



Review Article

The uplift of the East Africa - Arabia swell

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ABSTRACT

The East Africa - Arabia topographic swell is an anomalously high-elevation region of ~4000 km long (from southern Ethiopia to Jordan) and ~1500 km wide (from Egypt to Saudi Arabia) extent. The swell is dissected by the Main Ethiopian, Red Sea, and Gulf of Aden rifts, and characterized by widespread basaltic volcanic deposits emplaced from the Eocene to the present. Geochemical and geophysical data confirm the involvement of mantle processes in swell formation; however, they have not been able to fully resolve some issues, e.g., regarding the number and location of plumes and uplift patterns. This study addresses these questions and provides a general evolutionary model of the region by focusing on the present topographic configuration through a quantitative analysis and correlating long and intermediate wavelength features with mantle and rifting processes. Moreover, the isostatic and dynamic components of topography have been evaluated considering a range of seismic tomographic models for the latter. When interpreted jointly with geological data including volcanic deposits, the constraints do imply causation by a single process which shaped the past and present topography of the study area: the upwelling of the Afar superplume. Once hot mantle material reached the base of the lithosphere below the Horn of Africa during the Late Eocene, the plume flowed laterally toward the Levant area guided by pre-existing discontinuities in the Early Miocene. Plume material reached the Anatolian Plateau in the Late Miocene after slab break-off and the consequent formation of a slab window. During plume material advance, buoyancy forces led to the formation of the topographic swell and tilting of the Arabia Peninsula. The persistence of mantle support beneath the study area for tens of million years also affected the formation and evolution of the Nile and Euphrates-Tigris fluvial networks. Subsequently, surface processes, tectonics, and volcanism partly modified the initial topography and shaped the present-day landscape.

1. Introduction

Earth's topography is the expression of the interactions between the floating equilibrium of a density column, surface processes, flexure, lithospheric deformation (i.e., tectonics, volcanism, and magmatic underplating), and mantle dynamics (e.g., Braun, 2010; Flament et al., 2013; Faccenna and Becker, 2020). To unravel some of these processes, the contributions to topography can be inferred as being due to two main components: the isostatic and dynamic ones (e.g., Panasyuk and Hager, 2000; Gvirtzman et al., 2016). The first depends on the heterogeneities in lithosphere structure and density; the second represents the deflection of the surface in response to mantle tractions arising from density driven flow in the mantle and plate motions. In particular,

upwellings (i.e., mantle plumes), caused by the rise of hot mantle material, and downwellings, related to the descent of cold lithosphere during subduction, can cause uplift and subsidence at the surface, respectively (e.g., Hager et al., 1985; Panasyuk and Hager, 2000; Braun, 2010; Flament et al., 2013).

Mantle plumes show complex morphologies and dynamics (e.g. Heron, 2018; Koppers et al., 2021). One may classify them as "primary", consisting of whole-mantle structures rising from the core-mantle boundary as narrow, localized conduits or in the form of broader upwellings, or superplumes. "Secondary" plumes are confined to the upper mantle as the result of stagnation of the primary plumes at the base or within the mantle-transition zone (Ritsema et al., 1999; Courtillot et al., 2003; French and Romanowicz, 2015; Cloetingh et al., 2022). Regions

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affected by primary mantle plumes are often characterized by the emplacement of a huge volume of igneous rocks (Large Igneous Provinces – LIPs) typically attributed to the plume head arrival (e.g., Richards et al., 1989; Burke and Torsvik, 2004), and by broad, anomalously high-elevated regions (topographic swells; e.g. Richards et al., 1988; Ribe and Christensen, 1994; Heron, 2018). Despite successive isostatic adjustments and rifting events, topographic anomalies can be

preserved for tens of millions of years by processes such as crustal thickening (McKenzie, 1984; Cox, 1989) and/or by the persistence of a mantle anomaly below the crust (Faccenna et al., 2019).

Topographic swells associated with mantle plumes are documented worldwide on both oceanic and continental plates (Dietz and Menard, 1953; Ribe and Christensen, 1994a, 1994b; Ribe and Christensen, 1999; Gurnis et al., 2000; Sengör, 2001; Daradich et al., 2003; Roberts and

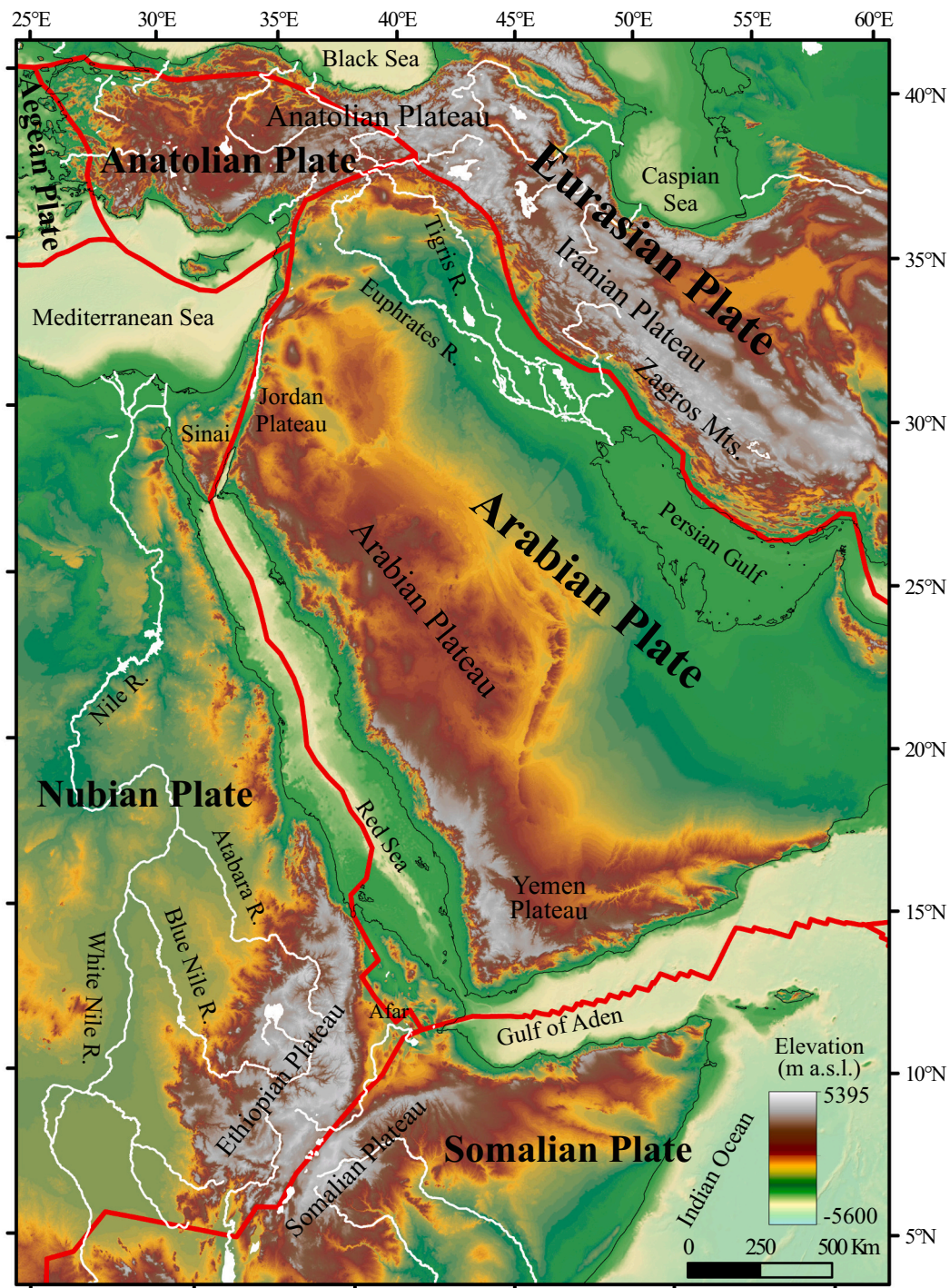


Fig. 1. Topographic configuration of the study area (ETOPO2022 global elevation model with resolution of ~500 m; www.ngdc.noaa.gov); the EAAS extends from south to north comprising the Ethiopian, Somalian, Arabian, and Jordan plateaux; solid red lines indicate plates boundaries. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

White, 2010; Jones et al., 2012; Roberts et al., 2012; Davila and Lithgow-Bertelloni, 2013; Rowley et al., 2013; Czarnota et al., 2014; Paul et al., 2014; Liu, 2015; Heller and Liu, 2016; Sembroni et al., 2016a, 2021; Faccenna et al., 2019; Friedrich, 2019; Clementucci et al., 2023; Molin et al., 2023). Their formation and evolution has been addressed with numerical and analog convection models (e.g. Houseman, 1990; Griffiths and Campbell, 1991; Farnetani and Richards, 1994; Ribe and Christensen, 1994a, 1994b; Ribe and Christensen, 1999; d’Acremont et al., 2003; Burov and Guillou-Frottier, 2005; Moucha et al., 2008; Braun, 2010; Moucha and Forte, 2011; Cramer et al., 2012; Burov and Gerya, 2014; Koptev et al., 2015; Kiraly et al., 2015; Barnett-

Moore et al., 2017; Koptev et al., 2017; Rubey et al., 2017; Sembroni et al., 2017; Cao et al., 2018).

One of the most studied ongoing swells is the East Africa-Arabia one (EAAS), also called the “Afro-Arabian dome” (Cloos, 1939; Almond, 1986; Camp and Roobol, 1992). It is an anomalously high-elevated region ~ 4000 km long (from southern Ethiopia to Jordan) and ~ 1500 km wide (from Egypt to Saudi Arabia; Cloos, 1939; Almond, 1986, Camp and Roobol, 1992; Avni et al., 2012; Bar et al., 2016; Fig. 1). The area is dissected by the Main Ethiopian, Red Sea, and Gulf of Aden rifts and is characterized, along its entire extent, by widespread volcanic deposits varying in age from Eocene to present-day (Coleman et al., 1983; Brown

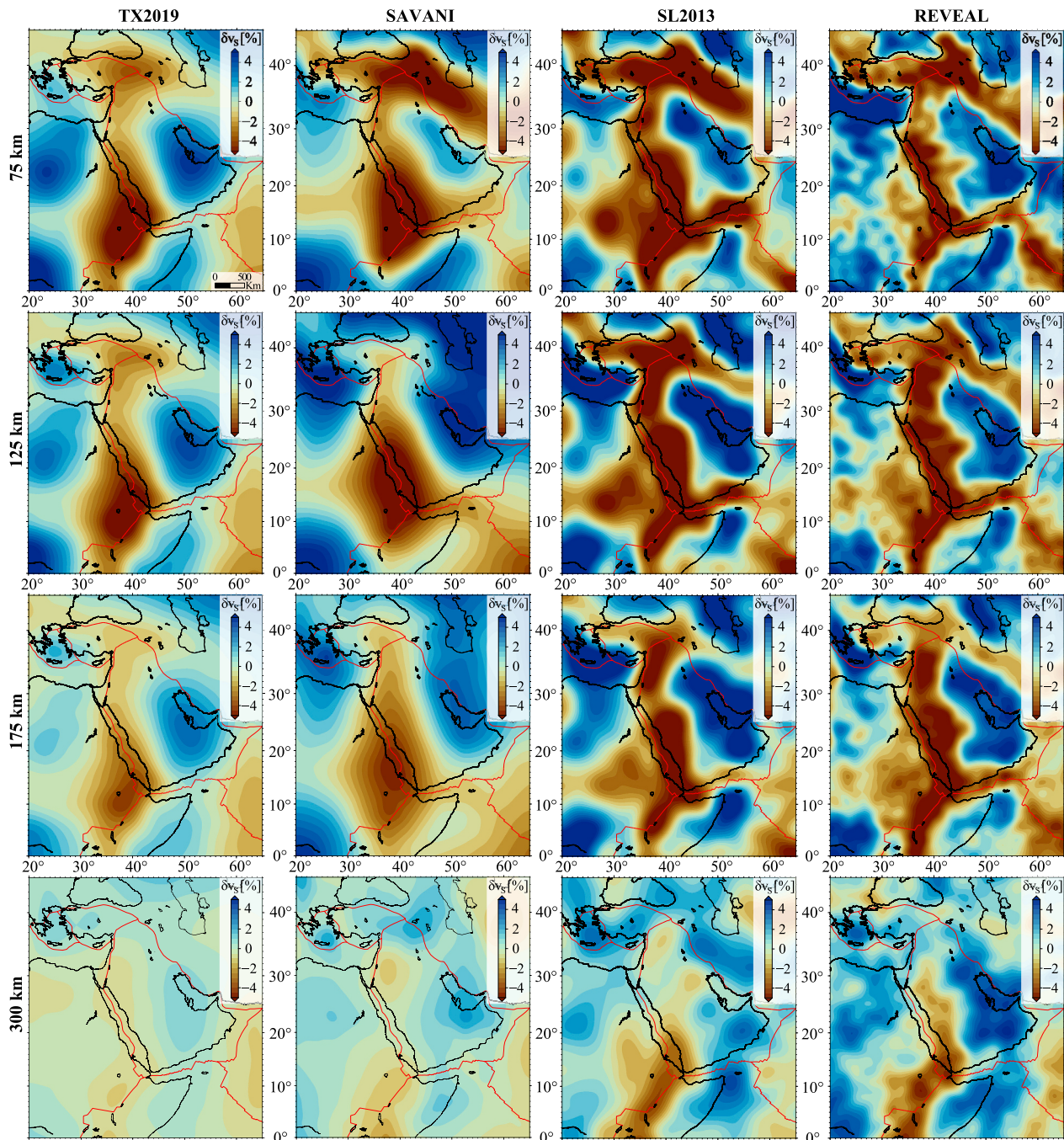


Fig. 2. Comparison of uppermost mantle structure from selected, global shear wave tomographic models: TX2019 (Lu et al., 2019), a body wave model, SAVANI (Auer et al., 2014) a radially anisotropic model based on body and surface waves, SL2013 (Schaeffer and Lebedev, 2013) a surface wave focused, higher resolution SV model, and REVEAL (Thrastarson et al., 2024), a full waveform inversion model. All tomographic models plotted at 75, 125, 175, and 300 km depth. Note the broad region of low seismic velocity below the area comprised between eastern Africa and Levant region. Overall anomalies are consistent across models, with the higher regional resolution models showing consistent northward slow structures, undulating underneath the Arabian plate.

et al., 1989; Camp and Roobol, 1992; Hofmann et al., 1997; Bosworth and Stockli, 2016; Purcell, 2017; Rooney, 2017; Fig. 1). The chemical composition (Baker et al., 1996; Kieffer et al., 2004; Rooney, 2017) and radiogenic isotope ratios (Krienitz et al., 2009; Hua et al., 2023) of the volcanics are consistent with a mantle source with lithospheric contamination.

Several seismic tomography studies show the presence of a broad

region of low seismic velocity below the EAAS, interpreted as the signature of hot mantle material (Lithgow-Bertelloni and Silver, 1998; Ritsema et al., 1999; Gurnis et al., 2000; Nyblade et al., 2000; Ritsema and van Heijst, 2000; Benoit et al., 2006a, 2006b; Montagner et al., 2007; Bastow et al., 2008, 2011; Chang and Van der Lee, 2011; Moucha and Forte, 2011; Nyblade, 2011; Faccenna et al., 2013; Hansen and Nyblade, 2013; Schaeffer and Lebedev, 2013; Auer et al., 2014; Emry

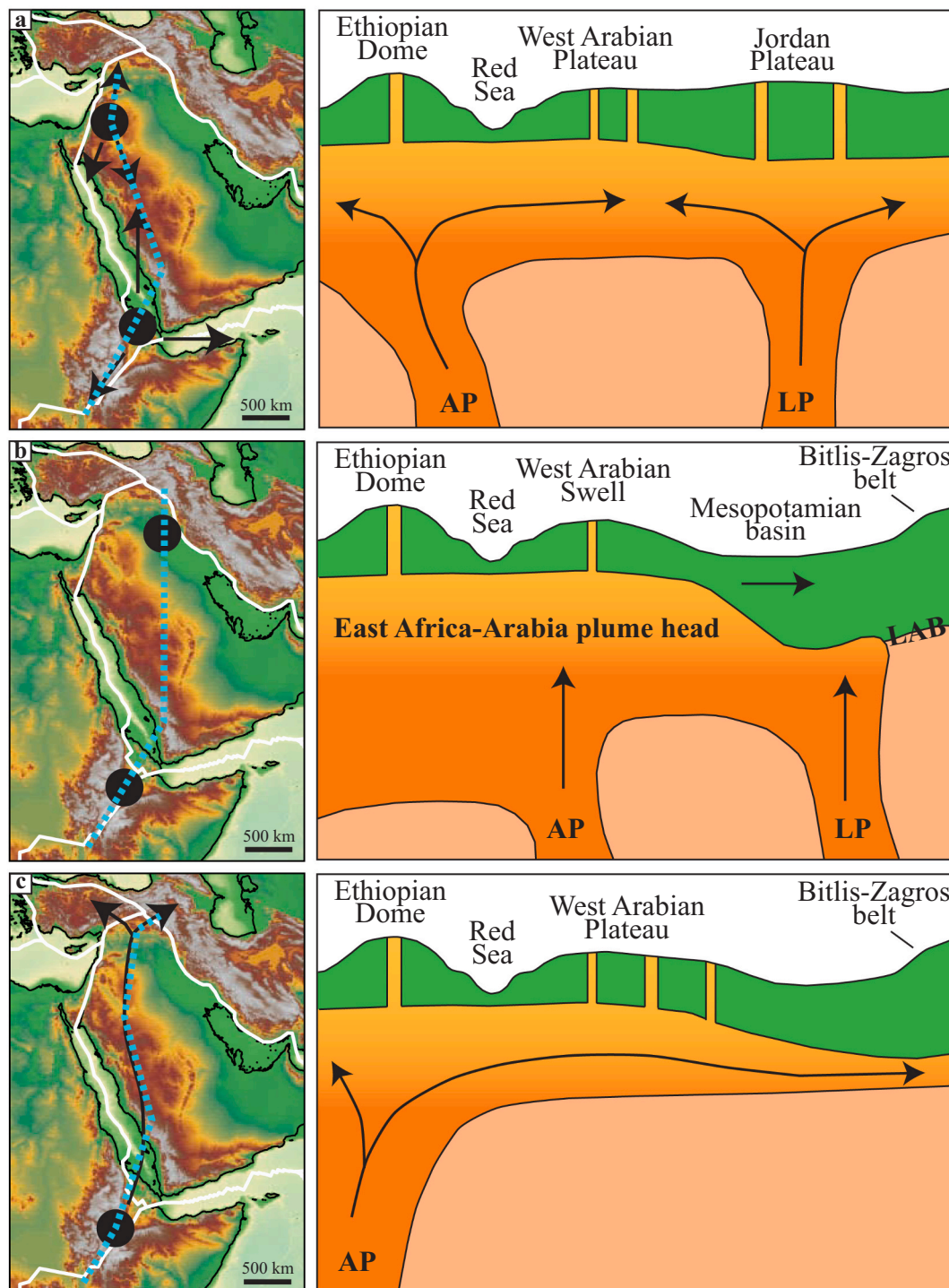


Fig. 3. Schematic maps showing locations of mantle plumes and mantle flow directions (left) with the relative conceptual models (right) according to (a) Chang and Van der Lee (2011), (b) Civiero et al. (2022), and (c) Faccenna et al. (2013). The dashed light blue lines on the maps indicate the traces of the schematic sections represented on the right. AP = Farar Plume; LP = Levant Plume. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

et al., 2019; Lu et al., 2019; Wei et al., 2019; Chang et al., 2020; Tsekhmistrenko et al., 2021; Thrastarson et al., 2024; see Fig. 2). However, although the role of mantle processes has been commonly accepted as the main cause of the current topographic and geologic configurations of the region, in past years there has been debate about the location and number of plumes (Dixon et al., 1989; Ebinger and Sleep, 1998; Chang and Van der Lee, 2011; Hansen et al., 2012; Faccenna et al., 2013; Koulakov et al., 2016). More recent whole-mantle studies (Chang et al., 2020; Tsekhmistrenko et al., 2021; Civiero et al., 2022; Boyce et al., 2021, 2023) seem to converge on the consensus that the region is underlain by a broad low wave velocity anomaly (primarily the African Superplume) with potential addition from other deep-seated plume tails.

In particular, three models can be recognized (Fig. 3). The first indicates the presence of two near-vertical mantle plumes rising below Afar and northern Arabia and flowing beneath the lithosphere (e.g., Debayle et al., 2001; Montagner et al., 2007; Sicilia et al., 2008; Chang and Van der Lee, 2011; Koulakov et al., 2016; Fig. 3a). The second assumes two stationary plumes (Afar and northern Arabia plumes) beneath moving lithospheric plates; in this case, the plume head is stagnating below the lithosphere (e.g., Civiero et al., 2022; Fig. 3b). The last model, in accordance with regional seismic velocity, anisotropy patterns, and seismic tomography (Hansen et al., 2006; Berk Biryol et al., 2011; Schaeffer and Lebedev, 2013; Auer et al., 2014; Qaysi et al., 2018; Lu et al., 2019; Wei et al., 2019; Thrastarson et al., 2024) considers only one mantle plume (Afar superplume) rising below eastern Africa, but flowing to eastern Turkey through pressure gradients and/or prior discontinuities (e.g., Camp and Roobol, 1992; Ebinger and Sleep, 1998; Ershov and Nikishin, 2004; Hansen et al., 2012; Faccenna et al., 2013; Wei et al., 2019; Lim et al., 2020; Agostini et al., 2021; Hua et al., 2023; Fig. 3c).

The uplift pattern of the area is also controversial. In Ethiopia, the uplift is believed to have occurred before (Sengör, 2001), during (Pik et al., 2003), or after (Gani et al., 2007) the emplacement of the volcanic deposits. Some studies (Sembroni et al., 2016a, 2021; Faccenna et al., 2019) suggest that the present topography of Eastern Africa is “a long-term, dynamically supported feature” initiated, at least, in the Oligocene which deeply influenced the present path of the Nile River system (Faccenna et al., 2019). Along the western portion of the Arabian Peninsula, the existence of two distinct groups of volcanic deposits (one dated between 30 and 20 Ma and one younger than 12 Ma), separated by a stasis of few million years in volcanic activity, led Camp and Roobol (1992) to constrain the “initiation of significant uplift”, associated with an active mantle upwelling, at the Miocene. Lastly, studies on the topography of the Levant area allowed much of the uplift and the topography to be attributed to the late Oligocene (Avni et al., 2012; Bar et al., 2016).

In this study, we intend to contribute to the discussion about mantle plume number, volcanism, and uplift of the east Africa-Arabia region by analyzing the present surface topographic signal.

Analogue and numerical models produced surface topography related to the impingement of a mantle plume in the form of a long wavelength bulge. While the modeled dynamic topography amplitude scales with asthenospheric density anomalies to first order, the details of geometry and uplift rates depend on mantle–lithosphere interactions, rheological structure, and intraplate stresses (Griffiths and Campbell, 1991; Burov and Gerya, 2014; Kiraly et al., 2015; Sembroni et al., 2017). However, the presence of multiple mantle plumes would suggest a topographic configuration that gives rise to multiple bulges separated by depressed areas. In the case of a single plume locally channelized, the expected surface topography would be characterized by a ridge with the highest elevation at the impingement zone and a gradual decrease to the distal portion.

Therefore, it is not enough to observe the mantle signal to define the geodynamic configuration of a given area. There is a need for the interpretation of that signal to be consistent with the observed surface

topographic configuration. To this end we performed a topography analysis (filtered topography, slope, swath profiles) and reevaluated isostatic (flexural isostasy) and dynamic components combining new data with a review of data from literature. Moreover, the pattern of topography along the entire area and the age trend of volcanic deposits of mantle origin have been compared to verify a common trend. The results allow the current topographic configuration of the East Africa-Arabia swell to be attributed to a single mantle plume that, flowing horizontally from East Africa to Turkey, caused uplift and intense volcanism from Eocene to present, influencing the formation and evolution of the major river networks of the area. Much of the present topography is the result of that uplift and of the successive modification by tectonics and surface processes.

2. Geological evolution

The EAAS extends within the Neoproterozoic Arabian Nubian Shield which is split between the Arabian and African plates and extends from Egypt (Sinai Peninsula), through Saudi Arabia, southward to Kenya (Fig. 4). It represents the northern portion of the East African Orogen that formed during the Pan-African Orogeny (900–550 Ma; Stern, 1994) when East and West Gondwana collided to form the supercontinent “Greater Gondwana” or “Pannotia” at the end of Neoproterozoic (Stern, 2002). In the southern part of the shield (Ethiopia and Kenya), the collision produced a pervasive north-trending shear zones; the central-southwestern sections were subjected to oblique and orthogonal east-west compression accompanied by north-south stretching; in the northern and northeastern parts (Arabian Peninsula), shearing and NW-trending thrusting, extension and tectonic escape, resulted in the NW-trending Najd Fault System (Johnson et al., 2017, and references therein).

In the Cambrian, an intensive erosional denudation affected the Arabian-Nubian Shield, forming an extensive low-relief surface over its northern part (e.g., Garfunkel, 1999; Johnson, 2003; Avigad and Gvirtzman, 2009) and in the Horn of Africa (the “pre-Ordovician planation surface” of Coltorti et al., 2007). Between the Precambrian and the Permian, the deposition of sediments at the Gondwana margin formed the Arabian Platform. While the deposits thickened toward the margin of the Arabian Platform, several erosion events exposed the Precambrian basement (e.g., Lebkicher, 1960; Powers et al., 1966; Murris, 1980; Weissbrod and Gvirtzman, 1989; Beydoun, 1991; Kohn et al., 1992; Alsharhan and Nairn, 1995; Ziegler, 2001; Garfunkel, 2002).

During the Permian and Early Mesozoic, continental fragments drifted away and migrated northward, leaving Arabia facing the newly born Neotethys Ocean (e.g., Dercourt et al., 1986; Beydoun, 1991; Alsharhan and Nairn, 1995). In the Horn of Africa this period is characterized by the formation of the Karoo-type rifts and the deposition of 2–4 km of clastic sediments (Beltrandi and Pyre, 1973; Davidson and McGregor, 1976; Davidson, 1983; Hunegnaw et al., 1998; Mège et al., 2015; Macgregor, 2018). Moreover, another denudation event originated a second planation surfaces (the “Late Triassic planation surface” of Coltorti et al., 2007, or the “paleotopography 1” of Sembroni and Molin, 2018) which indicates a pronounced uplift phase (from hundreds of meters to a thousand; see Coltorti et al., 2007). In the Jurassic, after the development of NE–SW and N–S striking structural basins, a main marine transgression took place from S and SE determining the deposition of thick sequences of marine marls, limestones, and evaporites (Guiraud et al., 2005; Davison and Steel, 2018; Macgregor, 2018). Since the Lower Jurassic, the extensional deformation caused the separation of the Madagascar-India block from Eastern Gondwana (Reeves and De Wit, 2000; Seward et al., 2004; Gibbons et al., 2013; Gaina et al., 2015; Macgregor, 2018). During the Late Cretaceous, the closure of the Neotethys Ocean (e.g., Robertson and Dixon, 1984; Garfunkel, 1998, 2004; Robertson et al., 2006) resulted in a horizontal compression and the formation of the S-shaped Syrian Arc fold belt, extending from Sinai to

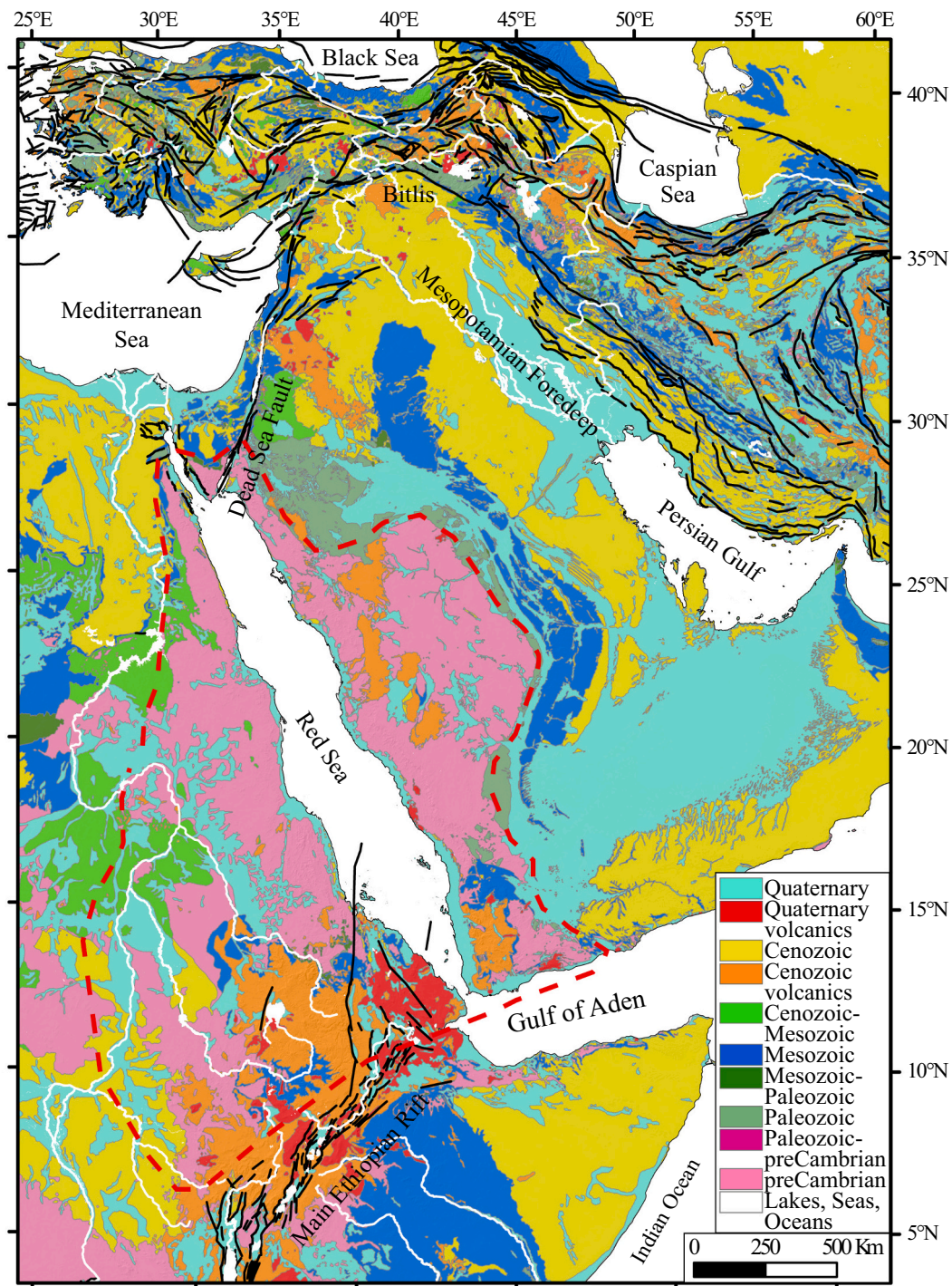


Fig. 4. Geological map of the study area (modified from the Lithologic Map of the World; (Hartmann and Moosdorf, 2012); active tectonic lineaments (solid black lines) are from the Global Active Faults database by Styron and Pagani (2020); dashed red line contours the Arabian-Nubian Shield (from Alemu, 2021). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

Syria where it is represented by the Palmyride inversion zone and fold belt (Krenkel, 1924; Chaimov et al., 1990; Brew et al., 2001), while in eastern Africa a new extensional tectonic phase formed several NW-SE narrow basins and a third denudation event originated the “Cretaceous planation surface” (Coltorti et al., 2007), “paleotopography 2” of Sembroni and Molin (2018). At the end of the middle Eocene the most extensive transgression over the Arabian Platform terminated (Ziegler, 2001; Gvirtzman et al., 2011; Avni et al., 2012). In the same period the separation of Arabia from Africa (6–13 mm/yr; Sella et al., 2002;

McClusky et al., 2003; Reilinger et al., 2006; Vigny et al., 2006; ArRajehi et al., 2010) occurred along with the development of the Gulf of Aden and Red Sea rift systems (Bosworth et al., 2005; Leroy et al., 2012). Continental rifting at the Gulf of Aden initiated between 38 and 33 Ma (Pik et al., 2013; Robinet et al., 2013; Purcell, 2017; Boone et al., 2021), concurrently with the onset of the Arabia-Eurasia collision (~23 mm/yr; ArRajehi et al., 2010; McClusky et al., 2003; Reilinger et al., 2006; Sella et al., 2002; Vigny et al., 2006). Thermochronology data indicate that collision took place at ~20 Ma along the Bitlis-Zagros zone (Okay et al.,

2010), though older ages have been proposed (Pirouz et al., 2017; Koshnaw et al., 2019).

During the Oligocene, geological and thermochronological data indicate a pronounced phase of uplift both in Ethiopia (Sengör, 2001; Pik et al., 2003; Sembroni et al., 2016a, 2016b; Sembroni et al., 2021) and along the Arabian plate (Gvirtzman et al., 2008; Avni et al., 2012; Bar et al., 2013; Turab et al., 2023), where the marine environments shifted toward the margins (e.g., Beydoun, 1991; Alsharhan and Nairn, 1995; Ziegler, 2001). The Arabian inland region and eastern Africa were subjected to denudation, which led to the formation of extensive low-relief surfaces (e.g., Leblich, 1960; Alsharhan and Nairn, 1995; Burke and Gunnell, 2008; Avni et al., 2012; Bar et al., 2016) termed “the Oligocene Peneplain” in northern Arabia (Picard, 1951; Quennell, 1958; Garfunkel and Horowitz, 1966; Avni, 1991; Zilberman, 1991; Avni et al., 2012; Zilberman and Calvo, 2013; Bar et al., 2016) and “Trap volcanics planation surface” (Coltorti et al., 2007) or “paleotopography 3” (Sembroni and Molin, 2018) in eastern Africa.

At ~27 Ma, rifting commenced along the western and southern Afar margins (Purcell, 2017). The western Afar faulting marks a continuation of the Red Sea rifting which, in the meanwhile, was opening starting from its southern portion (Bosworth et al., 2005; Wolfenden et al., 2005; Purcell, 2017; Boone et al., 2021). At the beginning of Miocene, the East African Rift (EAR) started the expansion both north and south from the Turkana region (Purcell, 2017).

Between 18 and 14 Ma, the Arabian plate separated from the Sinai sub-plate by the ~1000 km long Dead Sea Transform Fault (DSTF; Quennell, 1958; Freund et al., 1970; Garfunkel, 1981; Garfunkel et al., 1981; Joffe and Garfunkel, 1987; Bosworth et al., 2005) and began rotating counterclockwise. Recent dating of *syn*-faulting calcite has further constrained the formation of the DSTF plate boundary to 20.8–18.5 Ma in southern Israel and propagating northwards by 17.1 Ma (Nuriel et al., 2017). At the northern boundary of the Arabian Plate, the stress regime appears to have changed from compressional to strike slip, due to the oblique collision between Anatolian and Arabian plates (e.g., Beydoun, 1999). This caused the lateral escape of the Anatolian Plate, accommodated by a system of lithospheric scale dextral (North Anatolian Fault) and sinistral (East Anatolian Fault) strike slip faults (Şengör et al., 2005; Reilinger et al., 2006; Ballato et al., 2018). At ~13 Ma the EAR tectonic activity increased: the Western Branch and the northern segment of the Main Ethiopian Rift (MER) started to form (Macgregor, 2015; Purcell, 2017).

The central portion of the MER started joining the southern and northern segments between 8 and 5 Ma (Bonini et al., 2005; Abebe et al., 2010). In the Pleistocene, the deformation and the main magmatic activity abandoned the MER margin faults and shifted to the floor of the rift valley with the formation of the oblique Wonji Fault Belt (Ebinger, 2005; Corti, 2009; Purcell, 2017).

From the middle-late Miocene, to the north of the Bitlis collision zone, the Eastern Anatolian Plateau experienced a strong exhumation, probably due to the combined effects of a more advanced stage of the Arabia-Eurasia collision (Okay et al., 2010; Ballato et al., 2011; Cavazza et al., 2018, 2019; Gusmeo et al., 2021; Darin and Umhoefer, 2022) and the arrival of hot mantle material below the lithosphere (Molin et al., 2023, and references therein).

Plate reconstructions and geodetic data suggest that the present-day motion of the Arabian plate relative to the Nubian and the Eurasian plates stayed consistent since at least 11 Ma (e.g., ArRajehi et al., 2010; McClusky et al., 2010; Reilinger and McClusky, 2011; Viltres et al., 2022).

2.1. Volcanism

Intra-plate basaltic volcanism started in southern Ethiopia/northern Turkana at 46 Ma (“Eocene Initial Phase” of Rooney, 2017) and lasted for ~10 Ma. After a short hiatus, new basaltic eruptions (“Oligocene Trap phase”; Rooney, 2017) covered parts of Ethiopia, Eritrea, southern

Sudan, and western Yemen by 30–29 Ma (Hofmann et al., 1997; Fig. 4). This period of volcanism was coeval with initial faulting and deposition of syn-rift strata in the Gulf of Aden (Purcell, 2017; Rooney, 2017). In the same period, an intense magmatic activity occurred along the Western Arabian margin exploiting the NW-SE discontinuities associated with the Precambrian Najd Fault System (“Older Harrats”; Bosworth and Stockli, 2016; Fig. 4). Subsequently, a period of relative magmatic quiescence took place in the Horn of Africa with volcanism mainly concentrated in the Turkana area (Ukstins Peate and Bryan, 2008; Brown and McDougall, 2011), along the rift margins in Ethiopia and Yemen (Rooney et al., 2013), and in eastern Ogaden (Mège et al., 2016; Sembroni and Molin, 2018).

This period of volcanic quiescence is reflected in Ethiopia by the deposition of voluminous intratrappean sediments composed of red clays and sands between 29 and 27 Ma (Abbate et al., 2014). The last eruptions of the continental flood volcanics occurred at 26.5 Ma in Yemen and 25 Ma in Ethiopia. At 24 Ma basaltic dikes and igneous complexes formed northwest of Afar and along the present-day Red Sea margin of Yemen and Saudi Arabia (e.g., Coleman et al., 1983; Brown et al., 1989; Camp and Roobol, 1992; Ilani et al., 2001; Trifonov et al., 2011). At the same time, a large basaltic volcanic province developed in northern Egypt and in the Harrat Ash Shaam region of Jordan, while extensive NW-SE-trending diking, granitic intrusions and silicic volcanism occurred along the 1700 km length of the Western Arabian margin (Fig. 4).

In the middle-late Miocene voluminous volcanism affected the Eastern Anatolian Plateau (Keskin, 2003) causing the covering of large portion of the forming plateau (Fig. 4). This event seems to be coeval with both the collisional deformation along the Bitlis suture zone and the formation of the northern and eastern Anatolian faults (e.g. Keskin, 2003; Şengör et al., 2003; Faccenna et al., 2006; Faccenna et al., 2013; Schildgen et al., 2014; Memiş et al., 2020). In the same period (~13 Ma) a renewed magmatic phase, associated with the N-S movement along the DSTF, affected the Arabian Shield with the eruption of the so-called “Younger Harrats” (Bosworth and Stockli, 2016). Once initiated, volcanism at most locations continued until present.

3. Geomorphological setting

The East Africa-Arabia region displays spectacular signs of an ongoing mantle-plume impact on its surface. One of the most evident is the EAAS which extends for ~4000 km from Jordan to Ethiopia and ~1500 km from east to west (Almond, 1986; Dixon et al., 1989; Camp and Roobol, 1992; Fig. 1). To the south, the NW-SE trending Turkana depression, developed in the Early Cretaceous, separates the swell from the Kenya dome, while to the north the NW-SE trending Mesopotamian basin stands between the swell and the Bitlis Mts. (Fig. 4). While the low-lying nature of the Turkana region is the result of crustal stretching during the end of Mesozoic and Cenozoic (Kounoudis et al., 2023; Ogden and Bastow, 2022), the Mesopotamian lowland represents what remains of the foreland basin of the Zagros Fold and Thrust Belt and is ~900 km long and ~200 km wide (Berberian, 1995; Hessami et al., 2001).

Several studies demonstrated the presence of paleosurfaces both in the Arabia Peninsula (Avni et al., 2012; Bar et al., 2016) and in the Horn of Africa (Coltorti et al., 2007, 2015; Gani et al., 2007; Sembroni et al., 2016a; Sembroni and Molin, 2018; Sembroni et al., 2021). Avni et al. (2012) documented a regional truncation surface outcropping in the northern portion of the Red Sea and in the southern Levant region which separate middle Eocene – early Oligocene pre-rift deposits from late Oligocene – Holocene conglomerates and volcanic rocks. For this reason, the authors referred this surface to the Oligocene. Bar et al. (2016) extended such a surface to the western half of the Arabia Peninsula and defined it as a planation surface standing at elevation ranging between 800 and 1200 m (Arabian Plateau in Fig. 1). To the west and south, the Arabian Plateau is bounded by the elevated shoulders of the Red Sea and the Gulf of Aden rifts. To the north and east the plateau gently descends

toward the Persian Gulf and the Mesopotamian Basin. The northeastern portion of the Arabian Plateau comprises the Jordan Plateau, a large eastward-tilted area with summits at 1200–1700 m and elevation increasing southward (Fig. 1). The evolution of this structure seems to be linked to the formation of the Dead Sea depression (Salameh, 1997; Avni et al., 2012; Bar and Zilberman, 2016; Ben-Israel et al., 2020).

Similarly, at the opposite side of the Red Sea, the Horn of Africa presents remnant surfaces representing the preserved top of the basaltic plateau formed after huge eruptions occurred mainly in the Oligocene Trap phase (the “Trap volcanics planation surface” of Coltorti et al., 2007, or “paleotopography PT3” of Sembroni and Molin, 2018). These sub-horizontal ($< 3^\circ$) surfaces stand at an elevation comprised between 1000 m (in southern Ethiopia) and 2700 m (in central and northern Ethiopia; Sembroni et al., 2016a, 2016b) forming the so-called Ethiopian-Somalian Plateau (Fig. 1). This plateau is divided into the Ethiopian and Somalian portions by the NE-SW trending MER. Locally, large shield volcanoes rise 1000–2000 m above the average plateau elevation, the highest of which reach elevations > 4200 m. The Ethiopian Plateau consists of a flat to gently rolling landscape (Fig. 1). Conversely the Somalian Plateau is much less extensive and presents the flat top along the southeastern margin of the MER (Fig. 1). To the west and southeast of the high elevated plateaux, the topography gradually decreases down to 500 m respectively in the Sudan and Somalian lowlands (Fig. 1). According to several studies the incision of these uplands initiated by regional uplift in the Late Oligocene continuing up to present time (McDougall et al., 1975; Pik et al., 2003; Gani et al., 2007; Ismail and Abdelsalam, 2012; Gani, 2015; Sembroni et al., 2016b; Gani and Neupane, 2018; Sembroni and Molin, 2018; Sembroni et al., 2021; Gani et al., 2023). This long-lasting incision sculpted a landscape characterized by steep slopes bordering high-elevated low relief surfaces underlain by continental flood basalts (plateau remnants; Sembroni et al., 2016b).

The Ethiopian-Somalian and Arabian plateaux are separated by the NW-SE trending Red Sea which displays longitudinal and along-strike variations in rift flank morphology (Fig. 1). In particular, the western margin is characterized by the narrow, 1500 km long “Red Sea Hills” (mean elevation 500 m), while the conjugate Arabian escarpment presents highlands averaging 1000–1500 m in elevation.

Immediately to the north of the Mesopotamian basin, the East Anatolia Plateau consists of a dome-shaped feature characterized by a rolling low-relief topography standing at a mean elevation of ~ 2000 m (Molin et al., 2023; Fig. 1). Volcanoes rise from its surface, some of which reach 5000 m. To the north and south, the topography decreases down to sea level.

The hydrography of these high elevated region shows a typical radial pattern both in Ethiopia (Nile drainage system; Sembroni et al., 2021) and in Arabia (Fig. 1). An exception is the Eastern Anatolian Plateau where rivers integrate into the plateau in a disorganized way, locally following tectonic structures (i.e., the Euphrates River; Molin et al., 2023; Fig. 1).

4. Deep mantle processes beneath East Africa - Arabia

Uplift, rifting, and volcanism in the East Africa-Arabia region have been related to one or more plumes, based on geochemical, seismic, and other evidence (Ebinger et al., 1989; Camp and Roobol, 1992; Burke, 1996; Ebinger and Sleep, 1998; Courtillot et al., 1999; Rogers et al., 2000; Montelli et al., 2006; Rogers, 2006; Chang and Van der Lee, 2011; Fishwick and Bastow, 2011; Nelson et al., 2012; Chang et al., 2020; Civiero et al., 2022). Thin-lithosphere corridors have been proposed to assist in plume transport or channel plume material to volcanic fields hundreds of kilometers away (cf. Fig. 2; Camp and Roobol, 1992; Ebinger and Sleep, 1998; Faccenna et al., 2013; Agostini et al., 2021; Civiero et al., 2022; Hua et al., 2023).

The strongest geochemical evidence for the presence of plume material below the region is the lack of a prominent contribution from

shallow, depleted MORB mantle in the source material for pre-rift volcanism (Baker et al., 1996; Kieffer et al., 2004). Isotopic measurements in the basalts of southern Ethiopia, Afar, and Levant all indicate deep-mantle reservoirs, with a lithospheric component (e.g., Krienitz et al., 2009; Nelson et al., 2012).

Recently, Hua et al. (2023), according to a wide range of seismic and geochemical observations, inferred that the hot asthenosphere beneath Anatolia is fed by long-distance lateral transport of upper mantle from East Africa, with the lateral flow being driven by pressure gradients created by the buoyancy of the African mantle plume, consistent with the earlier inferences of Ershov and Nikishin (2004) and the mantle flow and seismic anisotropy modeling of Faccenna et al. (2013) (Figs. 2 and 5). In agreement with the suggested progressive northward motion of asthenospheric plume material, seismic body and surface wave imaging (Fig. 2) and seismic anisotropy of the upper mantle, e.g. as inferred from SKS splitting (Fig. 5) beneath East Africa and Arabia appear related to the combination of channelized flow from Afar along the Red Sea (e.g., Hansen et al., 2012; Bagley and Nyblade, 2013; Faccenna et al., 2013; Hammond et al., 2014; Wei et al., 2019), and oriented partial melt pockets of other rifting associated processed closer to the Ethiopian flood basalt province (e.g., Kendall et al., 2005; Bastow et al., 2010; Ebinger et al., 2024).

The associated low-velocity anomaly crosses the upper mantle under East Africa and continues propagating northeasterly to western Arabia (Figs. 2 and 5). Some tomography models show a low-velocity feature in the upper mantle extending from the Red Sea into the interior of Arabia (Debayle et al., 2001; Benoit et al., 2006a, 2006b; Fishwick, 2010; Emry et al., 2019) and a narrow anomaly aligned with the Red Sea but offset to the east (Chang and Van der Lee, 2011; Chang et al., 2011; Koulakov et al., 2016; Lim et al., 2020; Tang et al., 2018; Yao et al., 2017). In contrast, Civiero et al. (2022) proposed three mantle plumes beneath Kenya, Afar, and Levant feeding an integral East Africa-Arabia plume head. Low shear-wave velocities beneath the Gulf of Aden down to 150 km depth were also reported, suggesting that the plume below Afar may be also feeding the Gulf of Aden ridge (Sicilia et al., 2008).

If we assume that seismic anisotropy is due to shear in mantle flow, the updated SKS compilation of Fig. 5 is consistent with previous discussions about plume-related, and slab curtain modulated, asthenospheric channeling, and in particular the mantle flow and texture formation modeling of a more limited SKS dataset by Faccenna et al. (2013). However, the more extensive coverage of the Arabian Peninsula provides an opportunity to further explore details of channeling and deflection of flow by lithospheric structure, for example.

5. Methods and results

To describe the extension and geometry of the EAAS and to investigate the different components of topography, the present topographic configuration of the study area has been analyzed by several techniques including slope map, swath profiles, and filtered topography. To quantify the isostatic component of topography along the Red Sea rift shoulders, the flexural uplift has been calculated. The non-isostatic contribution to topography has been quantified by the realization of residual and dynamic topography maps. As elevation data source we used the ETOPO2022 global elevation model (resolution of 15 arc-second = ~ 500 m; www.ngdc.noaa.gov) because its relative fine resolution allows the analysis of topography at a regional scale. The data have been extracted and elaborated by means of ArcGIS and MatLab software.

To highlight a possible common trend between topography and volcanism age patterns, topography analysis results have been analyzed together with geochronological data. In detail, 1465 published ages of volcanic deposits from Ethiopia to Turkey have been collected and statistically analyzed.

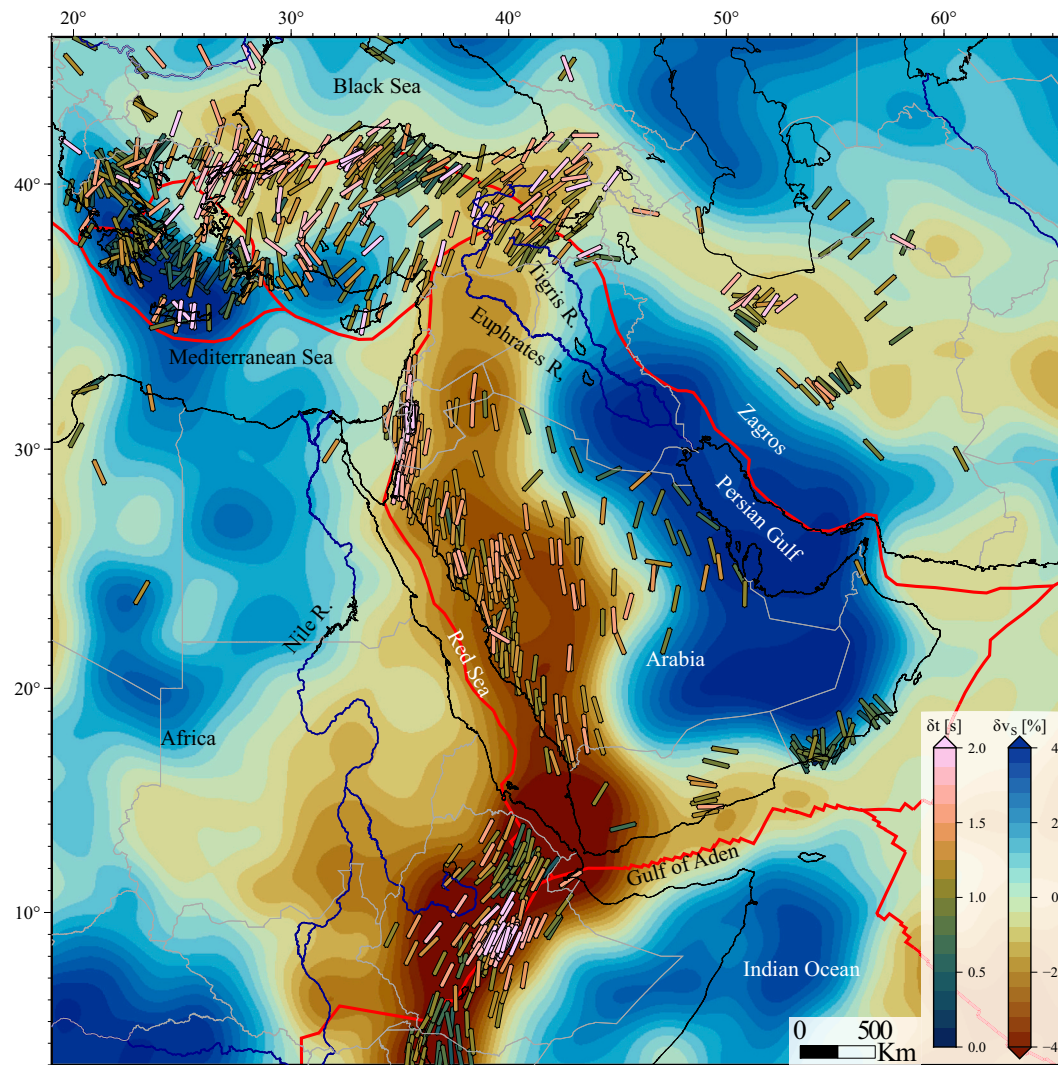


Fig. 5. REVEAL tomography model (Thrustarson et al., 2024) (see Fig. 2) averaged over the 100...400 km depth range (which dominates dynamic topography, cf. Fig. 10f) superimposed by SKS splitting observations aligned with “fast axes” and colored by delay time (compilation of Becker et al., 2012, updated as of 05/2024). Solid black show coastline, blue lines major rivers, and red line major plate boundaries. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

5.1. Slope map and filtered topographies

Slope maps have been widely used to mark morphologies, such as high-standing plateaux or steep escarpments, which are not immediately detectable from the observation of a DEM. In this study, a slope map from the ETOPO2022 DEM has been generated in ArcGIS environment by the “Slope” tool which identifies the variation in elevation over the distance (3×3 cell neighborhood) for each cell of the DEM. The tool, interpolating the gradient value of cell centers, generates a plane whose slope value is calculated using the average maximum technique (Burrough and McDonnell, 1998). In this study the values have been classified in four classes (Fig. 6): the first three classes describe flat or gently dipping portions of landscape with slope $< 15^\circ$; the last class covers the range 15° – 69° to emphasize escarpments.

The map shows the highest slope values ($> 15^\circ$) in coincidence with the margins of the Red Sea, the Afar and MER escarpments, along the DSTF, and in the mountainous regions in the southernmost (Ethiopia) and northernmost (Turkey, Armenia, Azerbaijan, Iran) portions of the study area (Fig. 6). The lowest values ($< 5^\circ$) concentrate on the remaining part and differentiate into low-elevated and high-elevated areas. The former ones correspond to the Nile, Tigris and Euphrates rivers valleys, the Persian Gulf surroundings, and the Horn of Africa

coastal zone. Conversely, the high-elevated sub-horizontal surfaces are in the inner sectors of Turkey (Anatolian Plateau) and Iran (Iranian Plateau), along the western margin of the Arabian Peninsula (Arabian Plateau), and in the Ethiopia (Ethiopian-Somalian Plateau).

Previous studies on the region (Almond, 1986; Dixon et al., 1989; Camp and Roobol, 1992; Sengör, 2001; Şengör et al., 2003; Forte et al., 2010; Faccenna et al., 2013; Sembroni et al., 2016a, 2016b, 2021) relate large-scale morphologies to mantle processes. To isolate and quantify this component, the present topography of the study area has been filtered in frequency domain by a circular low pass filter in ArcGIS environment. This methodology has been used in other parts of the world such as Yellowstone and Colorado Plateau (Wegmann et al., 2007; Roy et al., 2009), Eastern Africa (Sembroni et al., 2016b, 2021), Apennines and Carpathians (D’Agostino and McKenzie, 1999; Molin et al., 2004, 2012; Faccenna et al., 2011), and Easter Anatolia Plateau (Molin et al., 2023). To avoid flexural effects, the choice of the filter wavelength is important (Molin et al., 2012). In this study we used two wavelengths: 200 (Fig. 7a) and 400 km (Fig. 7b). Such values are effective in filtering out the topographic signals of crustal tectonics and fluvial wide valleys and to isolate the mantle component (D’Agostino and McKenzie, 1999; Molin et al., 2004, 2012; Wegmann et al., 2007; Roy et al., 2009; Faccenna et al., 2011; Sembroni et al., 2016b, 2021). In particular, the

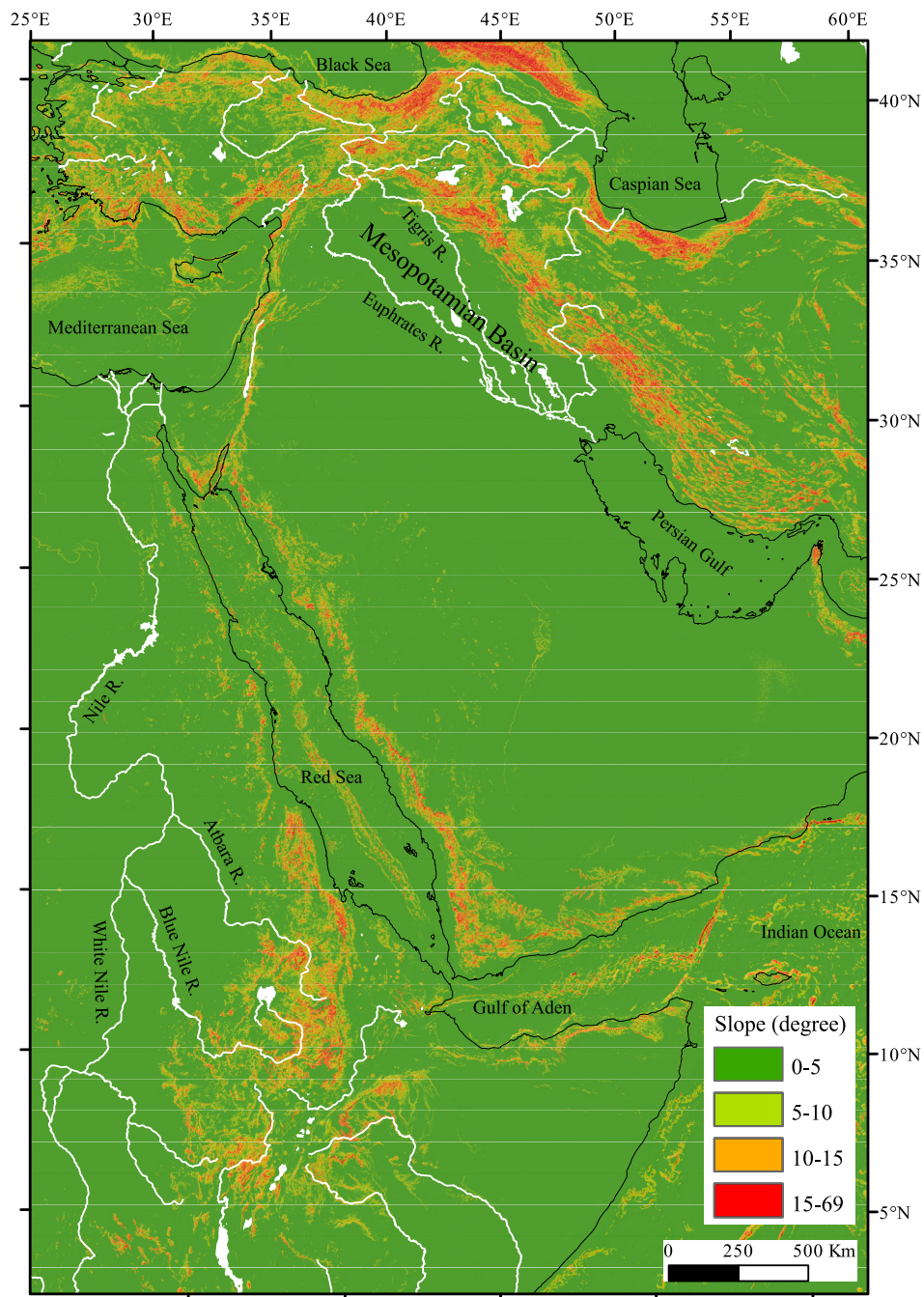


Fig. 6. Slope map of the study area elaborated from the ETOPO2022 global elevation model.

wavelength at 400 km allows to inhibit the topographic influence of the Red Sea, which presents a maximum width of ~ 300 km, and to discern the deeper component of the topography.

In the filtered topography at 200 km wavelength (Fig. 7a), values between 500 and 1000 m characterize a long strip of land from Ethiopia to Syria (NNW-SSE trend) interrupted only in coincidence with the Red Sea and the Gulf of Aden. Indeed, their maximum amplitude of 300 km exceeds the radius chosen for filtering the topography. We referred this topographic feature to the EAAS. Its highest elevation (1000–2500 m) falls in the Ethiopian-Somalian Plateau and in the southwestern corner of the Arabia peninsula. The swell is bordered to the west and east by lowlands at elevation comprised between 0 and 500 m represented, respectively, by the Nile River Valley and by the Mesopotamian Basin – Persian Gulf area. To the north, the filtered topography slightly

decreases in northern Syria and then dramatically increases up to 2500 m in coincidence with the Anatolian Plateau and the Zagros Mts.

In the 400 km filtered topography (Fig. 7b), the swell is still clearly visible presenting almost the same elevation pattern and geometry previously described. In this case the higher elevation (1000–2000 m) focuses exclusively on the Horn of Africa while the Arabian Peninsula shows a relatively homogeneous elevation between 500 and 1000 m.

5.2. Swath profiles

A swath profile is a distance vs. elevation plot in which the trend of maximum, minimum, and mean topography is represented within a specific rectangular observation window (swath). Swath analysis has been widely used to compare magnitudes of orogenic belts (Fielding,

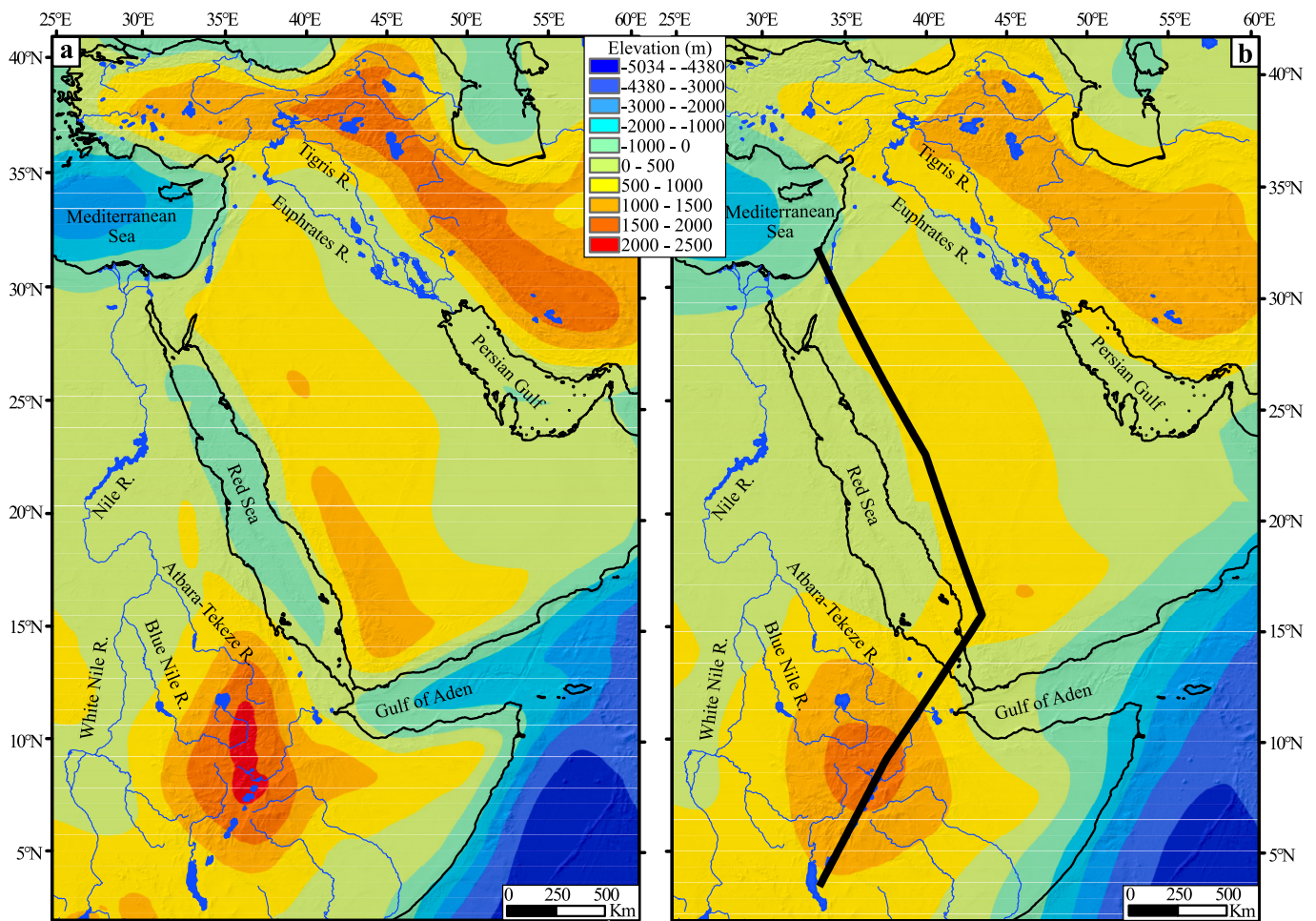


Fig. 7. Filtered topographies at 200 (a) and 400 km (b). The solid black line in (b) indicates the trace of the profiles represented in Fig. 13.

1996), to evaluate the relationship among erosion, precipitation and tectonic uplift (Thiede et al., 2004; Bookhagen et al., 2005; Hoke et al., 2005; Champagnac et al., 2009), to study the impact of rock erodibility on mountainous relief (Kühni and Pfiffner, 2001), and to catch drainage reorganization or fluvial terrace location (Godard et al., 2010; Stüwe et al., 2009; Wegmann and Pazzaglia, 2009). The difference between maximum and minimum topography at each point of the profile quantifies the local relief that, in a non-glaciated area, is a good approximation of fluvial incision (Isacks, 1992; Telbisz et al., 2013; see Fig. S2). Molin et al. (2004) showed that, in active tectonic landscapes, regions of high local relief often correspond with areas of incision in response to uplift.

To investigate the geometry of the EAAS, eight swath profiles have been extracted from the ETOPO2022 DEM by tracing 25 equally spaced topographic profiles into a 100 km wide observation window (Fig. 8). The elevation data along each profile have been sampled every 2 km. According to the configuration of the filtered topography at 200 and 400 km (Figs. 7a, b) we traced six profiles roughly orthogonal to the Red Sea (SW-NE trend), one orthogonal to the Main Ethiopian Rift and one along the crest of the swell, from Ethiopia to the border between Syria and Turkey. The first seven profiles clearly show a ridge-like topography interrupted by the Red Sea or the MER with a geometry varying from north to south (Fig. 8):

Profile 1. The ridge is ~800 km wide from Israel to the Euphrates-Tigris plain with a maximum elevation of ~800 m in coincidence with the Israel-Jordan Plateau. The higher peaks on the plateau correspond to the Harrat Ash Shaam volcanic province where volcanic edifices are 1600 m high.

Profile 2. The ridge considerably enlarges up to ~1000 km with a maximum elevation of ~1000 m occurring entirely on the Arabian margin of the Red Sea. As for Profile 1, the elevation gradually and smoothly decreases down to zero toward NE.

Profile 3. The ridge mostly extends in the Arabian margin with an amplitude of ~1100 km and an elevation of ~1000 m. In the Egypt margin, a small plateau (~200 km in width) at ~600 m of elevation may represent the northwestern portion of the ridge. The highest portion (~1000 m) is characterized by the edifices of three volcanic provinces named Harrat Qalib, Harrat Kura, Harrat Hutaymah arriving at ~2000 m of elevation (Coleman et al., 1983).

Profile 4. A pronounced ridge ~1200 km wide and ~1000 m high. ~200 km of the total width is observable in the Sudan margin while the remaining characterizes the Arabian one. The area with the maximum elevation is centered in the Arabian Plateau. Here the volcanoes of the Harrat Rahat and Harrat Al Kishb volcanic provinces tower up to an elevation of ~1500 m (Coleman et al., 1983).

Profiles 5 and 6. To the southernmost portion of the study area (Profiles 5 and 6) the ridge is first centered in the Red Sea (Profile 5) and then mostly in the southwestern margin (Profile 6) with a maximum width of ~1600 km and an elevation reaching ~2500 m in coincidence with the Ethiopian-Somalian and Yemen plateaux.

Profile 7. Orthogonal to the Main Ethiopian Rift and shows the southernmost portion of the swell. In particular, the bulge presents the highest elevation (~2700 m) and extends for >1300 km from the Sudan and Somalia lowlands, in accordance with previous studies (Sembroni et al., 2016b, 2021). The peaks seen in the maximum topography correspond with the major volcanoes that lie on the Ethiopian-Somalian

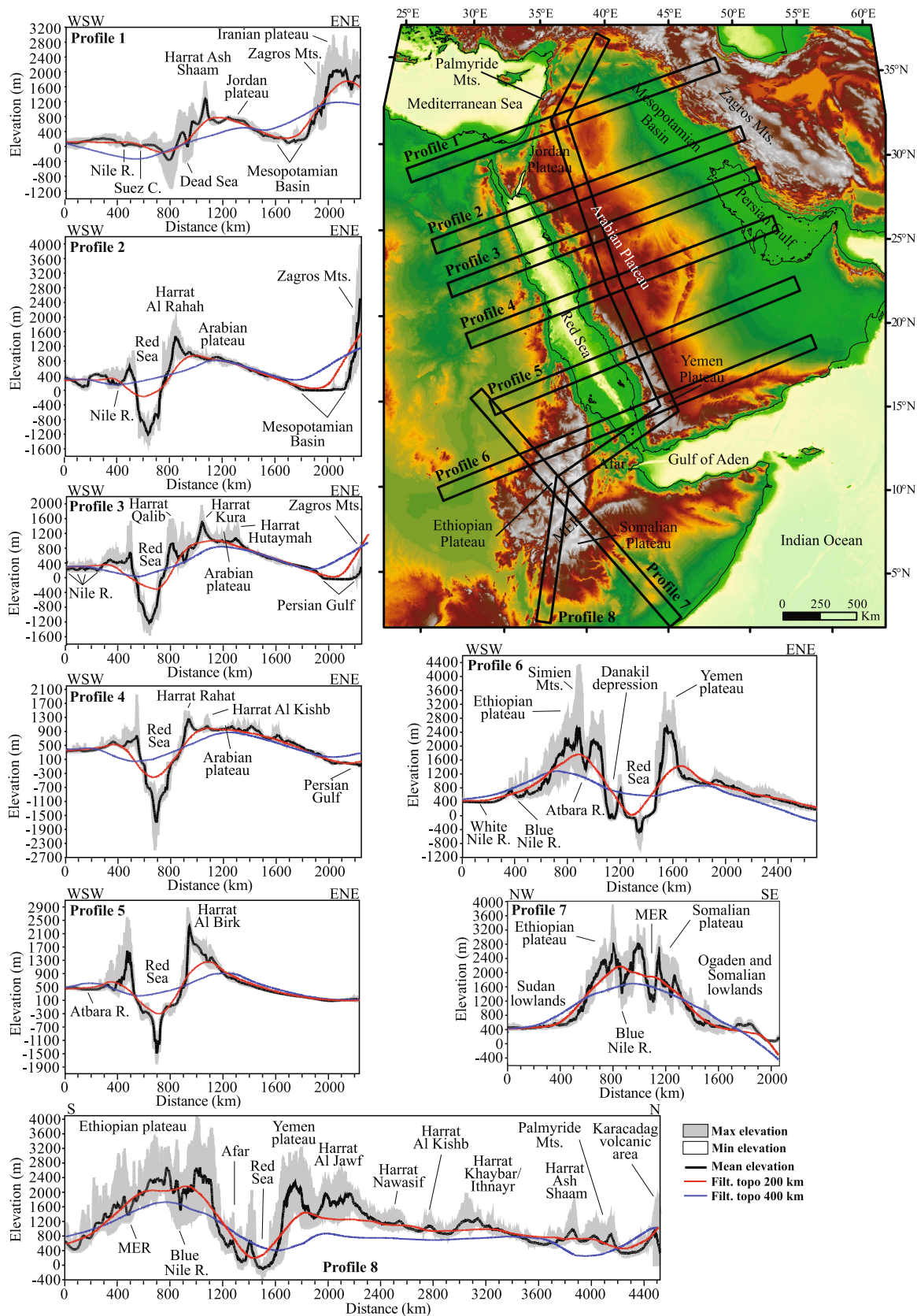


Fig. 8. Swath profiles extracted across the study area. To compare the general topographic configuration with the filtered topography, two curves extracted from the topography filtered at 200 and 400 km (Figs. 7a, b) along the middle line of each swath have been included in the plots. The trace of each swath is represented in the upper right map.

plateau.

Profile 8. Across the crest of the topography anomaly highlighted in the filtered topographies at 200 and 400 km (Figs. 7a, b). The EAAS extends for a total length of ~4000 km from Ethiopia to Jordan with a mean elevation gradually decreasing from ~2700 m (Ethiopia) to ~600 m (Jordan). The surface of the swell is characterized by the high-standing Ethiopian-Somalian and Yemen plateaux and by several volcanic provinces (harrats) in the Arabian portion which make its top surface irregular.

In all profiles, the local relief is low (< 200 m) on the ridge top and flanks, especially in the Arabian one, whereas is moderate to high (400–1000 m) in coincidence with the Red Sea and the MER and Afar margins (Fig. S2).

To better describe the topographic configuration of the swell, two curves extracted from the topography filtered at 200 and 400 km along the middle line of each swath have been included in the swath profiles (Fig. 8). The first curve (200 km in wavelength) roughly approximates the mean topography except for the Red Sea area where shorter wavelengths of topography are dominant. The second curve (400 km in wavelength) appears coincident with the mean topography of the Arabian (eastern) portion of the swell while deviates from the general pattern in the western part. It has negative values at the Indian Ocean coast.

In summary, the analysis of topography by swath profiles allows us to quantitatively define the geometry of the EAAS. It is confined between the Egypt-Sudan lowlands to the SW and the Euphrates-Tigris plain to the NE and characterized by an amplitude and an elevation gradually increasing from north (800 km and 800 m) to south (1600 km and 2700 m). The same trend is present in the topography at the Red Sea margins where the northern portion shows peaks approaching ~1000 m (southwestern margin) and ~2000 m (northeastern margin) of elevation which gradually increase at ~2500 m and ~2900 m respectively to the south. According to several studies (see Stüwe et al., 2022 and references therein) these high elevated portions along the margins of the Red Sea have been caused by the flexural uplift of the rift shoulders whose effect over time has been amplified by complex feedback between seafloor spreading and erosion. The same process has been studied by Weissel et al. (1995) and Sembroni et al. (2016a, 2016b) along the Afar escarpment and the Main Ethiopia Rift margins. The authors found increasing values of flexural uplift from ~500 m in the north to ~1200 m in the south.

5.3. The flexural uplift on the margins of the Red Sea

Although the EAAS represents a long wavelength signal, the opening of the Red Sea rift, with its maximum extent of ~300 km, may have modulated the expression of the swell partially hiding its topographic signal. To quantify such a topographic component, the uplift of rift flanks has been modeled as a flexural response to the unloading of the lithosphere due to extension.

Flexural deflections of the Earth lithosphere can support loads elastically on short wavelengths rather than having an isostatic floating equilibrium, which we expect to hold on long wavelengths, all relative to the effective elastic thickness. Elastic flexure with infill is governed by the equation (Turcotte and Schubert, 1982):

$$\nabla^2 [D(x,y)\nabla^2 w(x,y)] + \Delta\rho g w = q(x,y) \quad (1)$$

where $w(x,y)$ is the deflection of the surface, $D(x,y)$ the flexural rigidity, $\Delta\rho$ the density contrast between the crust and the mantle, g the acceleration due to gravity, and $q(x,y)$ the vertical load applied or removed from the lithosphere. The solution of eq. (1) for the case of a broken lithosphere is (Turcotte and Schubert, 1982):

$$w(x) = w_0 e^{-x/\alpha} \cos \frac{x}{\alpha} \quad (2)$$

$$D = \frac{ET_e^3}{12(1-\nu^2)} \quad (3)$$

$$\alpha = \left(\frac{4D}{g(\rho_m - \rho_a)} \right)^{1/4} \quad (4)$$

where w_0 is the maximum vertical displacement caused by the load, α the flexural wavelength, T_e is the equivalent elastic thickness of the lithosphere, ρ_m mantle, and ρ_a air density.

In this study, nine topographic profiles (see Figs. S3, S4), extracted from a smoothed topography (30 km radius smoothing circular window – to remove peaks related to the biggest volcanic edifices), have been fitted by eq. (2) using an iterative, nonlinear least-squares fitting algorithm (as implemented in MatLab). Confidence bounds for fitted coefficients and prediction bounds for the fitting curve reflect a 95% confidence interval (see Fig. S4). Lastly, the data extracted along each profile have been interpolated by a triangulation algorithm in ArcGIS environment obtaining a TIN surface (Fig. 9a).

The results indicate a progressive increase in flexural uplift from north to south on both the Red Sea margins, except for the northernmost portion of the African margin (Fig. 9a). This is in accordance with thermochronological data which dated the beginning of rifting at 26–20 Ma in the southern portion of the Red Sea (Boone et al., 2021 and references therein). On average, the western margin is characterized by lower values (415–3382 m) than the eastern one (1447–2771 m). The deformation related to flexural (flexural wavelength) is consistent up to a maximum of ~250 km from the Red Sea margins (Fig. 9a; Table S1). The inferred elastic thickness varies from 13 to 76 km on the African side (37.7 km average) and from 20 to 29 in the Arabian one (25.2 km on average; Table S1) in agreement with previous studies (Chen et al., 2015; Sreenidhi et al., 2023). In particular, Chen et al. (2015) correlate the relatively small T_e values and the low S-wave velocity anomaly with the elevated topography of the region concluding that the uplift is supported by hot mantle.

To distinguish the different components of topography across the Red Sea area, a topographic profile has been extracted along the southern portion of the Red Sea where the flexural uplift is higher (Fig. 9a). Together with it we plotted the flexural uplift extracted from Fig. 9a and the two curves from the filtered topographies (Figs. 7a, b). The graph shows that the present topography is dominated by the flexural component only in the first ~200 km from the rift shoulders. Beyond that distance, the topography is better approximated by the filtered topography at 400 km (Fig. 9b), consistent with the expectation of dynamic or isostatic control of topography at long wavelengths.

5.4. Residual, dynamic topography, and river networks

Filtered analysis of actual topography can be complemented by considering anomalous topography relative to what is expected isostatically from a given lithospheric model, i.e. residual topography (e.g., Panasyuk and Hager, 2000). Our approach for estimating residual topography follows the approach of Becker et al. (2014) and Faccenna and Becker (2020). As in Faccenna and Becker (2020), we use a modification of the CRUST1 (Laske et al., 2013) global crustal thickness model which we continuously update, e.g. earlier by merging data from the compilation of Gvirtzman et al. (2016) and other studies. Compared to Faccenna and Becker (2020), we here also include a few newer studies as compiled by Boyce et al. (2023) for an updated Moho map that is shown in Fig. 10a (see Boyce et al., 2023, for references). A smoothed version of that new crustal thickness model is then used along with CRUST1 density anomalies to estimate residual, i.e. non-isostatic, topography, ignoring any lateral lithospheric mantle density or thickness variations, for simplicity.

The resulting, updated residual topography map (Fig. 10b) is overall very similar to that of Faccenna and Becker (2020) and Stephenson et al.

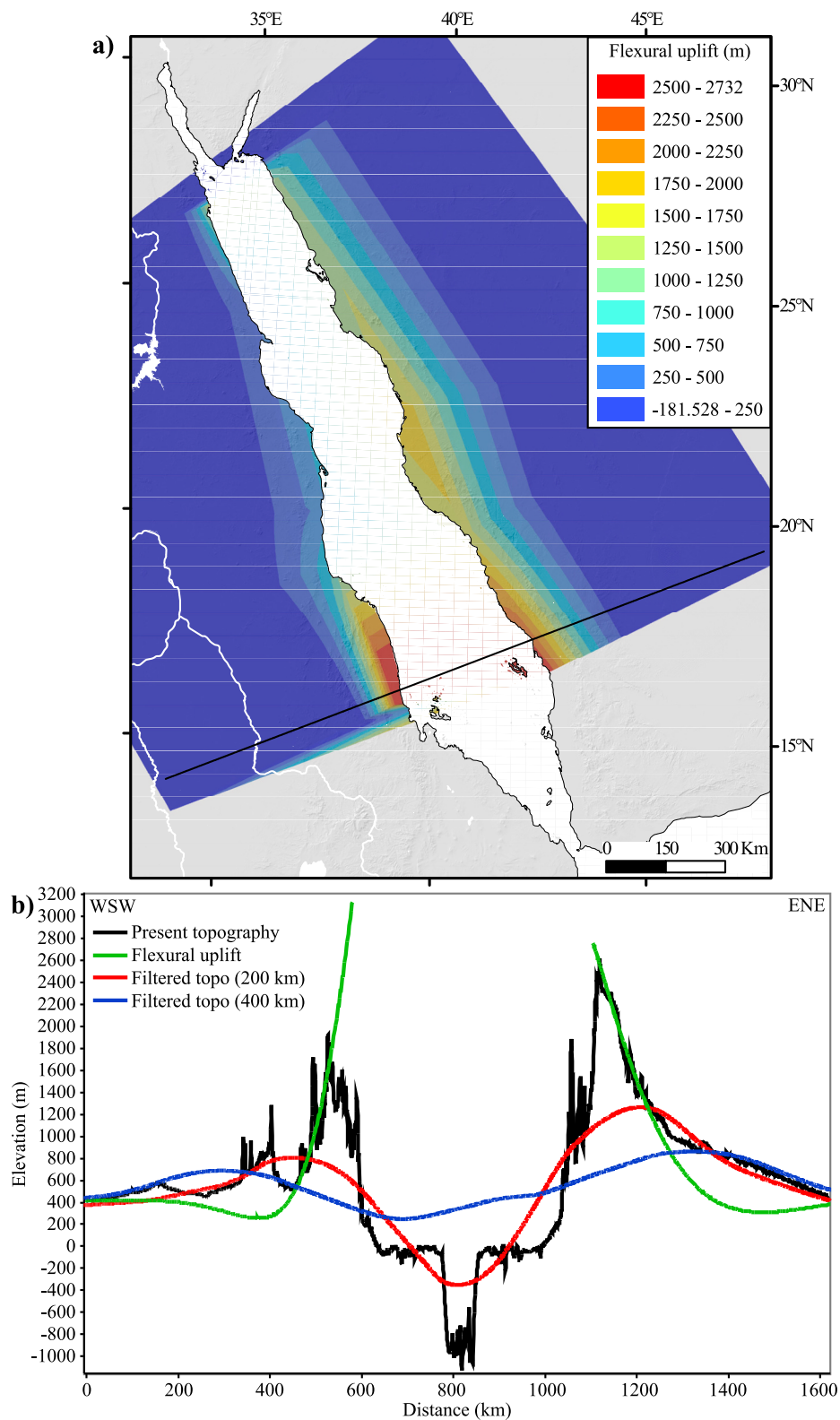


Fig. 9. a) Map of the flexural uplift calculated along the Red Sea region; the solid black line represents the trace of the profile shown in Fig. 9b; b) topographic profile extracted from the ETOPO2022 global elevation model. Along the same trace the flexural uplift and the filtered topographies at 200 and 400 km (Figs. 7a, b) have been extracted for comparison. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

(2024); it shows positive anomalies (0–2000 m) concentrated in eastern Africa and along the Red Sea and the western portion of the Arabia Peninsula. The highest values (>2000 m) characterize the Main Ethiopian Rift, the southwestern portion of the Arabian Peninsula, and the

Red Sea. Values between 0 and 2000 m are found to the west of the Nile River, in the central portion of the Arabian Peninsula, and in the western and central sectors of Anatolia. Negative values appear along the Mesopotamian Basin and the Persian Gulf area (–3000–0 m), in coincidence

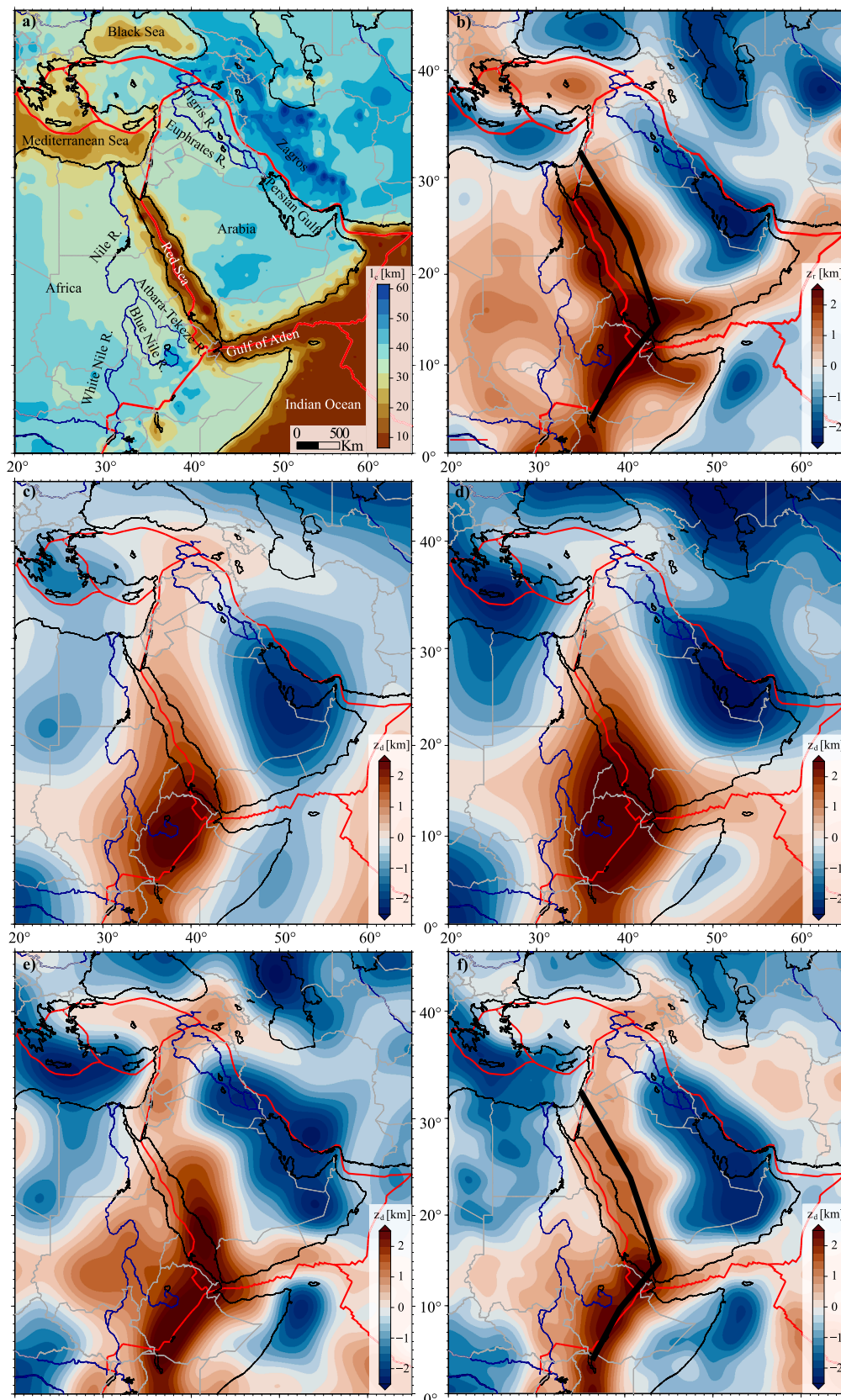


Fig. 10. a) Moho map based on a merger of CRUST1 (Laske et al., 2013) with constraints from regional studies (e.g. compiled in Gvirtzman et al., 2016), as in Faccenna and Becker (2020), but updated to also include the database of Boyce et al. (2023). b) residual topography based on crustal thickness and density variations, computed as in Faccenna and Becker (2020), but using a smoothed version of the new Moho map in a); c-e) dynamic topography based on mantle flow driven by scaling anomalies from TX2019 (c), SAVANI (d), SL2013 (e), and REVEAL (f, cf. Fig. 5), i.e. the seismic tomography models shown in Fig. 2. Residual and dynamic

topography computations follow Becker et al. (2014) and Faccenna and Becker (2020) but the wave speed anomaly to density scaling factors is reduced by 0.7 for models e) and f). The thick solid black line in b) and f) indicates the trace of profile represented in Fig. 13.

of the Zagros-Bitlis mountain belt (−1000–0 m), in the eastern sector of the Mediterranean Sea comprising the Nile River delta area (−1500–0 m), and in the western portion of the Indian Ocean (−1500–0 m), where we have not corrected for the effects of seafloor age dependent half-space cooling.

We can compare this lithospheric based estimate of anomalous residual topography with what deflections of the surface might be caused by mantle flow. At long wavelengths which allow ignoring flexural effects, those two anomalous topography estimates should match, assuming perfect knowledge of structure, density anomalies, and transport properties (see, e.g., discussion in Becker et al., 2014). We compute dynamic topography from a simplified, global mantle flow model with only radial viscosity variations by converting radial stresses to equivalent topography (e.g., Panasyuk and Hager, 2000), following the approach of Becker et al. (2014). On long wavelengths, this is a good approximation to the effects of mantle flow since dynamic topography of the surface is primarily sensitive to density anomalies in the uppermost mantle (e.g., Panasyuk and Hager, 2000), with only minor modifications due to lateral viscosity variations (e.g. Becker et al., 2014). Density anomalies are inferred by scaling seismic tomography with a constant scaling factor for all depths below 100 km, following Faccenna and Becker (2020), for simplicity. The seismic tomography models we use are, as in Fig. 2, TX2019 by Lu et al. (2019) (Fig. 10c), SAVANI by Auer et al. (2014) (Fig. 10d), SL2013 by Schaeffer and Lebedev (2013) (Fig. 10e), and REVEAL by Thrastarson et al. (2024) (Fig. 10f). Shear wave anomalies of models SL2013 and REVEAL have higher RMS than TX2019 or SAVANI and were scaled down by a factor of 0.7 to make overall anomalies comparable, sidestepping the general, unresolved issues of uncertain mineral physics scaling and uneven and model-dependent seismological anomaly recovery.

We therefore focus more on pattern than amplitudes, and the dynamic topography predictions based on the more recent, higher resolution models (Figs. 10e, f) indicate similar patterns to the residual topography one, implying that there is indeed a significant contribution from mantle processes to the non-isostatic, anomalous parts of topography. Moreover, comparison of the averaged shear wave velocity structure of Fig. 5 with the actual dynamic topography from flow prediction of the same model (Fig. 10f) highlights the aforementioned upper mantle density anomaly control of dynamic topography.

Positive values are concentrated along a roughly S–N trending strip from Ethiopia to Anatolia. The largest dynamic topography anomalies are found in part of Ethiopia, Eritrea, Djibouti, and Yemen. Positive anomalies up to 2000 m cover most of Sudan, the Red Sea, the western portion of the Arabian Peninsula, and the Middle East, up to Anatolia (Figs. 10c, d, e, f). Negative anomalies characterize the Somalian coast of Indian Ocean, northern Africa, eastern Mediterranean, and the Mesopotamian Basin-Persian Gulf area. This updated analysis broadly confirms earlier inferences (e.g. Daradich et al., 2003; Moucha and Forte, 2011; Chen et al., 2015; Faccenna and Becker, 2020), but Fig. 10 indicates an encouraging, improved regional match of residual and dynamic topography patterns.

The region represented in Fig. 10 is drained by two main drainage systems: the Nile and the Euphrates-Tigris ones. The Nile River network is characterized by three main trunks: the White Nile, the Blue Nile, and the Atbara-Tekeze. While the first joined the main river network only in the Late Pleistocene (Williams, 2019), the Blue Nile and the Atbara-Tekeze rivers have a long history started at least in the Oligocene (Pik et al., 2003; Sembroni et al., 2016a; Faccenna et al., 2019; Sembroni et al., 2021). Indeed, geological and geophysical data (Garzanti et al., 2006; Padoan et al., 2011; Sembroni et al., 2016a; Faccenna et al., 2019) suggest that the deep mantle processes related to the upwelling of the Afar superplume created a stable topographic gradient which made the

Blue Nile and the Atbara-Tekeze rivers stably connected to the Nile since at least 30 Ma.

Similarly, data from fluvial terraces (Demir et al., 2007; Stow et al., 2020) constrain the first appearance of the Euphrates River at the middle Miocene when the river flowed into the sea in southern Turkey. Successively, due to the regional uplift, the coast has shifted successively to the southeast resulting in the current path of the river (Demir et al., 2007; Stow et al., 2020). Demir et al. (2008), considering ~270 m of incision since the early Late Miocene (~9 Ma), estimated ~600 m of surface uplift on this timescale.

Plotting the drainage network on Fig. 10 we note that the Blue Nile-Atbara and the Euphrates-Tigris river networks source from areas of high positive residual and dynamic topographies (respectively the Ethiopian and the Eastern Anatolian plateaux) and end in regions of very low negative values (respectively the eastern Mediterranean Sea and the Persian Gulf).

Moreover, the area occupied by the swell (see Figs. 7 and 8) is characterized by the lowest values of lithosphere thickness in the region (Gvirtzman et al., 2016; cf. Fig. 2) and the highest residual and dynamic topographies (Fig. 10). P- (Benoit et al., 2006a; Hansen et al., 2012; Wei et al., 2019; Boyce et al., 2021, 2023) and S-wave (Benoit et al., 2006b; Priestley et al., 2008; Chang and Van der Lee, 2011; Schaeffer and Lebedev, 2013; Auer et al., 2014; Chen et al., 2015; Lu et al., 2019; Thrastarson et al., 2024) tomography indicates a low velocity anomaly at depths between 110 km and 300 km (Fig. 2). Moreover, the estimation of gravity effect of sediments all over the area results in a negative gravity anomaly region coinciding with the swell (Chen et al., 2015).

5.5. The geometry of the swell top surface

Previous studies demonstrated that during the Oligocene, several portions of the study area were characterized by low-relief surfaces which at present cover an area comprised of Ethiopia and the Levant (Coltorti et al., 2007; Gani et al., 2007; Avni et al., 2012; Bar et al., 2016; Sembroni et al., 2016a, 2016b; Sembroni and Molin, 2018; Sembroni et al., 2021). These surfaces are different in origin: the Levant and western Arabia ones are part of a wide planation surface separating the late Eocene – early Oligocene sediments from the late Oligocene - Holocene deposits (Avni et al., 2012; Bar et al., 2016) while the eastern Africa surface represents the top of the Oligocene basaltic plateau emplaced during or immediately after the impingement of the Afar superplume (Sembroni et al., 2021; Ebinger et al., 2024). At present, the surfaces stand at an elevation between 800 m (northern portion of the study area) and 2700 m (southern part).

To reconstruct the envelope of these surfaces, we mapped all the surfaces remnants by a spatial query in an ArcGIS environment (Fig. S1a). All surfaces with a slope lower than 3° and standing at an elevation ranging between 800 and 1200 m in the Arabian Peninsula and between 1000 and 2700 m in the Horn of Africa have been isolated (Fig. S1a). The surfaces have been then interpolated as a triangulated irregular network (TIN; Fig. S1b) and then smoothed by a circular low pass filter 100 km in radius to remove the artifacts related to the triangulation algorithm.

The result shows a surface whose trend changes from NE-SW in the southern portion to NNW-SSE in the northern one (Fig. 11). The elevation gradually decreases from south (2500 m) to north (900 m) for a total distance of ~4000 km from Ethiopia to Jordan. In Fig. 12 some images taken from Google Earth show examples of portions of the surface. This feature occupies the same region characterized by negative gravity anomaly (Chen et al., 2015), by P- and S-wave low velocity anomaly (Kendall et al., 2005; Benoit et al., 2006a, 2006b; Priestley et al., 2008; Bastow et al., 2010; Chang and Van der Lee, 2011; Hansen

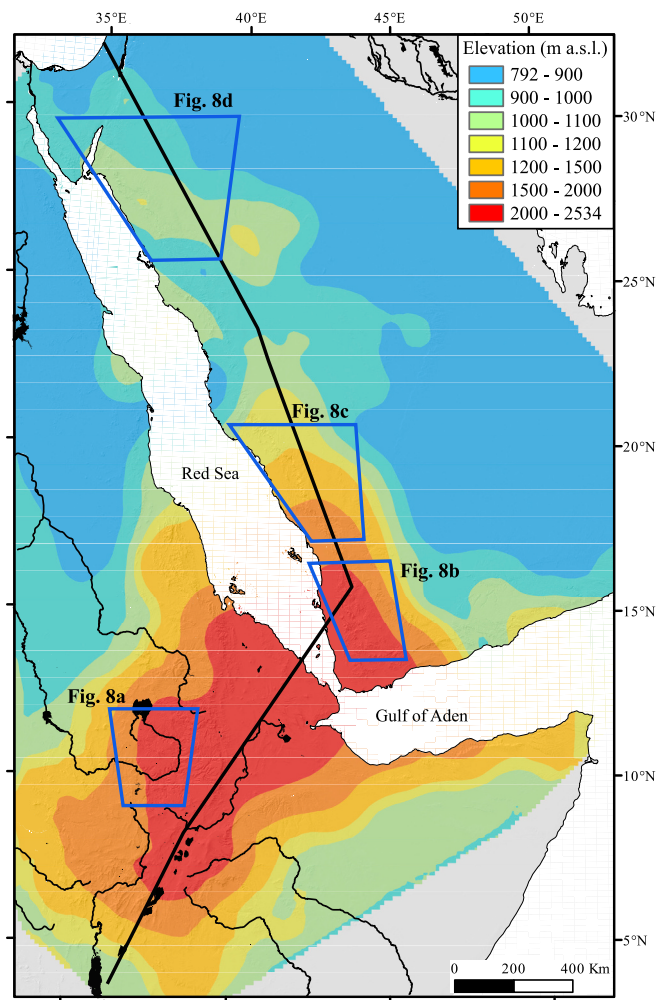


Fig. 11. a) The top surface of the East African-Arabia swell obtained by interpolating the isochronous high-standing surfaces mapped between Ethiopia and Jordan. The thick solid black line indicates the trace of the profile represented in Fig. 13.

et al., 2012; Schaeffer and Lebedev, 2013; Auer et al., 2014; Chen et al., 2015; Lu et al., 2019; Reiss et al., 2019; Wei et al., 2019; Chang et al., 2020; Hosseini et al., 2020; Andriampomanana et al., 2021; Boyce et al., 2021, 2023; Saeidi et al., 2023; Ebinger et al., 2024; Thrastarson et al., 2024; Fig. 2), and by high residual and dynamic topographies (Fig. 10).

Lastly, to compare the geometry of the swell top surface with the trend of the filtered, residual, and dynamic topographies a profile for each has been extracted by using the same trace represented in Figs. 7b, 10b, f, and 11 (Fig. 13). The swell top surface profile matches quite well the residual and dynamic topographies. The same decreasing trend, even if with lower values, characterizes filtered topography at 400 km where the minimum at ~1250 km coincides with the Gulf of Aden-Red Sea rift system whose signal cannot be detected in the swell top surface. In summary, all the considered topographies show a decreasing trend from south to north with maximum values concentrated in the area comprising Ethiopia, Eritrea, Djibouti, and Yemen (Fig. 13).

5.6. The distribution and age of volcanic deposits

Since the Eocene, the study area was affected by widespread volcanism. To investigate the distribution and a possible age pattern of volcanic deposits, a compilation of 1465 volcanic samples has been analyzed expanding the databases previously published (Camp and

Roobol, 1992; Bosworth and Stockli, 2016; Civiero et al., 2022; Hua et al., 2023; Fig. 14; see supplementary material for references – Table S2). Each sample is classified according to the country where it was sampled and the database has been cleared of all those deposits whose geochemical analyses showed a clear derivation from subduction processes (the “orogenic” deposits of Lustrino and Wilson, 2007). The study area has been divided into two parts: the southern portion (Ethiopia, Djibouti, Yemen, Saudi Arabia; Fig. 15a) and the northern one (Egypt, Israel, Jordan, Syria, Turkey; Fig. 15b).

The age frequency plots for both areas show a roughly bimodal distribution in accordance with previous studies (Camp and Roobol, 1992; Bosworth and Stockli, 2016; Civiero et al., 2022; Figs. 15a, b). In particular, the southern area presents the age ranges of 18–42 Ma and 0–16 Ma with peaks comprised between 28 and 32 Ma and 0 and 4 Ma respectively (Fig. 15a). These two ranges are separated by a brief period (16–17 Ma) of low volcanic activity well documented all over Saudi Arabia (Camp and Roobol, 1992; Bosworth and Stockli, 2016). In general, Ethiopia and Saudi Arabia are characterized by a quasi-continuous volcanic activity, while the ages of volcanic deposits in Yemen and Djibouti are comprised only between 32 and 20 Ma and 0 and 4 Ma, respectively (Fig. 15a).

In the northern area the bimodal distribution is shifted toward younger ages (Fig. 15b). In detail, the first range falls between 10 and 22 Ma while the second one is comprised between 0 and 8 Ma, with a remaining small number of samples with older ages (Fig. 15b). In contrast to the southern area, here the two ranges are separated by a sharp decrease in number of samples with ages comprised between 8 and 10 Ma. In general, except for Egypt, whose volcanic deposits ages vary between 18 and 32 Ma, in the remaining countries the volcanism is overall continuous (Fig. 15b).

The presence of young volcanic deposits in both areas may be in part because such deposits, outcropping at or near the surface, could cover the older ones. To investigate a possible pattern of volcanics age from the southern to the northern portions, the age data have been plotted against latitude following the example of previous studies (e.g., Bosworth et al., 2005; Bosworth and Stockli, 2016; Boone et al., 2021; Civiero et al., 2022). The result indicates a decrease in the maximum age of volcanic deposits from Ethiopia to Turkey with a plateau in coincidence with the Saudi Arabia data (Fig. 15c). By plotting on the same graph, the curve of filtered topography at 400 km extracted along the axis of the EAAS (Figs. 7b), it is possible to see a similar trend (excluding the Red Sea minimum) with elevation decreasing from ~1700 m in Ethiopia to less than zero in Jordan.

6. Discussion

The EAA swell covers the region from Ethiopia to Jordan crossing the Gulf of Aden and the Red Sea. Its configuration consists of a NNW-SSE trending ridge with amplitude and height decreasing from south to north. Most of volcanoes and basaltic lava fields are located along its axis. In general, the ridge appears asymmetric with respect to the Red Sea. This is possibly related to crustal thickening due to magmatic underplating as indicated by recent geophysical analysis in the central portion of the Arabian margin of the Red Sea which evidenced the presence of a magmatic body in the lower crust atop an underplated Moho (Mukhopadhyay et al., 2023). That portion is characterized at the surface by large volcanoes and basaltic lava fields dated between Tertiary and Quaternary (Coleman et al., 1983; Camp and Roobol, 1992). This configuration appears similar to the Main Ethiopian Rift one where the rift opened immediately to the east of the Ethiopian Plateau affected by a strong underplating during the Trap phase (Rooney, 2017), as evidenced by seismic refraction and receiver function data (Mackenzie et al., 2005; Tiberi et al., 2005).

The continuity of the swell is interrupted to the north in coincidence with the Mesopotamian Basin, considered as the foreland basin of the Bitlis-Zagros mountain belt. Here, a wide depression extending from

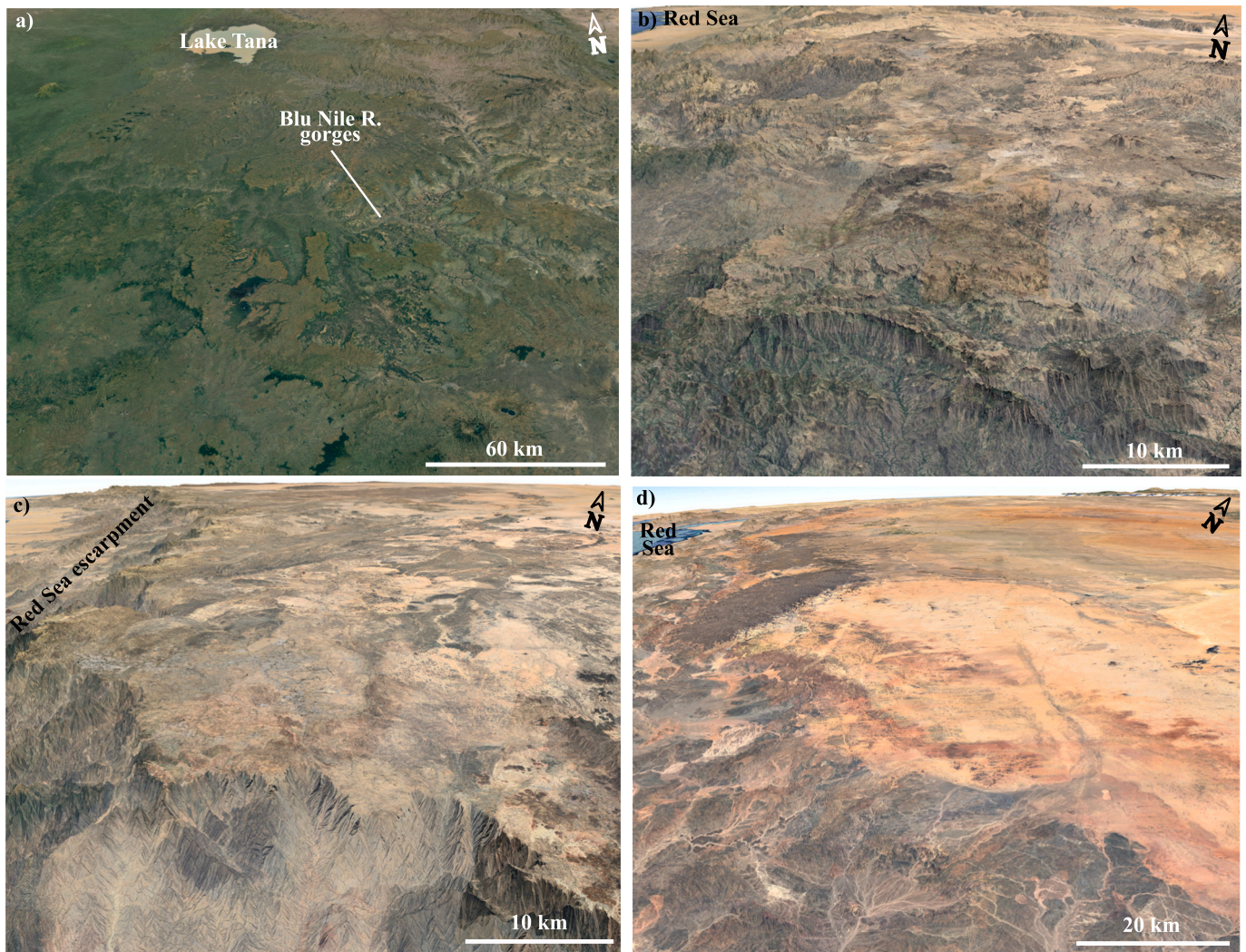


Fig. 12. Images from Google Earth showing portions of the Oligocene surface from Ethiopia (a), Yemen (b), western Saudi Arabia (c), and northern Saudi Arabia (d). Note the low-relief top surfaces in all figures which represent portions of the swell top surface. See Fig. 11 for the exact location.

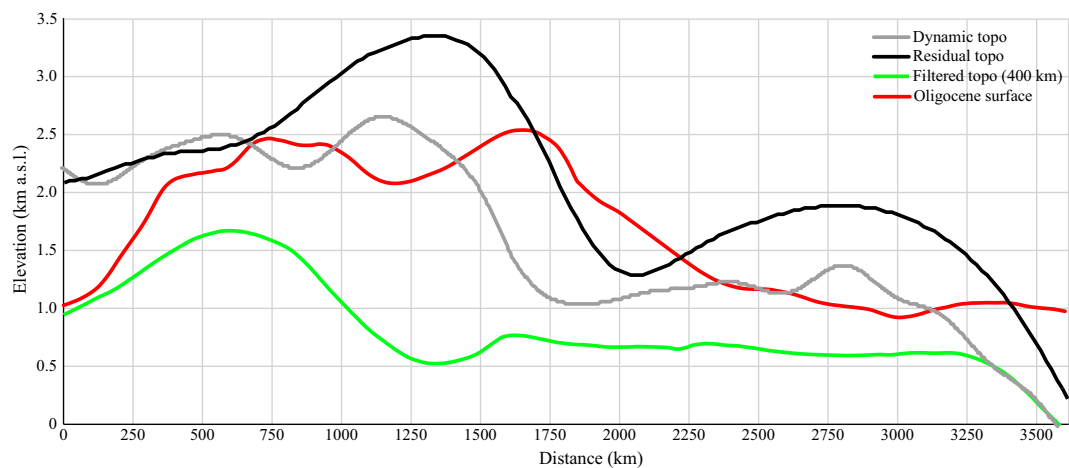


Fig. 13. Plot showing the trends of dynamic, residual, and filtered topographies (400 km) with respect to the Oligocene swell top surface along the solid black line drawn in Figs. 7b, 10b, f, and 11.

Syria to the Persian Gulf may be related to the depression of the lithosphere by subduction of the basement of the Arabian plate under the sedimentary rocks of the same plate (e.g., Yeats, 2012). This seems

confirmed by the maps of dynamic topography (Figs. 10c, d, e, f) which, in accordance with previous studies (Gvirtzman et al., 2016; Steinberger, 2016; Faccenna et al., 2019; Steinberger et al., 2019; Lu et al.,

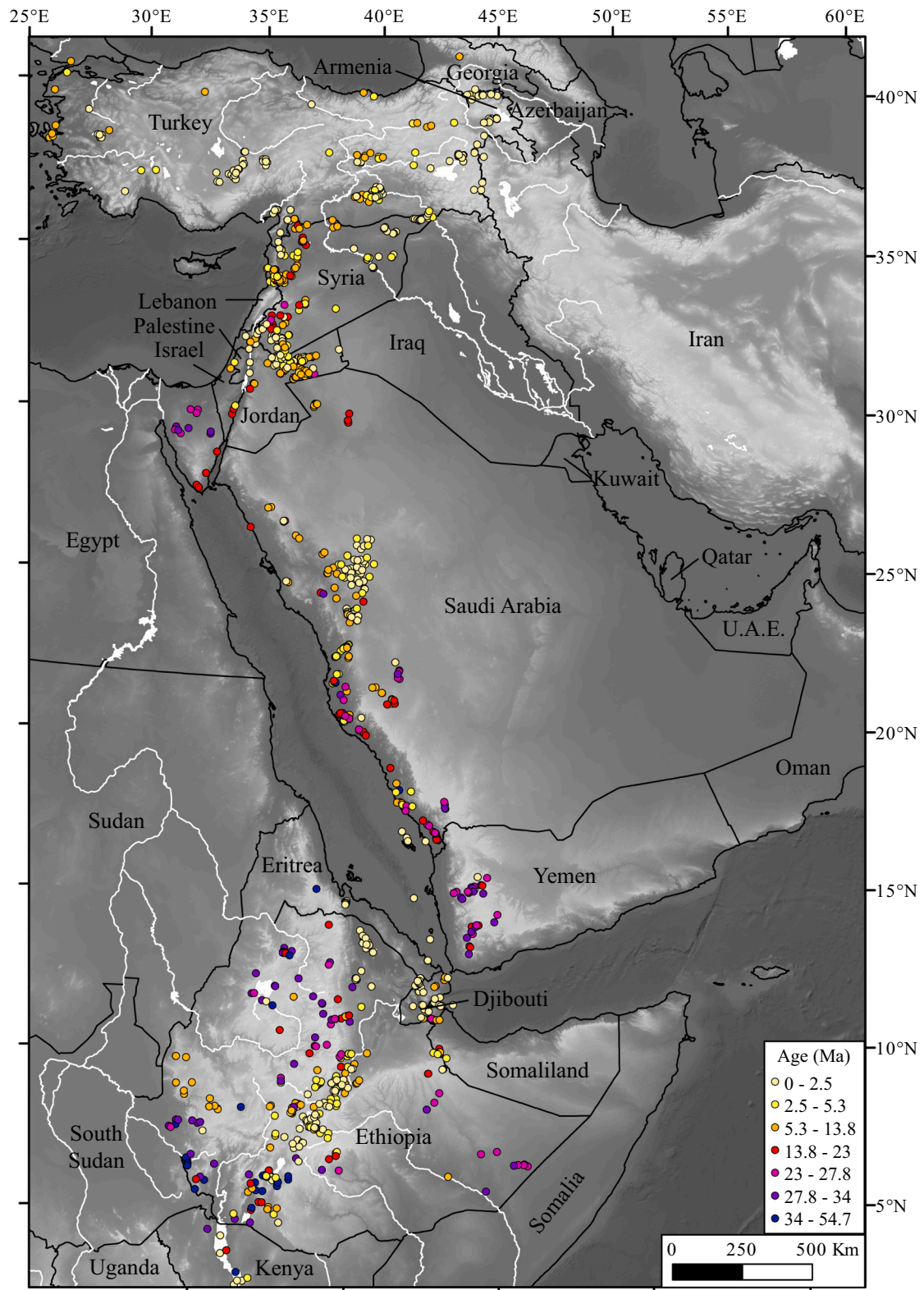


Fig. 14. Distribution of volcanic deposits along the study area distinguished by age.

2019), indicates negative values in the area. To the northwest, the swell is bounded by the easternmost portion of the Mediterranean basin characterized by negative values in filtered topography (Figs. 7a, b). As for the case of the Mesopotamian Basin, this area presents negative dynamic topography (Figs. 10c, d, e,f) related to the depression of the

lithosphere induced by slab sinkers (Forte et al., 2010; Gvirtzman et al., 2016; Steinberger, 2016; Faccenna et al., 2019; Steinberger et al., 2019; Lu et al., 2019).

The incision (local relief) is low both on the top and on the flanks of the swell with the exception of the areas close to the Red Sea and Main

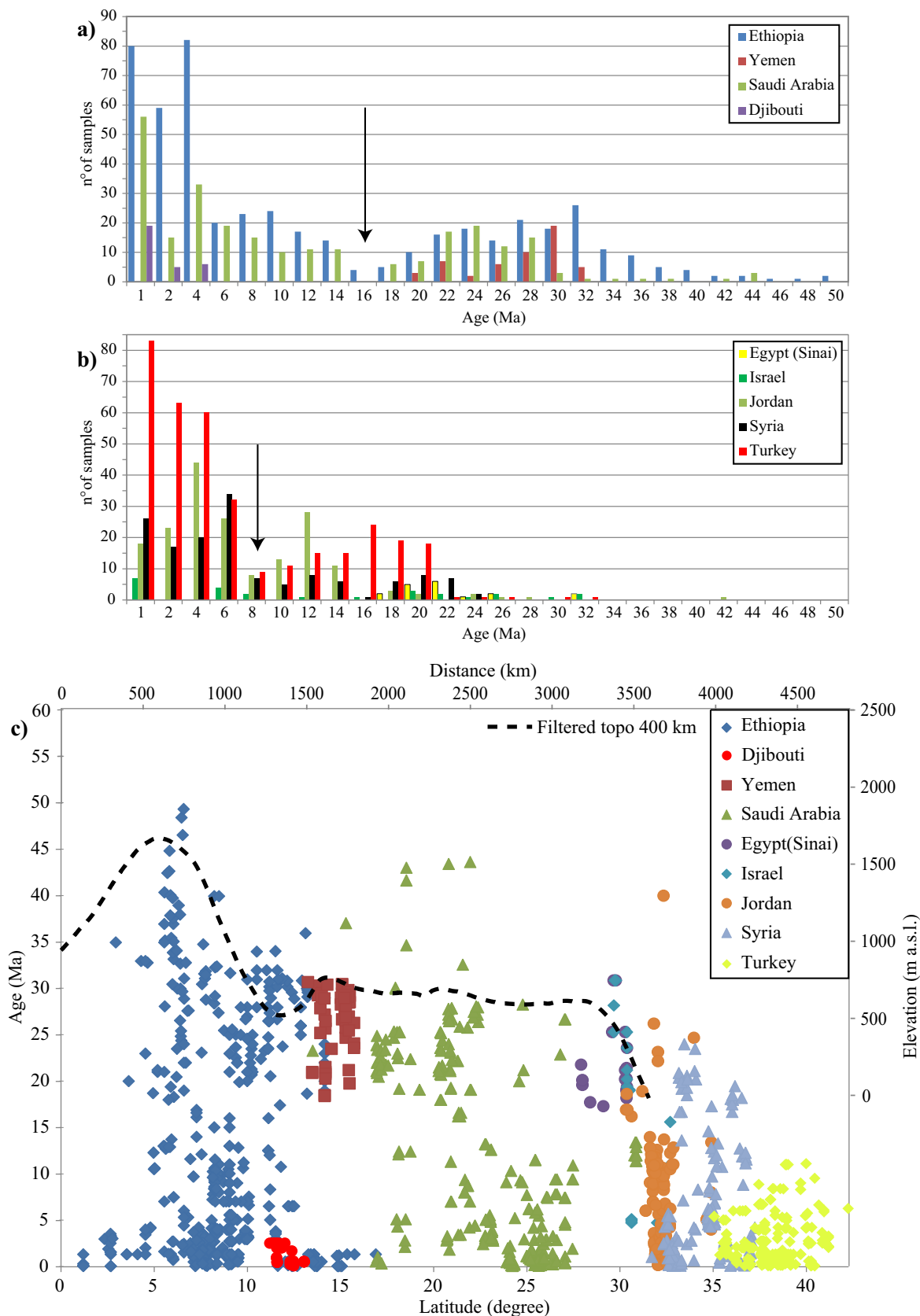


Fig. 15. a-b) Frequency plots of the volcanic deposits outcropping along the study area according to their ages; both plots show a bimodal distribution characterized by a marked minimum at 16–17 Ma in Fig. 15a and at 8–9 Ma in Fig. 15b; c) Distribution of the ages of the volcanic deposits according to latitude. Superimposed on the graph, the curve of the filtered topography at 400 km (the trace is shown in Fig. 7b) has been represented for comparison.

Ethiopian rift shoulders, where the highest values are related to the flexural uplift (Weissel et al., 1995; Sembroni et al., 2016a; Stüwe et al., 2022; Fig. S2), and the Ethiopian Plateau, because of the strong fluvial incision (Sembroni et al., 2016b, 2021; Fig. S2). The elastic thickness values obtained (Table S1) together with published seismic tomography data (Priestley et al., 2008; Chang and Van der Lee, 2011; Hansen et al., 2012; Chen et al., 2015) well correlate with the topographic pattern of the region and point to a mantle supported uplift (Chen et al., 2015). The feedback between erosion and seafloor spreading in the Red Sea area (Stüwe et al., 2022) could have enhanced the flexural uplift at the Red Sea margins inhibiting the capture of the swell top surface by rivers draining the inner part of the Red Sea shoulders. Portions of this surface has been locally studied in the past (Coltorti et al., 2007; Gani et al., 2007; Avni et al., 2012; Bar et al., 2016; Sembroni et al., 2016a, 2016b; Sembroni and Molin, 2018; Sembroni et al., 2021), but the topographic analysis performed here allows designating different fragments. They are located roughly along the axis of the swell ridge at an elevation between 800 m, in the north, and 2700 m, in the south (Fig. S1). The overall elevation pattern and geometry of the surface resembles the filtered, residual, and dynamic topographies (Figs. 7a, b and 10b, f). In turn, the filtered topography at 200 km approximates the present topography except for the Red Sea margins where the flexural component is dominant (Figs. 8 and 9). This suggests that the swell top surface and part of the present topography of the study area can be related to mantle processes (Sengör, 2001; Daradich et al., 2003; Forte et al., 2010; Moucha and Forte, 2011; Faccenna et al., 2013).

Recent studies on the uplift history of the Ethiopian-Somalian and Yemen plateaux (Sembroni et al., 2016a; Faccenna et al., 2019;

Sembroni et al., 2021) and the Jordan plateau (Bar et al., 2016) show that the southern portion of the swell (Horn of Africa) in the lower Oligocene was at elevations very close to those currently shown by the swell top surface, while the northern one (Levant region) reached the elevation of ~ 1000 m (the average elevation of the swell top surface in this region) at the end of upper Miocene (Fig. 16). This means that the bulk of the uplift in most of the study area occurred between the Oligocene and Miocene and that the present topography is very similar to the one reached in that period, partly modified by surface erosion, flood basalt emplacement, and flexural uplift along the Red Sea and MER margins. This is also confirmed by thermochronology data showing a rapid and significant exhumation phase during that period (Boone et al., 2021; Lanari et al., 2023) with rates between 0.1 and 0.3 mm/yr in the northern portion of the Red Sea (Lanari et al., 2023).

The opening of the Red Sea begun in its southernmost portion between the end of the Oligocene and the beginning of the Miocene (Boone et al., 2021, and references therein). This would relate, at least temporally, this event to the deep process underlying the formation of the swell but does not allow to define with certainty whether the same process also caused the opening of the Red Sea. Several studies (McQuarrie et al., 2003; Bellahsen et al., 2003; Bosworth et al., 2005; Koptev et al., 2018; Khalil et al., 2020) agree that the Red Sea and the Gulf of Aden rifts were mainly driven by slab pull occurred along the Bitlis-Zagros subduction front, with plume-related magmatism acting to generate extension in the Afar and Red Sea.

However, the thermo-mechanical weakening of the lithosphere related to plume impingement together with pre-existing lithospheric heterogeneities seem to have played a role in rift evolution (e.g., Koptev

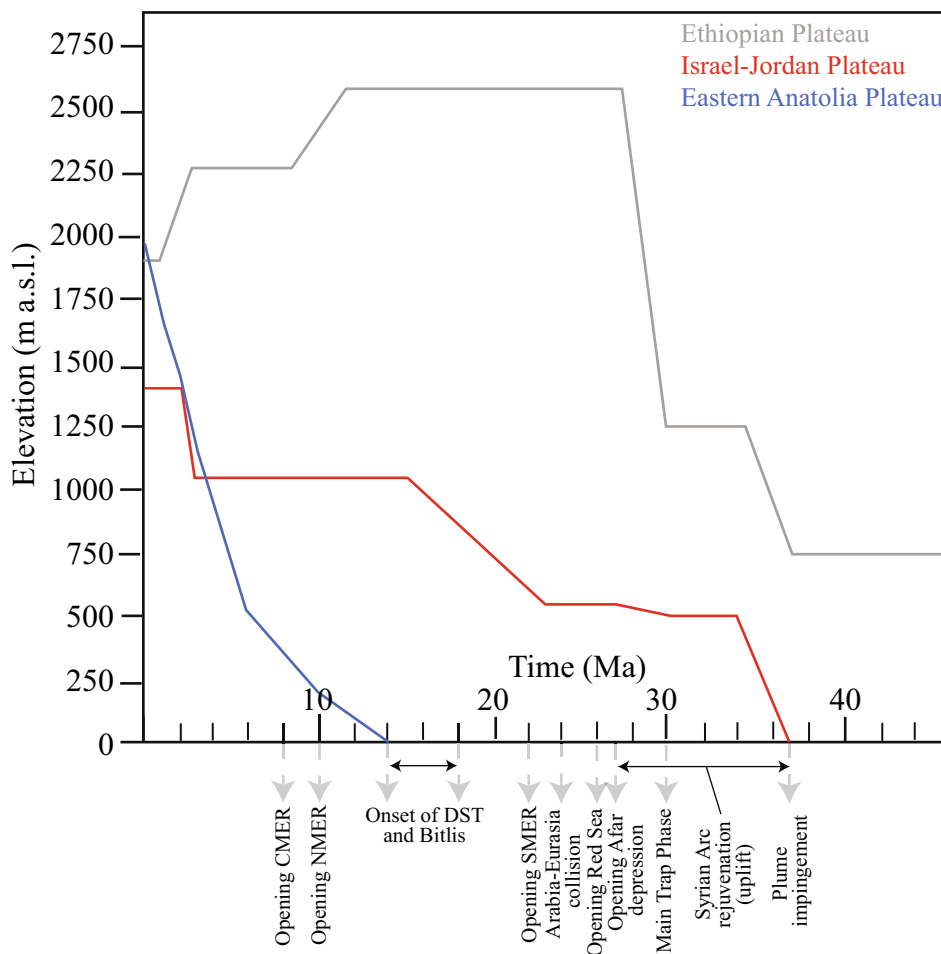


Fig. 16. Uplift histories of the Ethiopian, Jordan, and Eastern Anatolian plateaux according to recent studies (respectively Faccenna et al., 2019; Bar et al., 2016; Molin et al., 2023).

et al., 2018; Khalil et al., 2020). The analysis of flexural uplift along the Red Sea margins confirms the progressive northwards opening of the rift with an uplift increasing from north to south (Fig. 9a and Table S1). The greatest elevations (w_0) and wavelength (α) were found on the Arabian plate. Despite the high values (see Table S1), the flexural uplift marginally influenced the topographic configuration of the EAAS since its effect becomes close to zero at a maximum distance of ~ 250 km from the Red Sea margins (Fig. 9a). The same analysis performed in the Main Ethiopian Rift area suggests again a southward increase in flexural uplift (from 500 to 1200 m) with maximum wavelength of ~ 200 km from the rift shoulders (Weissel et al., 1995; Sembroni et al., 2016a).

The comparison between flexural uplift data and filtered topographies in the Red Sea region (Fig. 9b) clearly shows how the topography in this area is the result of the interaction between shallower (flexural uplift) and deeper (mantle plume) processes. Indeed, the topographic profile across the Red Sea presents neither a pattern typical of a rift margin subjected to flexural unloading (green curve in Fig. 9b) nor that of an area characterized by bulging due to rising hot mantle material (blue curve in Fig. 9b) but something in between: for the first 200 km the green curve is what best approximates the topography; beyond that distance the profile follows the blue curve.

Like topography and flexural uplift, the areal distribution of volcanism shows a marked asymmetry as much in the Horn of Africa as in the Arabian Peninsula (Fig. 14). Our analysis reveals that in the Horn of Africa most of the volcanic deposits outcrop to the west of the Main Ethiopian Rift (cf. Sembroni et al., 2016a). Conversely, in the Arabian Peninsula, most of volcanic fields follows the axis of the EAAS to the E of the Red Sea (Fig. 14). Along the strike of this axis, from Ethiopia to Turkey, there is an overall decrease in the age of volcanics and a parallel decrease in topography (Fig. 15c). This implies that a single main mantle upwelling is sufficient to explain the present topographic configuration and the chronology of volcanic deposits, as seismic tomography and geochemical data seem to confirm (Priestley et al., 2008; Hansen et al., 2012; Faccenna et al., 2013; Schaeffer and Lebedev, 2013; Auer et al., 2014; Gvirtzman et al., 2016; Lu et al., 2019; Thrastarson et al., 2024). If, as claimed in some studies (e.g., Chang and Van der Lee, 2011; Koulakov et al., 2016; Chang et al., 2020), there were more sources of volcanism (more arrivals of hot mantle material) or there was a different path, both the topography and volcanics age distribution would likely show a more irregular pattern. However, the concomitance of other local processes underlying Arabian volcanism cannot be ruled out. For example, the recent volcanism (< 12 Ma) in southwestern Arabia has been speculated to be related to decompression melting in the mantle lithosphere caused by flexural uplift (Stüwe et al., 2022).

The appearance of magmatism below the Arabian Peninsula seems to be spread over a very large area. This would imply the presence of a very low-viscosity asthenosphere as seems to be confirmed by seismic, petrological, and geochemical data (Hua et al., 2023).

The ages of the volcanic deposits are progressively younger toward the north, some of which may be due to younger deposits partly or completely covering older ones. Despite this limitation, the frequency of volcanic deposits ages shows a bimodal trend with a brief period (few millions of years) of partial quiescence characterized by few events (Fig. 15a, b). On the Arabian plate, the areal distribution of volcanics before and after this time span shows a marked change. In particular, the old volcanics were emplaced along NW-SE dikes inherited from Precambrian tectonic lineaments (Najd Fault System; Johnson et al., 2017 and references therein). Conversely, the new volcanic centers are aligned N-S (Fig. 14). The age of the partial quiescence shifts to younger ages from south (16–18 Ma) to north (8–10 Ma; Fig. 15a, b). The older quiescence period is coincident with the advanced stage of the Arabia-Eurasia collision at the Bitlis collision zone (18–14 Ma – Okay et al., 2010; Ballato et al., 2011; Cavazza et al., 2018, 2019; Gusmeo et al., 2021; Darin and Umhoefer, 2022) and the activation of the Dead Sea transform fault (17–20 Ma – Quennell, 1958; Freund et al., 1970; Garfunkel, 1981; Garfunkel et al., 1981; Joffe and Garfunkel, 1987;

Bosworth et al., 2005; Nuriel et al., 2017). These processes caused the counterclockwise rotation of the Arabian plate and the change in the direction of maximum compression (at the Bitlis-Zagros front) and extension (in the Red Sea and Gulf of Aden) from NE-SW to N-S (Boone et al., 2021, and references therein). This may have caused the closure of the NW-SE Precambrian lineaments and the opening of new N-S lineaments as was already proposed in the case of the Harrat Ash Shaam in Syria (Al Kwatli et al., 2012).

The pause and renewal of volcanism in the study area are followed, respectively, by increases and decreases in the number of mammal evolution lineages (de Vries et al., 2021). Although de Vries et al. (2021) cite also the climate factor as a contributing cause, the close correlation between volcanism pauses and renewal and the mammal evolution lineages trend, makes the geological factor predominant as already demonstrated in other parts of the world (e.g., Deccan and Siberia; Prave et al., 2016; Ernst and Youbi, 2017).

In summary, the present topography of the region spanning from Ethiopia to Syria shows an elongated ridge with a roughly NNW-SSE direction. It originates in Ethiopia, where it presents the greatest elevations and amplitude, and extends mainly into the Arabian plate with decreasing elevation and amplitude toward the north. The ridge is well described by the Oligocene low-relief surfaces and by the trend of filtered (200 and 400 km), residual, and dynamic topographies (Figs. 7, 10, 11). The same area has lower lithospheric thickness and higher residual and dynamic topographies respect to the surroundings and has negative P- and S-wave velocity and gravity anomalies (Fig. 10; Benoit et al., 2006a, 2006b; Priestley et al., 2008; Chang and Van der Lee, 2011; Hansen et al., 2012; Schaeffer and Lebedev, 2013; Auer et al., 2014; Chen et al., 2015; Gvirtzman et al., 2016; Lu et al., 2019; Boyce et al., 2021, 2023; Ebinger et al., 2024; Thrastarson et al., 2024). The decrease of filtered topography from Ethiopia to Syria is very similar to the trend of the volcanic deposits age, decreasing from south to north. All these data indicate the presence of a main single process which sculpted the past and present topographic configuration of the area. This process, having a wavelength of > 200 km, is likely subcrustal as it exceeds flexural wavelengths. As already stated in the previous works on the study area (Faccenna et al., 2013; Gvirtzman et al., 2016; Agostini et al., 2021; Hua et al., 2023) this deep seated process may be identified with the upwelling of the Afar Superplume below the Ethiopian lithosphere that, taking advantage of prior discontinuities and/or pressure gradients, channeled to the Levant region in a few million years causing the tilting to the SE of the Arabian Peninsula discussed in several studies (Sengör, 2001; Daradich et al., 2003). The northward progression of plume induced flow as discussed by Faccenna et al. (2013) is also broadly consistent with time dependent mantle convection reconstructions for the region (e.g. Moucha and Forte, 2011; Faccenna et al., 2019; Straume et al., 2024).

According to geological data (Avni et al., 2012; Bar et al., 2016) the planation surface detected between the southern Levant area and the northern Red Sea region can be dated at the Oligocene. However, the region rose from 500 m to 1000 m elevation only at the end of upper Miocene (Bar et al., 2016; Fig. 16). This means that the arrival of the superplume in this sector may be placed at the beginning of the Miocene (Fig. 16). If this is correct, the tilting of Arabia can be placed between the impingement of the plume beneath Ethiopia (Late Eocene; Ebinger and Sleep, 1998; Sengör, 2001; Ebinger et al., 2024) and the arrival of the same plume in the Levant region (upper Miocene). This is confirmed by stratigraphic data which indicate a strong subaerial erosion phase throughout the Arabian Peninsula between the end of Paleogene and the beginning of the Neogene, immediately after a long transgression phase which established shallow marine condition over eastern Arabia since early Paleocene (Alsharhan and Nairn, 1995; Sengör, 2001; Ziegler, 2001). This phase generated an unconformity between the Paleocene and Miocene sediments and the shifting of the eastern Arabia coastline by > 500 km to the northeast (Alsharhan and Nairn, 1995; Ziegler, 2001; Fig. 17). Looking at the coastline reconstruction of the eastern sector of

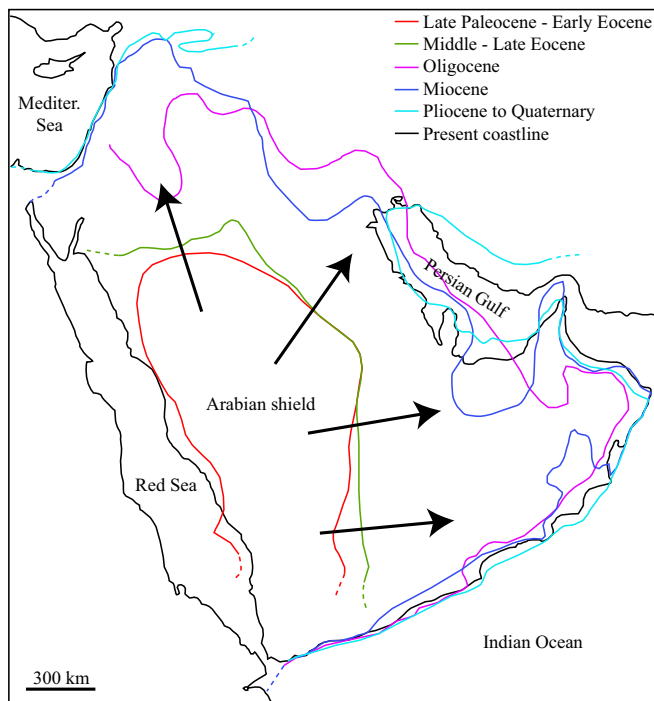


Fig. 17. Map showing the migration of the eastern Arabia coastline from Late Paleocene to Quaternary (modified after Alsharhan and Nairn, 1995 and Ziegler, 2001).

Arabia (Fig. 17), it is interesting to note that the northeastward shift appears to be parallel to the swell axis (Figs. 7, 8, 11). This suggests a direct involvement of the northward passage of the superplume in the coastline migration (cf. Straume et al., 2024). The presence of a new relief in the western side of Arabia is also confirmed by huge volumes of sands deposited from an easterly flowing river network between late Oligocene and early Miocene in an area comprising Saudi Arabia, Kuwait, and southeastern Iraq (Alsharhan and Nairn, 1995; Ziegler, 2001; Barrier and Vrielynck, 2008).

On the other hand, mantle flow took a longer time to reach Turkey. In fact, in the middle Miocene, the area was still under sea level (Molin et al., 2023 and references therein; Fig. 18). The uplift of Anatolia and in particular of the Eastern Anatolian Plateau is debated. Different models have been proposed: 1) mantle delamination (Göğüş and Pysklywec, 2008; Keskin, 2003; Bartol and Govers, 2014; Kounoudis et al., 2020); 2) slab break-off (Bottrill et al., 2012; Faccenna et al., 2006; Keskin, 2003; Schildgen et al., 2014); 3) mantle asthenosphere support (Faccenna et al., 2013; Keskin, 2007; Şengül Uluocak et al., 2021). Recently, Molin et al. (2023) tried to reconcile these three model in a single evolutionary scenario where at ~10–11 Ma the slab break-off drove the formation of a slab window which continued to widen until ~4–5 Ma (Faccenna et al., 2006, 2013; Schildgen et al., 2014). Such a corridor could have allowed the plume to reach the base of the Eastern Anatolian Plateau, as confirmed by the basaltic volcanism mainly Quaternary in age in that area (Figs. 14, 15a, c, 18). Recent upper mantle seismic tomographic

models (e.g. Kounoudis et al. 2020), seismic anisotropy studies (e.g. Paul et al., 2014; Merry et al., 2021), and residual topography calculations (e.g. Ogden and Bastow, 2022) seem to confirm the requirement for a mantle contribution to plateau uplift (cf. Straume et al., 2024).

The path of mantle flow throughout the region deeply influenced also the evolution of the main river networks: the Nile and the Euphrates-Tigris drainage systems. Indeed, both river networks source from areas of positive residual and dynamic topographies and have their base level in regions characterized by negative values (Fig. 10). Geological (Garzanti et al., 2006; Padoan et al., 2011; Sembroni et al., 2016a), geophysical (Faccenna et al., 2019, and references therein), and thermochronological data (Pik et al., 2003) indicate that the Nile River establish a connection between the Ethiopian plateau and the Mediterranean since at least Oligocene sustained by the upwelling of the Afar superplume at its source (Faccenna et al., 2019). Similarly, the first appearance of the Euphrates River is dated back to the middle Eocene when the subduction started at the Bitlis front (Okay et al., 2010; Ballato et al., 2011; Cavazza et al., 2018, 2019; Gusmeo et al., 2021; Darin and Umhoefer, 2022) and the whole region passed from marine to continental conditions (Molin et al., 2023, and references therein). The successive evolution of the river with the progressive migration of the coastline to the southeast and the consequent lengthening of the river course could be related partly to the ongoing subduction and partly to the arrival of Afar superplume in the northern Arabia region at late Miocene (Molin et al., 2023).

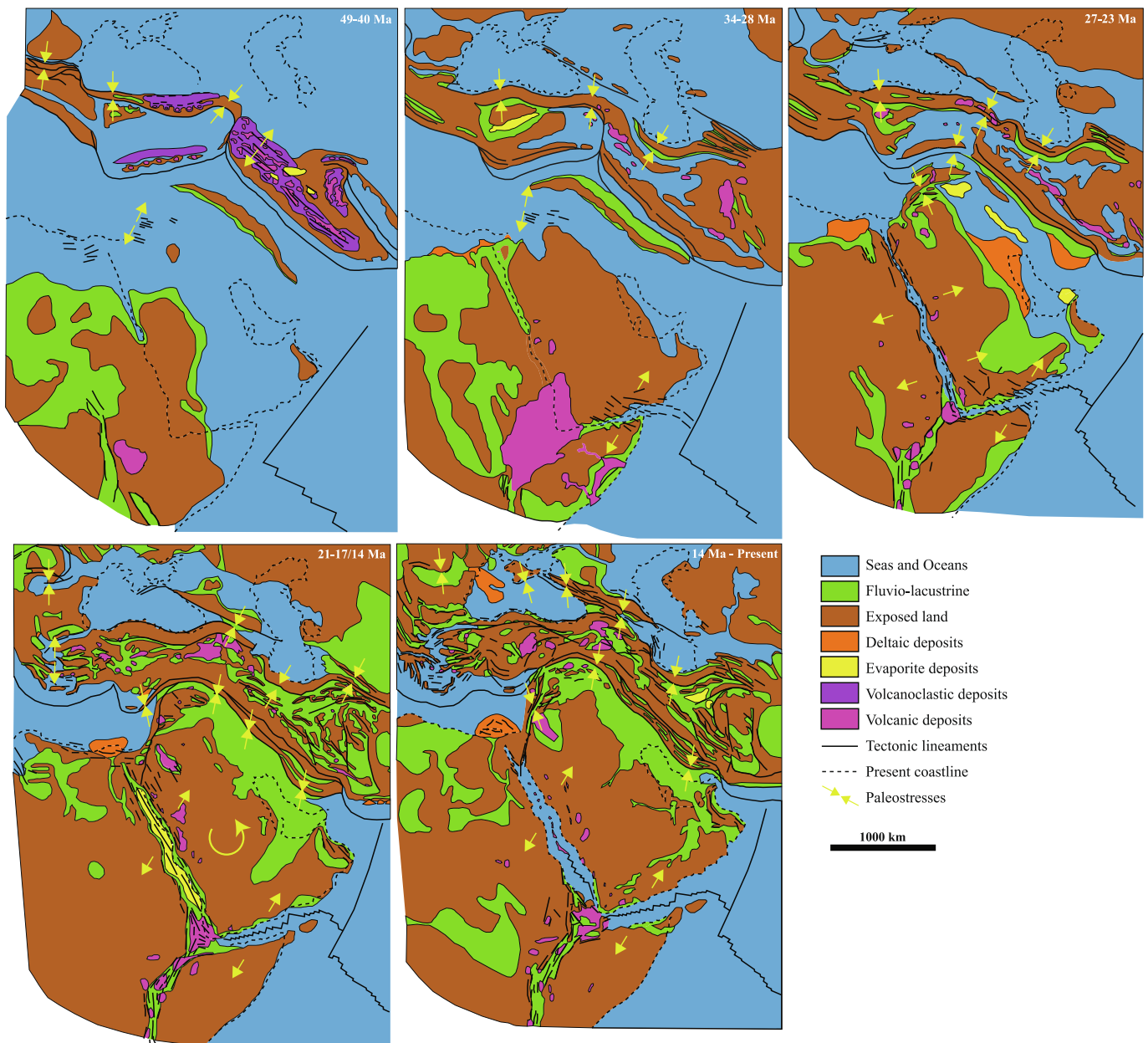


Fig. 18. Tectonic evolution of the study area (modified after Barrier and Vrielynck, 2008).

These data show how mantle processes can sculpt the surface of the Earth over tens of millions of years and support the idea expressed by Faccenna et al. (2019) that long-lived intra-continental rivers dynamics may provide indication of dynamic topography variations.

6.1. Tectonic evolution of the study area

The tectonic evolution of the study area can be summarized as follows (Figs. 18, 19):

45–35 Ma: Plume impingement occurred beneath Ethiopia, causing early uplift and volcanism mainly in southern Ethiopia.

31–29 Ma: Trap basalts are emplaced all over the Horn of Africa and the SW corner of the Arabian Peninsula forming an extended plateau. At about the same time (28–27 Ma) basalts poured out also in the western sector of Saudi Arabia from dikes and volcanoes aligned to Precambrian NW-SE structures. About 2/3 of the mammal fauna in the Afro-Arabia region becomes extinct (de Vries et al., 2021).

27–23 Ma: Volcanism continued in Arabia while a relative pause has

been registered in Ethiopia highlands (formation of intratrappean levels; Abbate et al., 2014). Parallel to this pause in volcanism an increase in mammal fauna occurred in the region until about 20 Ma (de Vries et al., 2021). During this time the Oligocene planation surface formed in the Sinai and Levant areas (Avni et al., 2012). The uplift in the Oligocene formed almost all the topography of the Horn of Africa and the SW corner of the Arabian Peninsula. Meanwhile, the Gulf of Aden opened, and the Red Sea began to propagate from south to north (as well as the related flexural uplift as suggested by thermochronology data; Boone et al., 2021, and references therein). The tilting to the SE of the Arabia Peninsula began, testified by a strong erosion phase throughout Arabia and the northeastward shifting of the eastern Arabia coastline (Alsharhan and Nairn, 1995; Ziegler, 2001). An easterly flowing river network carried huge volume of sands eroded from the uplifted western Arabia to the Iraq-Persian Gulf area where a large deltaic system formed.

21–17/14 Ma: At the end of upper Miocene the Levant region experienced strong uplift which increased elevation from 500 to 1000 m (Bar et al., 2016; Fig. 16). Volcanic activity on the Ethiopian-Somalian

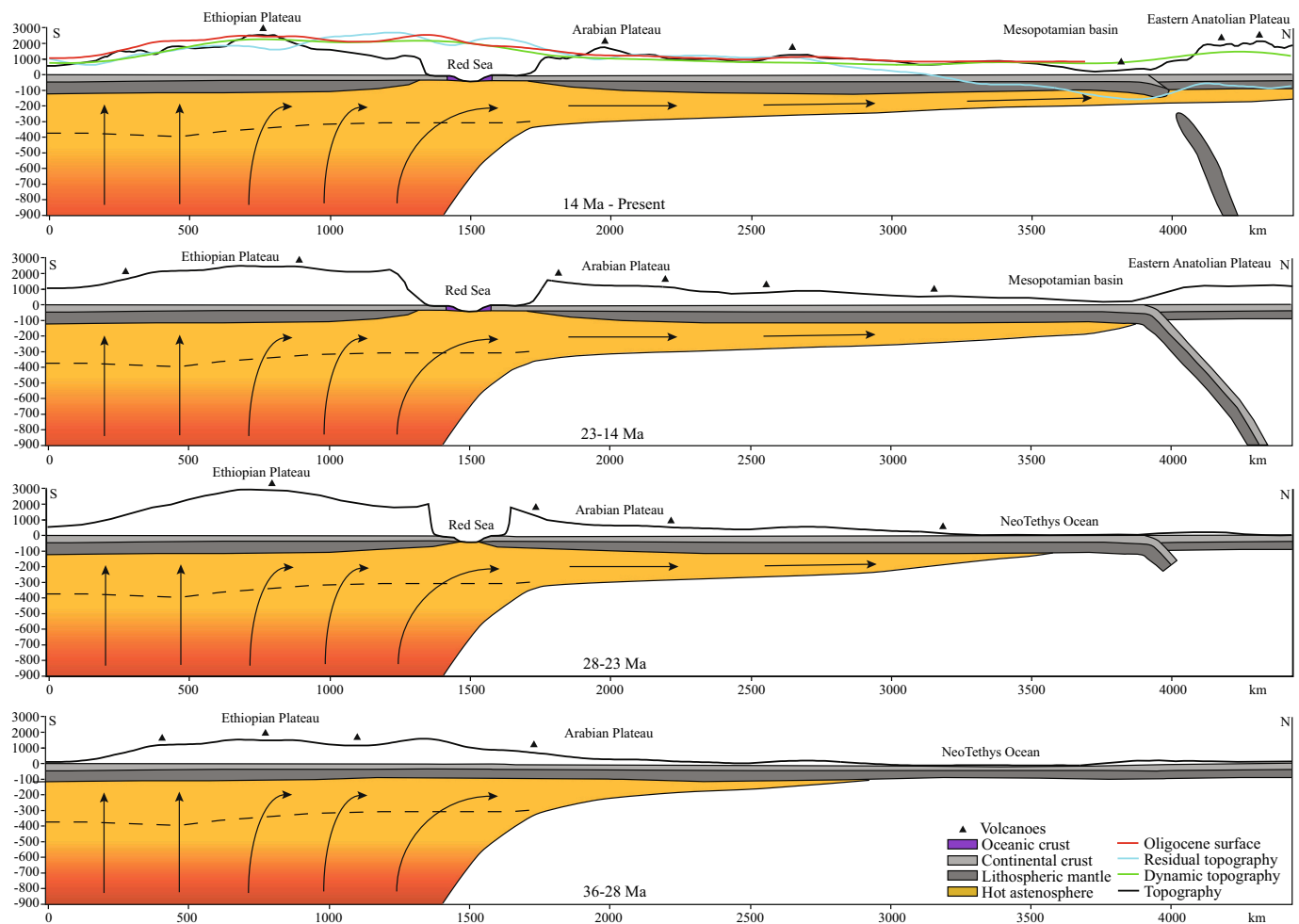


Fig. 19. Topographic evolution of the study area represented along a S–N trending topographic profile from Ethiopia to Eastern Anatolia. Note the lateral migration of the mantle flow and the parallel increase of surface topography (see text for further explanations).

plateau (shield volcanoes) and in the Ethiopian rift started over. In parallel, another dramatic decline in mammals occurred (de Vries et al., 2021). The stress regime changed: the Arabian plate rotates counter-clockwise, and the direction of maximum compression becomes ~N-S. At the same time, the Dead Sea Transform Fault developed. A brief pause in volcanism in both Arabia and Horn of Africa occurred in coincidence with this change in the stress regime.

14 Ma-Present: Volcanism resumed in most of the study area. As a demonstration of the change in stress regime, the new volcanic edifices in Arabia are aligned ~N-S. It is at this stage that the most extensive volcanic fields formed in Arabia Peninsula. The eastern portion of Anatolia is strongly uplifted as demonstrated by geological data (passage from marine to continental deposition; Fig. 18), dating of Euphrates R. fluvial terraces (Demir et al., 2007) and river network analysis (Molin et al., 2023).

7. Conclusions

The East Africa - Arabia region has long been studied because of its anomalously high topography, volcanism, and active rifting. There is broad consensus about the role of mantle plumes in generating this high elevation, but the number of plumes and the uplift patterns are debated. We contribute to this discussion by providing an integrative evolutionary model of the region which seeks to integrate the range of constraints we have reviewed here. The main results are the following:

1. The EAA swell is a NNW-SSE trending ridge extending from Ethiopia to Jordan with amplitude and elevation gradually increasing from north (800 km and 800 m) to south (1600 km and 2700 m). Its continuity is interrupted to the south by the Turkana tectonic depression and to the north by the Mesopotamian foredeep basin whose depression may be related to the flexure of the lithosphere by subduction as confirmed by dynamic topography modeling. The area occupied by the swell presents thin lithosphere, high residual and dynamic topography, underlain by low seismic velocity anomalies.
2. The swell top is characterized by low relief surfaces, Oligocene in age, located roughly along the axis. The envelope of these surfaces represents the swell top surface which extends for >4000 km from Ethiopia to Jordan with present elevation decreasing from south (2500 m) to north (900 m).
3. The elevation pattern and geometry of the swell top surface resembles the filtered (200 and 400 km), residual, and dynamic topography, suggesting that this surface is caused by mantle convective processes. Moreover, the uplift pattern of the southern portion of the swell shows that the surface was at elevation close to the present one since the beginning of the lower Oligocene, while in the northern portion it reached ~1000 m at the end of the upper Miocene. Such a shift in uplift histories could match the migration pattern of the mantle plume, and indicates that most of the present topography is mainly the surface expression of mantle plume impingement.
4. The migration of mantle flow from eastern Africa drove the tilting of the Arabian Peninsula between late Eocene and upper Miocene.

Stratigraphic data indicate an erosion phase in the Arabian Peninsula between the end of Paleogene and the beginning of the Neogene and a significant shifting of the eastern Arabia coastline to the northeast which could constrain such an event.

- The Nile and Euphrates-Tigris river networks source from areas of positive residual and dynamic topography and have their base level in regions characterized by negative values. This suggests that the formation and evolution of these drainage systems are influenced by mantle processes. In particular, the stable presence of a mantle upwelling beneath east Africa and the progressive migration of mantle flow to Turkey contributed to the formation of the Nile (Oligocene) and Euphrates-Tigris (Middle Miocene) river networks and to the maintenance of their path through tens of millions of years.
- The incision pattern (local relief) on the top and on both flanks of the swell is low except for the rift margins, where the higher values are related to the flexural uplift, and of the Ethiopian Plateau because of the strong fluvial erosion.
- Analysis of flexural uplift at the Red Sea margins shows an increase in uplift from south to north confirming thermochronological data from literature. In general, the western margin presents lower values than the eastern one, while the deformation related to flexure is visible up to ~250 km from the Red Sea margins. The comparison between flexural data and filtered topography confirms that the Red Sea region topography is the result of the interaction between shallower (flexural uplift) and deeper (mantle plume) processes.
- The distribution of ages of volcanic deposits shows a gradual decrease in maximum age from Ethiopia to Turkey which broadly follows the curves of filtered topography extracted along the axis of the swell.
- Along the swell axis volcanoes and basaltic lava fields are located. The frequency of volcanic deposits ages shows a bimodal trend with peaks separated by a brief period of low volcanic activity with a trend toward younger ages from south to north. In particular, the older quiescence period is coeval with the advanced stage of the Arabia-Eurasia collision at the Bitlis zone and the activation of the Dead Sea Transform Fault. These processes caused the counter-clockwise rotation of the Arabian plate and the change in the direction of maximum compression from NE-SW to N-S favoring the opening of N-S trending tectonic lineaments and the closure of old NW-SE ones. In confirmation of this, Arabian volcanics older than 16 Ma emplaced through NW-SE-trending dikes, while younger deposits appear aligned in a N-S direction.

All these results point to the presence of a single process which shaped the past and present topographic configuration of the region, the upwelling of the Afar superplume. Once the plume reached the base of the lithosphere below the Horn of Africa, it flowed laterally toward the Levant area by exploiting pre-existing lithospheric structure, and then reached the Anatolian Plateau, facilitated by slab break-off and the consequent formation of a slab window. The buoyancy of the mantle material below the lithosphere generated the East African-Arabia swell interrupted to the north by the Mesopotamian foredeep basin.

Declaration of competing interest

Claudio Faccenna reports financial support was provided by Italian Research Ministry. If there are other authors, they declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

Data availability

The original data supporting this research are available in the supplementary material or on request.

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Appendix A. Supplementary data

Supplementary data to this article can be found online at <https://doi.org/10.1016/j.earscirev.2024.104901>.

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