Group velocity tomography of the upper crust in the eastern Tennessee seismic zone from ambient noise data.

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Abstract

The eastern Tennessee seismic zone (ETSZ) is the second most seismically active area in the central and eastern United States after the New Madrid seismic zone, but the relatively weak seismicity and the absence of correlation between the seismicity distribution and the surface geology make its seismogenic potential controversial. In this work we investigate the structure of the upper crust in the ETSZ by means of group velocity tomography maps from seismic noise data. Results show that the seismic activity is associated with a relatively low velocity anomaly mainly located in one or more basement blocks. These blocks, bounded to the NW by the NY–AL lineament and to the SE by the Clingman lineaments, are buried beneath low velocity strata consistent with the presence of a relatively thick sedimentary cover. The imaged low velocity anomaly migrates towards the SE at increasing periods, suggesting a possible SE dipping weak structure where most of the seismic activity takes place.

The correlation between the NY–AL magnetic signature and the position of the seismic velocity anomalies supports the interpretation of the low velocity zone as a major basement fault projected to the surface as the NY–AL magnetic lineaments. The fault juxtaposes Granite–Rhyolite basement to the NW with Grenville southern Appalachian basement to the SE.

Introduction

The ETSZ is an area of increased seismicity approximately 300x50 km located in the Valley and Ridge province of the southern Appalachians (Fig. 1). The two largest events recorded within the zone, both Mw 4.6, occurred on 30 November 1974 near Maryville, Tennessee [Bollinger *et al.*, 1976] and on 29 April 2003 near Fort Payne, Alabama (Dunn and Chapman, 2006). However the ETSZ generated damaging earthquakes in historical time [Bollinger, 1973] thus its seismogenic potential, although widely debated, is of primary importance. Also, the well-known low attenuation of the central and eastern United States crust [Nuttli *et al.*, 1979], that causes even moderate earthquakes to be felt and potentially cause damage at far distance, makes it significant.

The Grenville orogeny is the oldest known episode of continental accretion in the eastern United States (~1 Ga). The opening of Iapetus Ocean (~530 Ma) that followed the break-up of Rodinia was accompanied by several rifting episodes (e.g. Reelfoot rift) within the Grenville terrain but according to palinspastic reconstruction rifting did not affect the ETSZ [Thomas, 2006]. The Paleozoic saw a sequence of several orogenic processes accreting the Grenville terrain, the last of which, the Alleghanian

orogeny (~330 Ma), represents the most recent episode of mountain building in the area.

Cambrian to Lower Ordovician rocks deposited in a passive margin setting and Middle Ordovician to Pennsylvanian rocks were emplaced during the Taconic and Alleghanian orogenic phases on the Appalachian forehand. The rocks compose of the strongly folded sedimentary cover beneath the Valley and Ridge province and beneath the ETSZ in particular. Thickness of the sediments overlaying the basement has been recently estimated by means of magnetic data inversion and ranges from 4 to 12 km, [Brandmayr and Vlahovic, 2016] thus locally thicker than previous results [Laske and Masters, 1997; Mooney and Kaban, 2010].

The Cumberland Plateau, which lies west of the Valley and Ridge, is widely agreed to represent the boundary between the stable North American craton and the continental margin, roughly marking the western Grenville deformation front [Thomas, 1982; Tesauero *et al.*, 2015]. The location of the Grenville front to the east and the westward extension of the Appalachians terrains at depth still remains uncertain but basement rocks beneath the ETSZ are presumed to be of Grenville age [Powell *et al.*, 2014 and references therein; Brandmayr and Vlahovic, 2016].

In the ETSZ, earthquakes occur at depths between 5 and 26 km [Vlahovic *et al.*, 1998] and faulting does not seem to involve Appalachian structures [Thomas, 2006]. Thus, the hypocenters do not correlate with surface geological features but mainly occur in a basement block known as Ocoee block [Johnston *et al.*, 1985]. This features are delimited by two NE–SW trending magnetic lineaments, namely the NY–AL lineament on the NW side and the less prominent Clingman lineament on the SE side. In seismology, spurious signals contaminating the quality of any seismic data have been traditionally referred to as a noise. However, this notion changed over the years

as it became widely accepted that ambient noise can bear important information about the Earth's structure, i.e. the Green's Function (GF). In active tectonic settings ambient noise serves as complementary seismic data; in other instances, where no or weak seismic activity is recorded, as is the case of the ETSZ, ambient noise can provide better resolution for crustal structures than most active source methods.

The cross correlation of continuous waveform signals (ambient noise) arriving at a pair of seismic stations gives an estimate of the GF of the medium between the stations [Aki *et al.*, 1965; Sabra *et al.*, 2005; Campillo and Paul, 2003; Bensen *et al.*, 2007]. Although the resulting signal should represent the full GF, thus should include body and surface waves, however, only the latter usually dominates ambient noise correlation, presumably due to attenuation and scattering. Surface waves emerge prominently provided that data have been collected for a sufficiently long time in order to improve the signal-to-noise ratio. Then, once dispersion curves have been extracted for the medium between each pair of stations, surface wave tomography algorithms can be applied to map the lateral variations of group velocity at relevant periods, i.e. at different depths. Tomography results can be further inverted to determine a three-dimensional shear wave velocity structure of the area [Pontevedo and Panza, 2006; Panza *et al.*, 2007a; Costanzo and Nunziata, 2014].

Liang and Langston [2008] investigated the crust in eastern North America by means of tomography from noise data, imaging ETSZ with a variable resolution between $2^{\circ} \times 2^{\circ}$ and $1^{\circ} \times 1^{\circ}$. Powell *et al.* [2014] used local earthquake tomography to study the same area providing a P and S wave velocity structure down to a depth of 24 km.

Both papers delineate the ETSZ as a transition zone between low velocity to the west and high velocity to the east. With improved resolution, we will show that the results

of this study are in general consistent with their findings and spotlight some features that deserve further investigation.

Data and Method

The data used in this study are obtained from 24 short period stations that are part of the eastern Tennessee (ET) seismic network deployed and managed by the Center for Earthquake Research and Information (CERI) at the University of Memphis, Tennessee. The network covers the ETSZ, an area about 300 km long and 50 km wide (Fig. 1).

Although broad band data are more frequently used for ambient noise tomography, several authors showed that short period data are suitable as well to investigate the upper crustal structure [e.g. *Arroucau et al.*, 2010; *Young et al.*, 2011; *Rawlinson et al.*, 2014]. In fact, the ET network is composed by high gain short period sensors (mainly L4C and S13) with corner frequency usually at ~ 1 Hz. Self noise of 1 Hz seismometers is well below the New Low Noise Model (NLNM) until 0.1 Hz.

This means that with adequate amplification, a short period seismometer should be able to record low amplitude 20 s surface waves [*Havskov and Alguacil*, 2004].

Continuous waveform ambient noise data for 110 days from August to December 2011 were spooled, processed and analyzed in this study. Most of the processing steps adopted in this work are those described in *Bensen et al.* [2007], which has become the standard for processing ambient noise data [e.g. *Idowu et al.*, 2008; *Nunziata et al.*, 2009; *Arroucau et al.*, 2010], with few modifications.

Following *Saygin and Kennett* [2009], we did not apply any normalization prior to the correlation process. This is because both time and frequency domain normalization

are non-linear operations which could introduce artifacts, thus alter the spectral content of the data. The primary purpose of normalization is in fact to remove transients caused by seismic activities, however normalization and spectral whitening may alter the amplitude content of the signal leading to biased group velocity measurements [Lin et al., 2011].

In this study, each simultaneously recorded seismogram for all station pairs was cross-correlated over overlapping 1-hour segments. We use normalized cross-correlation [Cox, 1973], that means the correlation function of a wave propagating with velocity c at frequency ω between stations 1 and 2 separated by distance r is:

$$1. \quad C_{1,2}(\omega) = \frac{\sin(\frac{\omega r}{c})}{\frac{\omega r}{c}}$$

As a consequence the amplitude of the correlation function is in the $[-1,1]$ range, thus the weight of portions of signal including earthquakes is small.

The results for all station pairs have been subsequently stacked: stacking involves averaging the cross-correlated hourly files over the entire recorded hours. This procedure significantly increases the signal to noise ratio (SNR) and suppresses incoherent signals like that generated by earthquakes.

The so-called symmetric component was then calculated by averaging the causal and acausal parts of the cross-correlograms. Some example of average quality, symmetric component tracks are shown in fig. 2 along with the spectrogram of the longest interstation distance pair.

The extracted waveforms exhibit a dispersed wave train (Fig. 2) that can be interpreted as the Rayleigh wave component of the GF and may thus be processed by means of conventional surface wave analysis tools. A method called Multiple Filtering Technique (MFT) was applied to the resulting waveform [Dziewonski et al., 1969]. The waveform was filtered around the periods of interest and envelope files

were generated for each period. The peaks of the envelopes correspond to velocity values that were selected. This process [known as Frequency-Time Analysis or FTAN, e.g. *Levshin et al.*, 1972] is automated for all periods of interest.

As additional control step to ensure data quality, both the noise correlation function (NCF) and its time derivative are used to obtain quasi-independent measurements of the group velocities [see *Arroucau et al.*, 2010]; data for which the NCF and its time derivative differ by over 3% are discarded. There is also an important trade-off between data quantity and quality. For this study, an SNR value of 10 ensures that the data quality is not compromised by quantity, thus data with $\text{SNR} < 10$ are also screened out.

A threshold of 2 wavelengths is adopted as the minimum interstation distance to be accepted. Although the general empirical rule is more conservative (minimum 3 wavelengths), the configuration of the stations network induced us to be less strict.

Anyway, *Luo et al.* [2015] showed that reliable tomography results can be obtained from ambient noise data using interstation distances as short as 1 wavelength.

Finally, the process of extraction of the dispersion curves is accomplished in two stages; in the first, preliminary dispersion curves are constructed and averaged in order to build a phase-matched filter [see *Levshin and Ritzwoller*, 2001] that is subsequently applied to the seismograms prior to a second round of travel time picking. This allows additional dispersion measurements to be included that were discarded during the first stage due to low SNR [*Arroucau et al.*, 2010].

We analyzed periods from 1 to 15 s, but most of the dispersion data extracted at periods ≥ 12 s did not satisfy the SNR requirement.

Interstation dispersion curves have been successfully extracted at relevant periods ranging from 1 to 10 s, i.e. periods sensitive to the structure of the upper crust, as can

be determined from the values of the partial derivatives of Rayleigh wave's group velocities with respect to the structural parameters [Urban *et al.*, 1993].

In this study we adopted the two-dimensional tomography approach developed by Ditmar and Yanovskaya [1987], Yanovskaya *et al.* [1988] and Yanovskaya and Ditmar [1990]. The method is a generalization to two dimensions of the classical one-dimensional method of Backus and Gilbert [1968; 1970]. The lateral resolution of the tomographic maps is defined through two quantitative parameters, namely the average size, L , of the equivalent smoothing area and its elongation ε [Yanovskaya, 1997]. The latter is a measure of the azimuthal distribution of the ray paths, in this case the interstation paths. A value of $\varepsilon > 1$ denotes paths elongated in a longitudinal dimension, thus the equivalent smoothing area approaches an ellipse. Values of $\varepsilon < 1$ denote good azimuthal distribution of the ray paths, thus the equivalent smoothing area approaches a circle and L gives a measure of its diameter. The resolution retrieved by means of L and ε provides a consistent measure for lateral discretization in the subsequent inversion for structural parameters [Brandmayr *et al.*, 2010]. Hence there is no need for checkerboard or similar tests that often optimistically mask the first order effect of the anomaly position on the data set resolution capability [Foulger *et al.*, 2013].

Results

Tomography has been performed twice, on a grid of 0.2° and 0.5° respectively. Main features of the two runs are comparable, nevertheless some fine details imaged on a 0.2° grid are smaller than the optimal resolution (Fig. 3) and can be considered artifacts. In the following we will thus describe and discuss the results obtained on a

0.5° grid, for the selected periods of 1 s, 2 s, 4 s and 7 s, 9 s, 10 s. The reasons of this selection will be further explained in the discussion, but we anticipate that the intermediate periods show similar features in terms of anomaly results and resolution. The average size of the equivalent smoothing area is < 25 km in the optimally covered area and < 50 km in most of the study area, out of which it decreases rapidly. The best resolved area tends to decrease with increasing periods ($T = 9$ s and 10 s, Fig. 3), likely due to decreased number of interstation paths satisfying the SNR threshold and the trade-off between wavelength content and interstation distance (min 2 wavelengths).

The azimuthal distribution is also good in most of the study area ($\varepsilon < 1$, Fig. 3), with the exception of a central portion of the map, roughly corresponding to the Blue Ridge province, where a NE–SW azimuthal distribution of the paths is prevailing.

To further investigate resolution aspects, like amplitude recovery, spike tests have been performed for both point-like and coarse (~ 50 km size) anomalies (Fig. S1).

Spike tests results show little smearing and almost no data inconsistency artifacts in model recovery. Amplitude recovery indeed improves for coarse anomaly, as shown by *Rawlinson and Spakman* [2016].

Considering the resolution parameters, we discuss in the following sections only the results enclosed in the 100 km lateral resolution isoline, bearing in mind that the actual resolution in most of the study area is < 50 km. The group velocity anomaly is computed with respect to the average value at each relevant period (Fig. 4). At selected periods anomaly amplitudes range within 12–13% of the average group velocity. At shortest periods we mainly observe two parallel, roughly NE–SW trending, negative and positive anomalies of increasing extent and amplitude from 1 to 4 s. At periods longer than 5 s, the low velocity anomaly shifts to E–SE and tends

to be surrounded by high velocity anomalies of increasing extent and amplitude. Although ε value close to unity and smoothing parameter could affect the shape of the anomaly, especially at 7 s period, the velocity contrast being perpendicular to the main azimuthal direction of the paths can be considered reliable everywhere.

Surface waves sample the shallow structure at short periods and deeper structure as period increases. The considered periods thus sample the sedimentary rocks ($T < 5$ s) and the crystalline crust down to a depth of about 11–15 km (T from 5 s to 10 s), as can be determined from the values of the partial derivatives of Rayleigh waves group velocities with respect to structural parameters [Urban *et al.*, 1993]. Low velocity anomalies are generally associated with sedimentary basins and outcropping sedimentary rocks, while high velocity anomalies are often associated with mafic intrusions. Nevertheless inference of the mechanical properties of the medium, in terms of P and S waves, from group velocity is not straightforward and can be rigorously achieved only by further inversion of the local dispersion curves.

Discussion

Seismicity in the ETSZ is mainly clustered along the NE–SW trending NY–AL magnetic lineament, a prominent magnetic anomaly which source lies in the bedrock beneath average 7–8 km thick sediments (Fig. S2), in the Valley and Ridge province [Brandmayr and Vlahovic, 2016]. Most of the hypocenters range from 8 to 15 km [Vlahovic *et al.*, 1998], thus they are located in the upper part of the crystalline crust that is sampled by the longest periods presented in this study.

At the shortest periods ($T < 5$ s), in the northern part of the study area the NY–AL magnetic lineament roughly delineates the boundary between a low velocity anomaly

(~2.7 km/s) to the NW and a high velocity anomaly to the SE (3–3.2 km/s; Fig. 5; 1 s, 2 s, 4 s panels). More to the south the lineament instead tends to bisect the low velocity zone in two. We thus link this low velocity to the sedimentary cover of the Valley and Ridge province. Higher velocities (3–3.2 km/s) are found to the east, across the Blue Ridge, where outcrops of crystalline rocks are well documented [Schedl *et al.*, 1997] and sedimentary rock thickness obtained from magnetic data inversion is less than 2 km [Brandmayr and Vlahovic, 2016]. A localized high velocity anomaly (3–3.2 km/s; Fig. 5; 1 s, 2 s, 4 s panels) is found west of the lineament (35.5°N, 85.5°W), in the southern portion of the Appalachian Plateau, where calcareous metasediments are reported and the top of the basement lies at depths from 2 to 4 km.

At periods > 5 s, which are sensitive to the crystalline crust, velocity generally increases but a low velocity zone (2.9–3.0 km/s; Fig. 5; 7 s, 9 s, 10 s panels) is still present. The low velocity anomaly migrates to the SE with increasing periods (i.e. depths), and appears more clearly bounded by high velocity anomalies (3.2–3.4 km/s; Fig. 5; 7 s, 9 s, 10 s panels).

A new high velocity anomaly prominently appears at 9 s and 10 s NW of the NY–AL lineament (36.5°N, 84°W), with group velocity > 3.4 km/s. We relate this anomaly with magnetic and gravity highs and a mafic intrusion associated with a proposed Proterozoic rift [Powell *et al.*, 2014].

To the east, beneath the Blue Ridge, group velocity markedly increases with increasing period (from 3.15 to 3.4 km/s) in the SE part, while lower velocities (2.8–2.9 km/s) are found beneath the NE segment of the Carolinian Blue Ridge, at the border with the Piedmont, the velocity here being remarkably constant at the investigated periods. This result is consistent with what Liang and Langston [2008]

found at comparable periods, although the interpretation of this low velocity anomaly is dubious. We tend to exclude that it represents the sedimentary cover (Fig. S2), which is < 2 km in this area [Brandmayr and Vlahovic, 2016]. This area of complex geologic features likely hosts terrains accreted during the Iapetus subduction and rich in metasedimentary rocks [Graw *et al.*, 2015].

Most of the seismicity is thus correlated to relatively low velocities in the uppermost basement, where the anomaly is roughly bounded between the NY–AL and the Clingman lineaments.

Previous studies of group velocity [Liang and Langston, 2008] and body wave velocity structure [Powell *et al.*, 2014] show that ETSZ is located in a transition zone where velocity is high to the east and low to the west. The latter also suggested that the low velocity zone beneath ETSZ represents a major basement fault that juxtaposes Granite-Rhyolite basement to the NW and Grenville southern Appalachian basement to the SE. The surface expression of this fault, according to the same study, would be the NY-AL magnetic lineament. Although the study carried out by Liang and Langston [2008] was not aimed at a detailed image of the ETSZ ($2^{\circ} \times 2^{\circ}$ resolution), their range of absolute group velocity values is consistent with our results. The details of the P and S-waves velocity structure obtained from local earthquakes data inversion by Powell *et al.*, [2014] cannot be easily compared with our findings, given the fact that qualitative correlation of group velocities with body waves velocity structure can by itself be misleading and inappropriate. Nevertheless the general patterns of velocity distribution at seismogenic depths agree, and both methods identify high velocity anomalies associated with mafic intrusion NW of the NY–AL lineament (36.5°N , 84°W) and in the southern Appalachian Plateau (35.5°N , 85.5°W) that thus appear robust features, which deserve further investigation.

Liang and Langston [2008] imaged a relative low velocity anomaly at the Tennessee-North Carolina border, which they interpret as the effect of large clastic sediment discharge in the Tennessee embayment at Iapetus ocean margin. We image this low anomaly too, even if apparently a bit more to the south. The low velocity anomaly is particularly prominent at 7 s and 10 s periods, but since it is located in the area where we experience low azimuthal coverage and due to its contiguity with the main low anomaly, we do not speculate on source. We must stress again that group velocities are only broadly related to specific lithologies, the assessment of which needs further inversion of local dispersion curves for shear waves velocity structure.

Conclusions

Ambient noise tomography results at periods < 5 s show relatively low group velocity that can be correlated with the thick (7–8 km) sedimentary rock cover of the Valley and Ridge province. The seismic activity in the ETSZ is also associated with a low velocity anomaly and mainly located in one or more basement blocks bounded to the NW by the NY–AL lineament and to the SE by the Clingman lineament. This finding supports the interpretation of the low velocity zone as a major basement fault, trending NE–SW and juxtaposing Granite-Rhyolite basement to the NW with Grenville southern Appalachian basement to the SE, of which the NY–AL magnetic lineament is the projection on the surface.

At basement depths relatively high velocities beneath the Appalachian Plateau and the SE part of the Blue Ridge province bound the ETSZ to the west and to the east, respectively. A prominent, localized high velocity anomaly is found NW of the NY–

AL lineament, possibly indicating a mafic intrusion related to the proposed Proterozoic rift in the area. In the NE part of the study area, relatively low velocities are found at the boundary between the Blue Ridge and the Piedmont provinces, likely representing crustal accreted terrains during Iapetus subduction. Better understanding of the seismic structure needs further inversion of the tomography results to obtain a 3D shear velocity structure of the study area.

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References

- Aki, K. (1965), A Note on the Use of Microseisms in Determining the Shallow Structures of the Earth's Crust, *Geophysics*, 30, 665–666
- Arroucau, P., Rawlinson, N. and M. Sambridge, (2010). New insight into Cainozoic sedimentary basins and Palaeozoic suture zones in southeast Australia from ambient noise surface wave tomography, *Geophys. Res. Lett.*, 37, L07303, doi:10.1029/2009GL041974

Backus, G. and J.F. Gilbert (1968), Resolving power of gross earth data, *Geophys. J. R. Astr. Soc.* 16, 168–205.

Backus, G. and J.F. Gilbert (1970), Uniqueness in the inversion of inaccurate gross Earth data, *Phil.Trans. R. Soc. A.* 266, 123–192.

Bensen, G. D., M. H. Ritzwoller, M. P. Barmin, A. L. Levshin, F. Lin, M. P. Moschetti, N. M. Shapiro, and Y. Yang (2007), Processing seismic ambient noise data to obtain reliable broad-band surface wave dispersion measurements, *Geophys. J. Int.*, 169, 1239-1260, doi: 10.1111/j.1365-246X.2007.03374.x

Brandmayr, E., and G. Vlahovic, (2016), The upper crust of the Eastern Tennessee Seismic Zone: Insights from potential fields inversion. *Tectonophysics*.

Brandmayr, E., Raykova, R., Zuri, M., Romanelli, F., Doglioni, C., and G.F. Panza, (2010), The lithosphere in Italy: structure and seismicity. *The Geology of Italy: tectonics and life along plate margins*.

Campillo, M., and A. Paul (2003), Long-range correlations in the diffuse seismic coda. *Science*, 299(5606), 547-549.

Costanzo, M.R., and C. Nunziata, (2014), Lithospheric VS models in the Campanian Plain (Italy) by integrating Rayleigh wave dispersion data from noise cross-correlation functions and earthquake recordings, *Physics of the Earth and Planetary Interiors*, doi: <http://dx.doi.org/10.1016/j.pepi.2014.05.002>

Cox, H. (1973), Spatial correlation in arbitrary noise fields with application to ambient sea noise. *The Journal of the Acoustical Society of America*, 54(5), 1289-1301

Ditmar, P.G. and T.B. Yanovskaya (1987), A generalization of the BACKUS-GILBERT method for estimation of lateral variations of surface-wave velocity, *Izv. AN SSSR, Fiz. Zemli (Physics of the Solid Earth)* 23, 6, 470–477.

Dunn, M. M., and M. C. Chapman (2006), Fault orientation in the eastern Tennessee seismic zone: A study using the double-difference location algorithm, *Seismol. Res. Lett.*, 77, 494–504.

Dziewonski, A. M., S. Bloch, and M. Landisman (1969), A technique for the analysis of transient seismic signals, *Bull. Seism. Soc. Am.*, 59, 427-444.

Foulger, G. R., Panza, G. F., Artemieva, I. M., Bastow, I. D., Cammarano, F., Evans, J. R., ... and T.B. Yanovskaya (2013), Caveats on tomographic images. *Terra Nova*, 25(4), 259-281.

Graw, J. H., Powell, C.A., and C.A. Langston (2015), Crustal and upper mantle velocity structure in the vicinity of the eastern Tennessee seismic zone based upon radial P wave transfer functions, *J. Geophys. Res. Solid Earth*, 120, 243–258, doi:10.1002/2014JB011516.

Havskov, J., and G. Alguacil (2004), Seismic recorders. In *Instrumentation in Earthquake Seismology* (pp. 113-149). Springer Netherlands.

Idowu, O.A., Frederiksen, A.W., and J.F. Cassidy (2011), Imaging the Nechako Basin, British Columbia, using ambient seismic noise. *Canadian Journal of Earth Sciences*, 48.

Johnston, A.C., Reinbold, D.J., and S.I. Brewer (1985), Seismotectonics of the Southern Appalachians: *Seismological Society of America Bulletin*, v. 75, p. 291–312.

Laske, G., and G. Masters (1997), A global digital maps of sediment thickness. *Eos, Trans. Am. Geophys. Union* 78 F483

Levshin, A. L., and M. H. Ritzwoller (2001), Automated detection, extraction, and measurement of regional surface waves, *Pure Appl. Geophys.*, 158, 1531–1545, doi:10.1007/PL00001233.

Levshin, A.L., Pisarenkov, F., and G.A. Pogrebinsky, (1972), On a frequency time analysis of oscillations. *Ann Geophys.* 28, 211-218.

Liang, C., and C. A. Langston, (2008), Ambient seismic noise tomography and structure of eastern North America. *Journal of Geophysical Research: Solid Earth* 113.B3

Lin, F. C., Ritzwoller, M. H., and W. Shen, (2011), On the reliability of attenuation measurements from ambient noise cross- correlations. *Geophysical Research Letters*, 38(11).

Luo, Y., Yang, Y., Xu, Y., Xu, H., Zhao, K., and K. Wang (2015), On the limitations of interstation distances in ambient noise tomography. *Geophysical Journal International*, 201(2), 652-661.

Mooney, W. D., and M. K. Kaban (2010), The North American upper mantle: Density, composition, and evolution, *J. Geophys. Res.*, 115, B12424, doi:10.1029/2010JB000866.

Nunziata C., De Nisco G., and G.F. Panza G.F. (2009), S-waves profiles from noise cross correlation at small scale. *Eng. Geol.* 105, 161-170.

Nuttli, O. W., Bollinger, G. A., and D.W. Griffiths (1979), On the relation between modified Mercalli intensity and body-wave magnitude. *Bulletin of the Seismological Society of America*, 69(3), 893-909.

Panza, G.F., Peccerillo, A., Aoudia, A., and B.M. Farina (2007a), Geophysical and petrological modelling of the structure and composition of the crust and upper mantle in complex geodynamic settings: The Tyrrhenian Sea and surroundings. *Earth-Science Reviews* 80, 1-46.

Pontevivo, A., and G.F. Panza (2006), The lithosphere-asthenosphere system in the Calabrian Arc and surrounding seas- Southern Italy. *Pure Appl. Geophys.* 163, 1617-1659.

Powell, C.A., Withers, M., Cox, R.T., Vlahovic, G., and P. Arroucau (2014), Crustal velocity structure associated with the eastern Tennessee seismic zone: Vp and Vs images based upon local earthquake tomography: *Journal of Geophysical Research*, v. 119, p. 464–489, doi:10.1002/2013JB010433.

Rawlinson, N., Salmon, M., and B. L. N. Kennett (2014), Transportable seismic array tomography in southeast Australia: Illuminating the transition from Proterozoic to Phanerozoic lithosphere. *Lithos*, 189, 65-76.

Rawlinson, N., and W. Spakman (2016), On the use of sensitivity tests in seismic tomography. *Geophysical Journal International*, 205(2), 1221-1243.

Sabra, K. G., P. Gerstoft, P. Roux, W. A. Kuperman, and M. C. Fehler (2005), Extracting time-domain Greens function estimates from ambient seismic noise, *Geophys. Res. Lett.*, 32, L03310, doi: 10.1029/2004GL021862

Saygin, E., and B. Kennett (2009), Ambient seismic noise tomography of Australian continent, *Tectonophysics*, 481, 116–125, doi:10.1016/j.tecto.2008.11.013.

Schedl, A., Fullagar, P. D., and J.W. Valley, (1997), Implications of geochemical and gravity data for the nature of strata beneath the Blue Ridge-Piedmont thrust sheet, USA. *Earth and planetary science letters*, 146(1), 165-179.

Tesauro, M., M. K. Kaban, and W. D. Mooney (2015), Variations of the lithospheric strength and elastic thickness in North America, *Geochem. Geophys. Geosyst.*, 16, 2197–2220, doi:10.1002/2015GC005937.

Thomas, M. D. (1982), Gravity studies of the Grenville province: Significance for Precambrian plate collision and the origin of anorthosite, in 1982 SEG Annual Meeting, pp. 109–123, Soc. of Explor. Geophys, Dallas, Tex.

Thomas, W. A. (2006), Tectonic inheritance at a continental margin, *GSA Today*, 16(2), 4–11, doi:10.1130/1052-5173(2006)016<4:TIAACM>2.0.CO;2.

Urban, L., Cichowicz, A., and F. Vaccari (1993), Computation of analytical partial derivatives of phase and group velocities for Rayleigh waves with respect to structural parameters. *Studia geophysica et geodaetica*, 37(1), 14-36.

Vlahovic, G., C. A. Powell, M. C. Chapman, and M. S. Sibol (1998), ET1D: New one dimensional P- and S-wave velocity models for the eastern Tennessee seismic zone, *Seismol. Res. Lett.*, 69(5), 441–449.

Yanovskaya, T.B. and P.G. Ditmar (1990), Smoothness criteria in surface-wave tomography, *Geophys. J. Int.* 102, 63–72.

Yanovskaya, T.B. (1997), Resolution estimation in the problems of seismic ray tomography. *Izv. Phys. Solid Earth* 33, 762-765.

Yanovskaya, T.B., Maaz, R., Ditmar, P.G. and H. Neunhofer (1988), A method for joint interpretation of the phase and group surface-wave velocities to estimate lateral variations of the Earth's structure, *Phys. Earth Planet. Inter.* 51, 59–67.

Young, M. K., Rawlinson, N., Arroucau, P., Reading, A. M., and H Tkalčić, H (2011), High- frequency ambient noise tomography of southeast Australia: New constraints on Tasmania's tectonic past. *Geophysical Research Letters*, 38(13).

Wessel, P., Smith, W.H., Scharroo, R., Luis, J.M., 2011. The Generic Mapping Tools (GMT) Version 5. AGU Fall Meeting Abstracts. vol. 1, p. 05.

Figure captions

Figure 1. The tectonic setting of the ETSZ: colors indicate geological provinces and black lines the NY–AL and Clingman (CL) magnetic lineaments. Triangles denote ET network seismic stations, with ASTN and SWET stations marked in red and green, respectively. Hypocenters are from the CERI catalogue (1974–2014, gray dots, <http://www.memphis.edu/ceri/seismic/catalog.php> last accessed 11/28/2015). For color legend please refer to the digital version of the paper.

Figure 2. a) Average quality stacks of positive correlation functions lags between

ASTN and other stations. b) Positive correlation function lag (i.e. symmetric component) and c) spectrogram for the largest station pair offset ASTN-SWET.

Figure 3. Map of the lateral resolution (L , contour, km) and elongation parameter (ϵ , color scale) of the tomography results at selected periods. Blue to cyan areas show the best azimuthal coverage, i.e. where the ray paths are not elongated and the lateral resolution can be regarded as the diameter of the averaging circle. Red lines denote interstation paths. For color legend please refer to the digital version of the paper.

Figure 4. Group velocity anomaly maps at selected periods. Anomaly is given as percentage relative to average value at each period. Shaded area is not resolved.

Figure 5. Absolute group velocity maps at selected periods. Black lines denote NY–AL (left) and Clingman (right) magnetic lineaments. Red dots denote epicenter locations from CERI catalogue (1974–2014): events < 8 km of depth are shown on 1 s, 2 s and 4 s maps and events > 8 km are shown on 7 s, 9 s and 10 s maps.

figure 3

Figure 4

figure 5