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Key Points:

- New crustal model for Anatolia derived from joint inversion of receiver functions and surface waves
- Central Anatolian Plateau has N–S asymmetry, with lithospheric removal developed most extensively in the south
- Eastern Anatolian Plateau has a thin lithosphere due to delamination, and is underlain by buoyant Afar plume material

Supporting Information:

Supporting Information may be found in the online version of this article.

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Lithospheric Seismic Structure of the Anatolian Plate and Its Implications for Plateau Uplift: Evidence From Joint Inversion of Receiver Functions and Surface Waves

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Abstract High topography (>1 km) on the Anatolian Plate is widely attributed to buoyant mantle support, not just the isostatic response of crustal shortening. However, uncertainties in lithospheric structure hinder attempts to discriminate between competing uplift mechanisms. We jointly inverted receiver functions with Rayleigh wave group-velocity dispersion curves to obtain 1D shear velocity profiles for 575 seismograph stations across Anatolia. Crustal thickness increases from ~25 km in the west to ≤44 km below Eastern Anatolia, with short length-scale Moho topography in places. Within the Central Anatolian Plateau (CAP), thicker crust, lower residual topography, higher residual Bouguer gravity anomalies, and faster upper-mantle wavespeeds north of ~39°N than to the south suggest that northern CAP lithospheric removal is less-developed. Processes like slab break-off, lithospheric delamination and/or dripping are therefore more likely to contribute to the uplift in the south, with additional support from buoyant mantle flow through the Cyprus slab tear. A ~5 km Moho step across the Eastern Anatolian Fault (EAF) results partly from dominant crustal shortening on the Anatolian Plate. Further east, no Moho depth contrast exists across the Bitlis suture; Arabia-Eurasia crustal shortening has thus been accommodated by both plates. An ENE–WSW-trending band of positive residual topography and negative residual Bouguer anomalies parallels the EAF below the northwestern tip of Arabia. This connects to a zone of peak residual topography and slow uppermost mantle wavespeeds below the magmatically active Eastern Anatolian Plateau. These observations are explained best by NE-flowing, buoyant mantle plume material from the Afar hotspot to the south.

Plain Language Summary This study addresses the long-standing question of why the Central (CAP) and Eastern Anatolian Plateaus (EAP) are elevated significantly above sea level, despite their relatively normal crustal thickness. To achieve this goal, we use seismological data to obtain crustal and mantle structure below 575 seismograph stations across the region. Our results confirm the requirement for a significant mantle contribution to Anatolia's high topography. Below the southern CAP, removal of mantle lithosphere (e.g., via dripping or delamination) is the likely uplift driver, with additional contribution from asthenospheric upwelling through a tear in the Cyprus slab. Such processes are comparatively less-developed below the northern CAP. The high elevation of the EAP is the result of detachment of the Bitlis slab and associated near-complete removal of the EAP mantle lithosphere. Hot, buoyant mantle plume, sourced from the Afar region to the south, also contributes to the EAP's high elevations, as well as the uplift of the northwestern tip of Arabia.

1. Introduction

1.1. Overview

The Anatolian Plate has developed in response to the ongoing collision between the African/Arabian plates to the south and the Eurasian Plate to the north (Figure 1). There is a long-standing consensus that the high elevations of the Central (CAP) and Eastern Anatolian Plateaus (EAP) are not the result of crustal shortening alone (e.g., Kounoudis et al., 2020; Ogden & Bastow, 2022; Vanacore et al., 2013; Şengör et al., 2003; Zor et al., 2003): mantle contributions must play a role, but precisely what tectonic/geodynamic processes dominate is debated. For example, Bartol and Govers (2014) invoke the hypothesis of a single large-scale lithospheric delamination event in early Miocene times to account for both CAP and EAP uplift. Hua et al. (2023) cite S-to-P receiver function data as evidence that the entire Anatolian Plate is underlain by uniformly hot, asthenospheric plume material from the Afar hotspot to the south. However, other studies have championed shorter length-scale processes to explain

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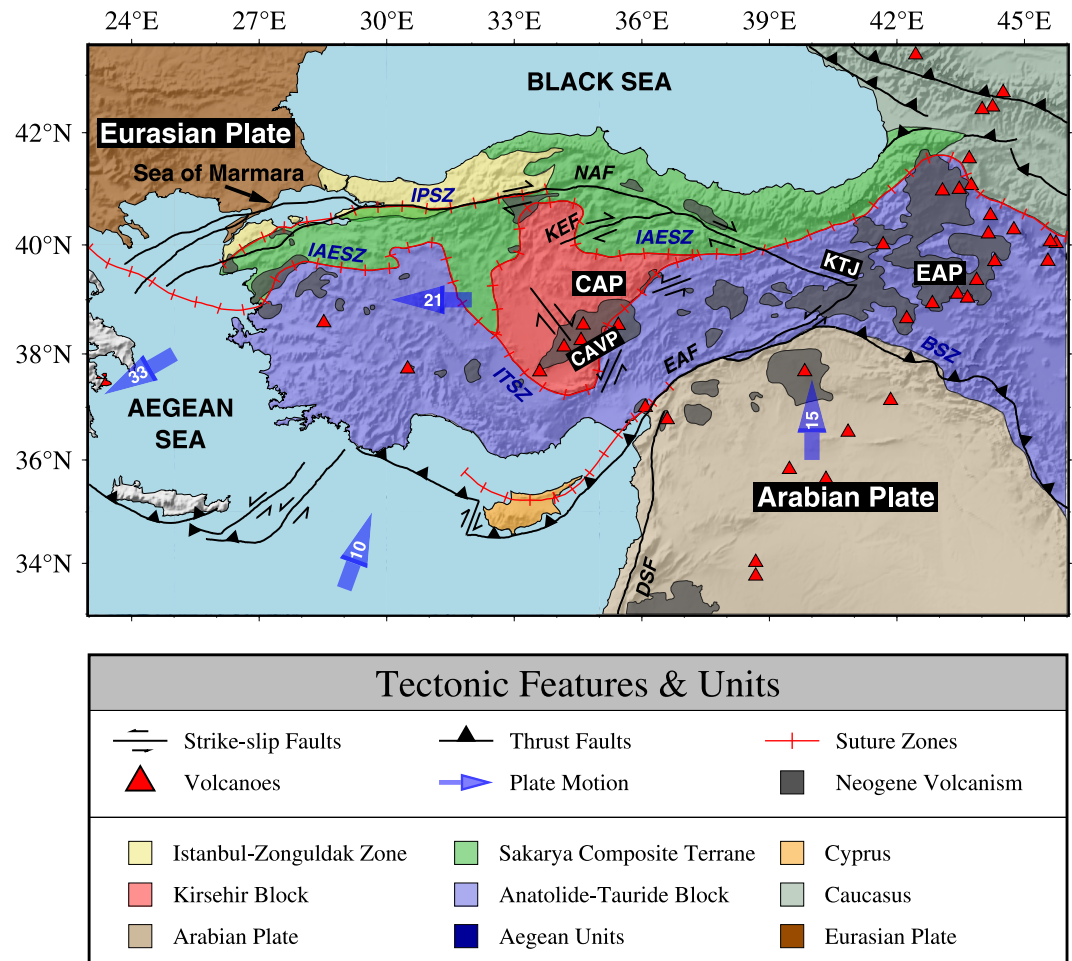


Figure 1. The tectonics of the Anatolian Plate (modified after Okay & Tüysüz, 1999). Blue arrows: plate velocities with respect to the Eurasian Plate in mm/yr (Reilinger et al., 2006). The Sakarya Composite terranes and the Istanbul-Zonguldak zone comprise the Pontides (e.g., Okay & Tüysüz, 1999). CAVP: Central Anatolian Volcanic Province; CAP: Central Anatolian Plateau; EAP: Eastern Anatolian Plateau; KTJ: Karlıova Triple Junction; NAF: North Anatolian Fault; EAF: East Anatolian Fault; KEF: Kirikkale-Erbaa Fault; DSF: Dead Sea Fault; BSZ: Bitlis Suture Zone; IAESZ: Izmir-Ankara-Erzincan Suture Zone; ITSZ: Inner-Tauride Suture Zone; IPSZ: Intra-Pontide Suture Zone. Volcanic data were compiled from multiple sources (Bertrand et al., 2003; Brew et al., 2001; Schildgen et al., 2014; Schleiffarth et al., 2018).

plateau uplift. For example, Göğüş and Pysklywec (2008) and Göğüş et al. (2017) follow a numerical modeling approach to champion the hypothesis of lithospheric delamination via Rayleigh Taylor instabilities below the EAP and CAP, respectively. Mantle seismic tomographic models for Anatolia illuminate complex upper-mantle heterogeneity, with evidence for slab gaps, horizontal tears, and lithospheric drips (e.g., Berk Biryol et al., 2011; Confal et al., 2020; Kounoudis et al., 2020; Portner et al., 2018), which SKS splitting studies of mantle seismic anisotropy demonstrate exert strong control on upper mantle convection patterns (Merry et al., 2021). Geological evidences for rapid uplift across parts of the Anatolian Plate have frequently cited these seismologically inferred observations of mantle flow as plausible mechanisms to explain rapid, recent, and localized uplift in parts of Anatolia (e.g., Cosentino et al., 2012; Racano et al., 2021; Radeff et al., 2017; Schildgen, Cosentino, Bookhagen, et al., 2012; Schildgen, Cosentino, Caruso, et al., 2012).

Key to evaluating different mechanisms for Anatolian plateau uplift is an accurate knowledge of Moho depth, since this exerts first-order control on estimates of residual topography (ϵ_{res}): the difference between observed elevation and that expected from crustal buoyancy alone (e.g., Gvirtzman et al., 2016; Lachenbruch & Morgan, 1990). However, a recent P-to-S receiver function study by Ogden and Bastow (2022) highlighted discrepancies between different studies of crustal thickness of ≥ 10 km below some seismograph stations on the

plate. These discrepancies suggest the need for further detailed constraints on crustal architecture to better understand the reasons for the Anatolian plateau's high elevations. Another purpose of this study is to investigate the lithospheric seismic structure across the Anatolian Plate for examining competing mechanisms. Large-scale mechanisms like a single plateau-wide lithospheric delamination or mantle plume scenario will yield relatively homogeneous upper-mantle velocity structure. In contrast, short-length scale mechanisms like lithospheric dripping and mantle upwelling through a slab tear and/or window will generate localized upper-mantle anomalies. To achieve these goals, we capitalized on the availability of teleseismic data from 575 broadband seismic stations from which we calculated P-to-S receiver functions for joint inversion with the recent 5–100 s-period surface wave model of Kaviani et al. (2020). Moho depths were determined by locating the steepest gradient at lower-crust to upper-mantle velocity range. Residual topography, upper-mantle velocity variations, and residual Bouguer gravity anomalies were then estimated based on seismically inferred Moho depths. Constrained by these observations of Moho depth, residual topography, and upper mantle velocity, we discuss the feasibility of competing proposed mechanisms, at different scales, responsible for the high elevations of the Anatolian Plate.

1.2. Tectonic Setting

The formation of the Anatolian Plate, which eventually concluded in the Oligocene, followed the closure of the Paleo- and Neo-Tethyan oceans and the subsequent collision between terranes on either side (e.g., Okay & Tüysüz, 1999; Robertson et al., 2012; Şengör & Yilmaz, 1981). These terranes of the Anatolian Plate are broadly sutured along two major lineaments: the Izmir-Ankara-Erzincan Suture Zone (IAESZ) and the Inner-Tauride Suture Zone (ITSZ; Figure 1). The IAESZ represents the closure of the northern branch of the Neo-Tethys ocean during Late Cretaceous-to-Paleogene times. It delineates the boundary between the Sakarya block (part of the Pontides) in the north, and Anatolide-Tauride and Kirsehir Blocks in the south (e.g., Okay & Tüysüz, 1999; Sarıfakıoğlu et al., 2017; Şengör & Yilmaz, 1981). The ITSZ separates the Anatolide-Tauride block from the Kirsehir Block in the heart of the Anatolian Plate (e.g., Parlak et al., 2013; Pourteau et al., 2010).

In the EAP, the continental collision between the Arabian and Anatolian plates initiated along the Bitlis Suture Zone during Eocene-Miocene times, signifying the onset of the diachronous closure of the southern branch of the Neo-Tethys ocean (e.g., Darin et al., 2018; Gülyüz et al., 2020; Hüsing et al., 2009; Rolland et al., 2012). However, beneath western Anatolia (WA), the subduction of Neotethyan oceanic lithosphere is ongoing along the Aegean Trench (Bocchini et al., 2018; Taymaz et al., 2007). Western Anatolia is undergoing N–S oriented extension at ~20 mm/yr (e.g., Aktug et al., 2009), as a consequence of the rollback of the subducting Aegean slab and/or Anatolian orogenic collapse (e.g., Jolivet, 2001; Jolivet et al., 2013; Schildgen et al., 2014; Seyitoğlu & Scott, 1996). Subduction below Cyprus and the CAP is slow (~18 mm/yr compared to ~35 mm/yr below the Aegean; e.g., Reilinger et al., 2006) and in the near-terminal stage.

The active tectonics of the Anatolian Plate are intimately linked to the evolution of major strike-slip faults and their interactions with surrounding plates. Following the mature continental collision between the Arabian and Anatolian plates since early Miocene (e.g., Darin et al., 2018), the Anatolian Plate began its westward extrusion and counter-clockwise rotation along the seismically active, dextral strike-slip North Anatolian Fault (NAF) and the sinistral strike-slip East Anatolian Fault (EAF; e.g. Reilinger et al., 2006), under the influence of the Arabian indentation as well as the slab pull of the Aegean slab (e.g., Faccenna et al., 2013).

The NAF developed westward after initiating 13–11 Ma in eastern Anatolia (e.g., Şengör et al., 2005). As the fault developed into the Sea of Marmara region, it split into two main strands displaying asymmetric strain distribution (e.g., Karabacak et al., 2022; Şengör et al., 2005). The total right-lateral displacement along the NAF relative to the Eurasian Plate is ~85 km with a slip rate of 26 mm/yr (e.g., Armijo et al., 1999; Hubert-Ferrari et al., 2002; Le Pichon & Kreemer, 2010; Sengör, 1979). The EAF marks the Arabia–Anatolia plate boundary. It has accommodated <33 km of left-lateral offset since the Late Miocene/Early Pliocene times (e.g., Arpat & Saroglu, 1972; Duman & Emre, 2013; M. Hempton, 1985; M. R. Hempton, 1987; Herece, 2008; Khalifa et al., 2020; Westaway & Arger, 1996; Whitney et al., 2023) with a GPS-derived slip rate of ~10 mm/year (Aktug et al., 2016; Reilinger et al., 2006). Near the southwestern section of the EAF, where it borders the Anatolian and Arabian plates, the deformation is dominated by compressive tectonics, manifesting as folds and thrusts (e.g., Lyberis et al., 1992; Yilmaz et al., 2006). In contrast, the interior of the Arabian Plate is largely seismically and tectonically inactive (e.g., Lyberis et al., 1992). West of the Karhova Triple Junction, where the EAF meets the

NAF (Figure 1), Sançar et al. (2015) identified a transpressional zone featured as oblique-reverse faults, compatible with high strain-rate compressional deformation (e.g., Kahle et al., 2000).

During the Middle Eocene to Oligocene, reduced magmatism occurred, possibly due to slab flattening (e.g., Schleiffarth et al., 2018), following the initial subduction of Arabian continental crust below the Anatolia/Eurasia Plate (e.g., Darin et al., 2018). Since the Early Miocene, mature Arabian-Anatolian collision has been accompanied by enhanced magmatism within the EAP and Central Anatolian Volcanic Province (CAVP), with distinct spatial and temporal patterns (Darin et al., 2018; Schleiffarth et al., 2018). CAVP magmatism is characterized by a NE-SW-trending distribution, while the EAP exhibits spatially diffuse and episodic Neogene-to-Quaternary volcanism. The potential role of mantle dynamics in driving these magmatic patterns is discussed in Section 1.3.4.

1.3. Previous Geophysical and Geodynamic Constraints

1.3.1. Crustal Thickness

The teleseismic receiver function H- κ stacking method of L. Zhu and Kanamori (2000) has been used to calculate crustal thickness and bulk crustal V_p/V_s ratio, both on a plate-wide (e.g., Ogden & Bastow, 2022; Tezel et al., 2013; Vanacore et al., 2013) and local scale (e.g., Abgarmi et al., 2017; Frederiksen et al., 2015; Karabulut et al., 2013; Özacar et al., 2010; L. Zhu et al., 2006). Additional constraints on Moho depth have been derived from receiver function common conversion point stacking (e.g., Abgarmi et al., 2017; Jenkins et al., 2020; Kahraman et al., 2015; Özacar et al., 2010; Tezel et al., 2013; L. Zhu et al., 2006), and inversion of receiver functions (e.g., Alkan et al., 2020; Karabulut et al., 2019; Tezel et al., 2013; Zor et al., 2003), including their joint inversion with surface wave dispersion data (e.g., Gök et al., 2007, 2011). From these studies, a consensus has emerged that Moho depths increase from ~25 km in WA to ~45 km in the EAP and that crustal thickening alone cannot explain the high elevations of the CAP and EAP (Özacar et al., 2010; Ogden & Bastow, 2022; Zor et al., 2003). However, in their recent assessment of Anatolian bulk crustal structure using a modified H- κ stacking approach (Ogden et al., 2019), Ogden and Bastow (2022) noted differences among earlier studies of up to 21 km in Moho depth at some stations.

1.3.2. Lithospheric Thickness Variations

P-wave receiver function analysis has the disadvantage that P-to-S conversions from the base of the lithosphere are expected to arrive within the same time window as crustal reverberations associated with the Moho P-to-S conversion phase. To overcome this problem, Anatolian Plate thickness has been constrained using teleseismic S-to-P receiver functions (e.g., Angus et al., 2006; Erduran et al., 2022; Hua et al., 2020; Kind et al., 2015), in which the lithosphere-asthenosphere boundary (LAB) phase arrives prior to S-to-P crustal phases, but after P-to-S converted energy. These studies indicate the Anatolian LAB is relatively shallow, at 60–100 km depth. For example, Kind et al. (2015) observed a relatively homogeneous LAB in the depth range 80–100 km across Anatolia; Hua et al. (2020), using an improved common conversion point stacking approach that included the physics of seismic scattering, reported depths of 60–90 km.

Anatolian Plate thickness has also been studied using joint inversion of receiver functions and surface waves (e.g., Delph et al., 2017; Erduran et al., 2022; Vinnik et al., 2014). In this approach, the body-wave conversions and surface wave dispersion data are inverted jointly to generate a 1D velocity model for a given seismic station. The LAB depth is then inferred via a velocity proxy (e.g., the steepest gradient, a specific shear-wave velocity value) that proceeds on the assumption that the lithosphere is a seismically fast wavespeed lid atop a slow wavespeed asthenosphere. Where successful, this approach has led to plate thickness estimates of <50 km in the northwestern margin of the Arabian Plate, and ~80 km in the northern Kirsehir Block and Eastern Taurides Mountains (Delph et al., 2017). However, at many stations across the Anatolian Plate, no single, unambiguous depth can be gleaned from the often-complex 1D velocity models in the region (e.g., Delph et al., 2017; Vinnik et al., 2014), leading to speculation that the lithospheric mantle is either negligible or absent below parts of the plate (e.g., Delph et al., 2017).

Thermal methods for LAB depth determination have yielded somewhat different results to seismological studies. For example, Artemieva and Shulgin (2019) employ a thermal approach to estimate a plate thickness of >125 km below the northern CAP and the eastern Tauride mountains. This contrasts sharply with <80 km thickness by Delph et al. (2017) through joint inversion of seismic data in the same region. Artemieva and Shulgin (2019) also

propose an 80–140 km thick lithosphere underlying the EAP. Seismic tomographic studies, however, reveal ubiquitously slow seismic wavespeeds in the upper mantle, interpreted as warm asthenospheric material below a thin lithosphere (e.g., Kounoudis et al., 2020; Portner et al., 2018).

1.3.3. Upper Mantle Anomalies

At ~100 km depth, slow wavespeeds in both body and surface wave tomographic models for the Anatolian upper mantle corroborate the conclusion drawn by the aforementioned receiver function studies that the Anatolian lithosphere is thin (e.g., Delph et al., 2017; Kounoudis et al., 2020; Zor, 2008). Kounoudis et al. (2020) converted seismic wavespeeds to temperature to reveal thermal conditions at 100 km depth that are consistent with convecting, not conductive mantle.

At greater depths, complex fast wavespeed anomalies have variously been interpreted as torn and/or broken-off oceanic lithosphere, and fragments of delaminated continental lithosphere (e.g., Berk Biryol et al., 2011; Delph et al., 2015, 2017; Kounoudis et al., 2020; Lei & Zhao, 2007; Portner et al., 2018; Salaün et al., 2012). For example, slow velocity anomalies separating fast wavespeed zones near ~29°E are readily interpreted as a gap/window between the Aegean and Cyprus slabs (Bakırcı et al., 2012; Berk Biryol et al., 2011; Kounoudis et al., 2020; Portner et al., 2018; Salaün et al., 2012). A horizontal tear in the Cyprus slab has also been imaged tomographically, with upwelling asthenospheric mantle likely penetrating the gap (Berk Biryol et al., 2011; Kounoudis et al., 2020; Portner et al., 2018). Apart from intricate high-velocity anomalies, continuous negative polarity common conversion point stacking amplitudes at 90–150 km depth are also observed (Hua et al., 2020, 2023), interpreted by Hua et al. (2023) as an indication for partially molten asthenosphere below the entire Anatolian Plate in response to the mantle flow from the Afar hotspot to the north. Although Hua et al. (2023) acknowledge that regional-scale processes (e.g., dripping or lithospheric delamination) may be impacting their results, they lack the short length-scale resolution to examine such hypotheses accurately.

1.3.4. Mechanisms for Plateau Uplift

Mechanisms of uplift for the CAP and EAP can be broadly divided into plateau-wide (e.g., Bartol & Govers, 2014) and localized uplift mechanisms (e.g., Göğüş et al., 2017). For example, following a geodynamic modeling approach, Bartol and Govers (2014) explained CAP and EAP uplift via a single delamination event of the continental mantle lithosphere in the Early Miocene, following the horizontal subduction and southward rollback of the northern Neotethys slab. Şengör et al. (2003) associated the topographic elevations of the EAP to the steepening and break-off of oceanic lithosphere, with subsequent in-flow of hot, buoyant asthenosphere. This scenario is supported by roughly north-to-south trends in the geochemical signature of young igneous rocks in the region (Keskin, 2003). Similarly, Schleiffarth et al. (2018) have attributed the roughly NE-SW trend of CAVP magmatism to Cyprus slab rollback and subsequent mantle upwelling. These hypotheses also accord with the region's slow upper mantle (e.g., Berk Biryol et al., 2011; Delph et al., 2015, 2017; Kounoudis et al., 2020; Salaün et al., 2012) and the occurrence of isolated high velocity anomaly bodies (e.g., Kounoudis et al., 2020; Lei & Zhao, 2007; Portner et al., 2018). Instead of associating Anatolian Plateau uplift to the loss of oceanic or continental lithosphere, some researchers have championed a hypothesis of northward flowing, hot, buoyant mantle plume material from the Afar hotspot to the southeast, aided by Aegean slab rollback (e.g., Ershov & Nikishin, 2004; Faccenna et al., 2013; Hua et al., 2023). This hypothesis is corroborated by the styles and spatial/temporal distribution of volcanism in western Arabia (e.g., Ershov & Nikishin, 2004; Hua et al., 2023) and the intraplate signature of magmatism (e.g., Nikogosian et al., 2018).

Localized tectonic processes (as opposed to plateau-wide mechanisms), such as lithospheric dripping or asthenospheric upwelling through slab tears, are also hypothesized to have played a role in both the uplift and volcanic activity during the tectonic evolution of the Anatolian Plate. Geodynamic modeling of Rayleigh-Taylor gravitational instabilities by Göğüş et al. (2017), for example, shows that dripping tectonics due to gravitational instability following folding of a Neotethyan arc may result in localized, Late Miocene–Recent uplift within the CAP. This mechanism may also account for the spatiotemporally heterogeneous distribution of EAP magmatism in Neogene times. In addition, rapid uplift of the southern margin of the CAP since the Middle Pleistocene has been attributed to various hypotheses, including localized slab break-off and/or tears along the Cyprus subduction zone (Cosentino et al., 2012; Racano et al., 2020, 2021; Radeff et al., 2017;

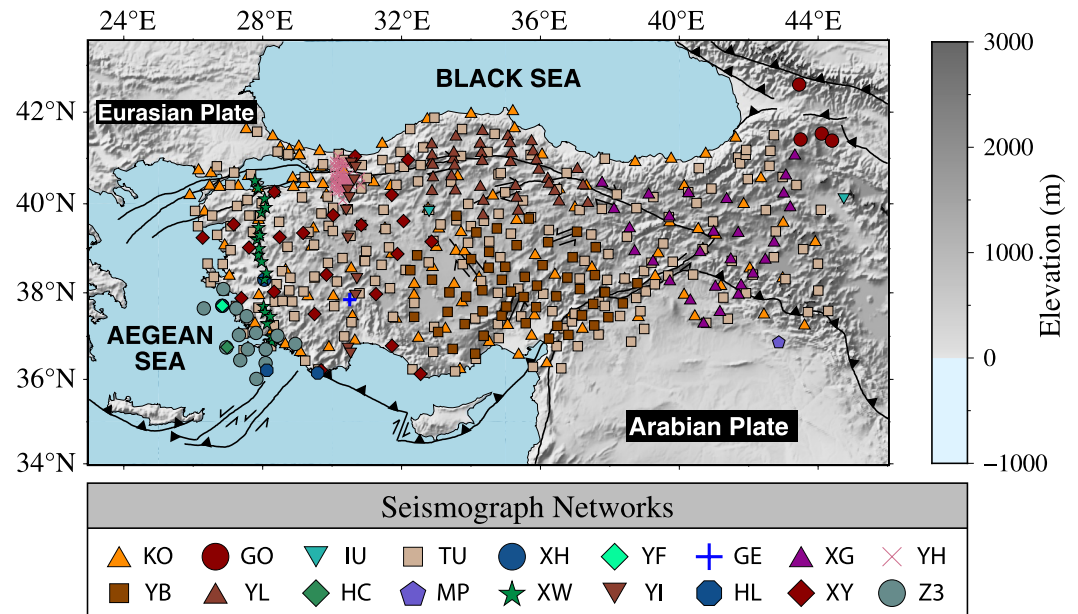


Figure 2. Seismograph stations and networks. Different symbols with different colors indicate stations from various seismograph networks. Bold black lines: major active faults. See Open Research Section for network details.

Schildgen, Cosentino, Bookhagen, et al., 2012; Schildgen, Cosentino, Caruso, et al., 2012; Öğretmen et al., 2018), and the underthrusting of African continental lithosphere beneath the central Anatolide-Tauride Block (e.g., McPhee & van Hinsbergen, 2019).

2. Method and Data

2.1. Data and Network

This study utilizes teleseismic data from 575 broadband seismograph stations across 18 permanent and temporary networks that have operated during the period 1994–2023 (Table S2; Figure 2). Teleseismic data were acquired through repositories including AFAD, EIDA, GEOFON, KOERI, IRIS DMC, ORFEUS, and RESIF. These stations provide dense sampling of the Anatolian Plate (Figure 2). Only seismograms from earthquakes of magnitude >4 in the epicentral distance range of 30° – 90° were used in the following processing (Figure 3). Prior to calculating receiver functions, seismograms were Butterworth bandpass filtered with corner frequencies of 0.04 and 3 Hz. Earthquake-station pairs were accepted for further analysis if the P-wave signal-to-noise ratio was >2 (e.g., Crotwell & Owens, 2005; Leahy et al., 2012; Ogden & Bastow, 2022; Xu et al., 2007). Subsequently, the seismograms were windowed from 20 s before to 100 s after the P-wave arrival, then rotated from the Z-N-E (Vertical-North-East) to the Z-R-T (Vertical-Radial-Transverse) coordinate system.

2.2. Receiver Functions and Surface Wave Dispersion Curves

P-wave radial and tangential receiver functions were calculated using the time domain iterative deconvolution method of Ligorria and Ammon (1999) with two different Gaussian filter factors: $\alpha = 1.0$ and $\alpha = 2.5$, corresponding to frequencies of $f \leq 0.5$ Hz and $f \leq 1.2$ Hz, respectively. Receiver functions with an iterative deconvolution variance (the cross-correlation of the original radial component with a recalculated radial component derived from the

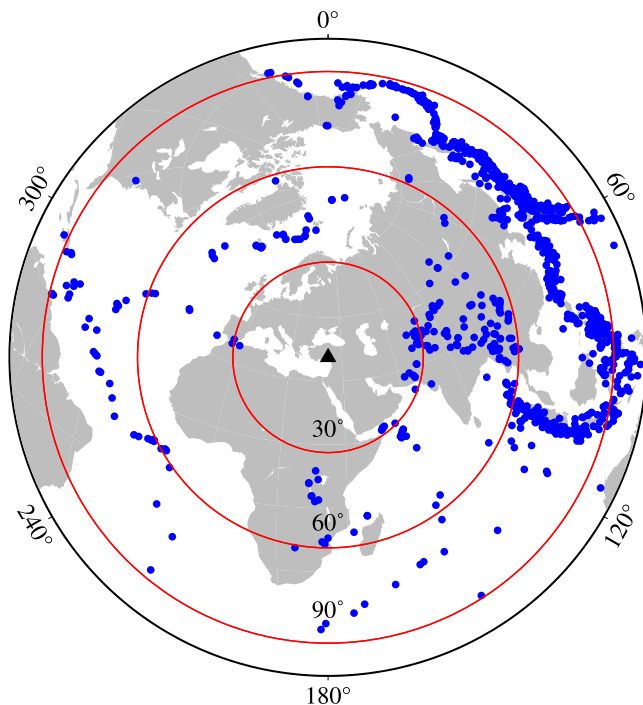


Figure 3. Locations of earthquakes (blue dots) utilized in this study. Red circles: 30° intervals from the center of the study area (black triangle). Epicentral distances are referenced to the center of the study area, so some earthquakes fall within 30° or beyond 90° .

receiver function) of $\leq 80\%$ were eliminated from the analysis, resulting in a $\sim 5\%$ reduction in receiver functions. Ultimately 36,254 receiver functions were selected for subsequent processing. To enhance the signal-noise-ratio (SNR), we stacked receiver functions based on their backazimuth (10° intervals) and epicentral distance (ray parameter intervals of 0.02 s/km). For each bin, we formed a receiver function trace defined by the upper values of the 68% confidence intervals of the stack. If the Spearman correlation coefficient (SCC) between this trace and the averaged receiver function of this bin exceeds a pre-determined threshold (0.65 when the Gaussian factor is 1.0; 0.5 when it is 2.5), this averaged receiver function is accepted. Subsequently, we stacked all grouped receiver functions for a given station to generate a reference receiver function and then calculated SCC between each individual stacked trace with this reference receiver function to obtain the mean SCC. Stations with a low mean SCC may indicate significant subsurface heterogeneity or anisotropy, rendering their data unsuitable for velocity structure inversion. Stations with backazimuth coverage of $< 50\%$ were also excluded from the final data set. The final receiver function data set comprised 9,829 grouped receiver functions from 486 seismograph stations.

Surface wave dispersion curves for the period range 5–100 s were sourced from the tomographic study of Kaviani et al. (2020), which combined Rayleigh wave group velocities and ambient noise measurements across the Anatolian Plate. Checkerboard tests in their study showed that an anomaly size of $1^\circ \times 1^\circ$ is well recovered across Anatolia (Kaviani et al., 2020). The full range of periods made available by Kaviani et al. (2020) was considered in our analysis to maximize the depth sensitivity of our study.

2.3. Joint Inversion Procedure

Inversion of receiver functions in isolation can lead to unrealistic velocity structures because of the strong trade-off between subsurface velocity and interface depth (e.g., Ammon et al., 1990). Conversely, surface wave dispersion data are sensitive to absolute wavespeeds, but lack the power to resolve abrupt velocity contrasts with depth. The joint inversion scheme developed by Julià et al. (2000) harnesses the complementary resolving power of receiver functions and surface wave dispersions to constrain S-wave velocity structure below a seismograph station (e.g., Bao et al., 2015; Dugda et al., 2007; Gilligan et al., 2014; Nemocón & Julià, 2021; Tang et al., 2022). Receiver functions in the two aforementioned frequency bands ($f \leq 0.5$ Hz and $f \leq 1.2$ Hz) are utilized in the joint inversion procedure. The purpose of overlapping these bands at low frequencies is to accentuate the significance of first-order structural features in the joint inversion. We windowed receiver functions from 4 s before to 16 s after the first arriving P-wave phase to ensure that both the receiver function and surface wave dispersion data sets have sensitivity to structures primarily at ≤ 120 km depth.

Because vertical resolution decreases with increasing period, we parameterized the starting model using a set of homogeneous horizontal layers with fixed velocity and thickness, but with thicker layers at greater depth. We adopted a layer thickness of 2 km in the upper 50 km, a 5 km layer thickness at 50–120 km depth, and a 10 km layer thickness at 120–400 km depth. Starting velocities were based on a modified ak135 velocity model (Table S1 in Supporting Information S1), but with a V_P/V_S ratio in the uppermost part of the model adopted from the H- κ stacking analysis of Ogden and Bastow (2022). The starting model employs a nearly constant velocity structure in the uppermost 120 km to avoid imposing predefined discontinuities in final S-wave profiles (Figure 4a). This approach ensures that all discontinuities are solely constrained by our data. In the joint inversion, depth-dependent weights were applied: layers shallower than 90 km were assigned relatively higher weights (> 0.1), while deeper layers were assigned lower weights (0.1 for 90–240 km, and 0.05 for 240–400 km depth). This strategy allows more flexibility in the shallow structure while limiting the deviation from the start model for the deeper model. It also prevents overfitting of the data at poorly resolved depths and avoids introducing artificial discontinuities.

Two primary parameters exert significant control over the final S-wave velocity structure for a given station: (a) the damping factor controls the balance between the model roughness and data fitting; (b) the influence factor determines the relative importance of the receiver functions and surface wave dispersion data sets (an influence parameter of 0 would imply 100% receiver function weighting; an influence factor of 1 implies 100% surface wave weighting). Following Julià et al. (2000); Nemocón & Julià (2021), in the case of station ANTO as an example, we first varied the damping factor from 0.05 to 2.0 with a fixed influence factor of 0.5. The preferred damping factor was chosen based on the shortest distance between the data point and the origin (Figure 5). Subsequently, with the selected damping factor of 0.15, the influence factor was varied between 0.05 and 0.95 and the optimal value was chosen from the elbow of the trade-off curve (Figure 5).

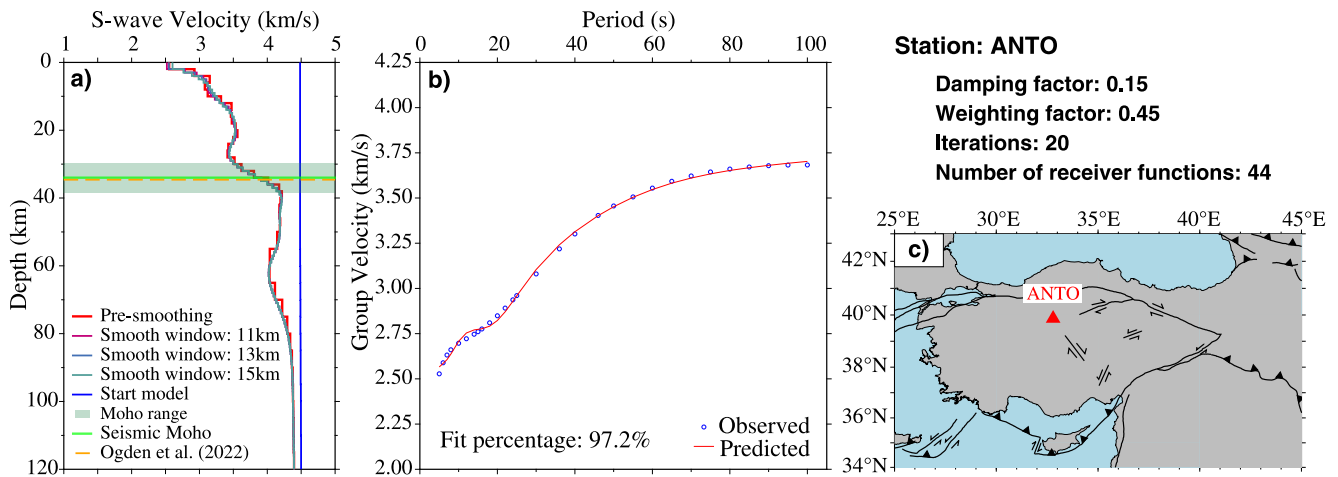


Figure 4. Joint inversion result for station ANTO. (a) 1D shear velocity profile (red line). The Moho range is 2.5 times the Moho depth uncertainty. (b) Rayleigh wave group velocities from Kaviani et al. (2020) (blue circles) at different period, and the predicted dispersion curve (red line) calculated from our joint inversion result. Associated observed and predicted receiver functions at different Gaussian factors and backazimuths are shown in Figure S1 in Supporting Information S1. (c) The location of station ANTO (red triangle).

Following the acquisition of 1-D velocity profiles beneath each seismic station, we employed seismic velocity proxies to determine the Moho depths beneath the Anatolian Plate. We performed spline interpolation to refine the velocity profiles, using a depth step of 1 km. Additionally, we applied a Savitzky-Golay filter (Savitzky & Golay, 1964) with various sliding window sizes (11, 13, 15 km) to smooth the velocity profiles and to preserve long-wavelength signals (Figure 4a). Subsequently, slopes of the segmented velocity profile were calculated via linear regression, utilizing a sliding window of 3, 5, and 7 km along the entire profile. The center of the window with the lowest slope, where the velocity ranges between 3.4 km/s and 4.2 km/s, was regarded as the Moho (Figure 4a). All combinations of window sizes for smoothing and linear regression were tested, and only stations with a stable Moho signal were retained for further analysis. After applying this procedure to all velocity profiles, crustal thicknesses were obtained for 399 stations. Uncertainties in the Moho depth were estimated as the depth variation when velocity at the Moho decreased by 0.2 km/s. More examples of joint inversion and trade-off analysis can be found in Figures S2–S16 in Supporting Information S1.

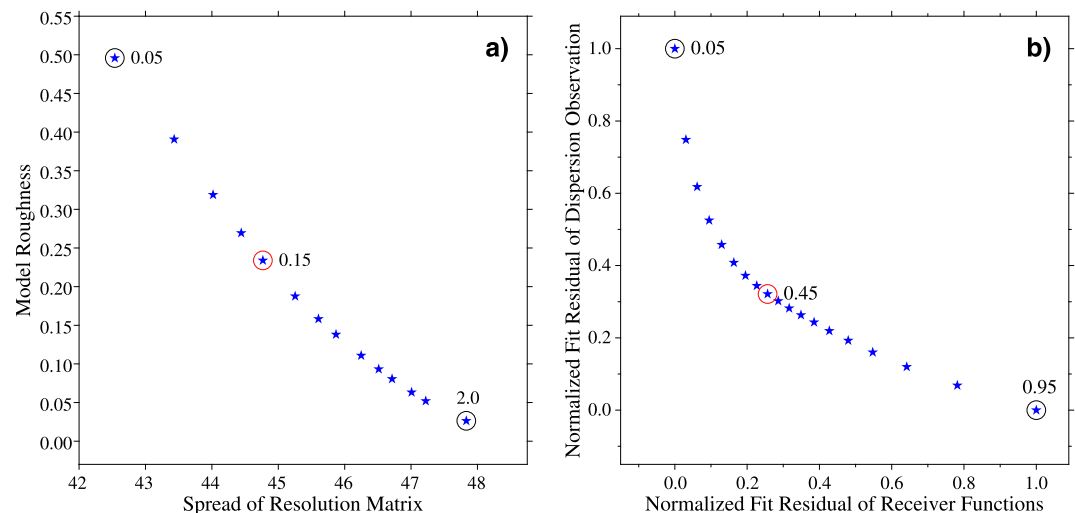


Figure 5. Trade-off analysis for (a) the damping factor and (b) influence factor for station ANTO. Red circle: the optimal inversion regularization parameters. Black circle: the starting and end points of the trade-off analysis.

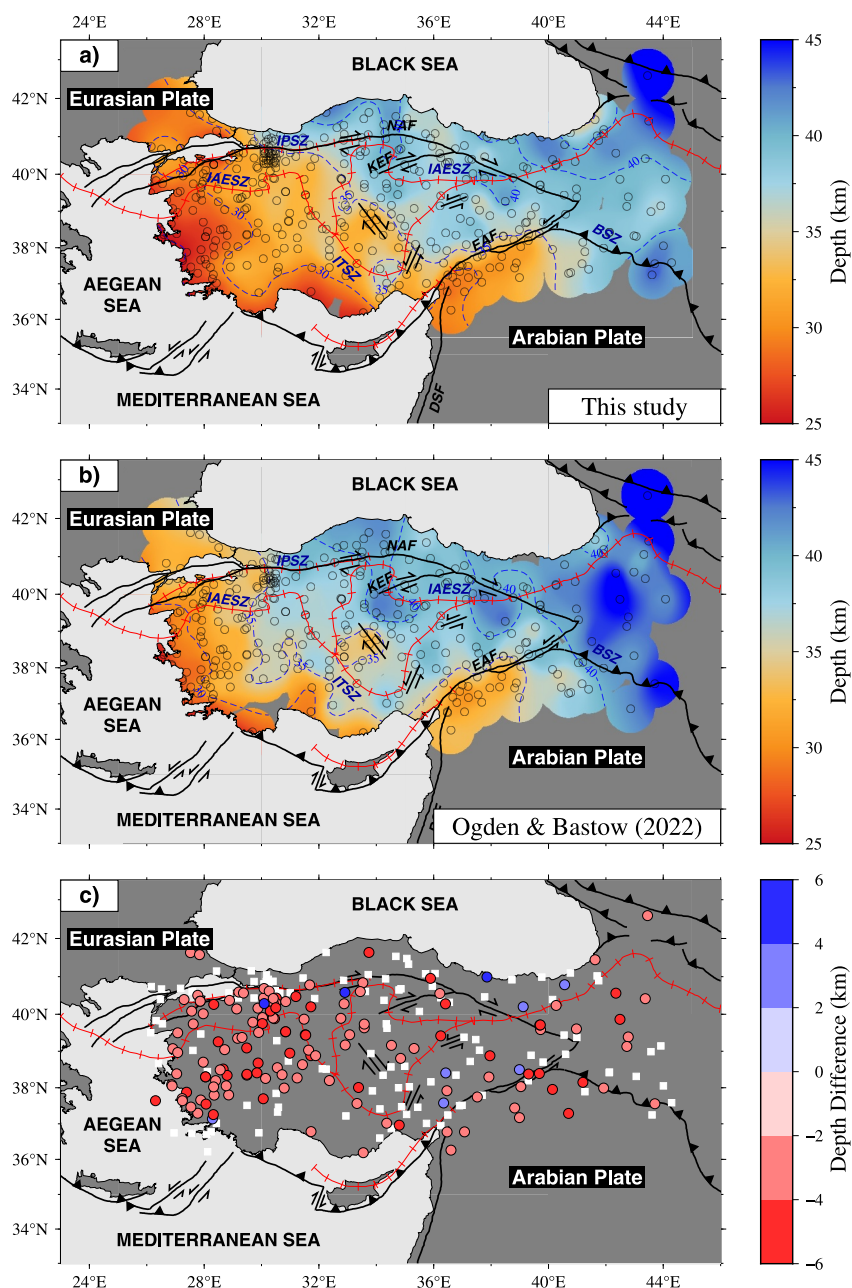


Figure 6. (a) Interpolated Moho depth map based on our joint inversion work. (b) Interpolated Moho depth map from the receiver function H- κ analysis of Ogden and Bastow (2022). In (a, b), open circles mark the locations of seismograph stations with reliable Moho depth estimates from this study and from Ogden and Bastow (2022), respectively. (c) Moho depth difference by subtracting the Moho depth of Ogden and Bastow (2022) from the Moho depth in our study (only stations with Moho depth differences >2 km are shown). White squares indicate locations of seismograph stations where Ogden and Bastow (2022) present no constraints. NAF: North Anatolian Fault; EAF: East Anatolian Fault; KEF: Kikkale-Erbaa Fault; DSF: Dead Sea Fault; BSZ: Bitlis Suture Zone; IAESZ: Izmir-Ankara-Erzincan Suture Zone; ITSZ: Inner-Tauride Suture Zone; IPSZ: Intra-Pontide Suture Zone.

3. Results

3.1. New Moho Depth Map for Anatolia

Generally, Moho depths derived from joint inversion increase from ~ 25 km in WA to ≤ 44 km below the EAP (Figure 6a), consistent with previous studies (e.g., Abgarmi et al., 2017; Karabulut et al., 2019; Özacar

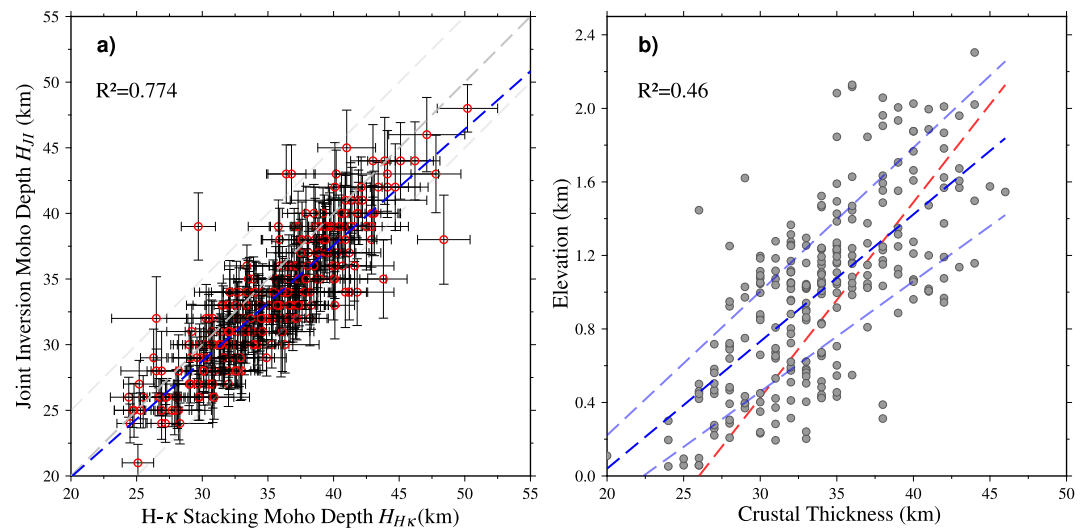


Figure 7. (a) Correlation analysis between the Moho depths derived from our joint inversion analysis (H_{JI}) and from the H- κ stacking study of Ogden and Bastow (2022), $H_{H\kappa}$. The blue dashed line of best fit is defined by $H_{JI} = -0.883H_{H\kappa} + 2.249$. Gray dashed lines: 1:1 reference line and ± 5 km offsets; (b) crustal thickness versus elevation for stations in the Anatolian Plateau (excluding the Pontides in the north) within the region 37–41°N and 27–45°E. Deep blue line is the best-fit line to the data, $\epsilon = 0.0692H - 1.343$, with an R^2 value of 0.46, by minimizing the squared differences between the observed and predicted elevation. Light blue dashed lines represent the 95 percent confidence intervals of the blue dashed line. Red dashed line from Karabulut et al. (2019) is also plotted, indicating the same relationship between the crustal thickness and topography.

et al., 2010; Ogden & Bastow, 2022; Tezel et al., 2013; Vanacore et al., 2013; Zor et al., 2003). Overall, our Moho depths match those obtained from the H- κ stacking study of Ogden and Bastow (2022) (mean difference ≈ 2.4 km; Figures 6 and 7a), with the largest discrepancies occurring in WA (similarly, comparison with the earlier study of Karabulut et al. (2019) reveals a mean difference of ~ 2.9 km; see Figure S17 in Supporting Information S1). This discrepancy is readily explained by the crustal P-wave velocity assumed in the H- κ stacking study. The average crustal velocity calculated from our results is 3.37 km/s, comparable to values reported by previous studies (e.g., Delph et al., 2015; Govers & Fichtner, 2016). Assuming a global average continental bulk-crustal V_P/V_S of 1.768 (Christensen, 1996), this translates to a P-wave velocity of 5.95 km/s, which is below the 6.2–6.8 km/s range explored by Ogden and Bastow (2022) (mean $V_P = 6.46$ km/s). A decrease of 0.5 km/s in assumed V_P , would have decreased the Ogden and Bastow (2022) H- κ stacking-derived crustal thicknesses by ~ 2.7 km, thus explaining most of the discrepancy observed in Figure 6c.

Across most of the NAF, a discernible N–S change in Moho depth is lacking, except for a section from $\sim 29.5^\circ\text{E}$ to 31°E in the western Pontides, where a Moho step, from <35 km south of the Intra-Pontide Suture Zone (IPSZ) to >35 km in the north is observed (Figure 6a). The Moho step is consistent with previous studies (e.g., Frederiksen et al., 2015; Jenkins et al., 2020; Kahraman et al., 2015; Taylor et al., 2016), corroborating the view that a lithospheric-scale, narrow shear zone characterizes the NAF's northern branch (Kahraman et al., 2015; Taylor et al., 2016), which likely developed along the weak IPSZ (Delph et al., 2015; Frederiksen et al., 2015; Şengör et al., 2005). In contrast, an obvious Moho step is observed across the EAF, where Moho depths increase from <35 km below the northwest tip of the Arabian Plate to 35–40 km below the Anatolian Plate.

3.2. Residual Topography (ϵ_{res})

Following the acquisition of the Moho depths presented in Figure 8a, we next use them to address the enigmatic issue of Anatolian plateau uplift by calculating residual topography. Residual topography (ϵ_{res}), is typically viewed as the difference between the mean elevation above sea level (ϵ) and the contribution to the elevation only considering crustal buoyancy in the asthenosphere (h_c ; Komut, 2015):

$$\epsilon_{res} = \epsilon - (h_c - h_0), \quad (1)$$

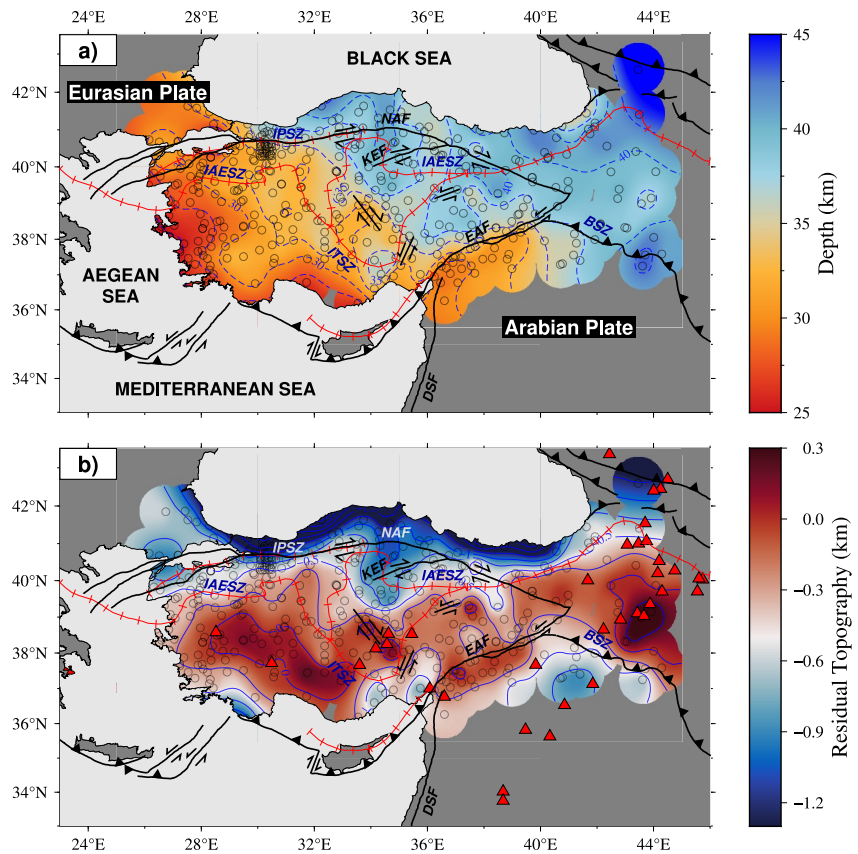


Figure 8. (a) Interpolated Moho depth map based on joint inversion results. (b) Residual topography map calculated from Moho depths. Open circles: the locations of seismograph stations with reliable Moho depth estimates. Red triangles: volcanoes that have erupted during the Quaternary. NAF: North Anatolian Fault; EAF: East Anatolian Fault; KEF: Kirikkale-Erbaa Fault; DSF: Dead Sea Fault; BSZ: Bitlis Suture Zone; IAESZ: Izmir-Ankara-Erzincan Suture Zone; ITSZ: Inner-Tauride Suture Zone; IPSZ: Intra-Pontide Suture Zone.

where h_0 is a constant, rendering sea level the reference datum. Lachenbruch and Morgan (1990) used quantities from the well-studied mid-ocean ridge systems to estimate h_0 as 2.4 km. The component of elevation attributed to crustal buoyancy is defined as:

$$h_c = \frac{1}{\rho_a}(\rho_a - \rho_c)H_c, \quad (2)$$

where ρ_a and ρ_c are the average density of the asthenosphere and continental crust ($\rho_a = 3,200 \text{ kg/m}^3$; $\rho_c = 2,850 \text{ kg/m}^3$), respectively. H_c is the crustal thickness. Following Yu et al. (2016), for a typical 35 km-thick continental crust (e.g., Molnar et al., 1993) in Airy-type isostatic equilibrium at zero elevation, the expected ϵ_{res} would be approximately -1.4 km . If that is entirely attributed to the negative buoyancy of mantle lithosphere, its thickness would be $\sim 90 \text{ km}$ with the density taken as $3,250 \text{ kg/m}^3$. An observation of $\epsilon_{res} < -1.4 \text{ km}$ may imply a thicker mantle lithosphere and/or convective draw-down of the surface. Observation of $\epsilon_{res} > -1.4 \text{ km}$ could imply an uplift contribution from a thinned mantle lithosphere and/or a buoyant asthenospheric mantle.

Figure 8b presents our new residual topography map for Anatolia. Residual topography across the Anatolia Plate ranges from -1.1 to $+0.3 \text{ km}$, implying the region's high elevations are, at least in part, the result of thinning of negatively buoyant mantle lithosphere, and/or a buoyant asthenosphere. We tested the sensitivity of our residual topography maps to variations in crustal density: the spatial variations in ϵ_{res} are consistent for crustal densities of $2,830 \text{ kg/m}^3$ (the continental global mean; Christensen & Mooney, 1995) and $2,900 \text{ kg/m}^3$ (Artemieva & Shulgin, 2019; see Figure S18 in Supporting Information S1). The EAP and northwestern tip of the Arabian Plate consistently exhibit $\epsilon_{res} > -0.5 \text{ km}$, while the region as a whole requires a mantle contribution to uplift. However,

we cannot preclude the possibility that some of the variation in residual topography may result from not incorporating heterogeneous crustal densities.

In agreement with Ogden and Bastow (2022), we find a poor correlation between Moho depth and elevation (Figure 7b), and thus corroborate their conclusion that crustal thickness variations alone cannot explain variations in Anatolia's topography. Our results provide evidence for relatively short length-scale changes in the importance of crust versus mantle contributions to uplift across the region. This contrasts, for example, the findings of Karabulut et al. (2019), who concluded that a close correlation between crustal thickness and elevation across much of Anatolia implied an isostatic control on topography, albeit with an additional, plate-wide contribution from a low-density uppermost mantle.

3.3. Residual Bouguer Gravity Anomaly

Our new crustal thickness map provides the opportunity to better inform discussions of the causes of elevated residual topography via calculation of regional Bouguer gravity anomalies, which reflect lateral heterogeneities in crust-mantle density contrasts. To forward model the gravity response of our Moho depth model, we parametrized the region 25.2°–45°E and 35.5°–43.3°N using a grid of prisms, spaced 0.1° in latitude and longitude. The vertical component of the gravitational response of these prisms at sea level was calculated following Nagy et al. (2000) and Fukushima (2020). We assumed a reference column under isostatic equilibrium with a crustal thickness of 35 km and a mean elevation at sea level (e.g., Molnar et al., 1993). A layer with a density of -448 kg/m^3 (a global average crust-mantle density contrast; Sjöberg & Bagherbandi, 2011) was assigned from 35 km depth to the Moho where the crustal thickness is $<35 \text{ km}$. Conversely, for Moho depths $<35 \text{ km}$, a layer with a density of $+448 \text{ kg/m}^3$ was applied from the base of the prism to 35 km depth. Residual Bouguer anomalies were then calculated by subtracting the gravity response of our model from the WGM2012 global Bouguer gravity anomaly model of Balmino et al. (2012) to evaluate the lateral variation of crust-mantle density contrasts at Moho depths (Figure 9b and Figure S19 in Supporting Information S1).

Following $\Delta g = 2\pi\rho GH_c$ (Δg is residual Bouguer anomaly; G is the gravitational constant; H_c is crustal thickness derived from joint inversion), the large contrast in residual Bouguer gravity anomalies (Figure 9c) between WA ($\sim -33.6 \text{ mGal}$) and the central Pontides ($\sim +133.4 \text{ mGal}$) would require a crust-mantle density contrast of $\sim 114.4 \text{ kg/m}^3$ —too large to be accounted for solely due to lateral variations in crustal density (e.g., Mooney et al., 2023). Variations in mantle density therefore likely exert strong control on the anomalies. Accordingly, a clear correlation is observed between the residual Bouguer gravity anomalies and average uppermost-mantle velocity, calculated as the average in a 50 km depth range starting 5 km below the Moho (Figures 9a and 9b). For example, negative residual Bouguer anomalies (Figure 9b) and slow uppermost-mantle wavespeeds ($<4.1 \text{ km/s}$, Figure 9a) in extensional WA, the southern CAP and the EAP suggest the presence of a buoyant asthenosphere and/or thin mantle lithosphere. In contrast, relatively fast uppermost-mantle wavespeeds and positive residual Bouguer anomalies of ($>50 \text{ mGal}$) in the Pontides, the northern CAP and northeastern Arabian plate reflect a denser and/or cooler mantle, likely associated with a thicker mantle lithosphere.

4. Discussion

4.1. Western Anatolia

Western Anatolia's thin crust ($\sim 25 \text{ km}$; Figure 8a) is consistent with the region's 25 Ma–Recent extensional tectonics. However, the lowest uppermost mantle wavespeeds (Figures 9a and 10a) are found somewhat inland from the Aegean Sea's east coast; residual topography is also higher (Figure 8b), and residual Bouguer gravity anomalies are more negative (Figures 9b and 9c) inland in WA. A simple explanation for these observations comes from seismic tomographic studies of the Anatolian upper-mantle (e.g., Berk Biryol et al., 2011; Kounoudis et al., 2020; Portner et al., 2018), and from studies of upper mantle anisotropy (e.g., Confal et al., 2020; Merry et al., 2021; Paul et al., 2014) that reveal strong horizontal mantle flow through a gap between the Aegean and Cyprus slabs. This buoyant mantle material is expected on the basis of petrological studies, which document a transition in WA from subduction zone-style volcanism to intraplate-style volcanism, with increased asthenospheric source, since ca. 10 Ma (e.g., Dilek & Altunkaynak, 2010; McNab et al., 2018). Western Anatolia's elevation is thus partially supported by the complex asthenospheric flow induced by the region's complex slab architecture.

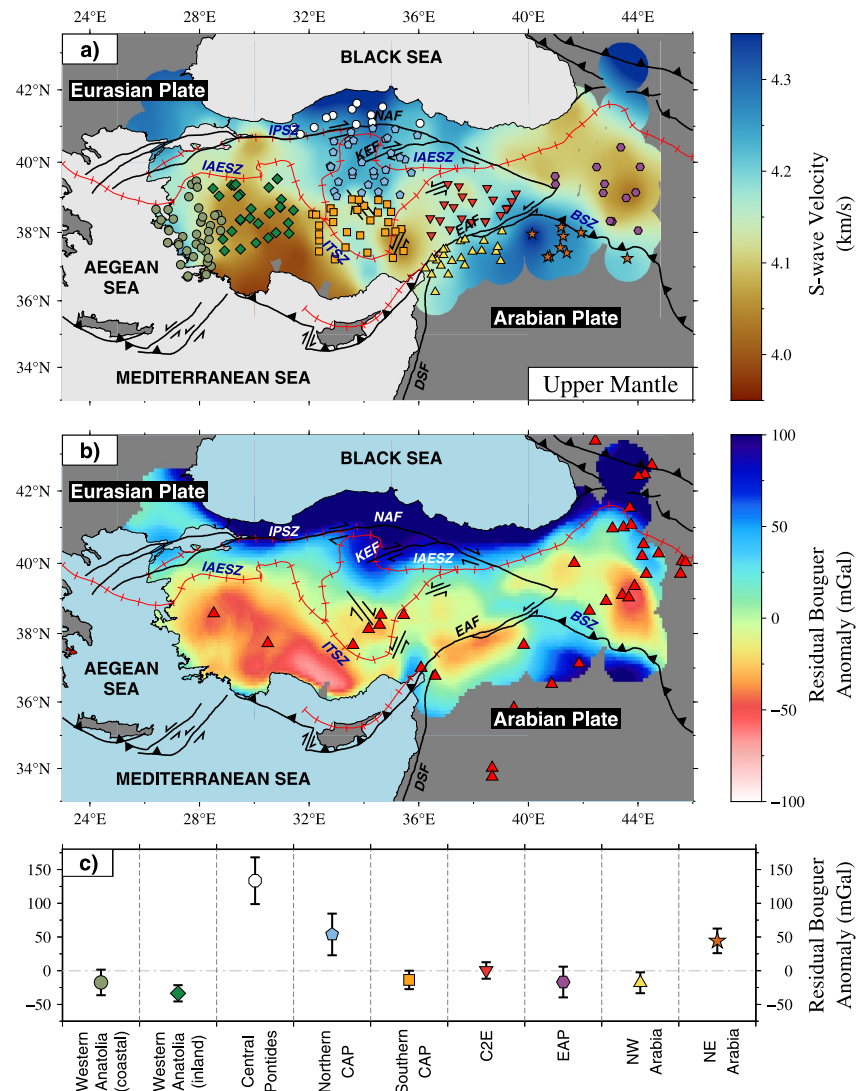


Figure 9. (a) Mean S-wave velocity in the mantle over a depth range of 50 km, starting from 5 km below the Moho to avoid contamination by velocities in the crust-mantle transition depth range. Various colored symbols highlight clusters of stations in different regions. (b) Residual Bouguer anomalies across Anatolia. (c) Mean and standard deviation of residual Bouguer Anomalies of each cluster of stations shown in (a). C2E: The region between the Central Anatolian Plateau and Eastern Anatolian Plateaus.

4.2. The North Anatolian Fault (NAF)

Moho depths vary little across the NAF (Figure 8a), suggesting an absence of lateral juxtaposition of crustal units of distinct thickness. This is unsurprising when reviewed in light of the fault's young age and relatively small offset (85 ± 5 km in 11–13 Ma; Şengör et al., 1985). However, other geophysical properties do show abrupt N–S contrasts across the NAF: negative residual topography ($\epsilon_{res} < -1.0$ km), and strong positive residual Bouguer anomalies (>80 mGal) characterize the Eurasian Plate north of the fault, while generally higher residual topography ($\epsilon_{res} \geq -0.5$ km) and lower residual Bouguer anomalies (<60 mGal) are observed below the Anatolian Plate south of the fault (Figures 8b and 9b). Additionally, faster mantle wavespeeds are observed below the Eurasian Plate north of CAP and WA (Figure 9a). These observations are consistent with full waveform inversion (e.g., Fichtner et al., 2013) and tomographic studies (e.g., Berk Biryol et al., 2011; Delph et al., 2015; Kounoudis et al., 2020), which identify pronounced mantle velocity contrasts across NAF at <100 km depth, corroborating the hypothesis that the NAF may have developed due to a contrast in lithospheric mantle, not

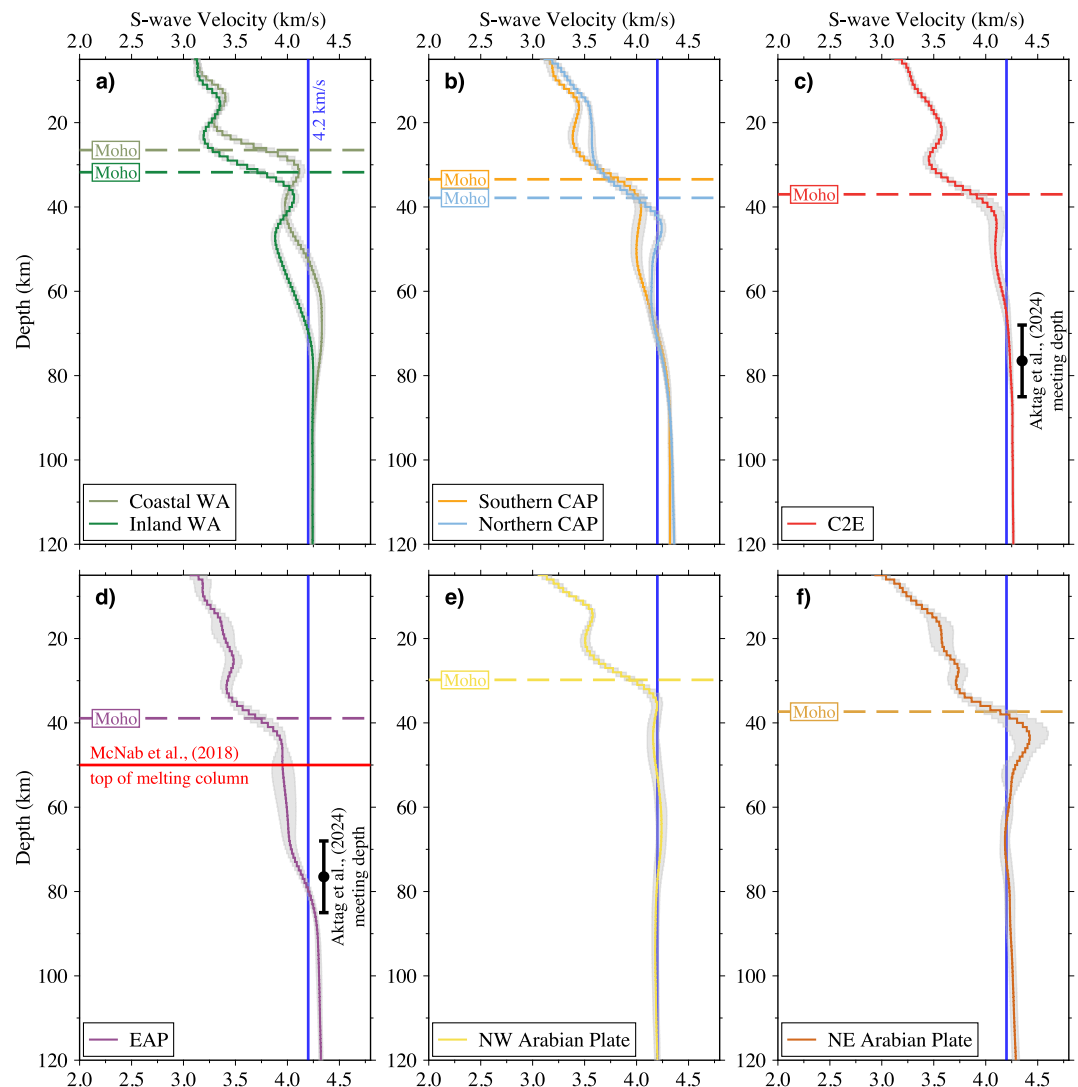


Figure 10. 1D clustered velocity profiles for (a) the coastal and inland part of Western Anatolia, (b) the northern and southern part of the Central Anatolian Plateau (CAP), (c) the region between the CAP and Eastern Anatolian Plateaus (EAP), (d) the EAP, (e) the northwestern tip of the Arabian Plate, and (f) the northeastern tip of the Arabian Plate, derived from clusters of stations in Figure 9a. In (b), the southern CAP profile excludes stations located in the central Taurides. Dashed lines in each subplot mark mean Moho depths obtained from all stations contributing to the stack. Blue vertical lines mark a reference shear-wave velocity of 4.2 km/s. The Aktag et al. (2024) melting depth in (c) and (d) marks the range of source-depths for Tunceli lavas erupted within the EAP. WA: western Anatolia.

crustal, properties: mechanically strong, thick Eurasian lithosphere to the north and mechanically weak, thin Anatolian lithosphere to the south (Kahraman et al., 2015; Kounoudis et al., 2020; Taylor et al., 2016).

One potential mechanism for the lithospheric contrasts is subduction-related hydration and magmatism, which may have weakened Anatolian upper plate lithosphere (e.g., Arcay et al., 2006), thereby promoting lithospheric thinning via dripping or delamination south of more refractory East European craton lithosphere. On the other hand, the NAF may mark the northern limit of metasomatic modification, regardless of any initial contrasts between Precambrian (Eurasian) and Phanerozoic (Anatolian) lithospheric structure (e.g., Kounoudis et al., 2020). Intriguingly, the northern CAP deviates from the N–S asymmetry pattern across the NAF, as evidenced by residual topography, gravity, and uppermost mantle velocity data (Figures 8b and 9). We re-visit this observation in Section 4.3.

4.3. Asymmetric Lithospheric Structure of the Central Anatolian Plateau

Residual topography of $\epsilon_{res} \geq -1.0$ km characterizing the entire CAP (Figure 8b) corroborates the widely held view that uplift in this region requires a mantle contribution associated with either loss of the negative-buoyancy lithospheric mantle, and/or upward flow of buoyant asthenospheric mantle (e.g., Cosentino et al., 2012; Bartol & Govers, 2014; Delph et al., 2017; Merry et al., 2021; Ögretmen et al., 2018; Ogden & Bastow, 2022; Schildgen, Cosentino, Bookhagen, et al., 2012; W. Zhu et al., 2023). Furthermore, a pronounced N–S asymmetry is observed in both residual topography ($\epsilon_{res} \geq -0.5$ km south of $\sim 39^\circ$ N; Figure 8b) and residual Bouguer gravity anomalies (negative anomalies concentrated south of $\sim 39^\circ$ N; Figures 9b and 9c), imply that the mantle contributes markedly more uplift in the southern CAP than to the north of 39° N, where the lithosphere is thicker. Consistent with this inference, Figures 9a and 10b demonstrate that a thin but discernible fast wavespeed “lid” (10–15 km mantle lithosphere) below the northern CAP contrasts slow mantle velocities south of $\sim 39^\circ$ N from which no estimate for plate thickness can be gleaned seismically. Similar shear-wave velocity structure was also found below the southern CAP by Delph et al. (2017).

Numerous previously postulated geodynamic mechanisms can be invoked to explain our observations for the CAP. For example, Göğüş et al. (2017) proposed that oroclinal folding of Kırşehir arc terranes by Mid-Eocene-to-Miocene times led to the development of a thickened arc root. Subsequent near-complete removal of the lithospheric mantle via Rayleigh-Taylor instability (“lithospheric dripping”) then triggered volcanism associated with asthenospheric decompression melting in the CAVP (e.g., Reid et al., 2017), where we observe positive ϵ_{res} (Figure 8b) but no seismically discernible fast wavespeed mantle lithosphere (Figures 9a and 10b).

Lithospheric dripping is not unique in its ability to explain our observations, however. Processes related to the terminal stage subduction of the Cyprus slab offer alternative mechanisms to thin the lithosphere. For instance, Bartol and Govers (2014) invoked the hypothesis of a single large-scale delamination model to explain the high elevation in the CAP and EAP. In this model, retreat of the northern Neotethys slab led to the delamination of the continental mantle lithosphere from the IAESZ to the Cyprus trench. The asymmetric lithospheric structure of the CAP that we observe (Figures 8b, 9 and 10b) suggests that refinement of the “single plateau” (CAP–EAP-wide) delamination hypothesis is warranted, with slab retreat-related lithospheric delamination below the CAP occurring predominantly south of 39° N. Moreover, several geological studies that observed particularly high (~ 3 mm/yr) uplift rates over the past 500 ka in the southern CAP (e.g., Racano et al., 2020; Ögretmen et al., 2018) favor a slab break-off hypothesis (e.g., Cosentino et al., 2012; Racano et al., 2020; Schildgen, Cosentino, Bookhagen, et al., 2012; Schildgen et al., 2014).

Even in the absence of lithospheric delamination in the manner proposed by Bartol and Govers (2014), flat-slab subduction of the Cyprus slab may have thinned the continental lithosphere of the southern CAP. For example, Axen et al. (2018) follow a thermo-mechanical modeling approach for the western United States to demonstrate that an advancing flat slab can scrape off the lowermost continental mantle lithosphere. Flat subduction of the Cyprus slab has been proposed below the CAP (e.g., Schleiffarth et al., 2018), meaning the Axen et al. (2018) hypothesis can reasonably be invoked below the CAP south of $\sim 39^\circ$ N.

In addition to removal of the mantle lithosphere, upwelling of buoyant asthenospheric mantle may also contribute to the CAP's north–south asymmetric lithospheric structure (Figures 8b, 9, and 10b). Such a flow-field would be expected to develop as the return-flow response of the aforementioned lithospheric dripping or delamination processes (e.g., Andersen et al., 2024; Bartol & Govers, 2014; Göğüş et al., 2017). Accordingly, on the strength of null SKS splitting measurements of mantle seismic anisotropy observed at stations within the southern CAP, Merry et al. (2021) invoked the hypothesis of an upward asthenospheric mantle flow. However, Merry et al. (2021) also noted that vertical flow through a tomographically imaged (e.g., Confal et al., 2020; Kounoudis et al., 2020; Portner et al., 2018) horizontal tear in the Cyprus slab could also reproduce the splitting observations.

Another way in which the asthenosphere could be contributing to Anatolian plateau uplift is if mantle potential temperatures are elevated. Faccenna et al. (2013), for example, advocate the hypothesis of westward-flowing hot, buoyant mantle, originating from the Afar hotspot to the south, as the driving mechanism for Anatolia's westward rotation, as well as its uplift and magmatism. Other workers have also championed the elevated potential temperature plume-flow hypothesis for the Anatolian Plate (e.g., Hua et al., 2023). However, petrological estimations of mantle potential temperatures (T_p) provide no clear evidence that the CAP mantle is markedly hotter than the global average: McNab et al. (2018), for example, report $T_p \approx 1,380^\circ\text{C}$ below the CAP—only slightly above the

global average of 1,350°C. Similarly, Reid et al. (2017) found that T_p values estimated from basalts in the CAVP are even colder than the global average, further supporting the notion that the mantle beneath the CAP is not significantly hotter. We therefore suggest that relatively short length-scale processes are the primary mechanism for the CAP's high elevations, not flow of hot mantle plume material from the south.

4.4. Northernmost Arabia and the EAP

A striking observation in Figure 8a is the abrupt change in Moho depth across the EAF: depths of <35 km in the northwest tip of the Arabian Plate contrast depths of 35–40 km on the Anatolian Plate. Considering the EAF's small lateral offset (~33 km; e.g., Şengör et al., 2005), slow slip rate (<10 mm/yr; e.g., Aktug et al., 2016) and young age (≤5 Ma; e.g., Westaway & Arger, 1996; Herece, 2008; Duman & Emre, 2013; Whitney et al., 2023), the observed Moho step cannot be attributed to the lateral (i.e., via strike-slip faulting) offset of crustal domains of initially different thickness. Instead, the observed Moho step may simply be the result of the collisional juxtaposition of two distinct crustal domains. Additionally, however, distributed crustal shortening during the inception of Arabia-Eurasia collision is considered to have occurred dominantly (~80%, or 195–415 km, of total shortening) on the upper Eurasian (now Anatolian) Plate (e.g., Darin & Umhoefer, 2022). The converse is true to the east of ~40°E across the Bitlis Suture, where shortening is thought to have been accommodated more evenly on the upper (Eurasian) and lower (Arabian) plates (e.g., Darin & Umhoefer, 2022). For example, the Arabian Plate accommodated >40% of plate shortening in the Zagros-Caspian Sea transect, which accords with a general westward decrease in the amount of crustal shortening along the Bitlis suture (e.g., Pirouz et al., 2017).

The observed asymmetric crustal structure of the northern tip of Arabia (Figure 8a) may also relate to the diachroneity of the Arabia-Anatolia/Eurasian collision. In the tectonic reconstruction of Darin and Umhoefer (2022), collision west of 40°E began ~5 Ma after initial contact between the Anatolian and Arabian plates. The subsequent establishment of EAF strike slip faulting during the Late Pliocene (~5 Ma, e.g., Yilmaz et al., 2006; Köküm & İnceöz, 2018) in response to the westward rotation of Anatolia means that the north-easternmost tip of Arabia has undergone more protracted shortening than terranes to the northwest.

The northernmost tip of Arabia also displays intriguing variations in residual topography, ϵ_{res} (Figure 8b). Specifically, the area below and directly to the south of the EAF, displays residual topography comparable to the Anatolian Plate to the north of it ($\epsilon_{res} > -0.25$ km). This observation is counter to expectations based on uppermost mantle wavespeeds: evidence for a fast-wavespeed “lid” that could be attributed to negatively buoyant lithospheric mantle is lacking below the Anatolian Plate (Figures 9a and 10c), while a thin (50–70 km) lithosphere (akin to that noted by Delph et al., 2017) characterizes the northern tip of Arabia (Figures 9a and 10e, 10f). The negative residual Bouguer gravity anomalies in the northwestern tip of the Arabian Plate are distinct from the northeastern Arabian Plate, but consistent with those in the EAP and the Anatolian Plate to the north of the EAF (Figure 9b). The larger residual topography measurements in northernmost Arabia may, therefore, require an asthenospheric origin (Figure 8b).

To the northeast, this belt-like large residual topography anomaly corresponds to the chain of Quaternary-Recent volcanoes situated north of the Bitlis suture (Figure 8b). In the EAP, petrological studies constrain the top of the mantle melting column to be ~50 km, interpreted as the base of the plate (McNab et al., 2018). Petrological studies in this region yield elevated mantle potential temperature estimates of >1400°C (e.g., Aktağ et al., 2024; McNab et al., 2018), with an increase in asthenospheric contribution to erupted lavas since Late Miocene times (Rabayrol et al., 2019). Intriguingly, seismically derived mantle temperatures at 100 km depth also peak below the EAF and EAP (Kounoudis et al., 2020). Joint inversion of petrological and S-to-P receiver function data also favors the elevated T_p hypothesis for the EAP (~1420°C; Hua et al., 2023). Mantle anisotropy measured via SKS splitting analysis suggests a northeast-oriented upper-mantle flow-field (Merry et al., 2021), interpreted to derive from the Afar hotspot to the south, as suggested by earlier studies of regional mantle flow (e.g., Chang & Van der Lee, 2011; Faccenna et al., 2013). Alternatively, some studies have suggested upwelling of plume material directly below northernmost Arabia (Civiero et al., 2022; Nikogosian et al., 2018), which could also explain our observations.

Below the EAP, widespread slow wavespeeds north of the Bitlis Suture (Figure 9a) are consistent with the view that Bitlis slab break-off, potentially including removal of EAP lower lithosphere via delamination (e.g., Göğüş & Pysklywec, 2008; Memiş et al., 2020), is almost complete (e.g., Berk Biryol et al., 2011; Kounoudis et al., 2020; Portner et al., 2018). The resultant thinning of the lithosphere largely explains the broad slow uppermost mantle

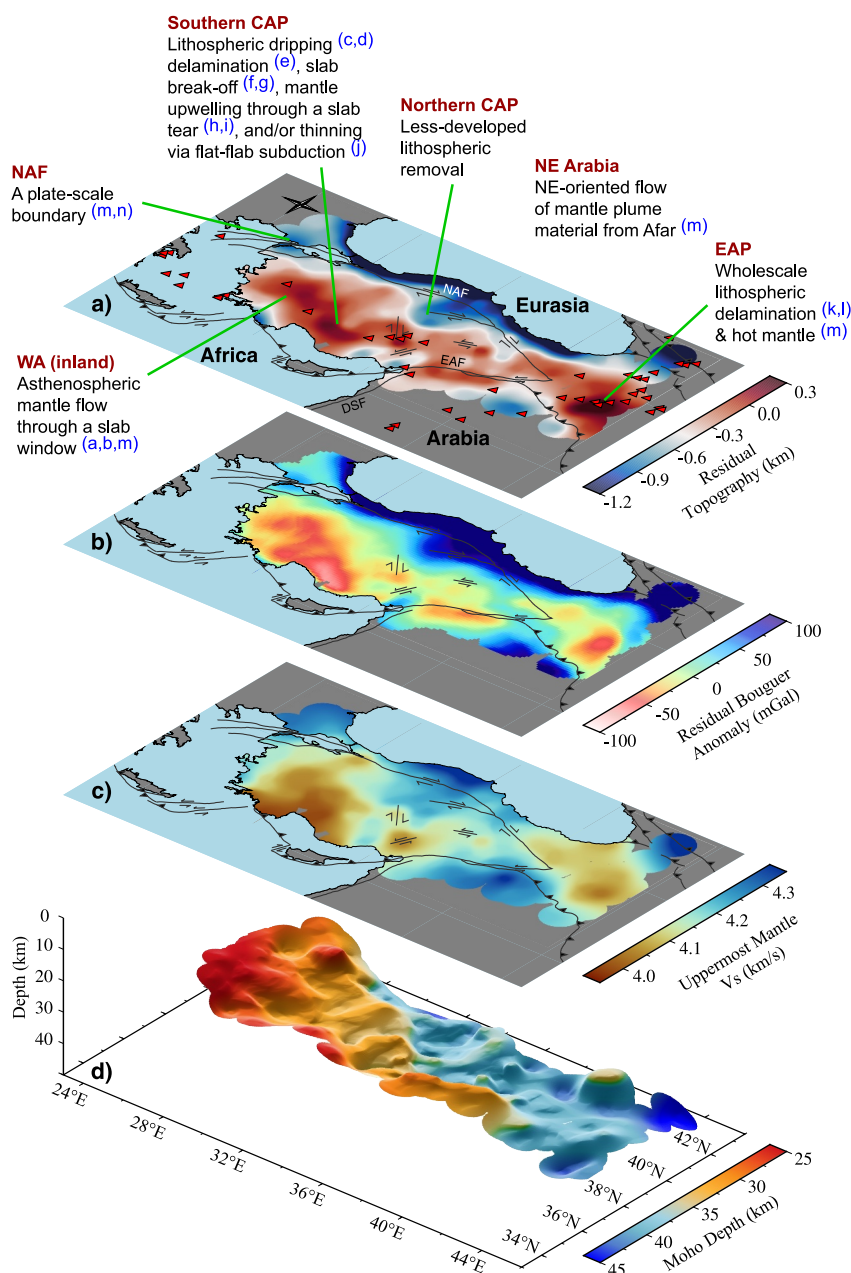


Figure 11. Summary of geophysical constraints and tectonic interpretations for Anatolia. (a) Residual topography map of the Anatolian Plate, including proposed mechanisms for observed anomalies. (b) Residual Bouguer anomalies. (c) Mean uppermost mantle S-wave velocities. (d) Moho depth surface. (a) Berk Biryol et al. (2011), (b) Confal et al. (2020), (c) Göğüş et al. (2017), (d) Andersen et al. (2024), (e) Bartol and Govers (2014), (f) Schildgen, Cosentino, Bookhagen, et al. (2012), (g) Racano et al. (2020), (h) Cosentino et al. (2012), (i) Öğretmen et al. (2018), (j) Axen et al. (2018), (k) Göğüş and Pysklywec (2008), (l) Memiş et al. (2020), (m) Merry et al. (2021), (n) Kounoudis et al. (2020).

beneath the EAP (Figure 9a). Additionally, aforementioned northward flow of hot mantle plume following the Bitlis slab break-off (e.g., Confal et al., 2018; Molin et al., 2023) will further depress the upper mantle wavespeed (Figures 9a and 10d) and elevate residual topography (Figure 8b).

5. Conclusions

Lithospheric seismic structure has been derived using joint inversion of surface wave dispersion and teleseismic receiver function data at 575 broadband seismograph stations operating during 1994–2023 across the eastern

Mediterranean. The resulting velocity models inform new maps of crustal thickness, residual topography (ϵ_{res}), uppermost mantle wavespeed, and residual Bouguer anomaly. An overall west-to-east Moho depth increase from 25 km to ≤ 44 km corroborates earlier studies in the region (Figure 11d). Residual topography of -1.1 – $+0.3$ km below Anatolia, confirms the high elevations of the CAP and EAP require a mantle contribution (Figure 11a).

The CAP exhibits marked N–S asymmetry: thin crust (≤ 35 km), slow uppermost mantle wavespeeds (< 4.2 km/s), negative residual Bouguer anomalies and high residual topography ($\epsilon_{res} \geq -0.5$ km) characterize the CAP south of $\sim 39^\circ\text{N}$ (Figure 11), where lithospheric dripping (e.g., Göğüş et al., 2017), slab-break-off (e.g., Öğretmen et al., 2018) and/or lithospheric delamination (e.g., Bartol & Govers, 2014) processes have been proposed to operate. Asthenospheric return flow associated with each of these processes and/or flow of asthenosphere through a seismically imaged horizontal tear in the Cyprus slab (e.g., Kounoudis et al., 2020), is expected to contribute to the region's high topography and rapid uplift rates (~ 3 mm/yr; Öğretmen et al., 2018). In contrast, negative residual topography ($\epsilon_{res} \leq -0.5$ km), positive residual Bouguer anomalies, and the persistence of a thin (10–15 km) but discernible lithospheric mantle characterize the CAP north of $\sim 39^\circ\text{N}$ (Figures 10b and 11a, 11b). Thus, removal of lithosphere below the CAP is non-uniform.

Across the EAF, Moho depths increase by ~ 5 km into Anatolia (Figure 11d). This Moho step has likely benefited from the greater degree of upper-plate crustal shortening in this part of the Arabia–Eurasia/Anatolia collision and its diachroneity. In contrast, across the Bitlis suture east of $\sim 40^\circ\text{E}$, where upper and lower plate shortening are known to be more equal, Moho topography is lacking.

Below the EAP, residual topography peaks ($\epsilon_{res} \geq -0.25$ km) where uppermost mantle wavespeeds are slow (≤ 4.1 km/s) and a fast wavespeed lid is lacking (Figures 10d, 11a, and 11c). Lithospheric delamination associated with Bitlis slab break-off is thus largely responsible for the EAP's high elevations (e.g., Göğüş & Pysklywec, 2008; Şengör et al., 2003). When reviewed in light of the elevated mantle potential temperatures derived by petrological studies of young lavas erupted within the EAP ($> 1420^\circ\text{C}$; e.g., Aktaş et al., 2024), and NE–SW-oriented mantle flow inferred via study of seismic anisotropy (e.g., Merry et al., 2021), high residual topography and negative residual Bouguer anomalies (Figures 11a and 11b) below the EAP and north-westernmost Arabia may indicate the EAP is underlain by hot mantle plume material derived from the Afar hotspot to the south.

Data Availability Statement

This study utilized seismograms from FDSN networks TU (Disaster and Emergency Management Authority, 1990), HC (Technological Educational Institute of Crete, 2006), HL (National Observatory of Athens et al., 1975), GE (GEOFON Data Centre, 1993), Z3 (Friederich & Meier, 2005), KO (Kandilli Observatory and Earthquake Research Institute, Boğaziçi University, 1971), IU (Albuquerque Seismological Laboratory/USGS, 1988), XW (Paul et al., 2007), XY (Paul et al., 2013), YI (Paul et al., 2008), GO (Ilia State University-Seismic Monitoring Centre of Georgia, 1988), MP (Seismological Laboratory of University of Basrah (SLUB) and Laboratory of Seismology and Spatial Data Analysis of University of Baghdad (LSDAUoB), Iraq, 2014), XG (Sandvol, 1999), XH (Mitchell & Zhu, 2002), YB (Sandvol, 2013), YF (van der Lee et al., 1999), YH (DANA, 2012), and YL (Beck & Zandt, 2005). Seismograms were obtained from: TU via AFAD repository (<http://tdvms.afad.gov.tr>); HC and HL via EIDA (<http://eida.gein.noa.gr>); GE and Z3 via GEOFON (<http://geofon.gfz-potsdam.de>); KO via KOERI (<http://eida-service.koeri.boun.edu.tr>); IU via ORFEUS (<http://www.orfeus-eu.org>); XW, XY, YI via RESIF (<https://ws.resif.fr>); GO, MP, XG, XH, YB, YF, YH, YL via the NSF SAGE data archive operated by EarthScope Consortium (NSF award 1724509) (<https://ds.iris.edu/ds/nodes/dmc>). Maps were generated using the Generic Mapping Tools (GMT; Wessel et al., 2019). Seismograms were analyzed using ObsPy (Beyreuther et al., 2010) and the Seismic Analysis Code (SAC; Helffrich et al., 2013). Theoretical arrival times of P-waves were calculated utilizing the TauP toolkit (Crotwell et al., 1999). Joint inversions were performed using the Joint96 module of the Computer Programs in Seismology suite (<https://rbherrmann.github.io/ComputerProgramsSeismology/>; Herrmann, 2013). Forward modeling of the Bouguer gravity response of our Moho model was performed utilizing the Harmonica package (Fatiando et al., 2024).

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