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Structure of the north-western North Anatolian Fault Zone imaged via teleseismic scattering tomography

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SUMMARY

Information on fault zone structure is essential for our understanding of earthquake mechanics, continental deformation and seismic hazard. We use the scattered seismic wavefield to study the subsurface structure of the North-Anatolian Fault Zone (NAFZ) in the region of the 1999 İzmit and Düzce ruptures using data from an 18-month dense deployment of seismometers with a nominal station spacing of 7 km. Using the forward- and back-scattered energy that follows the direct *P*-wave arrival from teleseismic earthquakes, we apply a scattered wave inversion approach and are able to resolve changes in lithospheric structure on a scale of 10 km or less in an area of about 130 km by 100 km across the NAFZ. We find several crustal interfaces that are laterally incoherent beneath the surface strands of the NAFZ and evidence for contrasting crustal structures either side of the NAFZ, consistent with the presence of juxtaposed crustal blocks and ancient suture zones. Although the two strands of the NAFZ in the study region strike roughly east-west, we detect strong variations in structure both north-south, across boundaries of the major blocks, and east-west, parallel to the strike of the NAFZ. The surface expression of the two strands of the NAFZ is coincident with changes on main interfaces and interface terminations throughout the crust and into the upper mantle in the tomographic sections. We show that a dense passive network of seismometers is able to capture information

from the scattered seismic wavefield and, using a tomographic approach, to resolve the fine scale structure of crust and lithospheric mantle even in geologically complex regions. Our results show that major shear zones exist beneath the NAFZ throughout the crust and into the lithospheric mantle, suggesting a strong coupling of strain at these depths.

Key words: Seismology, Teleseismic Scattering, Tomography, North Anatolian Fault Zone.

1 INTRODUCTION

The North Anatolian Fault Zone (NAFZ) is one of the longest continuous continental strike slip fault systems on Earth, posing considerable hazard to Northern Anatolia and especially the megacity of Istanbul towards its western end. Here, we use a novel imaging technique using the scattered seismic wavefield in a tomographic approach to resolve the lithospheric structure within and below the NAFZ with the aim to better resolve the crustal structure around the fault zone and find evidence for shear zones at depth.

The NAFZ is a 1500 km long, right-lateral strike slip fault running the length of northern Anatolia, separating the Eurasian plate to the north from the deforming Anatolian region in the south, accommodating the relative motion and deformation between these tectonic domains. Together with the East Anatolian Fault, the NAFZ accommodates the westward motion of Anatolia (McKenzie, 1972; Reilinger et al., 2006; Şengör et al., 2005; Barka, 1992) driven by the gradient of gravitational potential energy from the Anatolian plateau to the Hellenic Trench (England et al., 2016).

The NAFZ ruptured in a series of M$\geqq$ 6.7 earthquakes during the 20th century from east to west (Stein et al., 1997), interpreted as stress transfer along the strike of the NAFZ from one earthquake bringing the next segment closer to failure. The two most recent events in the current series occurred in 1999 with epicentres in İzmit (M= 7.6) and Düzce (M=7.2) (Barka et al., 2002; Gülen, 2002), with the fault rupture extending into the Sea of Marmara and the next anticipated event in the series posing a pronounced risk to the city of Istanbul.

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While the deformation at the surface is localized on faults (Hussain et al., 2016; Bürgmann & Dresen, 2008), the distribution of deformation throughout the crust and into the mantle remains unclear (Bürgmann & Dresen, 2008; Vauchez et al., 2012; Moore & Parsons, 2015). Understanding the structure and dynamics of fault zones, especially at depth, is essential for our understanding of continental deformation and seismic hazard.

Here, we aim to better understand the structure of the NAFZ, especially in the middle and lower crust and into the upper mantle, using data from temporary seismic stations and exploiting the scattered seismic wavefield following the *P*-wave arrivals of teleseismic events (Frederiksen & Revenaugh, 2004). We use data from the 18-month DANA deployment (DANA, 2012) across the NAFZ in the region of the 1999 ruptures (Fig. 1a). The *P*-wave coda contains energy from *P*-to-*P* and *P*-to-*S* scattering at small-scale heterogeneities along the ray-paths. Structure can be recovered from the scattered seismic energy through migration approaches ranging from common-conversion-point or common-scattering-point stacking (e.g. Dueker & Sheehan, 1997) to full depth migration (e.g. Ryberg & Weber, 2000). Here we are using a tomographic waveform approach based on linear inverse theory of the scattered wavefield (Ji & Nataf, 1998; Frederiksen & Revenaugh, 2004) to resolve the structure of the lithosphere and potential shear zones. The scattering tomography builds on different principles than the more established *P*-receiver function method previously applied to this dataset (Kahraman et al., 2015). With different resolution, limitations and trade-offs this study will contribute to our knowledge of the fine-scale structure of the lithosphere beneath this major fault zone.

We find that the two strands of the NAFZ evident in the shallow structure coincide with main interfaces and interface terminations throughout the crust and into the upper mantle indicating that the fault zone structure may extend to depths of at least $\sim$75 km in this region. We find evidence for small-scale variation of structure in the vicinity of the strands that might indicate the detection of heterogeneity related to past deformation along the present day fault.

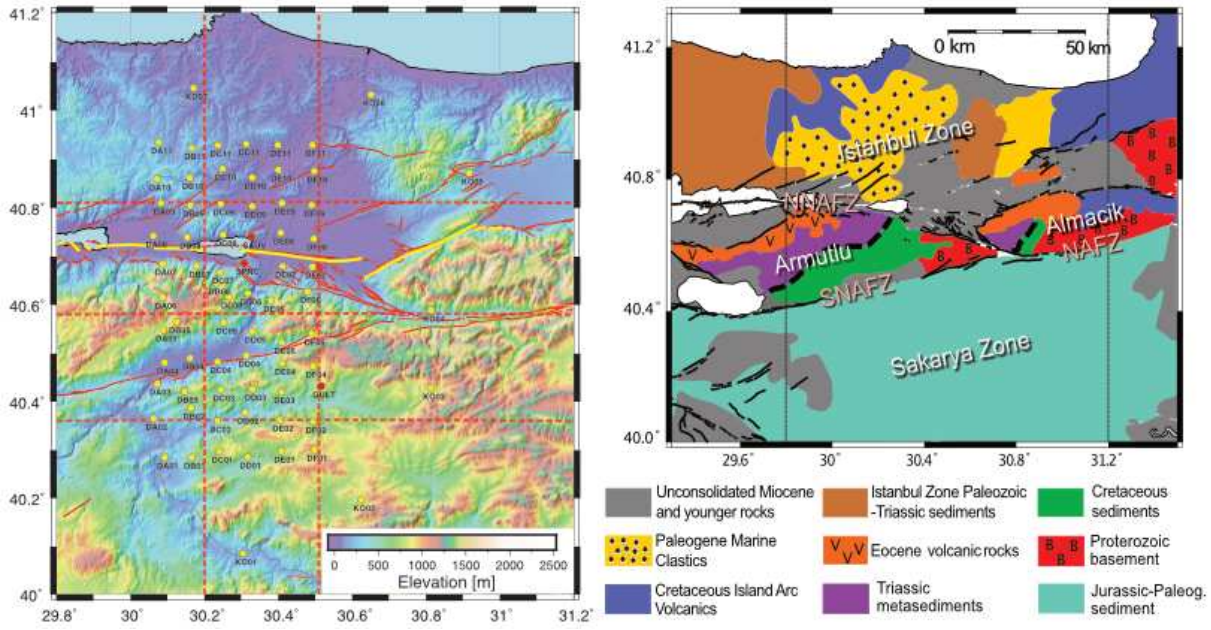

Figure 1. Study region. (a) Topographic map of the study region using SRTM data (Farr et al., 2007). Stations are indicated by yellow circles (permanent stations in red). Mapped faults (red lines) provided by Emre et al. (2018) and rupture of the 1999 İzmit and Düzce earthquakes (yellow) provided by Gülen (2002). Dashed north-south and east-west lines indicate location of depth profiles shown in Fig. 6 and 7 and are approximate locations of depth profiles provided by Kahraman et al. (2015). (b) Simplified geological map of the region outlining the three main tectonic blocks and geological areas. After Taylor et al. (2019a).

2 TECTONICS AND PREVIOUS GEOPHYSICAL SURVEYS

The study region (Fig. 1b) is an amalgam of continental and subduction-related oceanic fragments that remain after the closing of the Tethyan Ocean in the late Tertiary (e.g. Okay et al., 2008; Robertson & Ustaömer, 2004). The DANA network samples three tectonic blocks situated in the study region (i) The Istanbul-Zonguldak Zone (IZ) to the north of the northern strand of the NAFZ, (ii) the Sakarya zone (SZ) to the south of the southern strand; and (iii) the Armutlu and Almacık blocks (AA). The NAFZ splays into a northern (NNAFZ) and southern (SNAFZ) strand west of about 30.65°, with the NNAFZ and the SNAFZ following the northern and southern boundary of the AA. The AA is interpreted as the cause for the splay (e.g. Akbayram et al., 2016). The NAFZ is co-located with the intra-Pontide suture for at least half its onshore length and the boundaries between the SZ, AA, and IZ likely represent the suture in this locale (e.g. Okay et al., 2008). The NAFZ is believed to have developed ∼11 Ma ago in eastern Anatolia with strain localisation

propagating westward and reaching the Sea of Marmara, and our study region, before 3.9 Ma (Akbayram et al., 2016). Slip on the northern and southern NAFZ strands has been estimated to be approximately 16-25 mm/yr and 5-19 mm/yr, respectively (Stein et al., 1997; Flerit et al., 2003; Meade et al., 2002). The northern branch of the NAFZ in our study area last ruptured in the 1999 İzmit earthquake (Tibi et al., 2001; Barka et al., 2002) and still shows active slip at the surface (Hussain et al., 2016) although current seismicity is not focussing beneath either fault strand (Altuncu Poyraz et al., 2015).

The crustal structure in our study region has been characterised using a wide variety of geophysical techniques including seismic refraction (Karahan et al., 2001; Horasan et al., 2002; Bekler & Gürbüz, 2008), ambient noise (Taylor et al., 2016, 2019a), receiver functions (Vanacore et al., 2013; Kahraman et al., 2015; Jenkins et al., 2020), local earthquake tomography (Bariş et al., 2005; Koulakov et al., 2010; Beyhan & Alkan, 2015; Polat et al., 2016; Yolsal-Çevikbilen et al., 2012), magnetotellurics (Tank et al., 2005), and regional full waveform tomography (Fichtner et al., 2013; Çubuk-Sabuncu et al., 2017). Similarly, the upper mantle has been studied using receiver functions (Kind et al., 2015; Taylor et al., 2019b), seismic anisotropy (Biryol et al., 2010; Paul et al., 2014; Lemnifi et al., 2017; Legendre et al., 2021) and body and surface wave tomography (Berk Biryol et al., 2011; Bakırcı et al., 2012; Salaün et al., 2012; Fichtner et al., 2013; Govers & Fichtner, 2016; Papaleo et al., 2017, 2018).

The study region has been sampled by two seismic refraction experiments crossing all major tectonic blocks, finding crustal thicknesses of 32 ± 2 km at about 30.10° longitude in the east (Bekler & Gürbüz, 2008) and $\approx$38 km at about 29.25° longitude in the west of our study region (Karahan et al., 2001). The refraction data show evidence for a 5 km thick upper crustal layer with *P*-wave velocities (V_P) of 5.6 to 6.1 km/s and lower crustal velocities of V_P = 6.7 - 7.2 km/s (Bekler & Gürbüz, 2008). The upper mantle between 29.5°and 30.0° E is relatively slow (V_P = 7.6 km/s) (Bekler & Gürbüz, 2008), while Karahan et al. (2001) find higher mantle velocities further east (29.0°to 29.5° E) with velocities of $V_P \approx$ 8.1 km/s. There is evidence for seismic discontinuities at crustal depths of $\approx$17 km and $\approx$24 km (Karahan et al., 2001).

A study using local earthquake waveforms (Horasan et al., 2002) finds a Moho depth of 32 km in the Marmara region. They find crustal discontinuities at 4 km and 17 km depth with upper mantle velocities of 8.0 km/s and 4.6 km/s for V_P and V_S, respectively and mantle densities of 3.4 g/cm^3. Upper crustal velocities are found to be 5.8 km/s and 3.4 km/s, lower crustal velocities of 6.2 km/s and 3.6 km/s and near-surface velocities of 3.5 km/s and 2.2 km/s for V_P and V_S, respectively (Horasan et al., 2002).

Using fault zone head waves and fault zone reflected waves from seismicity and stations close to the fault zone it is possible to resolve the interface properties of the fault zone. Ben-Zion & Sammis (2003) and Ben-Zion et al. (2003) imaged a shallow fault zone extending about 3 – 4 km into the crust with velocity reductions of up to 50% but a width of only 100 m, well below the resolution of most other seismic methods. In the area of the 1999 Düzce and İzmit earthquakes Bulut et al. (2012) and Najdahmadi et al. (2016) image the material properties across the fault zone using trapped waves and find a bimaterial interface down to the base of the seismogenic crust with an average velocity contrast of 3.4% (Najdahmadi et al., 2016) and 6% with the southern block being fast (Bulut et al., 2012).

Studies of local earthquakes detect upper crustal anisotropy around the NAFZ, limiting it to the upper 8 km (Hurd & Bohnhoff, 2012) or 4 km (Peng & Ben-Zion, 2004), likely due to aligned cracks in the uppermost crust. Detected splitting times are on the order of 10 ms not likely influencing the results of this study. SKS analysis detects asthenospheric anisotropy (Biryol et al., 2010; Paul et al., 2014; Legendre et al., 2021) with lag times between fast and slow direction of typically 1.5$\pm$0.4 s with fast polarization directions smoothly varying from NNE–SSW in northern Turkey to NE–SW in eastern Turkey. A combined study of shear-wave splitting and anisotropic receiver functions show complex anisotropy especially in our study region (Lemnifi et al., 2017). Shear-wave splitting tomography using aftershocks of the 1999 İzmit and Düzce earthquakes show a 3 km wide anisotropic zone extending to 5 km depth with distinct asymmetry being related to damage from the unilateral eastward propagation of the 1999 İzmit rupture (Li et al., 2014).

P-wave receiver functions (PRFs) east of the Sea of Marmara indicate a deepening of the Moho from west (29 to 32 km) to east (34 to 35 km) (Zor et al., 2003; Vanacore et al., 2013). The

average crustal V_P/V_S in our study region is ∼1.75 (Vanacore et al., 2013). PRFs of the DANA dataset (Kahraman et al., 2015) find crustal thickness and V_P/V_S variation in both EW and NS directions with the crust deepening from 36.5 km (V_P/V_S = 1.73) to 40 km (V_P/V_S = 1.73) in the IZ, a constant crustal thickness of ∼ 37 km (V_P/V_S = 1.69 to 1.70) in the AA, and a slight thinning from ∼ 35 km (V_P/V_S = 1.73) in the west to ∼ 34 km (V_P/V_S = 1.85) in the east of the SZ (Fig. 1). Combining data from several permanent stations and temporary station deployments, including DANA data, Jenkins et al. (2020) determined Moho depths across the Sea of Marmara region finding thick crust of up to 41 km in the IZ, with a shallower Moho (32 – 34 km) in the AA and SZ with evidence of discontinuous structure across the NAFZ. The transition also shows complex Moho structure around the NNAFZ. Additionally, Jenkins et al. (2020) find east-west variation with a general deepening of the Moho towards the east.

Previous studies evidence strong crustal heterogeneity on scales of less than 10 km with sharp truncations of sub-horizontal interfaces coinciding with the surface locations of the northern and southern NAFZ strands. The northern strand seems to penetrate deeper into the crust and may extend into the upper mantle based on analysis of receiver functions and ambient noise cross-correlations (Kahraman et al., 2015; Taylor et al., 2016; Jenkins et al., 2020). Both structural changes in North-South and East-West direction have been reported (e.g. Kahraman et al., 2015; Çubuk-Sabuncu et al., 2017; Jenkins et al., 2020).

Using *P*-wave transfer functions and a grid-search inversion approach Frederiksen et al. (2015) detected a sharp change of crustal thickness across the northern NAFZ which is believed to follow the trace of the Intra-Pontide suture in this location and a change of the V_P/V_S ratio across the southern branch indicating a change in basement composition. The IZ shows thick crust (40 to 45 km) but low topography indicating that it is in isostatic disequilibrium or underlain by thicker lithosphere, a result supported by Jenkins et al. (2020). The transfer functions also provide evidence for thick sediments in the Sakarya and Pamukova basins in agreement with ambient noise analysis (Taylor et al., 2019a).

Tomographic studies using traveltimes from local, regional and teleseismic events (e.g. Bariş et al., 2005; Salah et al., 2007; Koulakov et al., 2010; Bakırcı et al., 2012; Beyhan & Alkan,

2015; Polat et al., 2016; Papaleo et al., 2017, 2018) and full waveform information (e.g. Fichtner et al., 2013; Govers & Fichtner, 2016; Çubuk-Sabuncu et al., 2017; Blom et al., 2020) find strong velocity contrasts in the crust along the NNAFZ and the SNAFZ. These are interpreted as the fault zone exploiting the sutures between the Istanbul zone, Armutlu block and Sakarya Zone as result of the closure of the Neo-Tethys ocean. Nonetheless, Fichtner et al. (2013) note the lack of a low velocity fault zone signature at depth west of $\sim$32° E due to the absence of a well-localised suture and insufficient strain localization due to the young age of the fault zone. In contrast, Koulakov et al. (2010) using local tomography note the juxtaposition of high-velocity, low attenuation blocks (e.g. Armutlu block) to lower velocity areas. A smaller scale, full waveform tomographic study has been performed by Çubuk-Sabuncu et al. (2017) noticing strong lateral and vertical velocity variations and strong radial anisotropy of the crust in agreement with the active tectonics of western Turkey. Salah et al. (2007) focus on the rupture area of the 1999 İzmit and Düzce earthquakes using local tomography to resolve *P*- and *S*-wave velocity as well as the Poisson ratio and detect prominent low velocity zones down to depths of 25 km as well as well as a high-velocity anomaly at a depth of 8 km between 30.0 and 30.4°E along the strike of the fault zone. Seismicity is more prevalent in the high-velocity region although it also occurs in low velocity regions. Similarly, traveltime tomography resolves narrow sub-vertical low velocity zones coinciding with the surface expression of the SNAFZ and the NNAFZ (Papaleo et al., 2017, 2018).

Magnetotelluric (MT) data show differences in the crustal conductivity from south to north across the NAFZ (Tank et al., 2005) with a high resistivity ($\geq$ 1000 Ωm) crustal basement in the IZ to the north and a less resistive crustal basement (500 Ωm) in the SZ. The MT data resolve a localized conductive zone (30 to 50 Ωm) within the AA that extends into the upper mantle that has been attributed to partial melts or pore fluid flow from the upper mantle beneath the NAFZ.

3 METHOD

We apply the teleseismic scattering tomography approach by Frederiksen & Revenaugh (2004) to the DANA dataset to resolve the small-scale structure beneath the array. The scattered seismic wavefield is more sensitive to short-wavelength variations in material properties than is the

path-integrated sensitivity of transmitted phases such as used in e.g. seismic traveltime tomography. The *P*-to-*p* and *P*-to-*s* scattered energy in the coda of teleseismic *P*-waves travelling along different paths to the main arrival can uniquely determine Earth structure if the sampling of the seismic wavefield is dense enough to avoid spatial aliasing. In the tomographic approach some aliasing can be accepted without introducing issues with non-uniqueness of the solution due to the regularization of the problem. Several approaches to use the scattered coda energy to image the subsurface have been developed, forming a continuous spectrum of method complexity. The common approach of receiver function analysis uses stacked records of *P*-to-*s* (or *S*-to-*p*) scattered (converted) energy (Vinnik, 1977; Langston, 1979) which may be binned according to their common conversion point to improve signal-to-noise ratio (Dueker & Sheehan, 1997) and mapped to depth. The method assumes a 1D stratified seismic structure which is often violated in practice (Rondenay, 2009). Lateral variation of structure leads to diffraction of the seismic wavefield and diffraction stacking, a backprojection of the diffracted energy along its traveltime hyperbola, can be used to image small-scale perturbations on the order of the seismic wavelength of the structure at depth. These methods are widely used in controlled-source applications (Yilmaz, 2001), and are commonly described as migration techniques (Rondenay, 2009) but implementation requires dense spatial sampling of the seismic wavefield. General improvements and densification of recent passive seismic deployments make the application of more complex methods, such as traveltime stacking of the scattered wavefield (Revenaugh, 1995) or the application of inversion or backprojection operators in a 2D or 3D model space (Bostock & Rondenay, 1999) possible and allow higher resolution of detail. For a full review of these methods see Rondenay (2009).

For a more complete treatment of the scattering problem, the scattering image problem can be formulated as a tomographic inversion (Ji & Nataf, 1998). Using a waveform inversion, Frederiksen & Revenaugh (2004) have developed a linear tomographic inversion of the scattered seismic wavefield which we apply here. A full description is given in Frederiksen & Revenaugh (2004) and we outline only the main points of this approach here.

The Born approximation, a common approximation of the full scattering process, assumes a

weakly scattering medium and single scattering (Sato et al., 2012), which is a good approximation for this application where the amplitudes of the scattered wavefield are much smaller than those of the direct wave. In this approximation the scattering properties are represented as perturbations in elastic parameters ($\delta\lambda$, $\delta\mu$, $\delta\rho$) to a background model (λ_0, μ_0, ρ_0). The seismic equation of motion for the displacement u in an isotropic medium is given by:

$$\rho\ddot{u}_i = (\lambda\nabla\cdot\mathbf{u})_{,i} + [\mu(u_{i,j} + u_{j,i})]_{,j} \tag{1}$$

with λ and μ being the Lamé parameters and ρ the density. Eq 1 can be expanded, using small perturbations to the elastic properties ($\delta\lambda$, $\delta\mu$, $\delta\rho$) around a background medium with elastic properties λ_0, μ_o, ρ_0 (Frederiksen & Revenaugh, 2004), to:

$$\begin{aligned}\rho_0\ddot{u}_i - (\lambda_0+\mu_0)(\nabla\cdot\mathbf{u})_{,i} = \\ &- \delta\rho\ddot{u}_i + (\delta\lambda+\delta\mu)(\nabla\cdot\mathbf{u})_{,i} + \delta\mu\nabla^2 u_i \\ &+ (\delta\lambda)_{,i}(\nabla\cdot\mathbf{u}) + (\delta\mu)_{,j}(u_{i,j}+u_{j,i})\end{aligned} \tag{2}$$

The wavefield can then be divided into a primary (background) and scattered component ($\mathbf{u} = \mathbf{u^0} + \delta\mathbf{u}$) with the unperturbed wavefield satisfying the unperturbed wave equation

$$\rho_0\ddot{u}_i^0 = (\lambda_0+\mu_0)(\nabla\cdot\mathbf{u}^0)_{,i} + \mu_0 u^0_{i,jj} \tag{3}$$

Assuming that the scattered wavefield is much weaker than the unperturbed wavefield this gives the first-order Born approximation by discarding higher-order terms:

$$\rho_0\delta\ddot{u}_i - (\lambda_0+\mu_0)(\nabla\cdot\delta\mathbf{u})_{,i} - \mu_0\nabla^2\delta u_i = Q_i \tag{4}$$

with Q_i being a term of the unperturbed wavefield and the perturbed model parameters which is given by equation 13.22 in Aki & Richards (2002):

$$Q_i = -\delta\rho\ddot{u}_i^0 + (\delta\lambda+\delta\mu)\left(\nabla\cdot\mathbf{u}^0\right)_{,i} + \delta\mu\nabla^2 u_i^0 + (\delta\lambda)_{,i}\left(\nabla\cdot\mathbf{u}^0\right) + (\delta\mu)_{,j}\left[u^0_{i,j}+u^0_{j,i}\right] \tag{5}$$

with $\mathbf{u}^0$ being a solution for the unperturbed medium.

Assuming Rayleigh scattering, where the wavelength of the incident wavefield is much larger than the scale of the heterogeneity, the scattering problem reduces to a point scatterer and the full scattered wavefield is approximated by that of an array of point scatterers. Following Wu & Aki (1985), it is possible to derive expressions for the equivalent point source in Rayleigh scattering. These expressions also contain the directivity of the radiation of the scattered wavefield, and are provided as equations (7) to (10) in Frederiksen & Revenaugh (2004). This gives us the ability to compute both the amplitude and radiation pattern of scattering from small-scale heterogeneities in solving the forward problem of the waveform inversion.

We assume the incident *P*-wave to be planar (Fig. 2) with a known slowness vector, a condition well met for teleseismic records. The scattered wavefield is derived from the seismic observations by computing the 3-component receiver functions relative to the first arriving *P* wave. The considered input seismic wavefield includes the direct incident *P* and the free-surface reflections (*Pp* and *Ps*), producing forward- and back-scattering in the volume, respectively. The forward and backscattering of the input wavefield produces six possible scattered phases (where • indicates the scattering event along the raypath) at small-scale elastic heterogeneities: *P•p*, *P•s*, *Pp•p*, *Pp•s*, *Ps•p* and *Ps•s*. In the forward modelling, we consider every possible combination of perturbed parameter (*P*- and S-wave velocity perturbation ($\delta\alpha$, $\delta\beta$) and density perturbation ($\delta\rho$)), incident wave (forward scattering *P* and backscattering free surface reflection *Pp* and *Ps*) and station location. The inclusion of the free surface backscattered energy as well as the forward scattered direct wave increases the resolution of the study volume and allows us to resolve a 3D perturbation model, here represented as a regular grid of perturbed cells. We use ray tracing in a 1D velocity model to determine traveltimes to and from the scattering heterogeneity and to calculate incidence and refraction angles. We use equations (7) to (10) of Frederiksen & Revenaugh (2004) including a geometrical spreading factor for a layered medium to determine the amplitudes of the scattered energy in an elastic velocity model.

The Born approximation prescribes that single scattered waves propagate in the unperturbed

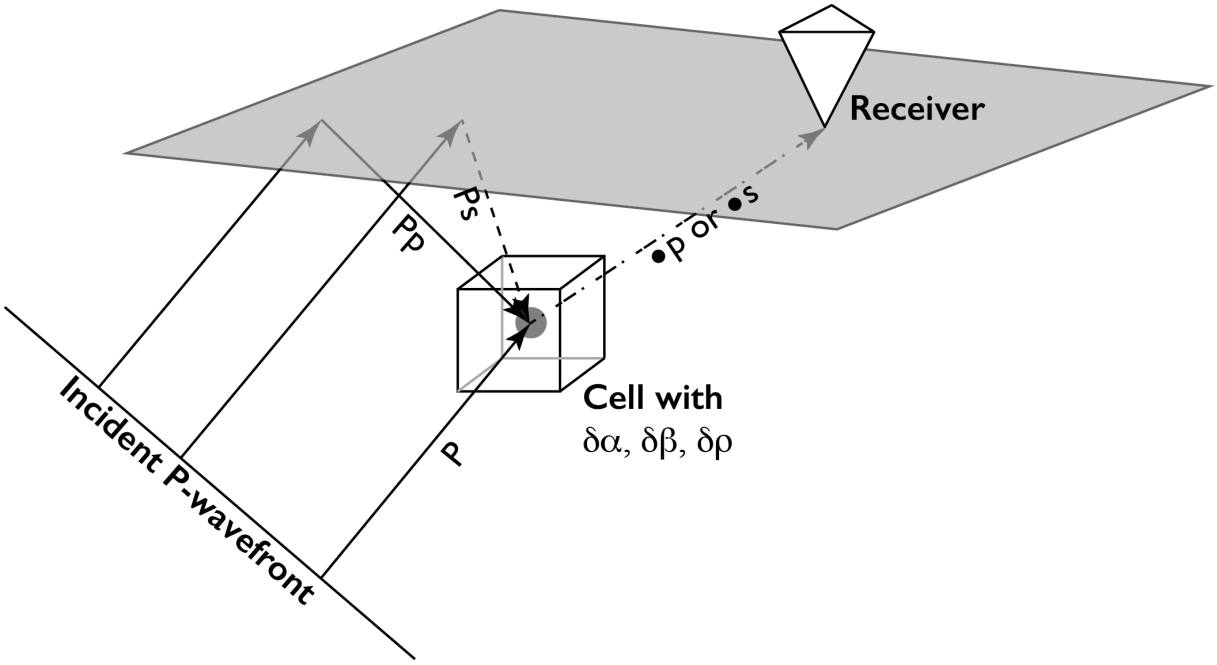

Figure 2. Sketch of the scattered phases included in the full waveform inversion. An incoming planar P-wave wavefront interacts with a cell with a parameter disturbance ($\delta\alpha$, $\delta\beta$, $\delta\rho$) either from the direct wavefront (forward scattered) or the back-scattered wave from the free-surface reflection. The wavetype can convert upon scattering from P to S.

medium and do not interact with heterogeneities again. Therefore, the scattered wavefields from individual heterogeneities are independent. The complete scattered wavefield T can therefore be obtained through the summation of the contributions of individual heterogeneities:

$$T = \sum_{i=1}^{N} \sum_{j=1}^{3} t_{ij} \tag{6}$$

with t_{ij} representing the time series representing the scattering contribution of the *j*th perturbed parameter of the *i*th scatterer (Frederiksen & Revenaugh, 2004).

The medium beneath the array is parameterised into a 3D grid of cells with each cell potentially containing a perturbation of elastic parameters. The perturbation for all cells can be collapsed in an M-element vector $\mathbf{m}$ with dimension M = # cells in [x, y, z] $\times$ properties [$\delta\alpha$, $\delta\beta$, $\delta\rho$]. Summing over all contributing elements we obtain the N-element vector $\mathbf{d}^i$ with the number of displacement samples depending on N = samples $\times$ stations $\times$ components $\times$ events. The dependence of the full scattered wavefield on arbitrary model $\mathbf{m}$ is then described as

$$\mathbf{d} = \mathbf{A}\mathbf{m} \tag{7}$$

$\mathbf{A}$ is an N $\times$ M matrix describing the sensitivity of each data point to each model parameter, i.e. each column of $\mathbf{A}$ represents a differential seismogram for a perturbation of a single parameter in a single cell of the perturbed model. Equation (7) is linear and can therefore be solved using linear inverse theory. To pose this problem as a damped inversion the inverse problem is formulated as a minimization:

$$\min \left\| \begin{bmatrix} \mathbf{A} \\ \lambda \mathbf{I} \end{bmatrix} \mathbf{m} - \begin{bmatrix} \mathbf{d} \\ 0 \end{bmatrix} \right\|_2 \tag{8}$$

with $\mathbf{I}$ being an MxM identity matrix and λ a weighting factor, representing uniform damping. We use the LSQR method (Paige & Saunders, 1982) to solve for the material properties in $\mathbf{m}$. For the inversion of real data it has been found that regularisation by smoothing is preferable to damping (Frederiksen & Revenaugh, 2004) as it provides results with higher coherence. Using LSQR, the model is smoothed by posing $\mathbf{m} = \mathbf{S}\mathbf{x}$ with $\mathbf{S}$ being a matrix containing a Gaussian smoother. We use $\mathbf{B} = \mathbf{A}\mathbf{S}$ and the minimization

$$\min \left(\|\mathbf{B}\mathbf{x} - \mathbf{d}\|^2 + \lambda^2 \|\mathbf{x}\|^2 \right) \tag{9}$$

in which we solve for $\mathbf{x}$ rather than $\mathbf{m}$ (Van der Lee & Nolet, 1997; Frederiksen & Revenaugh, 2004). For all recovery tests and real data inversions, we apply a moving Gaussian smoother with a standard deviation of one model element in the horizontal directions, but we do not smooth in the vertical direction. No smoothing is applied beyond three standard deviations. This choice biases the recovered model towards lateral coherence, making recovered lateral changes more coherent in our study region where we expect strong lateral changes across the NAFZ. The smoothing limits lateral resolution of structures to about 10 km. Vertical resolution of 2 km is defined by the model discretization.

The model space is parameterised as a regular grid with 5 km horizontal grid spacing and 2 km

vertical grid spacing with 30 (0-29) cells in horizontal directions and 60 (0-59) in vertical direction. Each cell is treated as a point scatterer with vertical and horizontal locations at depths $2 \cdot j$ km (j = 0, . . ., 59) and longitude/latitude location of $5 \cdot k$ (k = 0, . . ., 29), respectively. The maximum grid size is controlled by the maximum memory required to invert the dataset (see below). We tested the method with doubled lateral and vertical grid spacing and do not find noticeable differences in the general structure of the solutions except for obvious impacts on the maximum possible resolution of the solutions.

4 DATA

We use passive seismic data from stations of the Dense Array for Northern Anatolia (DANA) that were installed across the NAFZ in the region of the 1999 İzmit and Düzce ruptures (DANA, 2012). DANA was deployed between May 2012 and October 2013 and stations were arranged in a quasi-rectangular region of 35 km by 70 km with a nominal station spacing of 7 km (Fig. 1). Stations were aligned along seven north-south oriented lines (labelled A to F) and 11 east-west lines (labelled 01 to 11). Seven additional stations were installed in an eastern semi-circle with a radius of about 60 km. Three permanent stations (SPNC, SAUV, GULT) of Boğaziçi University and Kandilli Observatory and Earthquake Research Institute/National Earthquake Monitoring Center (BU-KOERI/NEMC) located within the DANA network grid were included in the analysis. Stations were equipped mainly with Güralp CMG-6TD and CMG-3T medium broadband and broadband three-component instruments (full information on the network can be found in DANA (2012)). Data were sampled at 50 Hz.

We use earthquakes within the deployment period with m_b >5.5 from the catalogue of the National Earthquake Information Centre (NEIC) and angular distances of 30° to 90°. For the permanent stations we add events from 2009 onwards (in total 47 additional events contributing typically a single 3-component seismogram (ZRT) to the dataset). Low frequency noise was suppressed by applying a 2-way, 2-pole high-pass filter with cut-off frequency of 0.1 Hz. We calculate 3-component receiver functions (RFs) with a maximum frequency of 1.2 Hz using the time domain iterative deconvolution approach by Ligorría & Ammon (1999) deconvolving the Z com-

ponent from the vertical, radial and transverse components. The calculated receiver functions were visually inspected to select events following these criteria: (1) transverse RFs show lower or comparable amplitudes than radial RFs, (2) the direct *P*-wave arrival is close to the predicted travel time for a 1D Earth model and (3) no evidence for large amplitude ringing. The pre-processing used to obtain the receiver functions is similar to the method used by Kahraman et al. (2015), but applied to all three components (vertical, radial, transverse) of the traces in our analysis. To remove the first arrival, which does not contain any additional structural information, we mute the first 2.5 s of each trace following the theoretical *P*-wave arrival.

In total, we use 1396 distinct source-receiver pairs (with 3 components) from 176 events in our analysis. The distribution of sources is shown in Figure 3. Traces were cut and tapered to 100 s and downsampled from the original 50 Hz sampling to 5 Hz . Despite the downsampling, the matrix to invert is very large which limits the achievable resolution and model depth. Typical storage requirements for the matrix inversion using the sparse storage method are $\approx$338 Gb for a model space dimension (x $\times$ y $\times$ z) 145 $\times$ 145 $\times$ 118 km^3 with an element size of 5 $\times$ 5 $\times$ 2 km^3 and 1396, 100 s long traces, sampled at 5 Hz. We are able to invert the full dataset without recourse to inverting subsets of data and stacking the resulting images (Frederiksen & Revenaugh, 2004; Zhang & Frederiksen, 2013) leading to improved image quality of our results.

5 RECOVERY TESTS

We tested several 1D background velocity models for data inversion and synthetic data recovery including models by Karahan et al. (2001), Bekler & Gürbüz (2008), and Horasan et al. (2002) and models including constant velocity and linear vertical gradients. The background models are used for raytracing to determine traveltimes of the incident and scattered wavefield. While timing of arrivals changes slightly for all realistic velocity models, the overall recovered structure in our tests does not depend significantly on the choice of background model, although depths of interfaces change due to changes in the traveltimes. We chose to use the model by Karahan et al. (2001) for all inversions presented here (Table 1, Fig. 4). This velocity model is derived from seismic

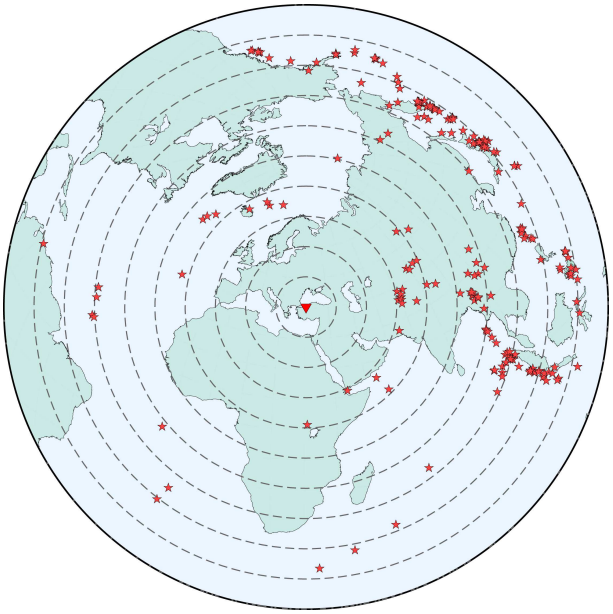

Figure 3. Earthquake locations relative to the center of the DANA array. Dashed circles give distance from DANA center in 10°steps. Earthquakes with m_b>5.5 occurring during the deployment and since 2009 for the permanent stations in an epicentral distance of 30° to 90°were used in the analysis.

experiments in the study area and has been used in previous studies using this dataset (Kahraman et al., 2015; Altuncu Poyraz et al., 2015).

Figure 5 shows the result of an inversion of the full (1396 individual source-receiver combinations) noisy synthetic data generated through the perturbation model (Fig. 4b)). A sub-set of the synthetic traces used in this inversion, i.e. the stations recording event 20123211812 and used in the data inversion, are shown in Fig. 4, c) and d). Synthetic data were generated using ray tracing through the background velocity model with the addition of the scattered wavefield (i.e. the sum-

Depth (km)	V_P (km/s)	V_S (km/s)	ρ (kg/m^3)
0 - 2	3.6	1.967	2.376
2- 14	5.900	3.225	2.814
14 - 26	6.500	3.552	2.955
26 - 40	7.000	3.691	2.975
⋮	8.055	4.347	3.326

Table 1. 1D velocity model used in inversion. Depth, *P*-wave, *S*-wave and density (ρ) following (Karahan et al., 2001)

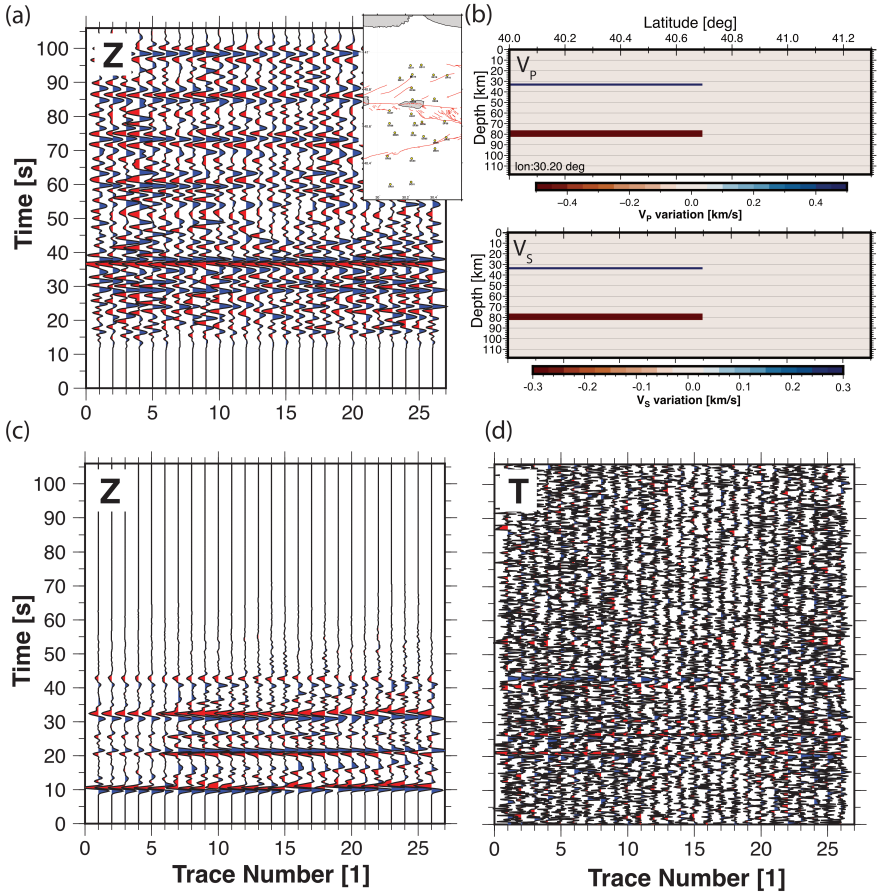

Figure 4. Example data and synthetics. (a) Recorded and deconvolved data example of vertical component of event 20123211812 (insert shows station configuration). Data are cut to 100 s and tapered. The direct P-arrival is suppressed. (b) North-South slices through perturbation model to calculate synthetics. (c) Noiseless synthetics (vertical) through perturbation model shown in b), including all scattering events from direct wave and free surface reflections. Synthetics represent the event-station configuration of the quality-controlled receiver functions for event 20123211812 shown in a). Although the perturbation is not continuous throughout the model, the scattered wavefield can be recorded across the network with distinctive moveout allowing the localisation of the perturbation. (d) Transverse component of the synthetic data calculated for model shown in (b) with added noise. We add Gaussian noise with a 10% standard deviation relative to the maximum signal amplitude to the synthetic data.

mation of all contributions of the single scatterers in the model). We use a 0.25 s wide Gaussian wavelet as the source time function. Synthetic tests use the source-receiver combinations for each event in the dataset, therefore recreating the same resolution as the recorded dataset. For comparison we show the recorded and deconvolved data in Fig. 4a) with the first arrival muted. Scattered phases can be seen coherently across the traces. The synthetic traces (Fig. 4c,d) show similar struc-

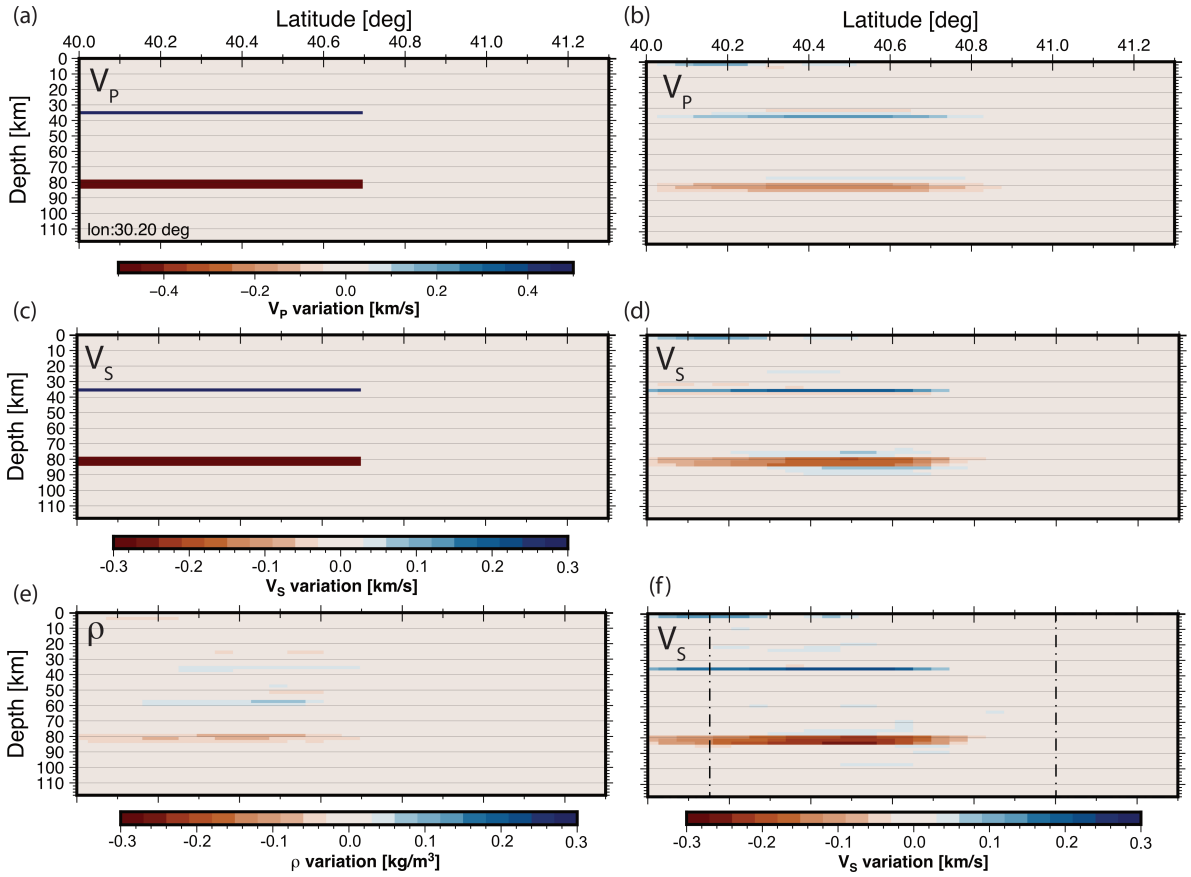

Figure 5. Recovery tests for model shown in Fig. 4 containing heterogeneous V_P and V_S structure. Slices were taken from the 3D model at longitude 30.2°. (a) V_P anomaly of input model containing terminating velocity anomalies at 32 km (negative) and 76 km (positive). Anomaly strength is ±0.5 km/s. (b) Recovered model after scattering inversion. Due to damping the velocity recovered is smaller than the input model but locations are well resolved. Horizontal smoothing with a standard deviation of a single lateral element (5 km) is applied. (c) as (a) but for V_S. Velocity anomaly is ±0.3 km/s. (d) As (b) but for V_S. (e) Recovered low amplitude density (ρ) anomaly after inversion. Input model does not contain density variations so recovered anomaly represents cross-talk between the different components. (f) Wiener filtered longitude slice of recovered model to fit recovered depth of anomaly as a zero-phase wavelet. Black dashed lines in f) indicate the perceived limits of the well-resolved region.

ture although clearly are not able to capture the full complexity of the data due to the simplicity of the model (Fig. 4b). Noise is added to the synthetic data through a random number generator (Marsaglia & Bray, 1964) using 10% RMS amplitude variation Gaussian noise compared to the direct wave amplitude to produce this noisy synthetic dataset (Fig. 4d).

The synthetic model is parameterised with 5 km cell spacing horizontally and 2 km vertically.

The model contains a V_P = +0.5 km/s and V_S = +0.3 km/s anomaly for a single depth element (2 km) starting at 34 km depth and a V_P = -0.5 km/s and V_S = -0.3 km/s anomaly with thickness of 6 km starting at 78 km (Fig. 5 a and c). No density variation was added to the model. The anomalous layers extend across the model in longitude but terminate 80 km into the model from the south (approximately at 40.7°N), leaving the part of the model that corresponds to the Istanbul zone free of a velocity anomaly. The inverted model in Fig. 5 b) and d) shows the recovery of the input model. Because the inversion uses the Born approximation, which generates signals from localized perturbation, the recovered model will be a band-pass filtered version of the input. We apply a Wiener optimum filter to minimize the effects of the inversion process, mainly to reduce sidelobes to aid interpretation. The optimization filter, as described for example by Gubbins (2004), is obtained by minimising the residual between the desired output g_t (Fig. 5a,c) and the signal obtained by convolution of the filter f_t^0 with the actual output x_t (Fig. 5b,d)

$$(e_t)^2 = \left(g_t - x_t * f_t^0\right)^2 \tag{10}$$

The effect of the inversion and the filter terms acting on a single trace of the synthetic model are shown in Supplemental Figure S1.

Although the input model in this test does not contain any density ($\delta\rho$) heterogeneity, Fig. 5e shows that the inverted model for the density structure is affected by cross-talk between the different parameters (more examples given in Supplemental Figure S2). However, relative amplitudes $\Delta\rho$ in this model are small and the effect is most prominent in areas with velocity anomalies. Tests with models including $\Delta\rho$ show that density structure can be resolved. Complete input and output models for this recovery test and further tests are shown in the Supplemental Figures S2 to S8.

These tests show that the recovery of velocity and density anomalies is variable within the model volume due to the relative sampling of the model volume by the ray configuration of the dataset. Peripheral regions are generally less well resolved than the center of the volume and areas of reduced resolution show poor recovery of amplitudes in the inverted model (Fig. 5). Within the central zone we do not observe strong depth or amplitude variations of the recovered model, adding confidence to our interpretation. Areas of the model space that are not well resolved are

masked in all following figures (and supplemental material) and the approximate limits of the well-resolved volume are shown in the N-S profiles (dot-dashed vertical lines in Fig. 5f), to which we limit our interpretation. These areas are estimated from our recovery tests as shown in Fig. 5 and the Supplemental Material.

The recovered model in Fig. 5b,d) shows some low-amplitude imaging artefacts around the recovered anomalies but also away from them, which are not fully removed by the Wiener filtering. These are likely due to the sampling of the volume by the dataset as well as the noise added to the synthetic data. Care has been taken when interpreting recorded data inversions to not interpret such artefacts as lithospheric structure.

Changing the depth extent of the inverted model space between 48 km and 118 km (in 20 km steps) does not lead to strong changes in the inverted model. A comparison between a 48 km and 118 km deep model containing the same structure for V_P and V_S is shown in supplemental Figure S3 a) and b). This holds even when synthetic traces were generated including structure below the inverted volume (Fig. S3c) showing that heterogeneities underneath the volume are not erroneously mapped into the model volume. In the following section we show models down to depths of 118 km (60 nodes with 2 km spacing) in a trade-off between achievable resolution, model size and required computer memory. The horizontal smoothing leads to some smearing of energy in horizontal directions. Nonetheless, Fig. 5 shows that terminating discontinuities can be accurately located within 1 to 2 horizontal elements (i.e. 5 to 10 km) in the central region of the model space. We also performed recovery tests using other structural models including velocity and density heterogeneities to better understand the performance of the method (for these further recovery tests please see Supplemental Material).

6 RESULTS

The results of the tomographic scattering inversion of the DANA dataset are shown in Fig. 6 and Fig. 7. Fig. 8 presents an interpreted section of the results. Slices in Fig. 6 and 7 were extracted from the three-dimensional inversion volume along North-South (Fig. 6) and East-West (Fig. 7) profiles at locations shown in Fig. 1. The locations of the profiles were chosen to be in similar

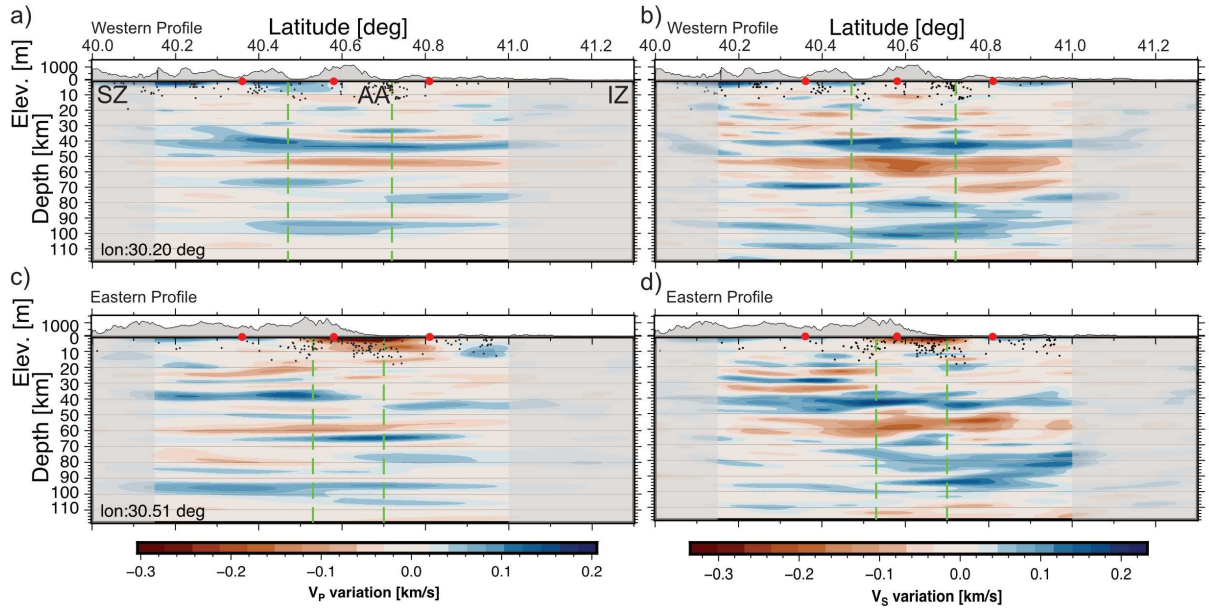

Figure 6. North-south oriented slices along 30.2°(top) and 30.51°(bottom) for the inversion for V_P (left) and V_S (right) structure. Profile locations are indicated in Fig.1. Approximate locations of the southern and northern branches of the NAFZ are shown as dashed green lines. Areas with limited resolution as determined from the recovery tests (Fig. 5) are masked in grey. Red dots indicate locations of EW slices shown in Fig.7. Black circles indicate local seismicity as determined by Altuncu Poyraz et al. (2015) within a ±5 km corridor projected onto the profile. The top panel in each subpanel shows SRTM topography along the profile (Farr et al., 2007).

locations to those shown in Fig. 6 of Kahraman et al. (2015) (for an equivalent display to Kahraman et al. (2015) see Supplemental Figure S9 , S11 and S13). Full solutions are presented in the form of animated GIFs in Figures S15 and S16 of the Supplemental Materials. A *kml* file is also provided to display the profiles in their correct geographical location. The model is filtered with the Wiener optimization filter as discussed above. In the following we report depths at the top of a heterogeneity in the filtered sections which gives the best agreement of input and recovered model on the recovery tests (e.g. Fig. 5)

Generally, the *S*-wave images show greater amplitude and are better constrained. The *S*-wave tomographic images also seem to show more fine scale structure likely related to the shorter wavelength. The density ($\Delta\rho$) profiles show some of the major structure and are shown in the supplemental Figs. S13 and S15 but suffer from cross-talk as shown in Fig. 5. As the interpretation of

the $\Delta\rho$ profiles is more difficult and there is no independent constraint on the density structure we do not discuss this parameter further in the text.

6.1 Western Profile

Profiles for V_P (Fig. 6 a) and V_S (Fig. 6 b) have been extracted along a longitude of 30.2° E. Areas with limited resolution as determined from the recovery tests (Fig. 5) have been masked in this profile in transparent grey. The V_P profile (Fig. 6a) is dominated by a velocity increase at ∼40 km depth for most of the profile, which we associate with the Moho. The Moho velocity increase bifurcates south of ∼40.4° with a shallower velocity increase located at ∼32 km depth deepening to 40 km at ∼40.4°N. The anomaly also seems to fade, i.e. showing less of a velocity anomaly, south of about 40.3°N. The point of bifurcation coincides with the surface expression of the southern strand of the NAFZ. A similar Moho signal is observed in the *S*-wave anomaly at ∼40 km, shallowing to about 38 km within the Armutlu block, which shows a thickening of this interface. The *S*-wave anomaly does not show the same shallow branch observed in the *P*-waves but shows lower amplitudes south of ∼40.4°N, i.e. south of the southern NAFZ strand.

Observed crustal structure includes a weak high V_P anomaly at ∼18 km in the Armutlu block with weak, complex V_S structure in the Sakarya zone. Complex structure starting at ∼32 km depth (positive and negative anomalies) in V_P and V_S can be seen in the vicinity of the northern strand (40.7°N) just overlying the Moho. The V_P model shows less structure in the upper crust except a fast anomaly to depths of ∼5 km around the southern branch and a slow (also seen in V_S) overlying fast anomaly between ∼10 km and ∼20 km depth in the Armutlu block.

The high velocity anomaly at 40 km depth is underlain by a strong low V_P and V_S anomaly at depths of ∼50 km. This anomaly shows lower amplitudes in the Sakarya Zone with the change coinciding with the surface expression of the southern NAFZ strand. We also identify a velocity increase in V_P and V_S at ∼64 km and ∼66 km depth, respectively, around 40.4°N (southern strand) and ∼74 km in V_P beneath the Istanbul zone (with a termination at the northern strand). The V_S anomaly shows a low velocity anomaly at ∼68 km depth just north of the northern strand changing to a high velocity anomaly at ∼74 km depth north of 40.9°N.

At greater depths we observe a fast anomaly in V_S at ∼78 km depth and a fast anomaly in V_P and V_S at ∼92 km but showing depth variation in V_S. The 78 km anomaly seems to merge with the deeper anomaly in the Istanbul zone.

6.2 Eastern Profile

The eastern North-South profile at 30.51° E (Figs. 6c and 6d) shows more structure, especially in the crust, than the western profile despite the close proximity of the two profiles.

We observe a strong, fast V_P anomaly at a depth of ∼34 km terminating halfway through the Armutlu block and re-emerging at a depth of ∼42 km just north of the northern strand in the Istanbul zone. In V_S we observe a more continuous structure with a high velocity anomaly at ∼36 km depth in the south, stepping to ∼42 km at ∼40.7°N coinciding with the northern strand. The V_P anomaly is weak in the Armutlu block on this profile and seems to terminate at 40.6° E, while the V_S anomaly is more continuous, but also weakens in this region. The amplitude variation of these anomalies cannot be explained by the limitations of the sampling (see Fig. 5).

Especially striking in this profile is the complex V_S structure in the Sakarya Zone down to depths of about 30 km manifesting as series of fast and slow anomalies between ∼10 km and 32 km (see supplemental Figure S7). The V_P structure is similar but weaker than V_S. The structure terminates abruptly at the southern strand with little crustal structure in the Armutlu block. The Adapazarı basin (centred at about 40.7° N) is representing as a low velocity anomaly between 40.6°N and 40.7°N to depths of about 6 km (V_S).

Similar to the western profile we identify a slow anomaly in both V_P and V_S at depths of ∼56 km and ∼52 km, respectively. The V_S anomaly seems to show more complexity. We identify a weak slow anomaly at ∼76 km depth in the Sakarya zone in V_P which appears detectable but much weaker in V_S. This anomaly seems to terminate at the southern branch. Fast anomalies are detected at ∼92 km in V_P and V_S across the profile with shallower fast anomalies for V_P and V_S at ∼76 km depth beneath the Istanbul zone and the Armutlu block. In V_P there is evidence of this interface splitting into a deeper interface deepening to ∼102 km across the southern strand.

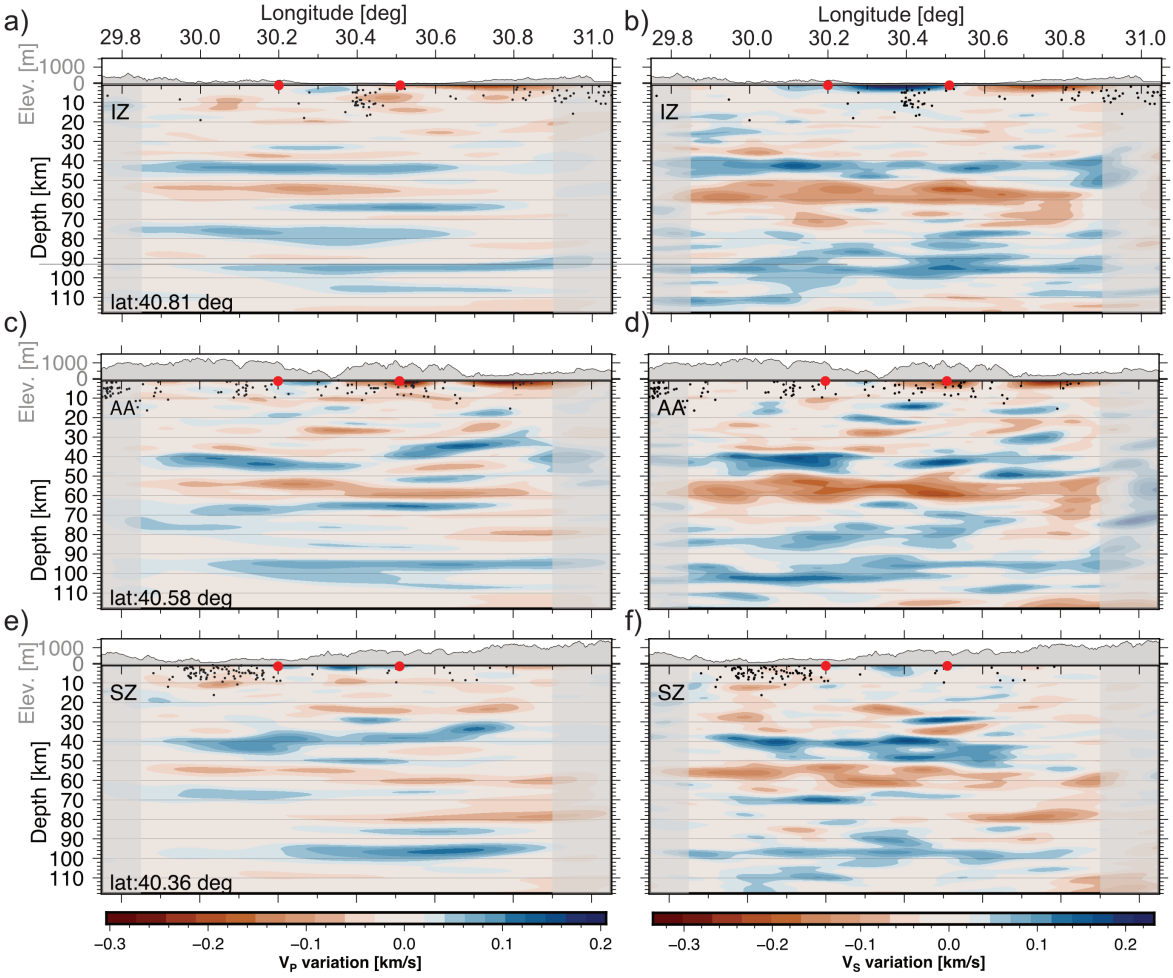

Figure 7. As Fig.6 showing west-east oriented slices through the inverted V_P (left) and V_S (right) structure. Slices are located in the Istanbul zone (IZ - a,b) at latitude 40.81°, the Armutlu-Almacık block (AA - c,d) at 40.58°and Sakarya zone (SZ - e,f) at 40.36°. Areas with limited resolution as determined from the recovery tests (Fig. 5) masked in grey. Black circles indicate local seismicity as determined by Altuncu Poyraz et al. (2015) within a ±5 km corridor projected onto the profile. The top panel in each subpanel shows SRTM topography along the profile (Farr et al., 2007).

6.3 Sakarya Zone

The West-East profile for V_P and V_S (Figs. 7e and 7f, respectively) has been extracted along 40.36°N and is fully located within the Sakarya zone. The Sakarya zone is the southernmost tectonic block in the study region. The inverted scattering tomography model shows a positive anomaly at depths of ∼38 km. In V_P this interface shallows to ∼32 km around 30.6°E. This anomaly seems rather complex and might be discontinuous. We also identify a laterally limited fast anomaly at ∼30 km between 30.4°E and 30.6°E. A deeper slow anomaly at about 54 km

depth can be seen that shows a slight step down to about 60 km (V_P) at about 30.4°E and seems complex in V_S. The western part of the profile shows a fast anomaly at ∼68 km, with a slow anomaly at ∼78 km in the east. A fast anomaly at ∼98 km depth (94 km in V_P) is identified which terminates at 30.2°E in V_P.

6.4 Armutlu Block

In contrast to the Sakarya Zone, the Armutlu Block (Figs. 7c and 7d for V_P and V_S, respectively) shows more structure down to depths of 40 km. A fast anomaly at ∼40 km terminates around 30.4°E and appears as shallow as 30 to 32 km further east in V_P. V_S also shows the termination but a less pronounced step. The step around 30.6° E seems to coincide with the profile moving from the Armutlu block to the Almacık mountains. West of ∼30.4°E, this interface is underlain by a slow anomaly at ∼50 km showing a step to ∼58 km at 30.4°E in V_P. Overall V_S seems more complex. We identify several small scale fast and slow anomalies in the crust, the strongest at ∼14 km around 30.4°E in V_S. Slow anomalies shallower than 40 km are indicated between 30.5°E and 30.9°E.

A fast anomaly at ∼ 94 km stretches across most of the profile in V_P, with comparable but more complex structure in V_S. the V_S section also shows more localised structures at depths greater than 80 km.

6.5 Istanbul Zone

The Istanbul zone (Figs. 7a and 7b for V_P and V_S, respectively) shows very little structure down to depths of about 40 - 42 km where a strong fast anomaly can be detected in V_P and V_S . This fast anomaly seems to terminate around 30.6°E for V_P but continues across the IZ in V_S. A slow anomaly is visible in V_S at depths less than 10 km between 30.6°E and 30.9°E and a fast anomaly between 30.2°E and 30.5°E.

The strong Moho signal is underlain by a slow anomaly around 52 km depth again terminating around 30.6°E for V_P. V_P shows a fast anomaly at ∼74 km depth, which like the Moho signal in this block, terminates at about 30.6.°E; the corresponding structure in V_S is weaker and dis-

continuous. A strong fast anomaly at ∼92 km depth can be seen in V_P and V_S, and again the V_S structure is complex.

7 DISCUSSION

The scattering tomography results show changes in the structure over distances of 10 km in the lithosphere. The smoothing process implemented in the inversion means that more abrupt changes present in the actual structure would also appear smoothed over that distance. These changes can be related to the different structure of the tectonic blocks (Ben-Zion et al., 2003; Bulut et al., 2012) and manifest in e.g. the North-South profiles but can be detected in the full data volume (see Supplemental Material). Nonetheless, we observe structural changes also in East-West direction similar to those detected earlier (e.g. Bariş et al., 2005; Beyhan & Alkan, 2015; Çubuk-Sabuncu et al., 2017) where more continuous structure within the tectonic blocks might be expected

The profile across the Armutlu block follows the strike of the NAFZ east of about 30.7°E where it leaves the Armutlu block (Fig. 1). The depth slices through the model shown in Fig. 6 and 7 show strong changes between the two north-south trending profiles despite their close proximity. Interpreted NS sections are shown in Fig. 8. We have performed recovery tests for the dominant interpreted structure in the tomographic model (Fig. 9). We include complex crustal structure in the Sakarya zone, a Moho step and lithospheric structure in this complex model. We are able to recover the input structure very well in V_S and V_P with stronger recovered anomaly amplitudes in V_S (Fig. 9a & b). Some low amplitude spurious signals due to noise and the inversion volume sampling are visible in these models as discussed earlier, but are generally of lower amplitude than the recovered model. These are mainly visible close to the input anomalies (Fig. 9 a). The difference in V_P and V_S recovery indicates that a joined interpretation of V_P and V_S might be necessary for robust interpretation of the scattering tomographic images. To highlight the most coherent part of the model we stack the depth profiles in longitude and divide these at 30.4°E to show western and eastern stacks in Fig. 10 for both V_P and V_S. Fig. 11 shows a schematic of the dominant structure in the tomographic images.

Comparing the individual slices and the stacked velocity-depth profiles shows that many features

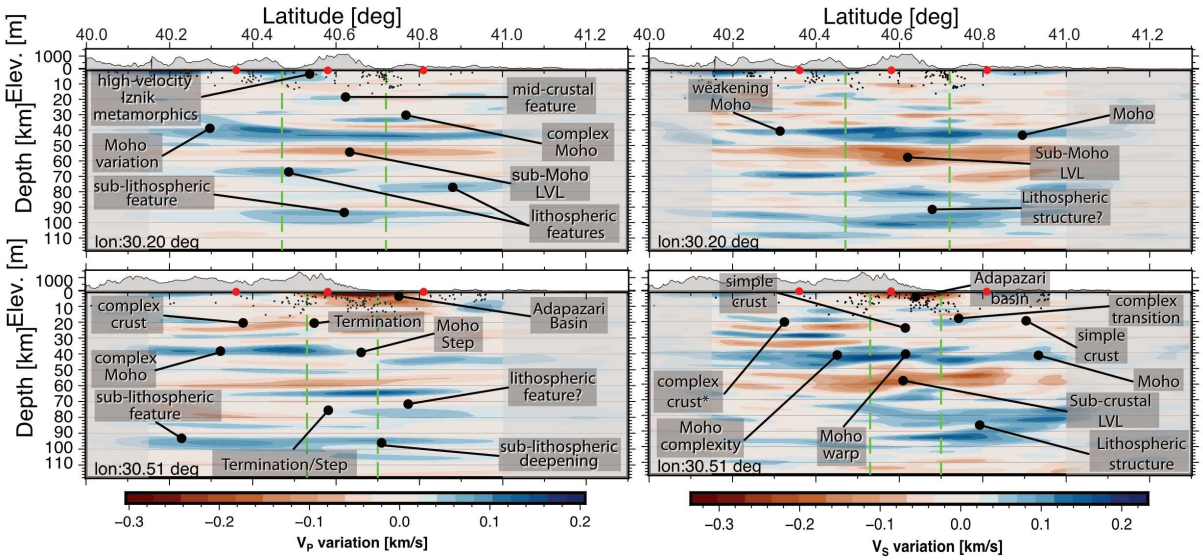

Figure 8. Interpreted NS cross sections for V_P (left) and V_S (right). Shown are North-South oriented slices as in Fig. 6. Black circles indicate local seismicity as determined by Altuncu Poyraz et al. (2015) within a $\pm$5 km corridor projected onto the profile. The top panel in each subpanel shows SRTM topography along the profile (Farr et al., 2007).

are coherent along stretches of the profile, but can change on short scale-lengths in both N-S and E-W directions.

7.1 Mohorovičič discontinuity

In the west, the Mohorovičič discontinuity (the Moho) is visible in both V_P and V_S as a dominant fast velocity at depths of $\sim$40 km with variations in V_P in the south and in V_S in the Armutlu block. In the east the Moho is shallower in the south (34 km (V_P), 38 km (V_S)) but shows a step to greater depths (42 km) between 40.6° N to 40.7° N at 30.51°E.

The deepening of the Moho might indicate the existence of a shear zone at the location of the northern branch which has also been indicated in teleseismic tomography in this region (Papaleo et al., 2018, 2017). The Moho at 30.2°E is overlain by a slow anomaly in V_P between 40.7°N and 40.9°N (Fig. 8). In the V_S model there is weak evidence for a similar, but weak and intermittent, structure between 40.8 and 41.0°N .

In the east, the Moho seems much weaker and discontinuous across all three tectonic blocks. The strongest change in Moho depth can be identified around 40.8°N in the eastern profile where

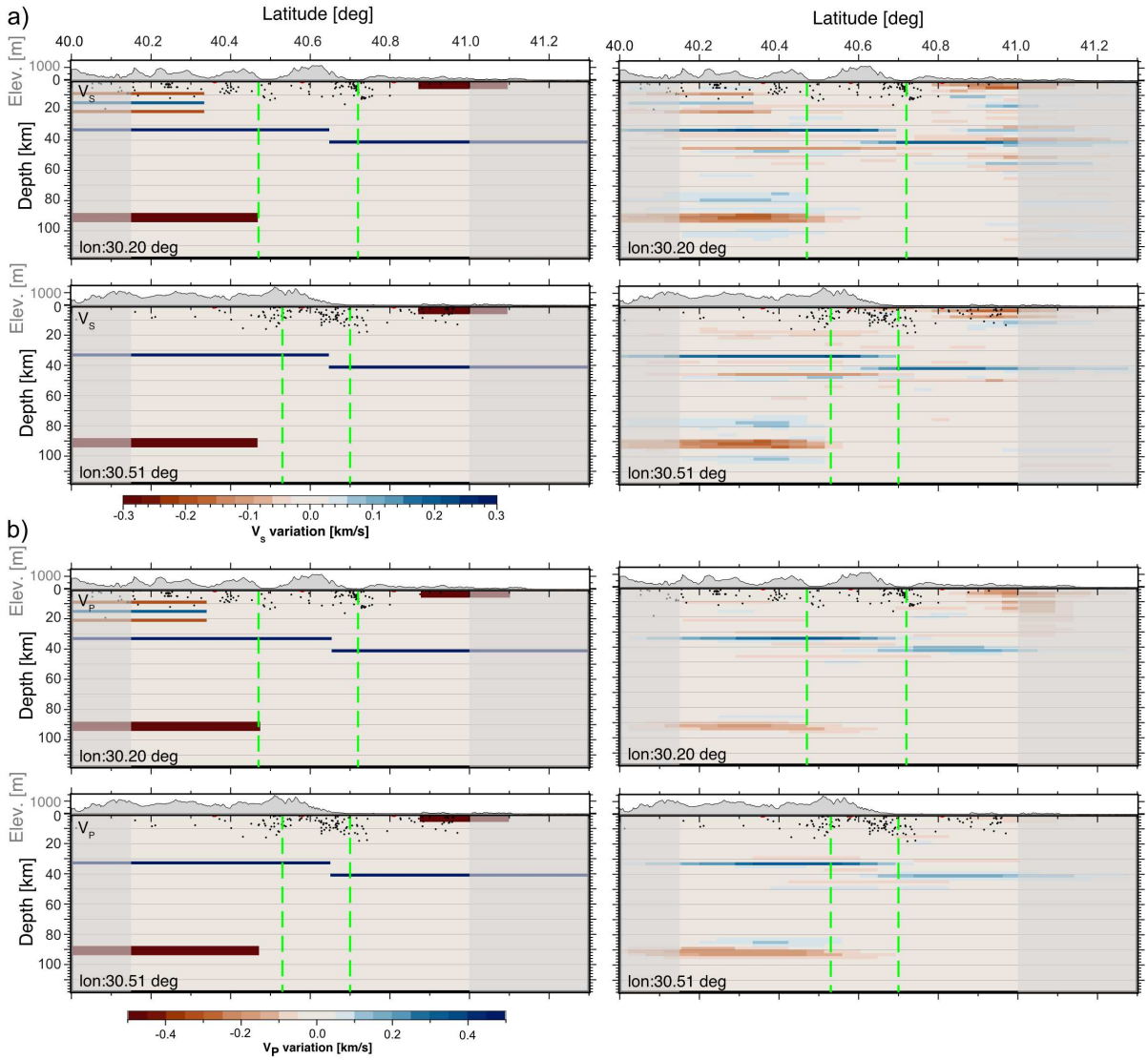

Figure 9. Recovery test for complex synthetic model containing V_P and V_S anomalies in the crust, a stepped Moho like structure and a deeper low velocity anomaly. This model is similar to the dominant structures interpreted in the discussion. a) V_S structure along NS profiles along 30.2° E (top) and 30.51° E (bottom). The input model is shown on the left, the recovered model on the right. b) as a) but for V_P.

we observe a step from 32 km to 40 km coinciding with the surface expression of the northern branch. For the northern branch the discontinuous structure seems to extend into the mantle as discussed later.

While there are strong north-south changes in the profiles in Figs. 6 to 10 we also observe strong east-west changes, e.g. in the Sakarya zone with a complex Moho structure around 30.5°E and the weakening of the Moho east of 30.4°E or the change in the Armutlu block at ∼30.5°E. The latter might be related to the step-over structure of the NAFZ related to the differential movement

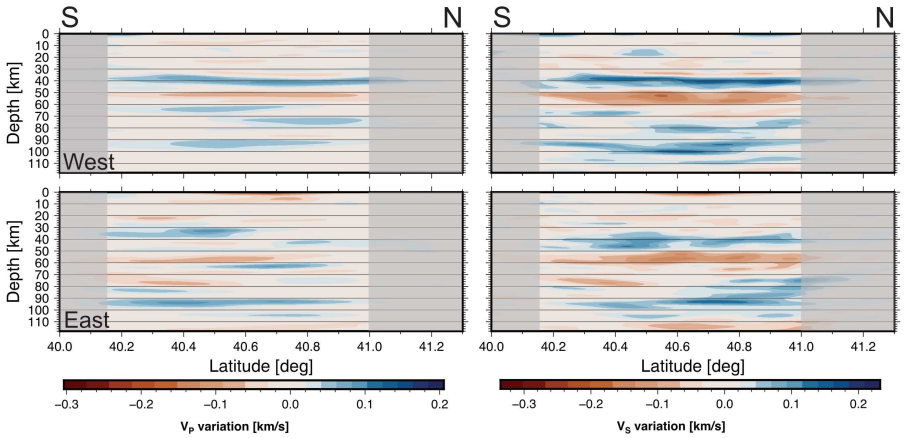

Figure 10. North-South depth slices of the model stacked in longitude (a) for dV$_P$ and (b) dV$_S$. Top figure for each velocity variation is for the western section (west of 30.4° E) of the model with the bottom the eastern part (east of 30.4° E).

of the Armutlu and Almacık blocks and the trend of the suture zones between the tectonic blocks. We also observe a pronounced change of the Moho depth between the Armutlu block and the Almacık mountains at around 30.6° E (Suppl. Fig. S14, S15), indicating strong contrasts in crustal structure from the Armutlu block to the SNAFZ shear zone.

7.2 Crustal structure

We find evidence for strong crustal structure variation along some of the profiles. The most striking structure is the apparent strong crustal layering south of the southern branch in the Sakarya zone (Fig. 6 c,d) for both V_P and V_S (best visible in V_S). The crustal heterogeneity is clearly truncated by the surface location of the southern branch and forward models indicate that it consists of a series of high and low velocity anomalies (e.g. Fig. 9) perhaps related to emplacement of magmatic sills during the Tethys closure (Karabulut et al., 2003). The crust in the Armutlu block on the other hand is relatively homogeneous, adding to the stark difference across the southern NAFZ branch.

Overlying the Moho in the area of the northern strand along the eastern profiles we detect small-scale, complex Moho structure. Modeling indicates that it could be related to a heterogeneity with limited extent approximating a point scatterer perhaps related to the material property changes in the fault zone. The lateral smoothing inherent to our inversions leads to a lack of resolution in this case.

We detect evidence for the Adapazarı basin as low velocity anomalies between 40.6°N and 40.8°N in the eastern profiles. Our method does not allow the necessary depth resolution at these depths for conclusion on the depth of the basin. The high velocity Iznik metamorphics (Taylor et al., 2019a) can be detected between 40.4°N and and 40.6°N in the western profile.

Areas in the proximity of the surface expressions of the northern and southern strands show more heterogeneous structures than areas further away, perhaps related to increased damage around the fault zone (Ben-Zion & Sammis, 2003). We detect a few localized crustal heterogeneities in the Sakarya zone and Armutlu block. There is evidence for a more continuous low velocity anomaly at ∼10 km and ∼25 km depth in the Armutlu block and the Sakarya zone that is best visible in the V_P models (Fig. 7). The scattering tomography shows less heterogeneity in the Istanbul Zone than in the neighboring tectonic units of the Armutlu block and the Sakarya Zone, which could be related to the reported absence of metamorphism and the lack of major deformation (Okay, 1989).

7.3 Sub-crustal structure

Below the Moho we identify a dominant low velocity layer at depths between ∼50 and ∼60 km in the north-south profiles for V_P and V_S (Fig. 10). The low velocity layer weakens but remains observable around 40.6°N and is possibly linked to the surface expression of the northern strand. The weakening is more pronounced in V_P than V_S. The interface to the anomaly is slightly deeper (52 km) in the stacked eastern profile but also shows changes in the extent of the reduced seismic velocities from traveltime tomography (Papaleo et al., 2017, 2018). The continuity of this structure beneath all tectonic blocks, although with possible depth and structural variations, indicates that it is related to lithospheric structure post-dating the amalgamation of northern Anatolia and the development of the suture zones. It is similar to a signal detected by Kahraman et al. (2015).

The fast anomalies detected at depths greater than 60 km show changes in depth and structure in the vicinity of the surface locations of the NAFZ branches although slightly offset to the north, possibly consistent with shear zones that dip to the North (Kahraman et al., 2015; Papaleo et al., 2017, 2018). The tectonic implications of of such a northerly dip remain unclear.

There is little evidence for a coherent deep low velocity anomaly in our model that can be

583 interpreted as a Lithosphere-Asthenosphere boundary (LAB). The lower part of the models seems
584 dominated by high velocity anomalies, although there is weak evidence for a low velocity anomaly
585 in V_S between 110 and 120 km depth. Results from previous studies suggesting shallow LAB
586 depths between 80 and 100 km are confirmed in the entire region outside the subduction zones
587 (Kind et al., 2015). Therefore we cannot confirm a detection of the LAB in our models. The LAB
588 might be too gradational to show up as signal in the *P*-wave coda and to be imaged using our
589 method.

590 **7.4 Shear zones**

591 In our scattering tomographic model we see the strongest evidence for the NAFZ shear zone in the
592 abrupt changes of crustal and sub-crustal structures. We see crustal structures that terminate on or
593 near both fault strands, most clearly in the changes of the crustal structure transitioning from the
594 Sakarya Zone to the Armutlu block (i.e. across the SNAFZ) which we can trace to Moho depths
595 (e.g. Fig. 6c,d). In general, the AA shows almost no heterogeneity in the crust. At the northern
596 boundary of the AA, coinciding with the NNAFZ we detect energy that seems consistent with the
597 existence of a laterally very limited heterogeneity (Fig 8) that might be related to the damage zone
598 of the fault zone (Ben-Zion & Sammis, 2003).

599 The Moho step detected in the eastern profiles (e.g. Fig 6c, d)) seems to coincide with the
600 surface expression of the NNAFZ and might indicate a localized subvertical shear zone extending
601 deeper than the Moho and into the mantle. Some interfaces in the lithospheric mantle (e.g. Fig
602 6a,c) also show terminations coinciding with the NNAFZ indicating sub-Moho structure related to
603 the shear zone. Willis et al. (2019) showed that localization of a shear zone in the lower crust can
604 be produced by thermal activation in this tectonic setting if the crust has a rheology comparable to
605 that of dry plagioclase. Furthermore, thermally activated shear zone localization of the upper 10
606 km or so of the mantle is also possible for a dry peridotite rheology.

607 Other continental transform faults such as the San Andreas Fault system (SAF), the Alpine
608 Fault (AF) and the Dead Sea Transform (DST) show similar structures (e.g Stern & McBride,
609 1998; Weber et al., 2004; Mohsen et al., 2005; Ford et al., 2014) indicating localized shear through-

out the crust . The SAF in southern California (Yan & Clayton, 2007) and the DST along the Aravia fault (Mohsen et al., 2005) seem to offset the Moho in close proximity to the surface expression of the fault similar to the eastern profiles across the NAFZ. There is evidence that the SAF also offsets the LAB even in the upper mantle (Ford et al., 2014).

Due to the intra-Pontide suture zone that juxtaposes tectonic blocks of different provenance in the study area, it is difficult to separate the residual signature of a suture zone from the shear zone. The NAFZ seems to exploit a crust weakened by the presence of sutures. Nonetheless, our results provide first evidence that the southern branch might extend throughout the crust (Fig. 6). We also see evidence in the crust indicating small-scale heterogeneity coinciding with the location of the southern and northern strands .

7.5 Comparison to receiver function structure

Our results allow a comparison with crust and upper mantle structures resolved in the region using other approaches. Direct comparison with the *P*-wave receiver function study of (Kahraman et al., 2015) using the same receiver array combines the comparative strength of scattering tomography in imaging lateral and vertical changes with the receiver function sensitivity to vertical discontinuities. Kahraman et al. (2015) noted pronounced variations in crust and upper mantle structure and properties both in north-south and east-west directions in agreement with this and other studies (e.g. Çubuk-Sabuncu et al., 2017; Beyhan & Alkan, 2015). Kahraman et al. (2015) imaged lateral terminations in key sub-horizontal discontinuities beneath the southern and northern fault zones. Here we constrain: 1) complex crustal layering in the Sakarya zone towards the east; and 2) a mid-crustal feature in the Armutlu block in the west of the study region, which appears to be confined by NAFZ fault branches extending deeper into the crust. Kahraman et al. (2015) find a deepening of the Moho from north to south in the east which is not as clearly seen in this study and they do not show evidence for a step in Moho depth roughly along the surface expression of the northern NAFZ, although there is a high-amplitude receiver function signal beneath the Moho in the Istanbul Zone, which may indicate a sub-crustal anomaly. Kahraman et al. (2015) show evidence for structure in the lithospheric mantle in some parts of their profiles that is broadly in

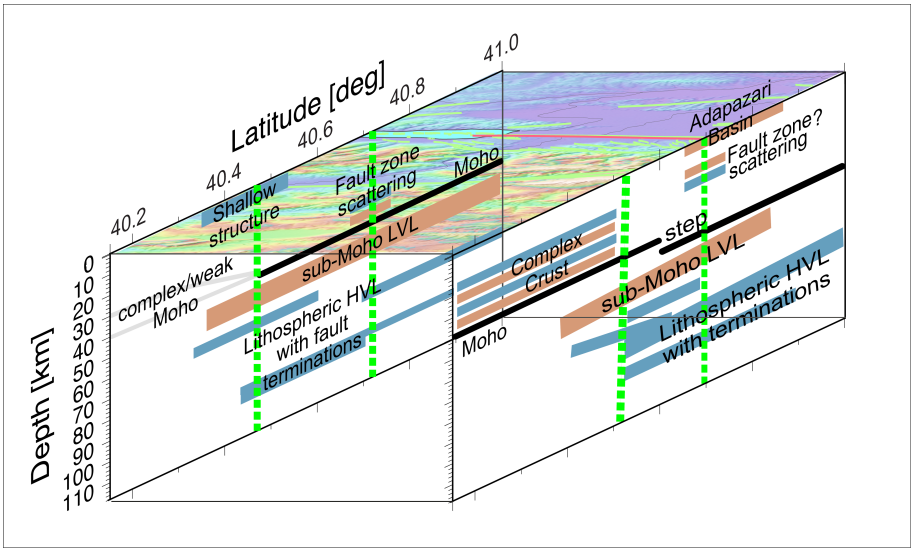

Figure 11. Cartoon of the dominant structures identified in the scattering tomography. We detect crustal and mantle lithospheric structure as well as the crust-mantle interface which all seem to be affected by the North-Anatolian Fault Zone. Top shows shuttle radar topography (SRTM) from Farr et al. (2007).

agreement with structures resolved in this study. In particular, their anomaly at $\sim$ 60 km depth beneath the Istanbul Zone is co-located with a sub-Moho low-velocity zone seen here (Fig. 8). A northwards dipping shear zone, interpreted through the termination of interfaces in the crust and lithospheric mantle in Kahraman et al. (2015), is not evident in the scattering tomography results. Instead, we observe a Moho step beneath the northern NAFZ surface expression and terminations of lithospheric features beneath the NNAFZ could indicate a sub-vertical extension of the fault zone into the lithospheric mantle. In contrast, the SNAFZ appears to terminate at the Moho in the scattering tomography results, in agreement with Kahraman et al. (2015).

8 CONCLUSION

We have used data from a dense deployment of seismometers over the actively deforming North Anatolian Fault Zone in the region of the 1999 İzmit and Düzce ruptures to analyse the scattered seismic wavefield following teleseismic *P*-wave arrivals. Extending the analysis of the scattered seismic wavefield to a tomographic inversion (Frederiksen & Revenaugh, 2004) we detect crustal and mantle heterogeneities that can be linked to the structure and tectonics of the lithosphere beneath the North Anatolian Fault Zone in north-west Turkey. Our high-resolution images from the

scattering tomography down to depths of 120 km allow unprecedented insight into the lithospheric-scale structure of a major continental strike-slip fault. We show complex structure in crust and lithospheric mantle that can be linked to modern active tectonic processes as well as the structure of the crustal terranes that form the region (Fig. 11).

Our tomographic models show complex crustal structure in the Sakarya zone terminating at the southern branch of the NAFZ and terminations of crustal discontinuities at the northern branch. The terminations of crustal structure are sharp within the resolution of our approach. We observe a step in Moho depth coinciding with the surface location of the northern branch of the NAFZ across most of the study region. Terminations of sub-horizontal structures beneath the Moho might indicate that the shear zone extends into the upper mantle to depths of at least 75 km. We detect changes in lithospheric structure perpendicular and parallel to the NAFZ indicating the imprint of complex tectonic history of the region onto the lithospheric structure.

We show that scattering tomography in conjunction with dense recordings of the seismic wavefield is able to provide deeper insight into crustal and mantle structure and the fine-scale structure around fault zones adding to our knowledge of lithospheric structure around fault zones. Strain associated with the NAFZ seems to be localized through the crust and into the mantle. The NAFZ likely exploits weaknesses due to old sutures in this region following the northwards subduction of the Tethys during the amalgamation of Anatolia.

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9 DATA AVAILABILITY

All data for this study (DANA, 2012) are freely available from the IRIS DMS under the network code YK. More information on the data can be found under the data DOI `doi:10.7914/SN/YH_2012`. `http://ds.iris.edu/gmap/YH?timewindow=2012/5/01-2013/10/01`

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