (Greene and Tominaga, 2019). (d) RTP magnetic map calculated for length of ENAM (Behn and Lin, 2000). Fracture zone extrapolations of Klitgord and Schouten (1986) are shown on all maps with interpreted inner magnetic quiet zone offsets of Green et al., (2017).
Figure 4a. Data gather from OBS A402 showing the hydrophone channel at a reduction velocity of 8 km/s. Upper image is uninterpreted and lower image is interpreted with arrival picks. Inset is a zoom on boxed region at a reduction velocity of 6 km/s showing the anomalously early, fast arrival.
Figure 4b. Upper image is picked (multicolor) and calculated (red) arrivals shown at a reduction velocity of 7 km/s for OBS A402. Lower image is equivalent raypaths for each arrival through the model space.
Figure 5a. Data gather from OBS A407 showing the vertical channel at a reduction velocity of 8km/s. Upper image is uninterpreted and lower image is interpreted with arrival picks.
Figure 4b. Upper image is picked (multicolor) and calculated (red) arrivals shown at a reduction velocity of 7 km/s for OBS A402. Lower image is equivalent raypaths for each arrival through the model space.
Figure 6a. Data gather from OBS A416 showing the vertical channel at a reduction velocity of 8km/s. Upper image is uninterpreted and lower image is interpreted with arrival picks.
Figure 6b. Upper image is picked (multicolor) and calculated (red) arrivals shown at a reduction velocity of 7 km/s for OBS A16. Lower image is equivalent raypaths for each arrival through the model space.
Figure 7a. Data gather from OBS B411 showing the vertical channel at a reduction velocity of 8km/s. Upper image is uninterpreted and lower image is interpreted with arrival picks.
Figure 7b. Upper image is picked (multicolor) and calculated (red) arrivals shown at a reduction velocity of 7 km/s for OBS B411. Lower image is equivalent raypaths for each arrival through the model space.
Figure 8a. Data gather from OBS B420 showing the vertical channel at a reduction velocity of 8km/s. Upper image is uninterpreted and lower image is interpreted with arrival picks.
Figure 8b. Upper image is picked (multicolor) and calculated (red) arrivals shown at a reduction velocity of 7 km/s for OBS B420. Lower image is equivalent raypaths for each arrival through the model space.
Figure 9. One dimensional (1D) starting velocity models for lines 4A and 4B taken from ENAM CSE lines 1 and 2. Solid lines are exact 1D profiles at the intersection points, dashed lines are smoothed to removed velocity inversions and high velocity lower crust.
Table 1. Summary table of picked arrivals per line, phase, and specific OBS.

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Line 4A

Line 4B
Figure 10. Final velocity models for line 4A (upper) and line 4B (lower). Key points including the line 1 and 2 intersections, the line 4A-4B intersection, and the Kane, Northern, and Norfolk fracture zone intersections are noted.
Figure 11: Ray coverage of line 4A (upper) and 4B (lower) both displayed with every 3rd ray only shown for clarity.
Figure 12. (upper top) Along-strike magnetic anomaly plot from EMAG2v3 in black and the calculated RTP anomaly of Behn and Lin, (2000) in red. (lower top) Along-strike gravity anomaly plot from the global free air anomaly in black, the calculated isostatic anomaly of Behn and Lin, (2000) in red, and the long wavelength filtered isostatic anomaly of Behn and Lin, (2000) in green. The profile is line 4A from the south end to the 4A-4B intersection, and line 4B north of the intersection.

(lower) Along-strike view of crustal structure beneath the ECMA. The profile is line 4A from the south end to the 4A-4B intersection, and line 4B north of the intersection. Bolded lines overlain on model are magnetic source body top and bottom from Greene et al., (2018).
10. Data Supplement

Supplement Figure 1. Data gather from OBS A401 showing the hydrophone channel at a reduction velocity of 8km/s. Upper image is uninterpreted and lower image is interpreted with arrival picks.
Supplement Figure 2. Data gather from OBS A403 showing the vertical channel at a reduction velocity of 8km/s. Upper image is uninterpreted and lower image is interpreted with arrival picks.
Supplement Figure 3. Data gather from OBS A404 showing the vertical channel at a reduction velocity of 8km/s. Upper image is uninterpreted and lower image is interpreted with arrival picks.
Supplement Figure 4. Data gather from OBS A405 showing the hydrophone channel at a reduction velocity of 8km/s. Upper image is uninterpreted and lower image is interpreted with arrival picks.
Supplement Figure 5. Data gather from OBS A406 showing the hydrophone channel at a reduction velocity of 8km/s. Upper image is uninterpreted and lower image is interpreted with arrival picks.
Supplement Figure 6. Data gather from OBS A408 showing the hydrophone channel at a reduction velocity of 8km/s. Upper image is uninterpreted and lower image is interpreted with arrival picks.
Supplement Figure 7. Data gather from OBS A410 showing the hydrophone channel at a reduction velocity of 8km/s. Upper image is uninterpreted and lower image is interpreted with arrival picks.
Supplement Figure 8. Data gather from OBS A412 showing the hydrophone channel at a reduction velocity of 8km/s. Upper image is uninterpreted and lower image is interpreted with arrival picks.
Supplement Figure 9. Data gather from OBS A414 showing the hydrophone channel at a reduction velocity of 8km/s. Upper image is uninterpreted and lower image is interpreted with arrival picks.
Supplement Figure 10. Data gather from OBS A415 showing the hydrophone channel at a reduction velocity of 8km/s. Upper image is uninterpreted and lower image is interpreted with arrival picks.
Supplement Figure 11. Data gather from OBS A418 showing the hydrophone channel at a reduction velocity of 8km/s. Upper image is uninterpreted and lower image is interpreted with arrival picks.
Supplement Figure 12. Data gather from OBS B409 showing the hydrophone channel at a reduction velocity of 8km/s. Upper image is uninterpreted and lower image is interpreted with arrival picks.
Supplement Figure 13. Data gather from OBS B413 showing the hydrophone channel at a reduction velocity of 8km/s. Upper image is uninterpreted and lower image is interpreted with arrival picks.
Supplement Figure 14. Data gather from OBS B415 showing the hydrophone channel at a reduction velocity of 8km/s. Upper image is uninterpreted and lower image is interpreted with arrival picks.
Supplement Figure 15. Data gather from OBS B417 showing the hydrophone channel at a reduction velocity of 8km/s. Upper image is uninterpreted and lower image is interpreted with arrival picks.
Supplement Figure 16. Data gather from OBS B421 showing the hydrophone channel at a reduction velocity of 8km/s. Upper image is uninterpreted and lower image is interpreted with arrival picks.
Supplement Figure 17. Data gather from OBS B422 showing the hydrophone channel at a reduction velocity of 8km/s. Upper image is uninterpreted and lower image is interpreted with arrival picks.