Lithospheric and asthenospheric anisotropic gradients along the Eastern North American Margin imaged by frequency dependent quasi-Love wave scattering

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Abstract

The eastern North American margin has experienced a wide array of plate-scale tectonic deformational events, including the breakup of Pangaea. The margin may also host complex patterns of active asthenospheric mantle dynamics. Several studies have observed a strong change in anisotropy across the margin that have been interpreted variously as active asthenospheric flow or past lithospheric deformation. Separating these candidate processes has proven difficult. To constrain the likely source of the change in anisotropy across the margin, we examine scattered quasi-Love waves over three frequency ranges with peak sensitivities in the lithosphere (~75km), the uppermost mantle (~150km), and the asthenosphere (~250km). We observe strong quasi-Love wave scattering along the margin in the lowest two frequency bands but far fewer scatterers in the lithosphere-dominated highest frequency band. The clear frequency dependence suggests a change in anisotropy across the margin is likely located in the asthenosphere and related to active mantle dynamics.

Introduction

The modern configuration of the eastern North American passive margin (ENAM) dates to the breakup of Pangea, with ~200 Ma rifting followed shortly by the onset of seafloor spreading [e.g., Withjack and Schlische, 2005; Thomas, 2006]. Deformation resulting from the early stages of continental breakup may still be preserved in the margin lithosphere allowing us to probe the circumstances related to continental breakup [e.g., Lynner and Porritt, 2017; Shuck et al., 2019; Worthington et al., 2021]. The ENAM may also host complex patterns of deeper asthenospheric mantle flow. Several previous studies of the margin have argued for complex mantle flow patterns including various styles of edge-driven convection [e.g., King, 2007; Ramsay and Pysklywec, 2011; Savage et al., 2017; Evans et al., 2019; Gao and Li, 2021; Savage, 2021], localized upwellings and downwellings [e.g., Rowley et al., 2013; Byrnes et al., 2019], pressure driven flow [e.g., Conrad and Behn, 2010; Liu, 2015], or combinations thereof [Long et al., 2021]. Several of these predict a change in mantle flow direction and/or dip near the continent-ocean boundary and are primarily driven by variations in lithospheric thickness across the margin, which changes from ~175km thick beneath the Appalachian Mountains to <90km thick at the coast [Brunsvik et al., 2021; Eilon et al., 2023].

These mechanisms may explain the pattern of seismic anisotropy along the margin, which is counterintuitively margin-parallel offshore and somewhat complex onshore [e.g., Long et al., 2016; Lynner and Bodmer, 2017; Yang et al., 2017; Brunsvik et al., 2021]. This differs from a simplistic expectation of extension-parallel frozen-in fabrics in the plate [Vauchez et al., 2000] and may require depth-dependence. Distinguishing between processes associated with the initial rifting of the margin in the lithosphere and those related to mantle dynamics is essential to understanding the evolution of the ENAM.

Measurements of seismic anisotropy provide direct constraints on patterns of deformation in both the lithospheric and asthenospheric mantle [e.g., Barruol et al., 1997; Deschamps et al., 2008; Yuan and Romanowicz, 2010; Wagner et al., 2012; Long et al., 2010; 2016; Yang et al., 2017]. Mantle seismic anisotropy originates from the lattice preferred orientation of olivine. As the upper mantle, lithospheric or asthenospheric, undergoes shear strain, crystal slip system limitations result in rotation of olivine grains yielding a bulk preferred orientation of anisotropy [Karato et al., 2008]. Several olivine fabric types exist, each with slightly different seismic anisotropy signals, and the dominant fabric depends on the prevailing thermodynamic and hydration conditions [e.g., Karato et al., 2008; Lynner et al., 2017]. Likely fabric types for ENAM mantle conditions predict fast anisotropic directions parallel to the (time-integrated) maximum strain axis. Once formed, bulk seismic anisotropy should largely persist until subsequent deformation realigns fast directions or high-temperature annealing occurs [Boneh et al., 2017]. For the cold lithospheric mantle, widespread deformation is infrequent and tied to significant tectonic events, such as the formation or breakup of Pangaea, and high temperature annealing would be unlikely. In the asthenosphere, anisotropy orientations arise from ever evolving mantle flow and are indicative of modern mantle dynamics. The ENAM is characterized by both significant past lithospheric deformation and complex modern mantle dynamics leading to potential anisotropy in both regions.

Perhaps the best-known method for examining seismic anisotropy in the mantle is shear wave splitting [see reviews by Savage, 1999; Long and Silver, 2009]. Shear wave splitting provides constraints on both the orientation and strength of anisotropy beneath seismic stations. Many studies have examined shear wave splitting along the ENAM [e.g., Barruol et al., 1997; Levin et al., 1999; Fouch et al., 2000; Long et al., 2010; 2016; Yang et al., 2017; Lynner and Bodmer, 2017; Aragon et al., 2017; Brunsvik et al., 2021]. Recently, Long et al. [2010; 2016] measured shear wave splitting at stations along the east coast and found that stations nearest the coast exhibited extremely weak or absent shear wave splitting while stations farther inland showed strong orogen-parallel splitting. Aragon et al. [2017] examined shear wave splitting along a transect of stations that ran from the coast through the Appalachian Mountains and observed a similar pattern. Both interpreted splitting patterns as asthenospheric flow dominating the anisotropic signal beneath the coast, while the orogen parallel splitting was lithospheric sourced. Yang et al. [2017] examined splitting at similar stations as Long et al. [2016] but in a different frequency range and found stronger splitting along the coast that is aligned margin parallel. They too interpret their splitting results as a combination of lithospheric and asthenospheric anisotropy.

Lynner and Bodmer [2017] examined shear wave splitting at stations located offshore ENAM allowing for a margin crossing view of seismic anisotropy. The offshore splitting is remarkably consistent with margin parallel anisotropy seen as far as ~500km away from continent-ocean transition (COT). Combining the onshore and offshore shear wave splitting studies suggests a transition in anisotropy across the margin. Lynner and Bodmer [2017] interpreted these results as complex edge-driven flow in the asthenospheric mantle driven by a change in lithospheric thickness across the margin. While there are various interpretations of lithospheric and asthenospheric deformation in each of the splitting studies along the margin, they are somewhat tenuous as shear wave splitting is a path-integrated measurement and offers no constraints on the depth of anisotropy.

Other methods of examining seismic anisotropy that can provide depth constraints have also been employed along the ENAM. Wagner et al. [2018] developed a surface wave model of azimuthal anisotropy along the coast. They found weak horizontal anisotropy at asthenospheric depths and mixed lithospheric anisotropy directions and strengths along the margin. Like most of the shear wave splitting studies, however, their model was restricted to the land portion of the margin and misses the transition between continental interior and the oceanic realm. Depth varying anisotropy in the oceanic portion of the margin was examined by Russell et al. [2021] using ambient noise surface waves. They found that the upper lithosphere in the offshore is characterized by margin-parallel-fast anisotropy similar to the orientations seen by Lynner and Bodmer [2017]. They argue that the splitting seen in Lynner and Bodmer [2017] could be explained by lithospheric anisotropy if their observations persist throughout the entire lithosphere. Ambient noise inversions provide excellent resolution in the upper ~50km, but quickly lose sensitivity at deeper depths leaving most of the oceanic lithosphere and the entirety of the asthenosphere unresolved. Brunsvik et al. [2021] created a 3D anisotropic model of the ENAM that crossed the COT and found margin-parallel anisotropy in the asthenospheric mantle and margin-perpendicular anisotropy in the lithosphere beneath the oceanic portion of the plate, contradicting the findings of Russel et al. [2021]. The conflicting results leave open the question of lithospheric or asthenospheric sourced anisotropy along the margin. It is clear from the shear wave splitting results and surface wave models that a change in anisotropy exists across the margin. The sharp anisotropic gradient may be indicative of a change in asthenospheric mantle dynamics producing variable anisotropic orientations [e.g., Lynner and Bodmer, 2017] or due to a change in frozen-in lithospheric anisotropy between the oceanic and continental components of the North American plate [e.g., Russell et al., 2021].

We can constrain such abrupt lateral gradients in seismic anisotropy using observations of quasi-Love wave scattering [e.g., Rieger and Park, 2010; Chen and Park, 2013; Margheriti et al., 2017; Eakin, 2021; Cheng et al., 2021]. When a Love wave encounters a strong anisotropic gradient, a component of the Love wave's transverse energy is scattered into a quasi-Love (qL) wave that travels on the vertical and radial components. This qL wave inherits the shape and character of the originating Love waveform at the point of scattering and thence propagates as a Rayleigh wave. By measuring the time difference between the Love and qL waves, we can back-project the location of the originating anisotropic gradient.

Servali et al. [2020] observed quasi-Love wave scattering along the ENAM concentrated along the COT, supporting previous anisotropy observations. The frequencies used by Servali et al. [2020] span the entire upper mantle, preventing them from discerning the depth extent of the anisotropic gradient. In this study, we separate qL scattering by frequency to identify the depths of strong anisotropic gradients and elucidate the source of inferred fabrics. We examine three frequency bands that provide sensitivity to: 1) the lithosphere (peak sensitivity at ~75km); 2) the entire upper mantle, similar to what was used by Servali et al. [2020], and 3) the uppermost mantle representing the shallow asthenosphere (peak sensitivity at ~150km depth).

Methods and Data

We examine qL scattering at all broadband seismic stations that operated at any time along the east coast of the US (Fig. 1) between 2012 and 2018. We restrict our analyses to earthquakes with magnitudes greater than 6.5, at least 60º distance from the margin, and which produced surface waves that travel across the ENAM COT from the oceanic portion of the plate onto the continental region. We measured qL scatterers from 14 events that met these criteria. Not all stations were operating continuously between 2012 and 2018. Most of our measurements were made between 2013 and 2015 when the EarthScope Transportable Array (TA; IRIS Transportable Array, 2003) was active along the east coast. In addition to TA stations, our dataset was augmented by the many additional temporary and permanent stations deployed along the margin (Supplemental Table S1).

We follow the methodologies of Chen and Park [2013] and Servali et al. [2020] to process quasi-Love scattering in our dataset. We first apply a series of bandpass filters to our data. We use: 110 – 100s, 80 – 70s, and 60 – 50s for the low, mid, and high frequency bands, respectively (Fig. 2). In each case we apply a 2-pole Butterworth bandpass filter 4 times sequentially to smooth the waveforms following previous methodologies [Chen and Park, 2013; Servali et al., 2020]. We rotate waveforms into radial and transverse components to isolate the qL and Love waves, respectively. We visually inspect each waveform for a qL phase. Quasi-Love scattering is most readily identified on the vertical component of the seismic traces (Fig. 3), but we require that qL energy is also visible on the radial component of the seismograms, (Fig. S1).

The maximum time delay between the Love and qL waves retained in our dataset is 80s. Earthquake-generated direct Rayleigh waves arrive far later than this 80s window for all the events used in this study, so they do not contaminate possible qL phases. Scattered qL waves travel on the vertical and radial components at Rayleigh wave speeds with retrograde-elliptical particle motion, Fig. S1. We therefore require that any qL detection produce such particle motion. This criterion incorporates the radial component of the seismogram and provides confidence in our observations since it necessitates that the qL is not just present on the vertical component. A potential complication in qL identification is the presence of multiple scatterers along the path. Multiple qL phases that overlap will interfere with each other leading to low correlations between the qL and Love waveforms [e.g. Chen and Park, 2013]. These arrivals will thus fail our quality control criteria, and not be included [see Servali et al., 2020 for greater methodological discussion]. If an event with multiple scatterers passes these criteria, it can only be because the scatterers are sufficiently time-separated to arrive without interference. In this case, only the most recent qL phase (meaning that closest to the station) is included in our data.

When qL scattering is identified, we measure delay times between the qL arrival and the primary Love wave as well as the scattering intensity. Delay time is calculated as the time difference between the Love and qL waves based on time shift between the vertical (the qL phase) and transverse (Love wave) components that results in the maximum correlation coefficient. We use this time delay to back-project along the great-circle raypath (Fig. 1) to locate the anisotropic gradient responsible for the emergence of the qL phase using a surface wave speed differential of 0.4 km/s [following Servali et al., 2020 for the ENAM region]. The choice of wave speed differential directly controls the back-projected location of the qL scatterers. Increasing the differential, for example, shifts scattering locations farther away from the stations. We employ the velocities used by Servali et al. [2020] for the ENAM region as a conservative estimate but recognize that this may skew the scatterers towards the margin.

Scattering intensity is defined as the RMS of the vertical component in a 200s window from the start of the qL arrival divided by the RMS of the transverse component in a 400s window containing the Love wave, multiplied by $\sqrt{2}$ to account for the difference in window lengths (see Chen and Park [2013] and Servali et al. [2020] for more detailed descriptions of the measurement processes). The minimum scattering intensity retained in our dataset is 0.05. Scattering intensity is related to a variety of factors such as the strength of the anisotropic gradient, the length scale of the change in anisotropy, and the direction of the incoming Love wave in relation to the orientation of anisotropy [Chen and Park, 2013]. It is, therefore, difficult to directly ascribe a percent change in anisotropy to any individual scattering intensity measurement, but the presence of qL scattering nonetheless necessitates an anisotropic gradient.

We implement the above procedure for each frequency band used in this study and treat each as an independent dataset. The dispersive nature of Love waves is such that the different frequency bands are sensitive to different depths within the Earth. We constrain the depth range over which each frequency band is sensitive by calculating the coupling between the spheroidal and toroidal modes within those frequencies [Park, 1993; Chen and Park, 2013] (Fig. 2). Although the frequency bands have different depths of maximum sensitivity, there is some overlap. The highest frequency band (50 – 60s) has the shallowest sensitivity, primarily < ~150km depth, with peak sensitivity around ~75km depth. The mid frequency (70 – 80s) is principally sensitive to depths between 75km and 300km with peak sensitivity at ~150km depth. Finally, the low frequency data (100 – 110s) is sensitive to the entire upper mantle (~100km to > 410 km).

The ENAM Community Seismic Experiment (ENAM-CSE) was deployed along the margin between 2014 and 2015 and notably included ocean-bottom seismometers (OBSs) as well as onshore stations [Gaherty, 2014; Lynner et al., 2020]. We attempted to measure qL scattering at the ENAM-CSE OBS stations. We found that the noise on the horizontal components was too strong to accurately measure qL phases, which have low amplitudes. We observe potential qL scattering on the vertical components of the OBS stations, but we were unable to extract sufficiently coherent Love waves needed for scattering intensity calculations or to accurately measure time delays. We were able to make qL observations from the onshore stations of the ENAM-CSE that met our quality control criteria. Data from the onshore ENAM-CSE stations are included our analyses, but none from the OBS stations were of adequate quality for inclusion.

Results

We observe qL scattering in all three frequency bands (Figs. 3, 4, and 5). In total, we made 297 observations in the low frequency, 398 in the mid frequency, and 179 in the high frequency band (Table S1). Scattering intensities vary from 0.05 to 0.25. We interpret all qL scatterers above 0.05 as indicative of an anisotropic gradient; scatterers with intensity below 0.05

do not pass our conservative quality controls. Average scattering intensities are similar in all three frequency bands: high (Average of 0.12 ± 0.04), low (0.11 ± 0.03) and mid (0.12 ± 0.04). All three frequency bands show some degree of variability in scattering intensities between nearby qL scattering points.

In the low and mid frequencies, we observe qL scattering that is regionally coherent with many measurements with comparably strong (>0.08) intensities. This finding is similar to Servali et al. [2020]. We measure relatively few scattering points in the high frequency band along the margin, and scatterers measured in this frequency band have much less consistency in intensity than seen in the other two frequencies. Back-projected scattering locations for the low and mid frequencies are primarily found along the margin, with the majority of scatterers occurring offshore (Figs. 4 and 5). Scattering intensity along the margin is stronger in the mid frequency band than in the low. At the margin, both frequencies show a gradient towards higher scattering intensities moving oceanward. The high frequency band is distinctive for a lack of scatterers south of ~37˚N, and particularly no clear association between scatterers and the margin.

In all three frequency bands, we find a cluster of qL scatterers onshore between ~37°N and ~39°N, which we term region C1 (Fig. 4). This is the only region along the ENAM with consistent qL scattering in the high frequency data. In the mid and low frequency bands, this feature blends into the widespread measurements that span the margin. The presence of this feature demonstrates that the high frequency band is capable of detecting scatterers in the margin region, including the onshore. This adds weight to our interpretation that a paucity of high-frequency scattering observations elsewhere on the margin truly reflects a lack of scatterers, rather than a detection limitation.

Low and mid-frequency scattering in the southern portion of our study area is focused along a linear feature (L1, Fig. 4) that begins along the coast near ~32°N and extends ~500km SE into the oceanic plate. This feature is absent at high frequencies. L1 exhibits the highest intensity scattering in the mid frequency (greater than 0.18); in fact this is the strongest consistent qL scattering seen in this study. Another linear feature (L2) with strong scattering exists farther east that aligns closely with the Kane Fracture Zone (Figs. 1, 4). L2 shows strong qL scattering but is present primarily in the highest frequency band, with only a few qL scatterers in the mid frequency and none in the low frequency data. Outside of C1, this is the only area with coherent high intensity high frequency qL scatterers.

In all three frequency bands, dispersed qL scatterers are seen throughout the oceanic portion of the plate north of ~30°N. These scatterers are mostly far weaker than those seen along the margin and along other linear features. They lack the consistency in scattering intensities and measurement density seen in other areas (Fig. 5). These diffuse qL measurements are difficult to attribute to any potential source or mechanism. They may be related to the Bermuda Rise or the New England seamounts, but the correlation with either feature is poor. We restrict our interpretations to regions with strong, consistent, coherent qL scattering.

Discussion

Three key features stand out in our results: 1) widespread qL scattering along the ENAM in the mid and low frequencies, 2) a linear region (L1) of strong scattering offshore of ~32°N in the mid and low frequencies, and 3) a separate linear feature (L2) far offshore of the margin, strongest in the high frequencies. We discuss each of these features in turn.

Along the length of the ENAM, we observe strong and regionally consistent qL scattering in the low and mid frequencies (Fig. 5). Indeed, qL scattering measurements in the low frequency are very similar to those seen by Servali et al. [2020], where the majority of scattering was concentrated along the COT. Unlike Servali et al. [2020], our bandpass filters at higher frequencies provide depth sensitivity to scattering sources. We recommend this approach in future qL studies. Scattering in the high frequency along the margin is far less pronounced. In detail, scattering intensities are stronger in the mid-frequency than in the low frequency band. These bands differ in sensitivity at depths greater than ~300 km, below which mid-frequency sensitivity rapidly declines (Fig. 2). Meanwhile, the absence of high-frequency scattering rules out strong anisotropic gradients shallower than ~150 km. Taken together, these observations suggest an anisotropic gradient in the shallow asthenospheric mantle (~150-300 km) of ENAM.

The highest concentration of qL scatters (and inferred anisotropic gradient) in the mid and low frequencies along the ENAM aligns with results shear wave splitting studies [Long et al., 2016; Yang et al., 2017; Lynner and Bodmer, 2017], which depict a stark change in anisotropy across the margin. This feature should readily produce qL scattering. Unlike the shear wave splitting results, however, our qL results can constrain the depth range over which the strongest change in anisotropy exists: ~150km to ~300km depth. That depth range is consistent with images from the anisotropic tomography model of Brunsvik et al. [2021], who use SKS and direct S phases in a tomographic framework to add depth-sensitivity to azimuthal anisotropy. Brunsvik et al. [2021] see a substantial anisotropic gradient at the margin between strong (>1.5%) anisotropy offshore between ~200km and ~300km depth and far weaker (~0%) anisotropy beneath the continent. They argue that this transition near the COT is tied to asthenospheric dynamics.

Complex mantle dynamics along the ENAM has been suggested by several studies. King and Anderson [1998] proposed that relief of the lithosphere-asthenosphere boundary (LAB) across the margin drives small scale edge-driven convection (EDC) in the asthenosphere. Subsequent seismic studies offer supporting evidence for EDC. Savage et al. [2017] imaged several low velocity anomalies along the margin, interpreted as distinct EDC cells. Lynner and Bodmer [2017] suggested the ENAM is characterized by 3D edge-driven convection following the geodynamic models of Kaislaniemi and van Hunen [2014]. Both 2D and 3D EDC models involve gradients in mantle flow direction with implied lateral changes in anisotropy. Within downwelling limbs of the convection cells, vertical flow fabrics should dominate. Adjacent to the downwellings, conservation of mass requires either horizontal flow towards the downwellings (if convection cells are large) or vertical upward flow (if convection cells have small horizontal length scale). In the regions surrounding the EDC cells, background mantle flow and associated shear-driven orientations should persist. Reorientation of flow from horizontal shear-driven flow to vertical downwelling would produce different anisotropic orientations over a short (less than ~200km) length scale and create a strong anisotropic gradient sufficient for qL scattering.

Our results cannot rule out lithospheric contributions to anisotropy along the margin. All three frequency bands have some sensitivity to the upper ~100km, which includes the lower lithosphere where anisotropy may be prevalent and aligned with present-day plate motion (due to shear at the LAB) or past plate deformation (such as continental extension). However, quasi-Love wave scattering only arises from sharp lateral gradients in anisotropy. If lithospheric anisotropy were strong but uniform, or only varying slowly, qL scattered phases would not be observed. On this basis, the anisotropic model of Brunsvik et al. [2021] argues against a lithospheric source for qL scatterers. They find the bulk of the lithosphere across the margin shows consistent anisotropic directions and strength, preserving fabrics from continental extension. Although they do find some evidence for lithospheric gradients in anisotropy direction, these are ~400 km offshore, in the least well resolved part of their model, and not overlapping with the majority of qL scatterer sources we observe.

In addition to measurements along the COT, there is a localized cluster of qL scattering that extends into the continental portion of the plate between ~37°N and ~39°N, region C1. This cluster is present in all 3 frequency bands but is most easily seen in the high frequencies since it is the only consistent qL scattering feature along the margin in this frequency band. Scattering in the same region is present in the low and mid frequencies as well but overlaps with the wide-spread COT-related qL measurements, and so is harder to identify as a distinct region. C1's presence across all frequency bands suggests a source that spans both asthenospheric and lithospheric mantle.

The C1 qL feature occurs near a known seismic low velocity anomaly, the Harrisonburg anomaly (also termed the Central Appalachian Anomaly; CAA), that spans the upper mantle [e.g., Porter et al., 2016; Wagner et al., 2018; Brunsvik et al., 2021]. Several mechanisms have been put forth to explain the Harrisonburg anomaly [see review by Long et al., 2021] including lithospheric loss [King and Anderson, 1998; Magni and Kiraly, 2020], Rayleigh-Taylor instability [Mazza et al., 2014], thermal ablation [Evans et al., 2019], or a plume-like upwelling [Chu et al., 2013]. Any of these mechanisms may result in modification to uppermost mantle anisotropy. Tomographic models largely agree that this feature is relatively localized, which might explain the relatively isolated C1 qL scattering. Scattering across all frequency bands implies a vertically extensive source – if this were the Harrisonburg Anomaly, our data might imply it is deeply rooted. The precise location of the tomographically-imaged Harrisonburg Anomaly, however, is farther inland the majority of our nearby qL scattering points. The deviation may occur due to the anomaly's velocity structure altering the relative Love and Rayleigh wave speeds used to backproject qL locations. Or the scatterers may depict the eastern edge of the anomaly, where anisotropic gradients are likely stronger than in the anomaly's center. Alternatively, previous studies have suggested the Harrisonburg anomaly may excite shear driven flow at the base of the lithosphere (Brunsvik et al., 2021; Li and Gao, 2021; Long et al., 2021). Shear driven flow could create a localized convective system that is offset from the anomaly and may produce a strong anisotropic gradient.

We observe a linear stretch of strong qL scattering (L1) in the mid and low frequencies offshore of ~32°N extending out into the Atlantic (Fig. 4). These qL measurements are some of the strongest seen anywhere in our dataset while also being completely absent in the high frequency, again suggesting an asthenospheric source. The western edge of L1 bleeds into the margin-related qL observations. Since L1 is unique as the only feature to extend far out into the oceanic realm, we suggest L1 is likely tied to LAB structure of the oceanic plate and not directly to dynamics along the margin. Scattering associated with L1 may result from an anisotropic change at the edge of a small-scale convective cell. Intriguingly, L1 aligns quite well with the Blake Ridge, a noticeable bathymetric and gravity lineation protruding from the North American continental shelf southeastwards into the Atlantic. However, since our measurements imply an asthenospheric anisotropy gradient and the Blake Ridge represents an uppermost crustal feature, it is hard to conceive of a mechanism that connects them that does not produce an anisotropy gradient in the lithosphere. The most likely scenario connecting the crust with the asthenosphere is topography along the lithosphere-asthenosphere boundary (LAB). LAB structure could lead to changes in crustal structure due to variable buoyancy and excite small scale asthenospheric flow, yet the similar formational history of the lithosphere would produce no deviations in anisotropy.

Our preferred explanation for L1 is the presence of an elongated small-scale convective (SSC) flow cell driven by instabilities at the base of the oceanic lithosphere. SSC would produce a localized anisotropic gradient restricted to the asthenospheric mantle capable of exciting qL scattering. Additionally, SSC may account for the above-average scattering intensities seen along L1 in the mid frequency data. Because SSC cells can be excited by structures at the LAB, deformation—and therefore anisotropy—is focused at the top of the asthenosphere. This is where the mid frequency data has peak sensitivity, leading to stronger scattering. SSC-related features have been proposed to exist at large scale (>1000 km) in at least one global model near our study region [French et al., 2013]. In that tomography model, a slow velocity channel east of the mid-Atlantic ridge extends towards ENAM, roughly coinciding with L1, which is potentially indicative of a small-scale convective cell. The resolution of this model, however, is not sufficient to draw a definitive link with L1. More recent imaging in the Pacific [Eilon et al., 2022] demonstrates that elongated convective cells in the oceanic asthenosphere (150-250 km) can be smaller in wavelength, which is more consistent with the detection frequencies of, and sharp gradients required for, scattering along L1. Higher resolution imaging offshore in the region of the Blake Ridge is needed to better assess this scenario.

An alternative explanation for L1 is that it represents the edge of a 3D edge-convection cell that stems from the continent-ocean transition along the margin. Savage et al. [2017] suggested multiple small-scale EDC cells exist along the ENAM based on low velocity anomalies seen in tomographic models. One of these EDC-related low velocity anomalies exists at the landward extent of the L1 feature. The convection cell may extend offshore along L1 producing anisotropic changes over short lateral distances and yielding qL scattering. Further, similar to the scenario above, the strong scattering intensities seen along L1 may be attributed to the convective cell beneath the LAB. None of the other proposed EDC low velocity anomalies imaged by Savage et al. [2017], however, have associated linear qL features. This mechanism cannot explain why only one of the low velocity features would extend far east of the COT and yield strong qL scattering, while the others do not, or why L1 is unique along the margin.

Finally, linear feature L2 closely coincides with the trace of the Kane Fracture Zone (KFZ) (Figs. 1 and 4). The KFZ is the result of an offset transform along the mid-Atlantic ridge at ~24˚N [Tucholke and Schouten, 1988; Dannowski et al., 2010]. Scattering along L2 is strongest in the high frequency band and is absent in the low frequency band suggesting a shallow (upper ~150 km), lithospheric gradient in anisotropy. This is consistent with a connection to the fracture zone, a definitionally shallow feature. Frequency dependent qL scattering at the KFZ gives us confidence in the data's ability to constrain the depth of anisotropic gradients, especially in the lithosphere.

L2 scattering indicates that fabrics associated with the fracture zone penetrate into the lithospheric mantle. Fracture zones juxtapose regions of a plate that have different ages and shear histories, with one side having bordered a transform fault and accumulated significant shear strain. Our observation may thus provide indirect evidence for highly concentrated shear extending well into the ductile portion of the oceanic lithosphere below the transform. Moreover, scattering along the KFZ falls between -62°E and -68°E corresponding to ~135 Ma and ~84 Ma aged oceanic plate. The KFZ experienced pronounced changes in offset during this time span [Tucholke and Schouten, 1988]. Resultant shearing and lithospheric deformation may have produced, or accentuated, anisotropic gradients that make L2 a standout feature.

Conclusion

We measure and locate frequency dependent quasi-Love (qL) wave scattering along the eastern North American margin. We measure qL phases in three frequency bands that highlight anisotropic gradients in the whole upper mantle, the shallow asthenosphere, and the lithosphere. We observe several features along the margin that exhibit frequency dependence in our qL measurements. Along the ENAM, we see strong, consistent scattering in mostly the mid and low frequencies, linked to anisotropy gradients in the asthenospheric mantle. We interpret these as signatures of edge-driven convection driven by lithospheric relief at the site of continental extension. Offshore, we see strong, apparently asthenospheric, qL scattering in a lineament that may be associated with small-scale convection. Coherent scattering associated with the Kane Fracture Zone, seen only in the high frequency data, indicates transform-fault-related deformation penetrates into the lithospheric mantle, but not deeper. Our results show that frequency-dependent qL scattering is useful for constraining the depth of anisotropy and elucidating mantle dynamics.

Figure Captions

Figure 1. Maps of the stations (left) and events (right) used in this study. Left: Stations (red triangles) used to examine qL scattering across the ENAM. The Kane Fracture Zone, Bermuda Rise, and New England Seamounts are also shown. The continent-ocean transition is highlighted by the dashed gray line. Right: Earthquakes (stars) used to measure qL scattering throughout our study region. Red lines show great-circle paths of the surface waves that provided measurable qL phases that met our quality control criteria.

Figure 2. Depth sensitivity kernels for the coupling of modes for the high frequency (50s to 60s; top), mid frequency (70s to 80s; middle), and low frequency (100s to 110s; bottom) qL datasets calculated following Chen and Park [2013]. Kernels are plotted for P (dotted line) and S (dashed line) wave anisotropy sensitivity. Quasi-Love wave scattering can occur where the sensitivity parameters diverge.

Figure 3. Examples of qL scattering recorded in all frequency bands at 4 stations from the June 27th, 2014 event in the Scotia Arc. The transverse component is shown in blue and the vertical in

green. All examples have strong, clear Love waves. qL scattering (highlighted with arrows) is present in all 4 cases in the low and mid frequencies and is absent in 2 examples from the high. Measured time delays and scattering intensities are shown in the upper right and left corners of the panels, respectively.

Figure 4. qL scattering results throughout our study region in all three frequency bands. Anisotropic scatterers (circles) are located by back-projection along ray paths using the measured time delays. Colors give scattering intensity. The Kane Fracture Zone is shown by the white dashed line. Scattering is focused along the ENAM and linear feature L1 in the mid and low frequencies. In the high frequency data, measurements are concentrated around linear feature L2 and the onshore region C1.

Figure 5. Averaged qL scattering observations in all three frequency bands. Scattering measurements are averaged in 1° bins. The color of each circle represents the average scattering intensity for each cell, and each circle is scaled by the number of qL scatterers located in that bin.

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Author Contributions

C.L.: Conceptualization, Investigation, Writing, Funding Acquisition. Z.E.: Investigation, Writing, Funding Acquisition.

Data Availability

All seismic data used in this study is publicly available and was accessed via the Data Management Center of the EarthScope Consortium.

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