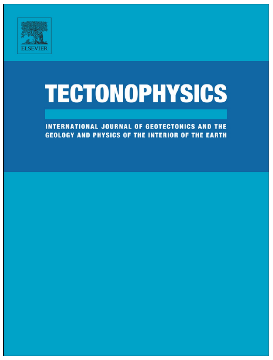

Preparing the ground for plateau growth: Late Neogene Central Anatolian uplift in the context of orogenic and geodynamic evolution since the Cretaceous

Peter J. McPhee, Ayten Koç, Douwe J.J. van Hinsbergen

PII: S0040-1951(21)00413-3

DOI: https://doi.org/10.1016/j.tecto.2021.229131

Reference: TECTO 229131

To appear in: *Tectonophysics*

Received date: 28 August 2020

Revised date: 4 August 2021

Accepted date: 21 October 2021

Please cite this article as: P.J. McPhee, A. Koç and D.J.J. van Hinsbergen, Preparing the ground for plateau growth: Late Neogene Central Anatolian uplift in the context of orogenic and geodynamic evolution since the Cretaceous, *Tectonophysics* (2021), https://doi.org/10.1016/j.tecto.2021.229131

Preparing the ground for plateau growth: Late Neogene Central Anatolian uplift in the context of orogenic and geodynamic evolution since the Cretaceous

Peter J. McPhee[1,3], Ayten Koç[2], Douwe J.J. van Hinsbergen[1]

Affiliations

1. Department of Earth Sciences, Utrecht University, Utrecht, The Netherlands
2. Department of Geological Engineering, Van Yüzüncü Yıl University, Van, Turkey
3. Institute of Geological Sciences, Freie Universität Berlin, Berlin, Germany

Corresponding Authors

Peter J. McPhee, peter.mcphee@fu-berlin.de; Douwe J.J. van Hinsbergen, d.j.j. van hinsbergen@uu.nl

Abstract

Central Anatolia (Turkey) is a small and nascent example of a high orogenic plateau, providing a natural laboratory to study processes driving plateau rise. The 1-km-high plateau interior uplifted since c. 8-5 Ma, with a further phase of kilometre-scale uplift affecting the southern plateau margin since 0.45 Ma. Several causes of plateau rise have been proposed: peeling or dripping delamination of the lithospheric mantle; asthenospheric upwelling through slab gaps created by slab fragmentation or break-off, and; continental underthrusting and crustal shortening below the southern plateau margin. The Neogene history of the plateau has not been diagnostic of the causes of plateau rise. We thus evaluate proposed uplift causes in the context of the Anatolian orogenesis, which formed the plateau lithosphere during subduction since the Cretaceous. We combine this analysis with available constraints on uplift, and geophysical data that illuminate the modern mantle (and crustal) structure. Our analysis suggests that lithospheric dripping, which followed arc magmatism and shortening in the Kirsehir Block (eastern Central Anatolia), is the most likely cause of plateau interior uplift. Lithospheric dripping is, however, an unlikely sole driver of multi-phase uplift along the southern plateau margin. There, underthrusting of the African continental margin, recorded by c. 11-7 Ma thrusting on Cyprus, is a viable cause of uplift since 0.45 Ma, but cannot account for earlier uplift since c. 8-5 Ma. Instead, slab break-off below the southern plateau margin is likely in light of geophysical data. On the SW plateau margin, small-scale peeling delamination of the Central Taurides by the Antalya slab since early Miocene times accounts for >150 km slab retreat with no corresponding upper-plate deformation. A southwest-travelling wave of subsidence and uplift signalled this retreat and may have contributed to coeval oroclinal bending of the western Central Taurides and southeastward thrusting of the Lycian Nappes.

Keywords:

Central Anatolian Plateau; Plateau Uplift; Tectonic Reconstruction; Subduction Evolution; Lithospheric dripping; Peeling Delamination; Orogenesis.

1. Introduction

High orogenic plateaus are broad high elevation regions, which have low topographic relief, have at least one steep outer-edge, and typically contain internal drainage systems. These regions, such as the Tibetan Plateau or Altiplano-Puna Plateau, are important physiographic features on Earth's continents: They form topographic barriers to atmospheric circulation (e.g., Ruddiman & Kutzbach, 1989), affect regional climate (e.g., Sobel et al., 2003), and contribute to continental deformation within plateaus and in surrounding regions via their gravitational potential energy (e.g., Coleman & Hodges, 1995; Molnar & Lyon-Caen, 1988).

Central Anatolia's physiography is typical of a high orogenic plateau (Figure 1): It has an average elevation of 1 to 1.5 km across an area of approximately 250,000 sq km, low topographic relief (Figure 1B), and contains an internal drainage system (Figure 1A). The steep southern edge of the plateau comprises the Central Taurides mountain range, which reaches 2 to 3.5 km elevation, and slopes southward towards the Mediterranean Sea. The northern edge comprises the Pontides mountain range, which reaches 2 km elevation and slopes northward towards the Black Sea.

The southern Central Taurides are covered by the Miocene-Pleistocene Mut Basin (Figure 1A), which contains marine sedimentary rocks that recorded at least two phases of km-scale uplift in late Miocene and Plio-Pleistocene times (e.g. Cosentino et al., 2012; Öğretmen et al., 2018). Channel incision and long-wavelength knick zones in northern Central Anatolia (Doğan, 2011; Çiner et al., 2015; McNab et al., 2017) and stable isotopes from continental basin rocks (e.g., Meijers et al., 2018) point to uplift since the late Miocene.

The specific geodynamic driving forces of Central Anatolian plateau rise remain debated, and several competing or perhaps complementary hypotheses have been proposed: 1) Miocene peeling delamination of a flat slab that would have existed below Central and East Anatolia (Bartol & Govers, 2014; Govers & Fichtner, 2016); 2) Paleogene thickening and late Neogene removal of lithospheric mantle by lithospheric dripping (Göğüş et al., 2017); 3) Upwelling of hot asthenosphere between segmented or detached slabs (e.g., Schildgen et al., 2014; Portner et al., 2018); 4) Continental collision driving underthrusting and thickening of the buoyant African margin below the southern Central Taurides (e.g., Robertson et al., 1995; Delph et al., 2017; Meijers et al., 2018; McPhee & van Hinsbergen, 2019), or deep underthrusting by subducted oceanic sedimentary rocks (Fernandez Blanco, 2015; Fernandez Blanco et al., 2020); and 5) rebound following late Miocene slab break-off of a subducted slab below the Mut Basin (e.g., Portner et al., 2018; Cosentino et al., 2012). The young Central Anatolian plateau thus provides a natural laboratory for the study of the surface expression of deep geodynamic processes and may be analogous to the early history of other, older and larger high orogenic plateaus where the early evolution is overprinted during plateau maturation.

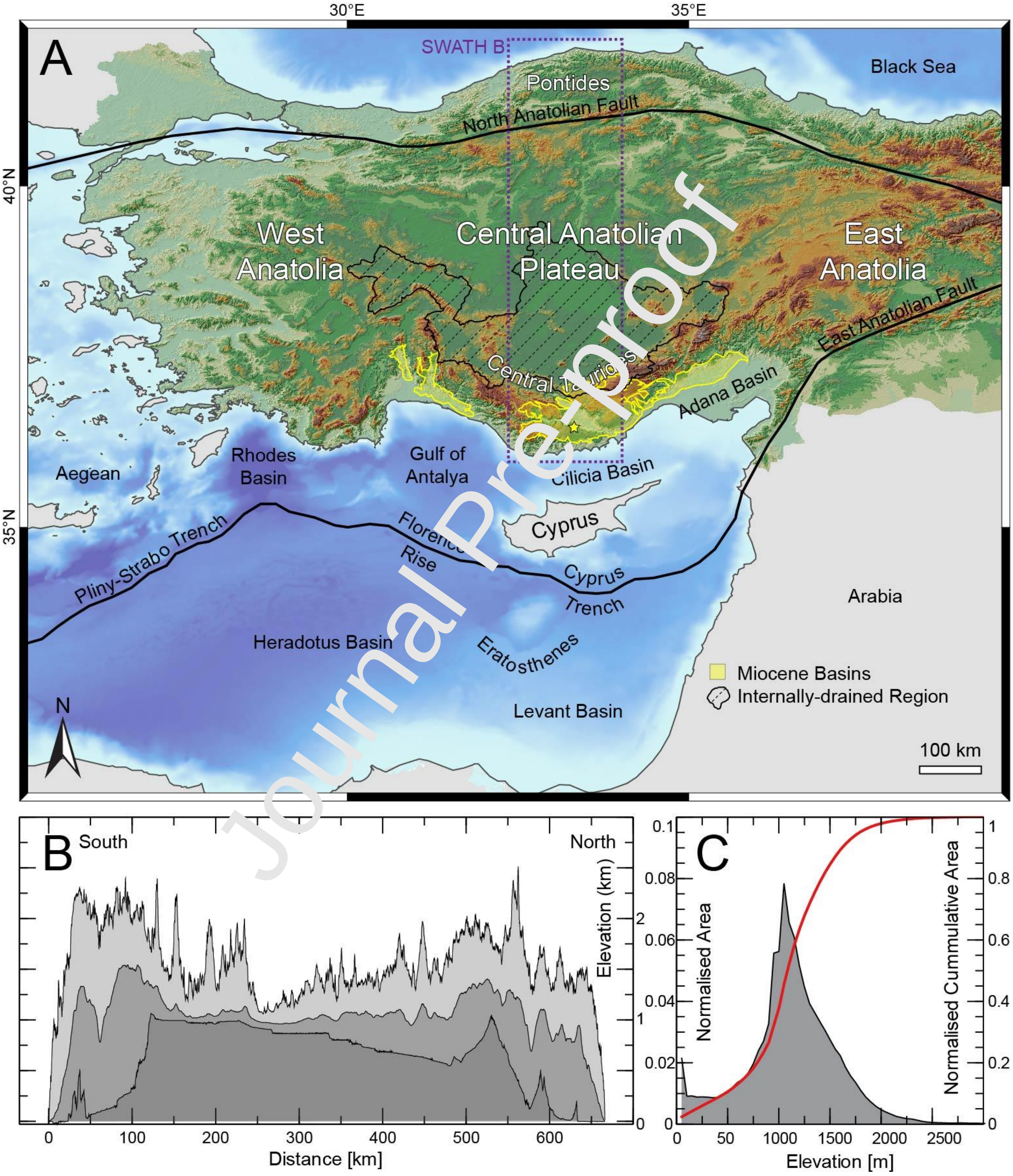

Figure 1: *A) Topography of Central Anatolia and surrounding regions, derived from ETOPO 1 data. The purple dashed-line box marks the area used in the generation of elevation swath profile shown in*

1B. The hatched area represents the internally drained plateau interior. The yellow star marks the Mut Basin. B) N-S elevation swath profile is taken from within the purple dashed-line box marked on 1A. C) Elevation frequency and hypsometry of the plateau calculated from SRTM 90 elevation data (calculated in the region between 30°E and 36°E).

Previous work on the causes of Central Anatolian plateau rise has focused on constraining the timing and spatial evolution of uplift, and/or insights from the modern crustal and mantle structure. These insights have inspired the above hypotheses, but the Neogene history alone has not been diagnostic in identifying the role and likelihood of the proposed geodynamic drivers. In this contribution, we therefore review evidence for the long-term kinematic record of crustal deformation in Central Anatolia, which, when combined with the history of Africa-Eurasia convergence (constrained by Atlantic Ocean reconstructions), and present-day mantle structure reveals the long-term evolution of subduction and collision of the African Plate. We combine this kinematic record with constraints on the spatial and temporal evolution of uplift to evaluate the geodynamic conditions that were present when plateau rise started, and use this to assess the viability and contribution of geodynamic mechanisms of Central Anatolian plateau rise.

2. Spatial and Temporal Evolution of Late Neogene Plateau Rise

We start by reviewing evidence for the timing and spatial extent of late Neogene uplift in Central Anatolia, which form important constraints when evaluating potential geodynamic causes of plateau rise.

2.1 Uplift of the Southern Plateau Margin (Taurides)

The Central Taurides form a high mountainous rim along the southern edge of the Central Anatolian plateau. Uplifted Neogene marine sedimentary rocks of the Mut, Adana, and Antalya Basin (including the Aksu, Köprüçay-Manavgat sub-basins) cover this margin, robustly constraining the timing and magnitude of late Neogene uplift.

The Mut Basin covers the southern Central Taurides and contains an Oligocene to Pleistocene stratigraphy. The lowermost stratigraphy in this basin consists of Oligocene-Burdigalian fluvial and lacustrine sedimentary rocks. These are covered by a Burdigalian – Tortonian sequence of marls, redeposited carbonates, and shallow-water ramp carbonates that have been mapped from sea level, up to a modern elevation of 2.2 km. The basin is tilted by a few degrees toward the south, forming an open, south-dipping monocline (Cosentino et al., 2012).

Pliocene-Pleistocene marine sedimentary rocks onlap the Tortonian strata, and are mapped up to 1.5 – 1.6 km elevation (e.g., Yildiz et al., 2003; Öğretmen et al., 2018) (Figure 2A). The youngest dated strata in this unit were deposited at c. 0.45 Ma, and are currently found at 1 km elevation. Benthic fauna in these strata indicate deposition at a water depth of 0.4 - 0.5 km, consistent with geological relics of a Pliocene-Pleistocene paleocoastline at 1.5 – 1.6 km elevation (Öğretmen et al., 2018).

Onlap of Pliocene-Pleistocene rocks onto the Miocene sequence suggests that initial uplift of the plateau margin occurred in pre-Pliocene times (late Miocene), and was followed by a period of stability, or gentle subsidence (Schildgen et al., 2012a; Cosentino et al., 2012). This was followed by rapid Pleistocene uplift, at rates of 3.21–3.42 mm/yr (Öğretmen et al., 2018), which, based on modelling of marine terraces, likely peaked between 0.5 – 0.2 Ma (Racano et al., 2020). The multi-phased uplift history may also be reflected by relicts of a pre-Pleistocene drainage system in the upper reaches of the modern Ermenek River (Schildgen et al., 2012a; Figure 2C). Incision rates of 0.52 to 0.67 mm/yr since c. 130 ka were calculated in this river system, based on the exposure ages of river terraces (Schildgen et al., 2012a), lending support to the idea that the most rapid rates of Pleistocene uplift were short-lived (Öğretmen et al., 2018).

Significantly lower rates of uplift affected the adjacent Adana Basin, which covers the south-eastern plateau margin. Pliocene (c. 5.3 Ma) marine sedimentary rocks deposited at up to 500 m water depth are exposed at 150 m elevation, constraining up to 0.65 km of uplift at rates of 0.02-0.13 mm/yr (Cipollari et al., 2013). Initial uplift of the eastern Central Taurides may have occurred at 5.45 Ma, based on a plateau-ward shift in sediment provenance during rapid deposition of a 1-km-thick package of conglomerates (Radeff et al., 2015).

The Antalya Basin (Figures 2A, 5C, & 5D) covers the Tauride fold-thrust belt on the southwestern margin of the plateau. This basin includes the Manavgat, Köprüçay, and Aksu sub-basins, and forms the on-land equivalent of the offshore Gulf of Antalya Basin, which reaches depths of 2.5 km below sea level (Figure 1 and 5C). The onshore basin contains a lower Miocene (locally Aquitanian and predominantly Burdigalian, c. 20 Ma) to Messinian sedimentary sequence of marine limestone, marine sandstone, marl, marginal marine conglomerates, and reefal limestones (Akbulut, 1977; Karabiyikoglu et al., 2000; Deynoux et al., 2005; Flecker et al., 2005; Çiner et al., 2008; Şiş et al., 2020). Deposition of continental rocks (tufas) constrains uplift and emergence of the Antalya Basin by late Pliocene times (c. 3.5 Ma) (Glover & Robertson, 1998). The basin was also deformed by late Miocene folding and thrusting, uplifting Messinian marine rocks of the eastern Köprüçay Basin up to 1.5 km elevation (Schildgen et al., 2012b), along the hinge of a west-verging asymmetric anticline (McPhee et al., 2018a).

On the eastern side of the Antalya Basin, in the western Central Taurides, low-temperature thermochronological data (apatite [U-Th]/He, apatite fission track, and zircon [U-Th]/He) constrain the thermal history of the plateau margin. An early to middle Miocene increase in cooling rate was identified, signalling a phase of exhumation. This was likely driven by erosion related to regional uplift, as no structural or climatic driver of erosional exhumation was identified (McPhee et al., 2019). These data show that in contrast to the southern Central Taurides, which was covered by the Mut Basin, the western Central Taurides were emergent and actively eroding throughout the Neogene.

2.2 Uplift of the Plateau Interior

In contrast to the plateau margin, sedimentary basins in the plateau interior have recorded terrestrial deposition since Oligocene times (e.g., Koç et al., 2012; 2016; 2017; 2018; Fernández-

Blanco et al., 2013; Ozsayin et al., 2013). These basins form the characteristic low relief of the plateau interior, and include the Central Tauride Intramontane Basins (Koç et al., 2012, 2016b, 2017) and the Tuz Gölü Basin and its sub-basins (Fernández-Blanco et al., 2013; Ozsayin et al., 2013; Görür et al., 1984; 1998), as well as the Ulukışla Basin (Clark & Robertson, 2005; Gürer et al., 2016b; Meijers et al., 2016) and basins overlying the Kırşehir Block (Gülyüz et al., 2013; Advokaat et al., 2014; Licht et al., 2017). The plateau interior basins are located at an average elevation of around 1 km: 1 km lower than Neogene marine sedimentary rocks preserved on the southern plateau margin. The southern plateau margin has therefore most likely experienced greater uplift (Koç et al., 2012; 2017; Schildgen et al., 2014).

Much of the plateau interior forms an internal drainage system (Figure 1A), which, based on stable isotope studies of lacustrine sediments, has existed since at least early Miocene times (Meijers et al., 2020). Analyses of oxygen isotope data from upper Oligocene lacustrine rocks in the south-eastern plateau interior indicate a low elevation depositional environment with no significant orographic barriers in northern or southern Central Anatolia (Lüdecke et al., 2013; Meijers et al., 2016). In contrast, middle to upper Miocene lacustrine rocks contain low $\delta^{18}O$ values, interpreted to reflect kilometre-scale plateau uplift and the formation of an orographic barrier along the southern plateau margin around c. 11–5 Ma (Meijers et al., 2018).

Longitudinal river profiles from the externally drained northern half of the plateau contain long-wavelength knick zones, indicative of regional uplift (see for example Figure 2C). Inverse modelling of the development of knick zones suggests that they formed in response to kilometre-scale uplift in the past c. 12 Ma, with highest rates of uplift in the past c. 6 Ma (McNab et al., 2017). Incision rates of 0.12 mm/yr between c. 5 - 2.5 Ma and 0.05 mm/yr since c. 1.9 Ma have been calculated based on exposure ages of abandoned fluvial terraces, and incision of well-dated volcanic rocks in the Cappadocia Volcanic Province (Çiner et al., 2015; Doğan, 2011; Aydar et al., 2013). These rates are lower than those calculated at the southern plateau margin.

2.3 Uplift of the Northern Plateau Margin (Pontides)

The Central Pontides form the mountainous northern plateau margin. Pleistocene to Present uplift is constrained by incised river terraces (0.27-0.29 mm/yr; Yildirim et al., 2013b; Berndt et al., 2018), by uplifted paleodeltas (0.2-0.3 mm/yr; Demir et al., 2004), and by uplifted Pleistocene marine terraces on the Sinop Peninsula (0.02 - 0.2 mm/yr; Yildirim et al., 2013a) (Figure 2B). Yildirim et al. (2013a, 2013b) showed that spatially variable Pleistocene to Present uplift rates and disequilibrium river profiles may be associated with active thrusting, linked to a restraining bend in the North Anatolian Fault (NAF) (Yildirim et al., 2011; Berndt et al., 2018).

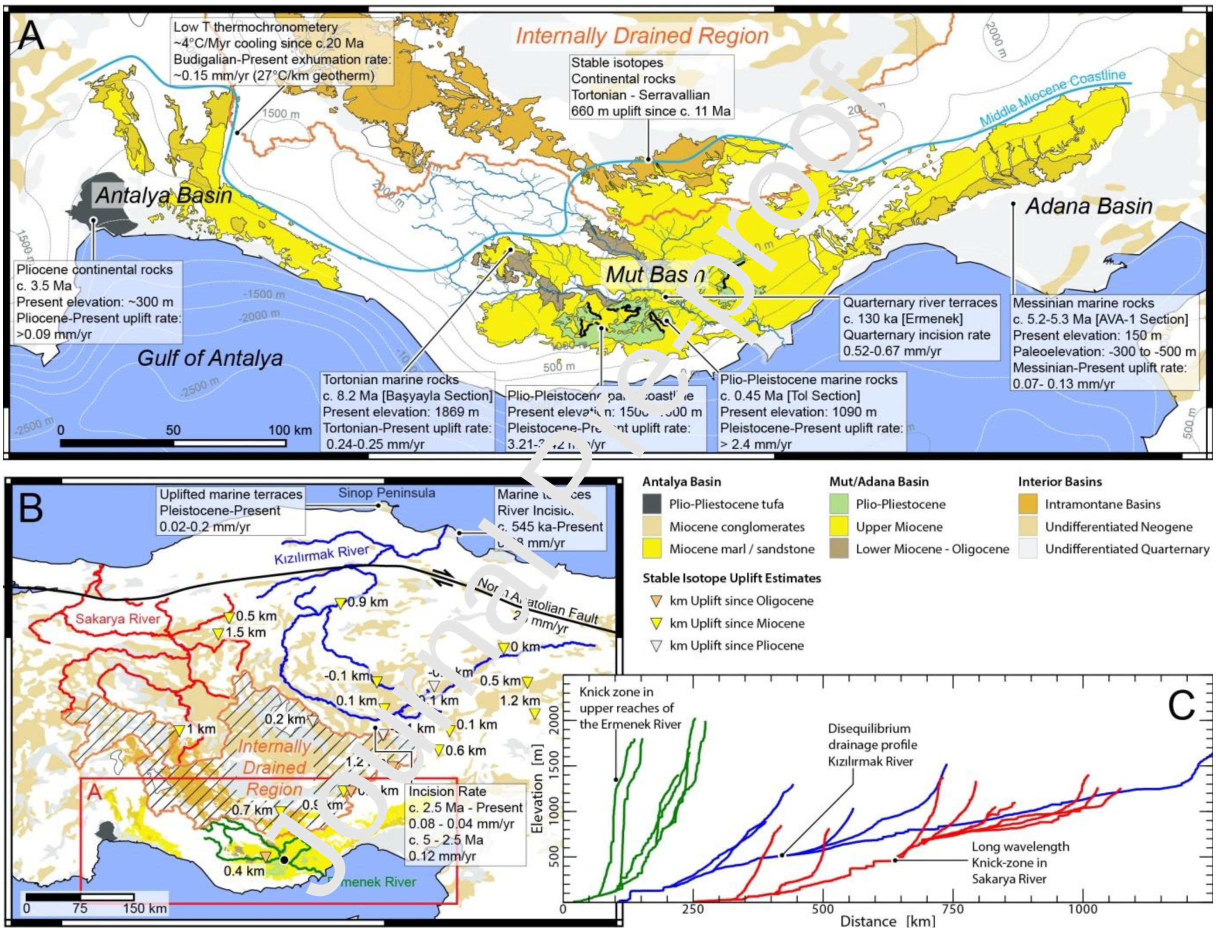

Figure 2: *Synthesis of uplift constraints across the Central Anatolian Plateau. A) Southern plateau margin, including the Antalya, Mut, and Adana basins. Contours are 10km smoothed elevation. See text for citations. B) Uplift constraints across the plateau interior and northern plateau margin. Stable isotope uplift estimates are based on data from Meijers et al. (2018; and references therein), assuming a lapse rate of −2.9 ‰/km. C) Longitudinal river profiles from three major rivers that drain Central Anatolia.*

3. Cretaceous to Present Geological Evolution of Central Anatolia

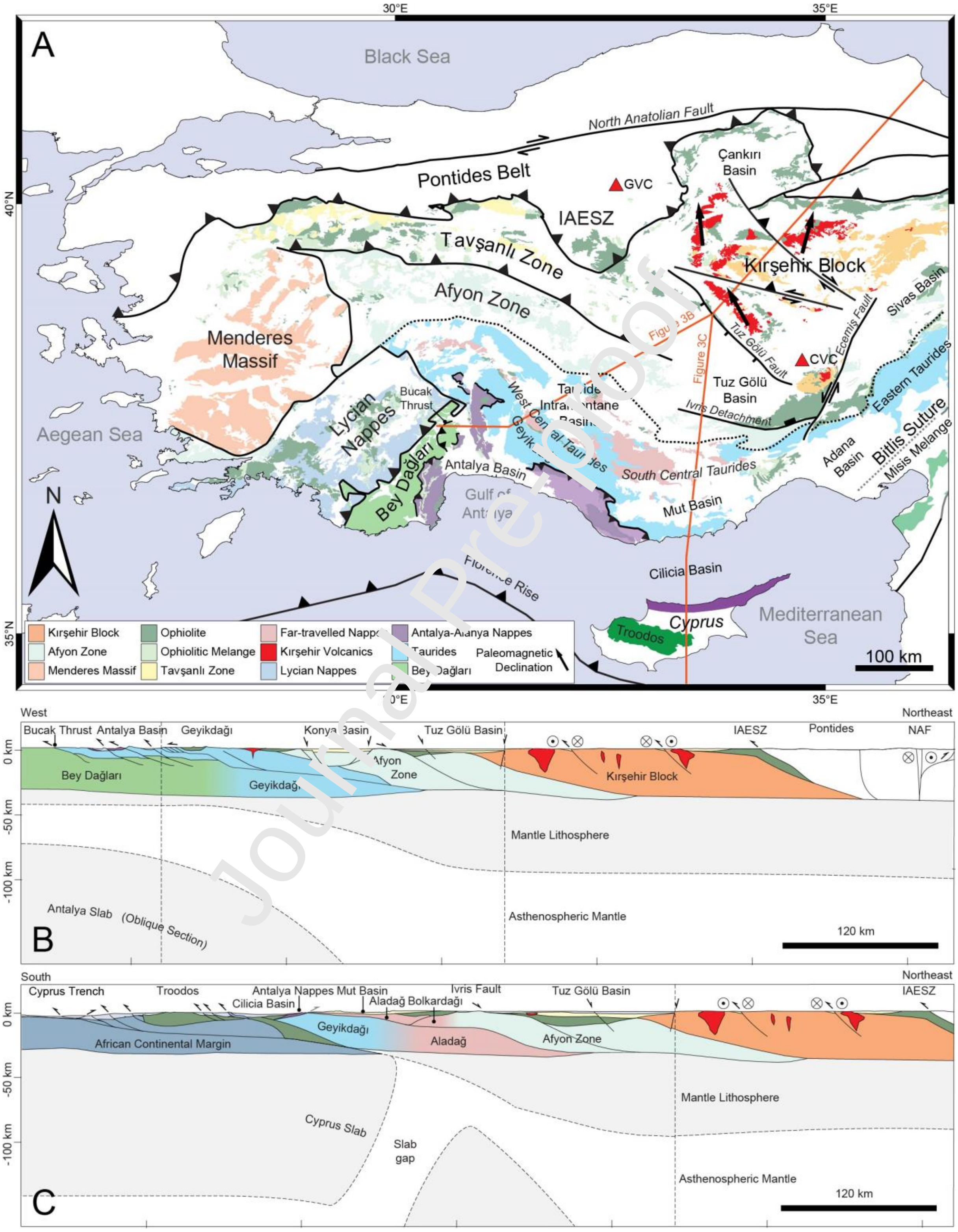

Figure 3: *A) Major tectonic units and contacts of Turkey modified from the MTA 1:500,000 geological map series (Senel et al., 2002). CVP = Cappadocia Volcanic Centre; GVC = Galatia Volcanic Centre. Far-travelled Nappes are the undifferentiated Bozkır, Aladağ, and Bolkardağı nappes.*
B) Conceptual cross-section across Anatolia (approximately to scale), extending southwest to the Bey Dağları Platform, incorporating the Bucak-Seydişehir cross-section of McPhee et al. (2018a). C) Conceptual cross-section across Anatolia (approximately to scale), extending southward over Cyprus and the Cyprus Trench. Geology of Cyprus, based on seismic sections of the Cyprus Trench from Reiche and Hübscher, (2015); Symeou et al., (2018); a geological map of south Cyprus by Bagnall (1960); and the cross-section of the Kyrenia range from McPhee and van Hinsbergen (2018).

3.1 Cretaceous to Eocene Orogenesis

The Central Anatolian crust comprises oceanic and continental nappes that were assembled by subduction and collision since the Mesozoic. We aim to use the long-term geological record of this orogenesis to evaluate hypothesised geodynamic causes of plateau uplift. We start by describing the first-order modern tectonic units and their interpreted tectonic and paleogeographic origin, from north to south and downward through the regional tectonostratigraphy. These tectonic units are shown in the geological map and conceptual cross-sections in Figure 3.

Pontides Orogen and the Southern Eurasian Margin

The Pontides orogen forms an east-west trending mountain range at the northern edge of the Central Anatolian plateau. In Early Jurassic times this orogen formed the southern continental margin of Eurasia (Şengör & Yilmaz, 1981; Okay & Nikishin, 2015; Dokuz et al., 2017; Topuz et al., 2014; van Hinsbergen et al., 2020), and was separated from the African continent by Tethyan ocean basins and Gondwana-derived continental blocks – the evidence of which we will describe in the following sections.

The Izmir-Ankara-Erzincan suture zone (IAESZ) marks the southern boundary of the Pontides orogen and contains Jurassic ophiolites with a so-called supra-subduction zone (SSZ) geochemical signature. These ophiolites were formed in the upper plate of a north-dipping subduction zone that existed from at least Early Jurassic time (Topuz et al., 2013; Hässig et al., 2013; Maffione & van Hinsbergen, 2018; Çelik et al., 2019). Sedimentary basins covering the IASZ suture demonstrate that in western and central Anatolia, subduction below the Pontides terminated in latest Cretaceous to Palaeocene times with the collision of the Kırşehir Block and Tavşanlı zone that we describe in the following sections (Kaymakcı et al., 2009; Meijers et al., 2010; Mueller et al., 2019). In East Anatolia however, oceanic subduction likely continued well into Paleogene times (Gürer & van Hinsbergen, 2019).

Late Cretaceous SSZ Ophiolites

South of the IAESZ, the Central Anatolian lithosphere comprises an overall east-west trending orogenic belt of continental and oceanic nappes that were accreted from Late Cretaceous to Eocene times (e.g., Şengör & Yilmaz 1981; Gürer et al., 2016; van Hinsbergen et al., 2016; 2020; Moix et al., 2008; Maffione et al., 2017; Okay, 1986; Plunder et al., 2013; Pourteau et al., 2018). All major nappes in this Anatolian orogen, which are rooted in the IAESZ, are overlain by klippes of Late Cretaceous (c. 94 – 90 Ma) SSZ ophiolites (Dilek et al., 1999; Robertson, 2004; Pourteau et al., 2010; Parlak, 2016; van Hinsbergen et al., 2016). The ophiolites are associated with metamorphic sole rocks that recorded subduction initiation at c. 105 Ma (Pourteau et al., 2018), and formed in the upper plate of an intra-oceanic subduction zone, below which oceanic and continental African Plate lithosphere subducted (e.g., Barrier & Vrielynck, 2008; Plunder et al., 2013; 2016; Menant et al., 2016; van Hinsbergen et al., 2016; 2020; Gürer et al., 2016; Gürer & van Hinsbergen, 2019). Based on paleomagnetic restorations of ophiolitic sheeted dykes, the SSZ ophiolites formed by NNE-SSW spreading (Maffione et al., 2017) This is interpreted to reflect ENE-dipping subduction, as part of a step-shaped subduction zone shown in Figure 4 (100 Ma) (van Hinsbergen et al., 2016; Maffione et al., 2017; van Hinsbergen et al., 2021).

Metamorphism and Exhumation of High-Grade Metamorphic Rocks

The structurally highest and northernmost units below the Cretaceous ophiolite klippes are the high pressure-low temperature (HP-LT) metamorphic Tavşanlı zone in western Central Anatolia (Figure 3A), and the high temperature-medium pressure (HT-MP) metamorphic Kırşehir Block in eastern Central Anatolia (Figure 3B and 3C). These units underwent burial, metamorphism, and accretion by c. 90-85 Ma (Whitney & Hamilton, 2004; van Hinsbergen et al., 2016; Pourteau et al., 2019).

The HP-LT metamorphic Afyon zone is located south of and structurally below the Tavşanlı zone and Kırşehir Block and consists of continent-derived metasediments that were metamorphosed at c. 70-65 Ma (Candan et al., 2005; Pourteau et al., 2013; Özdamar et al., 2013). There is no record of accretion of major rock units between c. 85 Ma accretion of the Kırşehir Block and Tavşanlı zone and c. 70 Ma accretion of the Afyon zone. In this time period a conceptual ocean basin that likely separated these continental units – the Intra-Tauride Basin – is thought to have subducted, producing a contemporaneous volcanic arc on the Kırşehir Block after it was accreted to the oceanic lithosphere of the Central Anatolian ophiolites (Şengör & Yilmaz 1981; Ilbeyli et al., 2004; Pourteau et al., 2010; Lefebvre et al., 2013; van Hinsbergen et al., 2016; 2020; Menant et al., 2016).

After burial and metamorphism, the Kırşehir Block, Tavşanlı zone, and Afyon zone were continually exhumed from below ophiolites and are now widely exposed on the plateau interior (Figures 3 and 5A). This started with east-west extensional exhumation of the Kırşehir Block by Late Cretaceous time (Gautier et al., 2002; 2008; Isik, 2009; Lefebvre et al., 2011; 2015; Advokaat et al., 2014; Genç & Yürür, 2010), and was followed by latest Cretaceous to early Eocene east-west extensional exhumation of the Afyon zone (Seyitoglu et al., 2017; Gürer et al., 2018b). Widespread exposure of these high-grade metamorphic rocks demonstrates several hundred kilometres of east-west upper plate extension, which was most likely driven by westward retreat of the subducting slab (Gürer et al., 2018b; van Hinsbergen et al., 2020).

Accretion of the Central Taurides Fold-Thrust Belt

The Central Taurides are a non-metamorphic fold-thrust belt that forms the high southern plateau margin to the south of the Afyon zone. The Late Cretaceous SSZ ophiolite-bearing Bozkır Nappes form the uppermost tectonic unit of the fold-thrust belt (Özgül, 1984; Andrew & Robertson, 2002; Çelik & Delaloye, 2006; Mackintosh & Robertson, 2012). The Bolkardağı and Aladağ nappes, which are interpreted as the southern non-metamorphic continuations of the Afyon zone (Özgül, 1984; Okay, 1986; Altıner et al., 2000), were accreted below the Bozkır Nappe in late Maastrichtian times (c. 72-66 Ma) based on the ages of underthrusted synorogenic sedimentary rocks (Özgül, 1984; Mackintosh & Robertson, 2012).

In middle Eocene times (c. 45-41 Ma) the Bozkır, Bolkardağı, and Aladağ nappes were thrusted at least 70 km south-westward over the Geyikdağı platform, which was subsequently deformed by thin-skinned folding and thrusting (Gutnic et al., 1979; McPhee et al., 2018a, b). In the western Central Taurides the Geyikdağı Platform formed a nappe that was thrust south-westward over the adjacent Bey Dağları platform (Figure 5D). Southward-increasing underthrusting of the Bey Dağları platform accommodated a paleomagnetically-constrained 40° CW rotation of the Geyikdağı Nappe experienced by late Eocene times, underpinning much of the western Central Taurides with continental Beydağları Platform lithosphere (Figure 4: 40Ma) (McPhee et al., 2018; 2019). In contrast, the southern Central Taurides were not affected by shortening after the middle Eocene accretion of the Geyikdağı Platform (McPhee et al., 2018b).

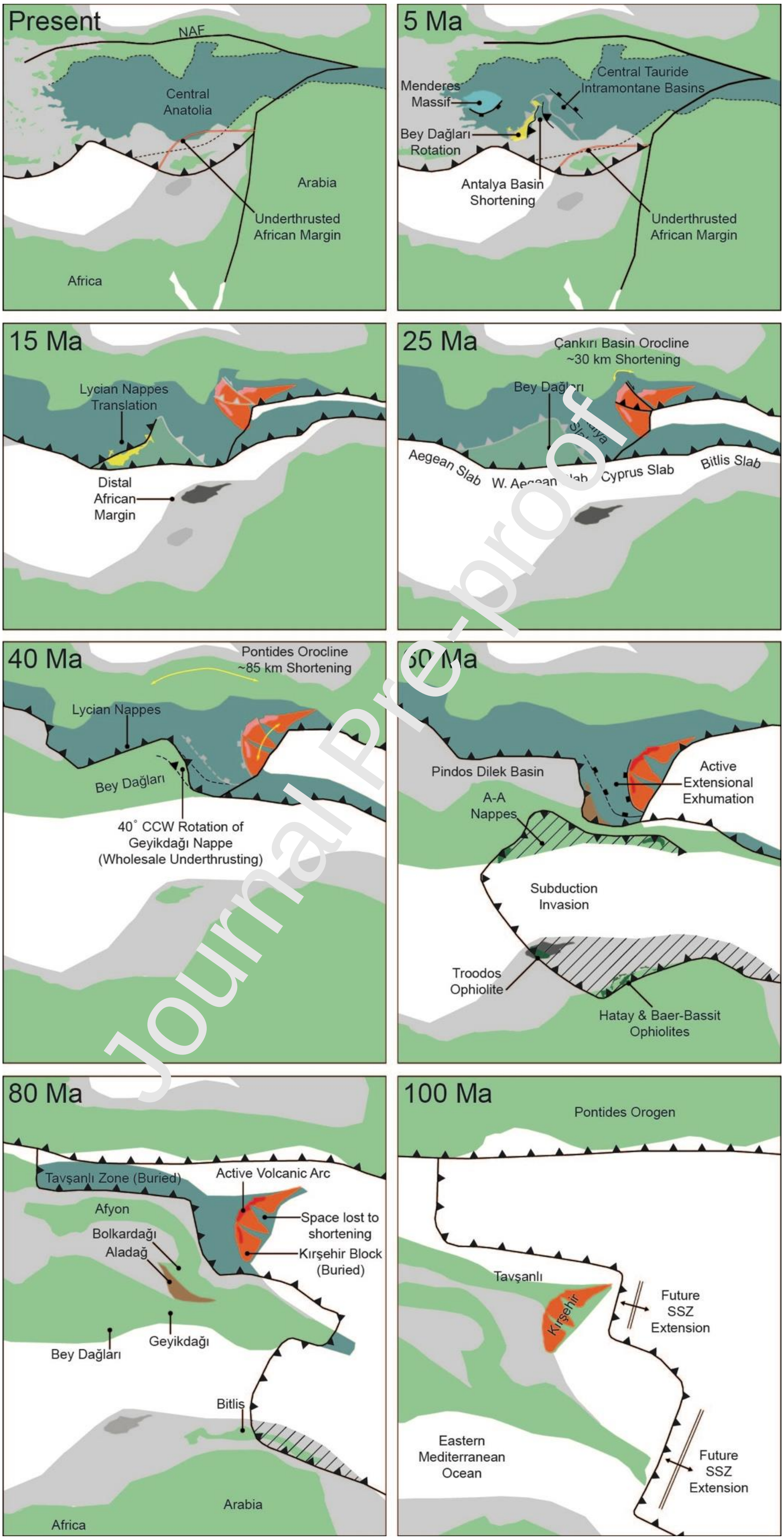
Present
NAF
Central
Anatolia
Arabia
Underthrusted
African Margin
Africa
5 Ma
Central Tauride
Intramontane Basins
Menderes
Massif
Bey Dağları
Rotation
Antalya Basin
Shortening
Underthrusted
African Margin
15 Ma
Lycian Nappes
Translation
Distal
African
Margin
25 Ma
Çankırı Basin Orocline
~30 km Shortening
Bey Dağları
Antalya
Slab
Aegean Slab
W. Aegean Slab
Cyprus Slab
Bitlis Slab
40 Ma
Pontides Orocline
~85 km Shortening
Lycian Nappes
Bey Dağları
40° CCW Rotation of
Geyikdağı Nappe
(Wholesale Underthrusting)
60 Ma
Pindos Dilek Basin
A-A
Nappes
Active
Extensional
Exhumation
Subduction
Invasion
Troodos
Ophiolite
Hatay & Baer-Bassit
Ophiolites
80 Ma
Tavşanlı Zone (Buried)
Active Volcanic Arc
Afyon
Bolkardağı
Aladağ
Space lost to
shortening
Kırşehir Block
(Buried)
Bey Dağları
Geyikdağı
Bitlis
Arabia
Africa
100 Ma
Pontides Orogen
Tavşanlı
Kırşehir
Future
SSZ
Extension
Eastern
Mediterranean
Ocean
Future
SSZ
Extension

Figure 4: *Palinspastic map-view restoration of Anatolia from 100 Ma to Present, incorporating restorations of Greece and western Turkey from van Hinsbergen & Schmid (2012); central and eastern Turkey from Gürer & van Hinsbergen (2019); the Central Taurides from McPhee et al., 2018b; Miocene oroclinal bending of the Central Taurides from Koç et al. (2018), and; Cyprus from McPhee & van Hinsbergen (2018) and Maffione et al. (2017).*

Antalya and Alanya Nappes

In a separate latest Cretaceous to Paleocene event, Late Cretaceous SSZ ophiolites were thrust from south(west) to north(east) over the southern margin of the Geyikdağı Platform (Özgül et al., 1984; McPhee et al., 2018a). Around the Gulf of Antalya, these ophiolites are associated with the far-travelled HP-LT metamorphic Alanya Nappes (Çetinkaplan et al., 2016), and the non-metamorphic Antalya Nappes that were derived from the southern margin of the Geyikdağı Platform (e.g., Robertson & Woodcock, 1981; Vrielynck et al., 2003). The Alanya and Antalya Nappes were later incorporated into the Eocene age (south)westward thrusting of the Central Taurides (McPhee et al., 2018a, b) that we describe above.

Correlation with West Anatolia

In West Anatolia (Figures 1A and 3A), a series of deeply underthrusted continental nappes equivalent to the Geyikdağı Nappe accreted until late Eocene time (Gessner et al., 2001; Lips et al., 2001; van Hinsbergen et al., 2010b; Schmidt et al., 2015), and are exposed in an extensional window in the Miocene age Menderes extensional province (e.g., Ring et al., 2003). The lowermost exposed nappe has a preserved Pan African crystalline basement and may be contiguous with the Bey Dağları Platform that forms the foreland of the western Central Taurides (Collins & Robertson, 1998; van Hinsbergen et al., 2010b). The Lycian Nappes are exposed between the Miocene-age Menderes extensional province and the Bey Dağları platform, and comprise a Late Cretaceous to Eocene nappe stack equivalent to the Bozkır and Bolkardağı nappes, including Late Cretaceous SSZ ophiolites (Collins & Robertson, 1997; 1998; 2003; van Hinsbergen et al., 2020; Plunder et al., 2016).

The structural trend of the West Anatolian fold-thrust belt was modified by a 75 km (and likely as much as 150 km) south-eastward translation of the Lycian Nappes over the Bey Dağları platform, coeval with c. 25 -15 Ma extensional unroofing of the Menderes Massif (Hayward & Robertson, 1982; van Hinsbergen, 2010; van Hinsbergen et al., 2010b). The Bey Dağları platform and Lycian Nappes were then also affected by a 25° CCW rotation from c. 15 – 5 Ma (Kissel & Poisson, 1986; Morris & Robertson, 1993; van Hinsbergen et al., 2010b).

3.2 Post-Eocene Subduction and Deformation

Subduction of the Eastern Mediterranean Ocean

Plate circuit reconstructions of the Atlantic Ocean show around 450 km of post-Eocene Africa-Eurasia convergence in the Central Anatolian region (e.g., Seton et al., 2012). The amount of convergence increased to the east, reaching 800 km of convergence north of Arabia (e.g., van der Boon et al., 2018). There is no record of post-Eocene crustal shortening in the Central Taurides associated with this convergence, and so subduction thrusts must have been located to the south of and structurally below the Central Taurides (van Hinsbergen et al. 2010; McPhee & van Hinsbergen, 2019; McPhee et al., 2018a).

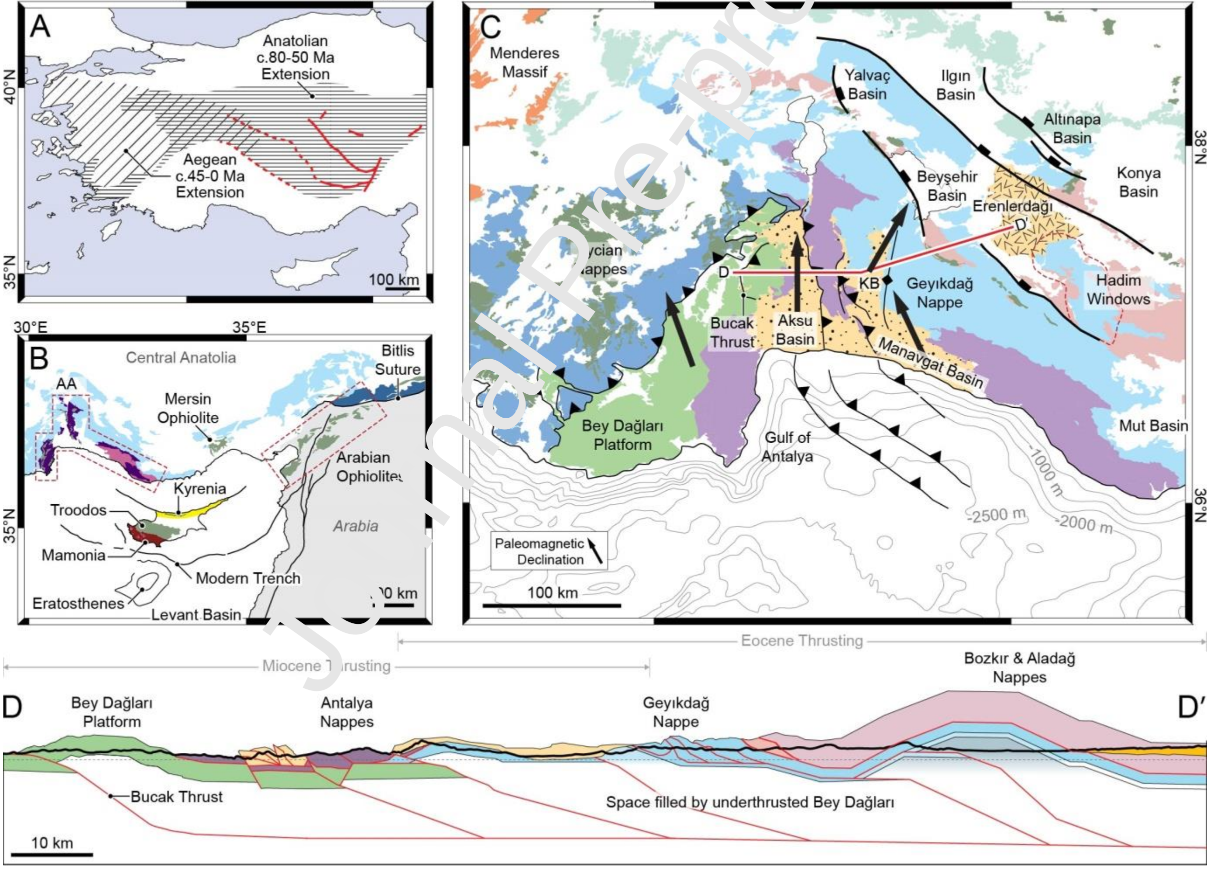

Figure 5: *A) The Aegean Extensional Province and Anatolian Extensional Province in Central and Western Anatolia, redrawn from Gurer et al. (2018). Red lines mark mapped (solid) and inferred (dashed) Central Anatolian extensional detachment faults. B) Late Cretaceous ophiolites of the Eastern Mediterranean Ocean that separated Anatolia from Africa and Arabia in late Eocene times. Redrawn from McPhee & van Hinsbergen (2018). AA = Antalya and Alanya Nappes, and associated*

ophiolitic rocks. C) Generalised geological map of the western Central Taurides, the Bey Dağları Platform, and the Lycian Nappes, showing the largest Miocene extensional and contractional structures (Koç et al., 2018; McPhee et al., 2018b). Offshore thrusts are from seismic interpretations of Hall et al. (2014). Miocene paleomagnetic data are summarised by black arrows in the Antalya Basin by Koç et al. (2016) and the Bey Dağları Platform from van Hinsbergen et al. (2010b). KB = Köprüçay Basin (sub-basin of the Antalya Basin). D) East-west cross-section along line D-D' (panel C of this figure), showing the internal structure of the western Central Taurides. Redrawn from McPhee et al. (2018a).

Upper Cretaceous oceanic lithosphere is preserved as ophiolites in the Antalya Nappes, on Cyprus (Troodos Ophiolite), and on northwest Arabia (Hatay and Baer-Bassit ophiolites; see Figure 5B) (Parlak et al., 1996; Moix et al.,2011; Morris et al., 2017; Aldanmaz et al., 2020). These ophiolites, and relics of Palaeozoic oceanic lithosphere preserved in sub-ophiolitic melanges (Moix et al., 2011; Granot, 2016), are remnants of an Eastern Mediterranean Ocean lithosphere that once separated the African margin from the Geyikdağı and Bey Dağları Platforms (Figure 4, 100 Ma) (e.g., Robertson et al., 2009; Moix et al., 2008; Menant et al., 2018; Maffione et al., 2017; Barrier & Vrielynck, 2008; van Hinsbergen et al., 2020).

The emplacement of Late Cretaceous ophiolites was likely caused by westward invasion of subduction zone that originated in East Anatolia and rolled back into the Eastern Mediterranean Ocean (Figure 4 and 10)(e.g., Maffione et al., 2017; van Hinsbergen et al., 2020). This replaced a Palaeozoic to lower Mesozoic oceanic lithosphere with a Late Cretaceous oceanic lithosphere that is partially preserved as ophiolites (Figure 4; 80 Ma and 60 Ma)(Moix et al., 2008; Barrier et al., 2018; Maffione et al., 2017; van Hinsbergen et al., 2020). After the Eocene accretion of the Central Taurides, the Eastern Mediterranean Ocean lithosphere was subducted without accretion (van Hinsbergen et al., 2010a; McPhee et al., 2018b), except for the accretion of the Misis Melange east of Adana (Figure 3A) (Robertson et al., 2004). This accounted for 400 km of Africa-Eurasia convergence (van Hinsbergen et al., 2010a; McPhee & van Hinsbergen 2019).

In East Anatolia, subduction of the Eastern Mediterranean Ocean ended with middle to late Miocene continent-continent collision of Arabia and Eurasia at the Bitlis Suture Zone (Figure 4, 15 Ma) (Şengör et al., 2003; Hüsing et al., 2009; Okay et al., 2010; Cavazza et al., 2010; 2018). In northern Cyprus, crustal shortening recorded the onset of a collision of the Central Taurides with the African distal continental margin and overlying ophiolites sometime between c. 11 to 7 Ma (McPhee & van Hinsbergen 2019). After that, the subduction plate boundary propagated to the south of and structurally below the Troodos Ophiolite, accreting the ophiolite, underlying African distal continental margin rocks, and overlying sedimentary basins to the Anatolian orogen (McPhee & van Hinsbergen 2019). Seismic stratigraphy across major faults, the development of flexural basins (e.g., Hall et al., 2005; Symeou et al., 2018), and structural and stratigraphic constraints on the onset of upper-plate contractional deformation in southern Cyprus (e.g., Kinnaird & Robertson, 2013) all suggest that the modern trench formed only in latest Miocene or Pliocene time. Furthermore, a long-lived subduction zone below Troodos is at odds with Late Cretaceous emplacement of the adjacent Hatay and Baer-Bassit ophiolites onto Arabia (McPhee & van Hinsbergen, 2018), which based on paleomagnetic and geochronological constraints, were part of the same microplate (e.g., Al-Riyami et al., 2002; Morris et al., 2006). Underthrusting of the African distal continental margin at the Cyprus Trench (Figure 4) (e.g., Robertson, 1998; Ben-Avraham et al., 2002) is ongoing at rates of 9 km/Myr, accommodating 5 km/Myr of Africa-Eurasia convergence (Reilinger et al., 2006).

Upper Plate Shortening in the Kırşehir Block

From the Kırşehir Block, to the east, a part of the reconstructed Late Cretaceous to early Miocene Africa-Europe convergence was taken up by shortening within the Anatolian orogen. Gurer & van Hinsbergen (2018) reconstructed c. 320 km of north-south shortening across Central Anatolia in Paleogene times (c. 60-25 Ma). Their reconstruction included c. 115 km shortening by restoration of oroclinal bending in the Central Pontides (Meijers et al., 2010) and Cankiri Basin (Espurt et al. 2014) (see Figure #). Approximately 200 km of Eocene-Oligocene shortening was reconstructed within the Kırşehir Block using paleomagnetically constrained restoration of vertical axis rotations (Lefebvre et al., 2013; Gurer et al., 2018a) and restoration of motion on major fault zones (Lefebvre et al., 2013; Gülyüz et al., 2013; Espurt et al., 2014; Advokaat et al., 2014; Gürer et al., 2016). A further 5 km of Eocene-Oligocene shortening deformed the Bolkar Mountains between the Mut Basin and Ecemiş fault (Figure 3A), forming a north-verging anticline and raising the Taurides to 1.5 km above the Ulukışla Basin (Gürer et al., 2016). This deformation of the Bolkar Mountains explains most of the modern topographic difference along the western Central Taurides, which rise from the 2 km elevation in the Mut Basin to 3.5 km elevation in the Bolkar Mountains (Figure 3A).

Formation and Deformation of Miocene to Present Basins

In early Miocene times, basins formed across southern Central Anatolia, including the Central Taurides Intramontane Basins (Koç et al., 2012a; 2016b; 2017) and Antalya Basin (Figure 5C) (Karabiyikoğlu et al., 2000; Flecker et al., 2005; Çiner et al., 2008; McPhee et al., 2018a). The lower

Miocene to Present Central Taurides Intramontane Basins formed by bidirectional extension, forming NW-SE and NE-SW basin-bounding faults that were parallel and perpendicular to western Central Tauride thrusts respectively (Koç et al., 2012a; 2016b; 2017). Koç et al. (2018) used paleomagnetic data to investigate vertical axis rotations of relay ramps between the basin-bounding faults, restoring up to 25 km of NE-SW Miocene extension. The Tuz Gölü Basin, which initially formed as a Paleogene sag basin, was affected by only a few hundred meters to a few kilometres of Miocene extension on steep normal faults (Fernández-Blanco et al., 2013; Ozsayin et al., 2013).

The Antalya Basin, which covers the western Central Taurides, was deformed by Miocene-Pliocene NW-SE-trending folds and thrusts (Çiner et al., 2008; Koç et al., 2016b; Poisson et al., 2003; McPhee et al., 2018a; Wasoo et al., 2020). These included a 70-km-long west-verging anticline that formed along the eastern margin of the Köprüçay Basin, uplifting Miocene marine sedimentary rocks to 1.5 km elevation (McPhee et al., 2018a). Shortening in the Antalya Basin was approximately equal to extension in the Central Tauride Intramontane Basins (McPhee et al., 2018a), leading to the development of a paleomagnetically-constrained westward-convex orocline that affected the eastern Antalya Basin and the underlying western Central Taurides in the Miocene (Figure 5C) (Koç et al., 2016a; 2018).

4. Geophysical Constraints

4.1 Modern Mantle Structure

Seismic tomography of the mantle below Anatolia (Figure 7 A-D) shows high-velocity anomalies that are interpreted as subducted slabs. A north-northeast-dipping high-velocity anomaly beneath the Gulf of Antalya and western Central Anatolia is well defined in the upper mantle (Figure 7B) and is interpreted as the Antalya slab. This slab anomaly is associated with a well-defined Wadati-Benioff zone that extends to 130 km depth (Howell et al., 2017; Kalyoncuoğlu et al., 2011) (Figure 7E). The Antalya slab anomaly is separated from the Aegean slab anomaly to the west by a well-resolved gap in the high-velocity anomaly (van Hinsbergen et al., 2010a; Biryol et al., 2011; Govers and Fichtner, 2016; van der Meer et al., 2018; Portner et al., 2018) associated with a conspicuous lack of a Wadati-Benioff zone (Bocchini et al., 2018) (Figure 7E).

A second high-velocity anomaly is resolved below Cyprus and southern Central Anatolia: this is associated with a diffuse zone of seismicity below Cyprus that reaches down to 60 – 70 km depth (Figure 7E) and is interpreted as the north-dipping Cyprus slab (Figure 7C). The upper part of the Cyprus slab anomaly may still be contiguous with the African plate (Biryol et al., 2011), or recently detached from the African plate (Portner et al., 2018; see also Gürer, 2017; van der Meer et al., 2018; Figure 6 and 7C). Most tomographic models (but not all, see Portner et al., 2018) suggest a vertical gap separates the Antalya slab from the Cyprus slab (de Boorder et al., 1998; Faccenna et al., 2006; Biryol et al., 2011; van der Meer et al., 2018; Figure Figure 6 and 7C.

Regional seismic tomographic models, which focus on Anatolian mantle structure, only resolve upper mantle tomographic features (Biryol et al., 2011; Portner et al., 2018). Global tomography models reveal that the majority of subducted lithosphere associated with Central Anatolian subduction resides in the lower mantle (Gürer, 2017; van der Meer et al., 2018). In the mantle transition zone and below, the Cyprus and Antalya slabs become indistinguishable in the tomography. The Cyprus slab appears to be overturned in the lower mantle, likely as a result of northward Cretaceous advance of the slab prior to closure of the IAESZ, and Paleogene Central Anatolian shortening (Gürer et al., 2016; Gürer, 2017).

There is no upper mantle high-velocity anomaly above the mantle transition zone (Figure 7D), and no reported earthquakes at depths greater than 50 km (Figure 6E) to the east of Cyprus, below the Miocene Bitlis Suture. This suggests that a slab associated with the Bitlis Suture broke-off and sank into the lower mantle (Faccenna et al., 2006; Hafkenscheid et al., 2006; Lei & Zhao, 2007; Biryol et al., 2011; Skobeltsyn et al., 2014).

Finally, the Pontides slab associated with the IAESZ reaches from the transition zone to the deep lower mantle (Gürer, 2017). Slab break-off along the IAESZ in western and central Anatolia probably occurred in Eocene time (e.g., Keskin et al., 2008), long before the uplift of the Central Anatolian plateau.

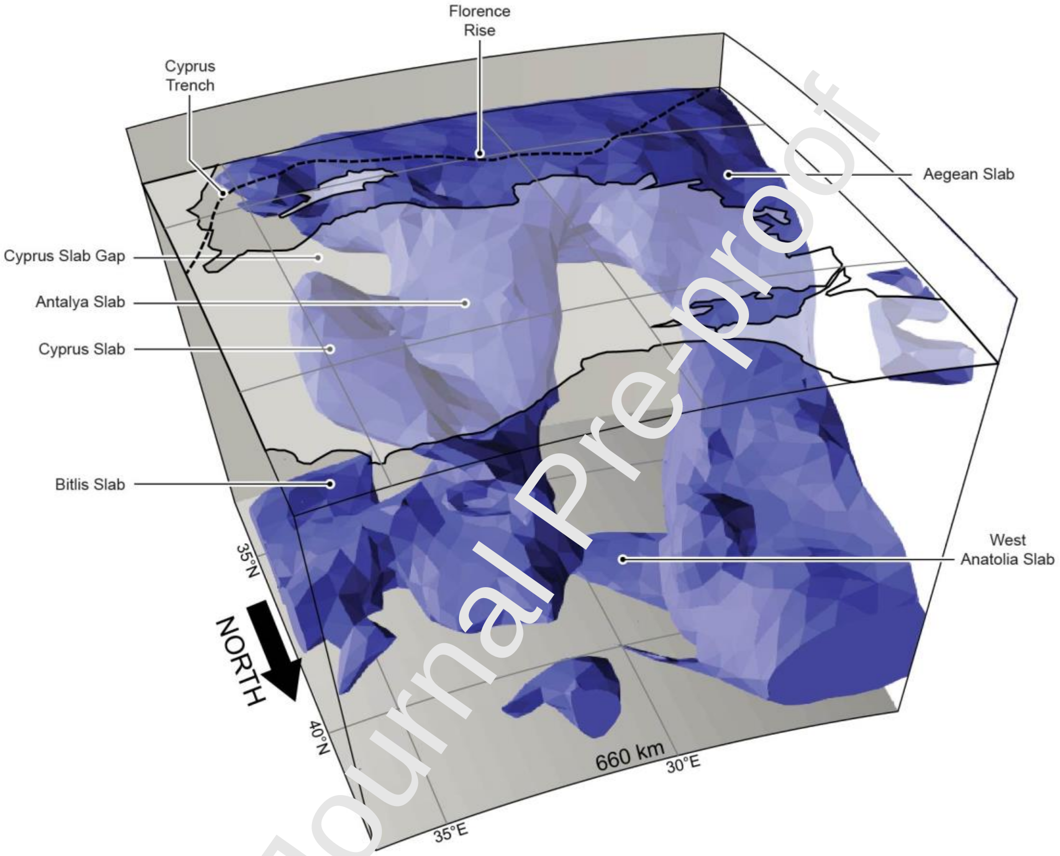

Figure 6: *3D rendering of the UU-P07 tomographic model in the study area. The blue mesh is a 0.4% isosurface extracted using marching cubes.*

4.2 Modern Lithospheric Structure

Geophysical and petrological data constrain the thickness of the Central Anatolian crust and lithospheric mantle. The Central Anatolian crust is between 30 and 45 km thick, based on receiver functions (Tezel et al., 2013; Vanacore et al., 2013; Abgarmi et al., 2017, Çivgin & Kaypak, 2017), seismic refraction data (Feld et al., 2017) and regional full-waveform tomography (Govers & Fichtner, 2016). On Cyprus and in the Levant Basin (Figure 5B), receiver functions (Vanacore et al., 2013), regional tomography (Koulakov and Sobolev, 2006), and wide-angle seismic data (Ben-Avraham et al., 2002; Feld et al., 2017) suggest that the crust is between 26 – 30 km thick.

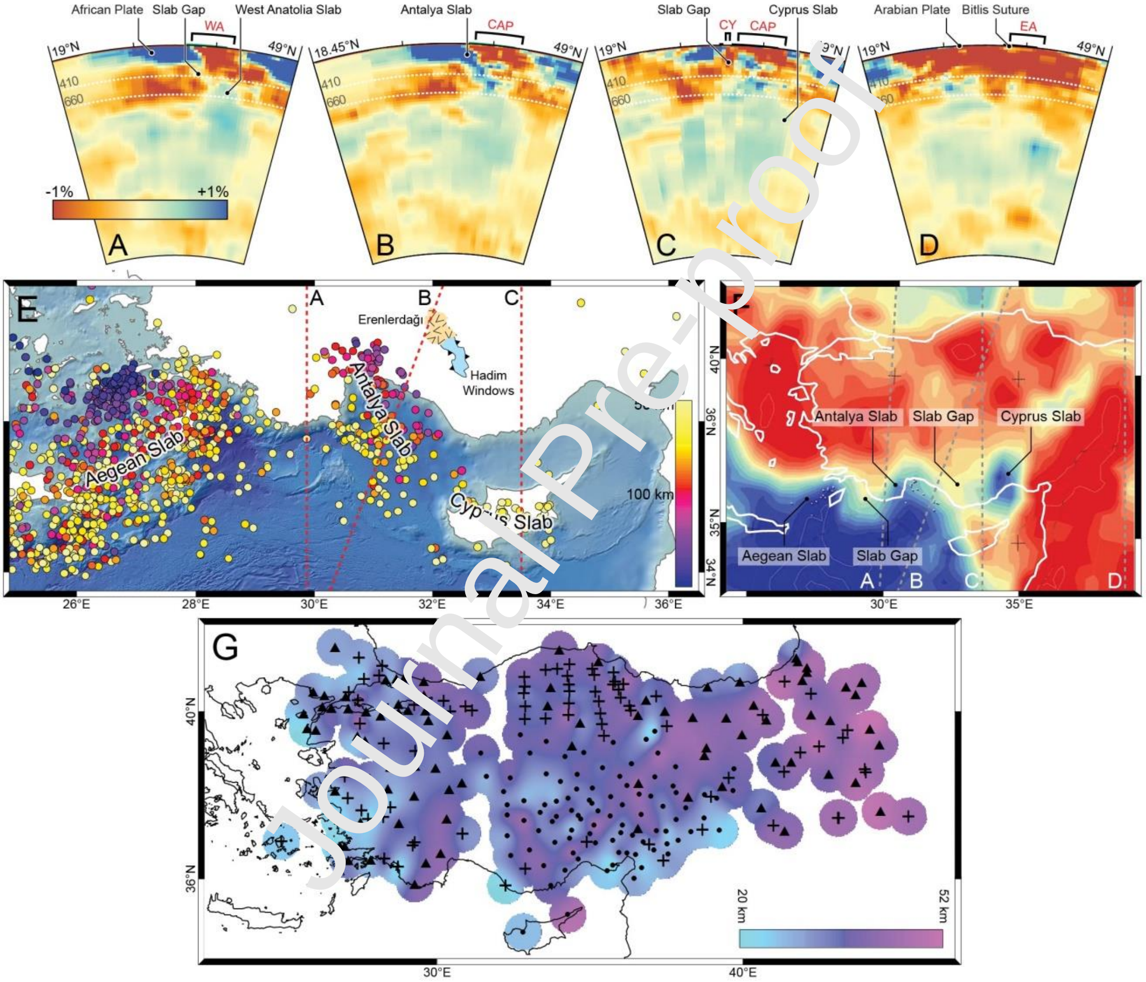

Figure 7: *A-E) Tomography slices from the UUP07 model (Amaru, 2007; Hall and Spakman, 2015) extracted using the Hades Underworld Explorer (van der Meer et al., 2018) WA = western Anatolia; CAP = Central Anatolian plateau; CY = Cyprus; EA = East Anatolia. The locations of the tomographic slices relative to Turkey are marked in Figure 5A. Note that these slices extend far to the north and south of Turkey (49°N to ~19°N). A) Detached West Anatolia slab; B) Antalya slab; C) Cyprus slab; D) Detached Bitlis slab. E) Earthquake hypocentres below 50 km depth, extracted from the USGS earthquake catalogue (1970-2016). Note that the colour depth scale is non-linear. F) Horizontal slice through the UUP07 model at a depth of 125 km, showing slab anomalies and gaps that separate*

them. G) Crustal thickness from receiver function data. Points are locations of seismic stations where receiver functions were calculated: triangles from Abgarmi et al. (2017); circles from Tezel et al. (2013); crosses from Vanacore et al. (2013). Points were interpolated using a spline algorithm with a barrier formed by a 50 km buffer around seismic data.

Estimates of the depth of the lithosphere-asthenosphere boundary (LAB) below Central Anatolia suggest that the region has an anomalously thin or even absent lithospheric mantle. The LAB has been estimated at <50 – 100 km based on S-receiver functions (Kind et al., 2015), and joint inversion of P and S receiver functions (Vinnik et al., 2014; Delph et al., 2017). In southeast Central Anatolia, the lithosphere is estimated to be around 55 km thick based on modelled melt equilibrium conditions (Reid et al., 2017).

5. Discussion

We have reviewed evidence for the rise of the plateau and its margins, and found a complex history. In a first phase, the southern plateau margins, including the Mut Basin, recorded uplift from c. 8-5 Ma at rates of around 0.15 mm/yr. This is consistent with uplift in the plateau interior that started at some time between c. 12 – 5 Ma based on stable isotope altimetry (e.g., Meijers et al., 2018) and the modelled and observed evolution of the plateau drainage system (e.g., McNab et al., 2017). We will evaluate different potential drivers of this plateau-scale uplift, and then address the multi-phase uplift of the plateau margins.

5.1 Crustal Thickening as a Driver of Plateau-scale Neogene Uplift

Crustal thickening has been an important contributor to surface uplift at various stages of Himalayan-Tibetan orogenesis (e.g., Kapp & DeCelles, 2019, and references therein) and in the Central Andes (e.g., Barnes & Ehlers, 2009, and references therein). We start our discussion by evaluating the potential role of crustal thickening as a driver of the c. 8-5 Ma plateau-scale uplift.

The 30-45-km-thick Central Anatolian crust may isostatically support the 1-km-high plateau surface because the dense underlying lithospheric mantle is only 5 – 55 km thick, as shown by seismological and petrological data (Figure 7G) (e.g., Reid et al., 2017; Kind et al., 2015). To drive Late Neogene plateau rise, crustal shortening should have occurred in response to Neogene convergence.

In the Central Anatolia region, Africa-Eurasia convergence averaged 8 km/Myr in the Neogene (e.g., van Hinsbergen et al., 2020, and references therein). One kilometre of uplift would require 7 km of crustal thickening (assuming 3300 kg.m^{-3} asthenosphere and 2800 kg.m^{-3} crust), equivalent to 20% shortening of the 550-km-wide plateau (35-km-thick crust), and consuming all Africa-Eurasia convergence from 17 Ma to the Present. Neogene basins across central and southern Central Anatolia contain no evidence for such shortening, with the exception of the Antalya Basin, where around 15 km of Miocene-Pliocene shortening occurred (McPhee et al., 2018a), and the Kırşehir Block where oroclinal bending ended in earliest Miocene times (Gurer et al., 2018). Instead, Neogene basins in the plateau interior formed by extension (e.g., Ozsayin et al., 2013; Koç et al., 2018), and Africa-Eurasia convergence was accommodated by subduction of the African Plate south of Central Anatolia (e.g., McPhee & van Hinsbergen, 2018). The 30 – 45-km-thick crust was developed by the Late Cretaceous to Eocene orogenesis that we reviewed in Section 3.

In the northern plateau margin the Central Pontides are presently affected by an additional 8 km/Myr NW-SE convergence within a restraining bend of the NAF (Yildirim et al., 2011). Schildgen et al. (2014) calculated that associated crustal shortening, which was limited to the Central Pontides, may have driven spatially variable Pleistocene uplift rates of 0.02 - 0.3 mm/yr (Yildirim et al. 2013a, 2013b; Berndt et al., 2018).

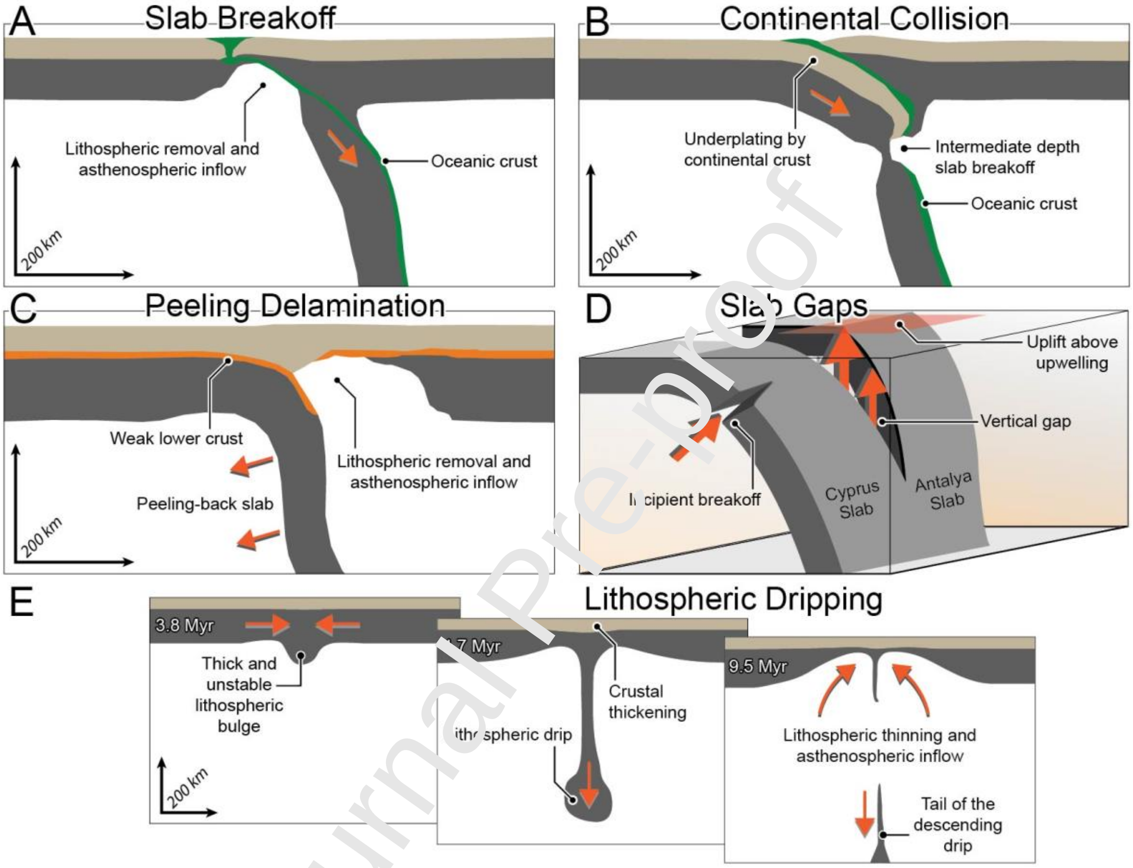

Figure 8: *Uplift mechanisms applied to Central Anatolia: i) Peeling delamination, based on modelling results in Memiş et al. (2020); ii) Continental collision (with incipient break-off), based on modelling results of Duretz et al. (2012); iii) Lithospheric dripping, based on modelling results of Göğüş et al. (2017); iv) Inflow of asthenosphere, between slab gaps, and during incipient break-off; v) Shallow slab break-off, based on modelling results of Duretz et al. (2011).*

5.2 Lithospheric Removal by Peeling Delamination or Dripping

Post c. 8-5 Ma plateau-scale uplift was therefore not driven by crustal shortening. Instead, we must look to modification or removal of the lithospheric mantle, and/or dynamic topography as drivers. Two general mechanisms explain regional mantle lithospheric removal on this scale: lithospheric dripping and peeling delamination (Figure 8C and 8E). Peeling delamination is the separation of the dense lower lithosphere from the crust by bending and sinking, with negligible internal shear strain (e.g., Bird, 1979), and is comparable to slab retreat (e.g., Göğüş & Ueda, 2018, and references therein). Lithospheric dripping is the sinking of the lower lithosphere while still attached to the crust, with high internal shear strain (e.g., Houseman & Molnar 1997; Beall et al., 2017).

Peeling delamination

Peeling delamination of flat slabs may be an important driver of uplift and upper crustal deformation in regions such as the Central Andes (e.g., Ramos & Folguera 2009), and has been proposed for the Central and East Anatolian plateaus. In a peeling delamination scenario (Figure 8C), a flat slab initially existed from the Eastern Mediterranean to the Pontides (Bartol & Govers, 2014; Govers & Fichtner, 2016). In Miocene times the slab steepened and retreated, peeling off the lithospheric mantle, and driving uplift by unloading the crust and exposing it to the asthenosphere. The process would be signalled by the enigmatic middle to late Miocene (c.16-9 Ma) onset of volcanism, including the late Miocene Galatia and the Cappadocia volcanic centres (Bartol & Govers, 2014).

Seismic tomography of the mantle below southern and western Central Anatolia contains slab anomalies that reach well into the lower mantle where they appear to be folded and thickened (Figure 7A-D)(Gürer, 2017; van der Meer et al., 2018; Portner et al., 2018). These slabs must have formed during a long history of African plate subduction because Cenozoic Africa-Eurasia convergence rates averaged only 15 km/Myr (e.g., Seton et al., 2012). The slabs are also much longer than the reconstructed paleogeographic width of the Eastern Mediterranean Ocean (450 km; van Hinsbergen et al., 2020) (Figure 4; 60 Ma). Subduction on these slabs therefore links back to Eocene and older continental subduction in Central Anatolia, and likely to Cretaceous subduction initiation (van Hinsbergen et al., 2016; 2020; Plunder et al., 2013; Pourteau et al., 2019).

Late Cretaceous to early Eocene subduction was recorded by the continual burial, accretion, metamorphism, and extensional exhumation of the Kırşehir Block and Tavşanlı zone, and then the Afyon zone. This history was analogous to the Miocene evolution of the Aegean region (Gurer et al., 2018b), where Aegean slab retreat was a driver, also exposing high-grade metamorphic rocks (e.g., van Hinsbergen et al., 2005; Faccenna et al., 2003). In Central Anatolia, several hundred kilometres of Late Cretaceous to early Eocene east-west upper-plate extension have been reconstructed, recording the westward retreat of an east-dipping Antalya slab, plus southward retreat of the Cyprus slab (Gürer et al., 2018b; van Hinsbergen et al., 2020) (Figure 5A).

A flat slab could only form in post-early-Eocene times, after extensional exhumation related to slab retreat. While Eocene-Miocene convergence may have been sufficient to transport a flat slab across Central Anatolia, based on the tectonic history we reviewed in Section 3, there is no structural or uplift-related expression of such transport (e.g., Ramos & Folguera, 2009) and no geodynamic argument to invoke slab-flattening such as a wide and unbroken slab (e.g., Schellart, 2020), ridge or oceanic plateau subduction (e.g., Gutscher et al., 2000), or rapid advance of the upper plate (e.g., Schepers et al., 2017).

Peeling-off delamination is also inconsistent with global tomographic models of the upper mantle (Figures 6C and 7). Across the Mediterranean region, slabs that steepened and retreated in Neogene times are now flat-lying on the mantle transition zone (e.g., Wortel & Spakman, 2000; Jolivet et al., 2009; see also the Aegean Slab in Figure 7). In contrast, the Cyprus and Antalya slab anomalies reach no farther than halfway across the plateau and dip steeply into the lower mantle (Gürer, 2017; van der Meer et al., 2018), ruling out recent plateau-scale slab steepening and retreat associated with peeling delamination.

Lithospheric Dripping

Lithospheric dripping is thought to be an important process in regions such as the Sierra Nevada (e.g., Zandt et al., 2004) and the Central Andes (e.g., DeCelles et al., 2015; Beck et al., 2014), and has been proposed as a driver of Central Anatolian uplift (Göğüş et al., 2017) (Figure 8E). Models of this phenomenon generally invoke localised thickening of the lithosphere (or eclogitisation of the lower crust), which becomes gravitationally unstable (e.g., Göğüş & Pysklywec, 2008; Houseman & Molnar, 1997; Marotta et al., 1998). The resulting instability grows and sinks, entraining the surrounding lithospheric mantle before finally detaching. The development of these drips drives uplift by thinning the dense lower lithosphere, and in Anatolia may have driven kilometre-scale uplift over a period of around 10 Myrs (Göğüş et al., 2017).

Göğüş et al. (2017) proposed that a lithospheric drip formed below the Kırşehir Block because after its accretion to the Anatolian orogen it was intruded by a Late Cretaceous magmatic arc and shortened by Paleogene oroclinal bending (e.g., Lefebvre et al., 2013). Also, the surrounding late Miocene Galatia and the Cappadocia volcanic centres (e.g., Kürkcüoglu et al., 2004; Reid et al., 2017) correspond to the edge of a high-velocity anomaly in the regional tomography of Fichtner et al. (2013), interpreted by Göğüş et al. (2017) as the remnants of a detached lithospheric drip.

The Late Cretaceous to Eocene evolution of the Anatolian orogen may have impacted the footprint of this process. Accretion of upper-crustal nappes, as seen in Central Anatolia or the Aegean was balanced by subduction of the lithospheric mantle (e.g., Tirel et al., 2013; Brun & Faccenna, 2008; van Hinsbergen et al., 2005). Subduction and the associated upper-plate extension in Central Anatolia thus left a thin or absent lithospheric mantle on the upper plate despite shortening relating to nappe accretion. Neogene lithospheric dripping likely affected a post Late Cretaceous lithospheric mantle – in contrast to conditions considered in numerical modelling. Paleogene oroclinal bending shortened the Kırşehir Block by around 40% (Gurer & van Hinsbergen, 2018), and so an instability may have developed even though the Kırşehir Block was likely delaminated during Late Cretaceous accretion. At least 45 km of lithospheric mantle removal is needed to drive one kilometre of uplift (e.g., Lachenbruch & Morgan, 1990), and so beyond the Kırşehir Block, uplift may have been limited by a thinly developed Paleogene-Neogene lithospheric mantle, possibly requiring additional uplift mechanisms.

Whilst lithospheric dripping below the Kırşehir Block is likely a kinematically-viable driver of Miocene uplift in eastern Central Anatolia, additional causes are needed to explain the uplift history of the plateau margin, including uplift relative to the plateau interior, the preceding and intervening marine transgressions in the southern plateau margin basins, and rapid Pleistocene uplift of Mut Basin.

5.3 Mantle Upwelling Through Slab Gaps

Ingress and upwelling of hot asthenosphere around the edges of subducted slabs has been proposed as a driver of Central Anatolian uplift by heating and thermal removal of the lithosphere, and as dynamic topography caused by changing mantle flow (Cosentino et al., 2012; Schildgen et al., 2014; 2016b). Flow has been proposed between the Cyprus and Antalya slabs (Figure 7F), through the Antalya and Aegean slab gap (Figure 7E & F and Figure 8D), and as a result of Bitlis slab break-off farther east (Figure 7D).

By late Eocene times, the extensional exhumation of high-grade metamorphic rocks in Central Anatolia had ended, and the Taurides fold-thrust belt was accreting. In the south Central Taurides, post-Eocene wholesale underthrusting of the African Plate left little or no accretionary record until the Miocene formation of a fold-thrust belt in northern Cyprus (Figures 3C and 4) (McPhee & van Hinsbergen, 2019). In the western Central Taurides, the Bey Dağları Platform had accreted to the Anatolian orogen by middle to late Eocene time (Figures 3B and 4).

In West Anatolia, the Bey Dağları Platform was accreted by late Eocene times (c. 35 Ma). The Bey Dağları platform is underpinned by a 30-35 km thick crust, leading van Hinsbergen et al. (2010a) to suggest that continuous subduction of the north-dipping West Anatolia slab, which is a coherent tomographic anomaly from 1400 km to 400 km depth (Figure 7A), was facilitated by delamination of the Bey Dağları Platform crust and its underthrusted equivalents in the Menderes Massif. The subduction thrust jumped from the deepest thrust of the Menderes Massif to the base of the Bey Dağları Platform, where it is currently piercing the surface along the Florence Rise, connecting the Aegean and Cyprus trenches. The Bey Dağları platform also formed the foreland of the western Central Taurides and was connected to the north-east-dipping Antalya slab in the Eocene. Trench jump associated with the West Anatolian slab subduction thus isolated the Antalya slab as a within-plate, passive body, still connected to the relict lithospheric mantle of the eastern Bey Dağları platform (Figures 9 and 10)(McPhee et al., 2018a). This history suggests that the Cyprus and Antalya slabs had been separated since Eocene or earlier times (McPhee et al., 2018a) – though these slabs may even have been separated since Late Cretaceous subduction initiation at E-W and N-S trenches (Figure 4, 100 Ma) (e.g., Maffione et al., 2017). The south and western Central Taurides experienced differing post-Eocene histories in terms of erosion, deposition, deformation, and uplift, likely relating to the subduction dynamics of the separate slabs (e.g., McPhee et al., 2018a; 2019).

Elsewhere, break-off of the Bitlis slab has been inferred at c. 13 – 10 Ma, based on a volcanic flare-up in East Anatolia, the rapid erosional exhumation of the overlying Bitlis region, and the end of deposition of deep marine rocks in the collision zone (Keskin, 2003; Şengör et al., 2003; Faccenna et al., 2006; Hüsing et al., 2009; Okay et al., 2010). The Antalya-Aegean slab gab formed at c. 15 Ma, based on c. 15 – 8 Ma clockwise rotation of the eastern Aegean region (external Hellenides; van Hinsbergen et al., 2005) at the edge of the retreating Aegean slab, coeval with alkaline and shoshonitic volcanism, and intrusion of granitic rocks (Jolivet et al., 2015 and references therein).

The slab gaps, and magmatism related to their formation, thus formed before the late Miocene uplift of Central Anatolia and were thus unlikely to have been a sole cause of uplift. Asthenospheric inflow may have contributed to processes such as thermal weakening of the Kırşehir Block lithosphere (Göğüş et al., 2017). It is also likely, based on evidence from generic modelling results (e.g., Király et al., 2020), that dynamic topography related to changes in slab configuration had some effect in Central Anatolia. Estimates of the contribution of dynamic topography in Central Anatolia, however, range from +2 km to -2 km (Gvirtzman et al., 2016; Howell et al., 2017; Şengül Uluocak et al., 2016): it is difficult to evaluate its effect on the plateau. Capturing modern vertical motions from the existing GPS network may allow testing of the importance of dynamic topography in the future, but dynamic topography does not seem to be a key cause of c. 8-5 Ma plateau rise.

5.4 Antalya Slab Evolution and the Uplift of the Western Central Taurides

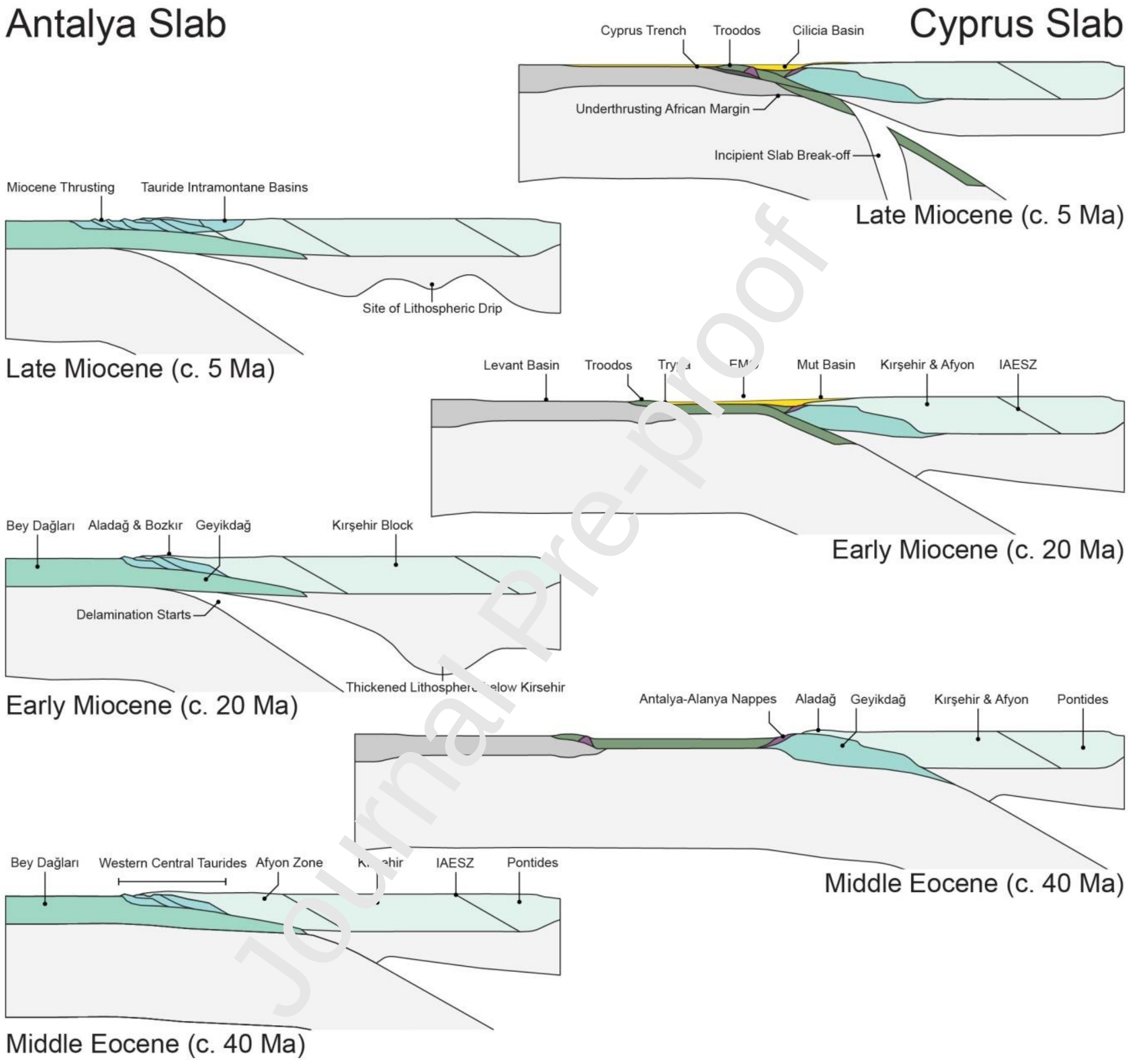

Figure 9: Conceptual early Eocene to Present evolution of subduction on the Cyprus slab along an approximate N-S section line (Figure 3B) and the Antalya slab along an ENE-WSW section line (Figure 3C). These cross-sections are approximately to scale, with 2x vertical exaggeration.

In the western Central Taurides, the Geyikdağı Nappe was connected to the northeast-dipping Antalya slab anomaly at the time of its Eocene accretion (Figure 9). This slab must have been located below or to the northeast of the Geyikdağı Nappe, which is presently exposed as far northeast as the Hadim tectonic window (Figure 5C). McPhee et al. (2019) noted that the modern Wadati-Benioff zone of the Antalya slab has been displaced westwards by at least 150 km relative to its Eocene position (Figure 6E). They thus inferred post-Eocene retreat of the slab to its present-day position below the Gulf of Antalya. There is no Eocene to early Miocene crustal deformation associated with this south-westward slab retreat (McPhee et al., 2018a), and only 25 km of NE-SW extension from the Miocene to the present (Koç et al., 2018), suggesting that most of this offset was accommodated by small-scale peeling delamination. Numerical models of this process predict a wave of vertical motions: subsidence at the point of peeling where the lithosphere is still connected to the crust, followed by uplift where the lithospheric mantle has peeled-off (e.g., Göğüş & Ueda, 2018; Memiş et al., 2020). We envisage peeling along deeply underthrusted Bey Dağları platform rocks that underpinned the western Central Taurides.

Low-temperature thermochronology data in the western Central Taurides suggest that Eocene erosion-related cooling was restricted to major thrust culminations (McPhee et al., 2019), rather than reflecting widespread uplift that could be related to peeling delamination. Low rates of cooling dominated until the early Miocene, and then increased across the western Central Taurides, likely signalling uplift and the erosion of the Bozkır and Aladağ nappes (McPhee et al., 2019). This inferred uplift was coeval with the enigmatic early Miocene subsidence and submergence of the Antalya Basin, which occurred without major deformation.

In middle Miocene times, kinematically-balanced extension in the Central Tauride Intramontane Basins (Koç et al., 2018) and shortening in the Antalya Basin region (McPhee et al., 2018a) formed the Central Taurides orocline (Koç et al., 2016a) (Figure 5C), signalling either gravitational sliding from the uplifted western Central Taurides into the Antalya Basin (McPhee et al., 2019), or retreat of the Antalya slab (Koç et al., 2016a). This may have been coeval with development of poorly-dated Miocene-Pliocene volcanism in the Erenler Dağı volcanic field (Figure 5C), which is interpreted to have formed in a volcanic arc to post collisional setting (Uyanık & Koçak 2017). Following this, the Antalya Basin was uplifted above sea level by c. 3.5 Ma (Glover, 1995), with subsidence of the Gulf of Antalya Basin continuing to the present day, defining a southwest-travelling wave of subsidence followed by uplift, as predicted by peeling delamination models.

The early Miocene subsidence of the Antalya Basin was coeval with both the subsidence of the Bey Dağları Platform (Hayward & Robertson, 1982; van Hinsbergen et al., 2010b) and southeastward gravitational sliding of the Lycian Nappes (Figure 4, 15 and 5 Ma)(Hayward & Robertson, 1982; Collins & Robertson, 1998; van Hinsbergen, 2010). This sliding was balanced by extension and unroofing of the southern Menderes Massif (van Hinsbergen, 2010; van Hinsbergen et al., 2010a).

Peeling-delamination-induced subsidence in the Antalya Basin region may have contributed to the subsidence of the Bey Dağları Platform. Lycian Nappes translation ended around c. 15 Ma (van Hinsbergen et al., 2010b) - around the same time as the inferred break-off/tearing of the West Anatolian slab (Figures 4, 6A, and 7) (van Hinsbergen et al., 2010a; Jolivet et al., 2015). Rebound of the Bey Dağları because of break-off may thus have decreased the topographic gradient, leading to an arrest in the motion of the Lycian Nappes. Shortening, oroclinal bending, and extension associated with the Central Tauride orocline, however, continued.

Finally, we consider it possible that following a phase of peeling delamination, Antalya slab subduction again caused underthrusting of the Bey Dağları Platform below the Taurides along the Bucak Thrust (Figure 5C). This would kinematically facilitate, and perhaps even drive the counter-clockwise rotation of southwest Anatolia and the southeast Aegean region (van Hinsbergen et al., 2010b; Koç et al., 2016a), including the rotation of the Bey Dağları Platform.

5.5 Cyprus Slab Evolution and the Uplift of the Mut Basin

By Eocene times, the Cyprus slab was associated with the accretion of the southern Central Taurides fold-thrust belt that was resting on a continental foreland – the accreting Geyikdağı Nappe (Figure 9). After the accretion of the Geyikdağı Nappe, Eastern Mediterranean Ocean lithosphere was subducted, accounting for 400 km of post-Eocene convergence between Anatolia and Africa, for which there is little or no accretionary record. This transition to subduction of oceanic lithosphere (Figures 4 and 9), which has a low bathymetry, may explain why the southern Central Taurides were apparently tectonically quiet throughout the Oligocene to middle Miocene, escaping erosional denudation such that even the structurally highest ophiolite-bearing Bozkır Nappes are widely preserved (McPhee et al., 2018b; 2019) (Figure 3A). Ultimately, oceanic subduction led to the transgression of the Mut Basin over the southern Central Taurides, possibly because of subduction-related process such as transient slab suction, as seen in generic models of forearc subsidence (e.g., Buiter et al., 2001; Bonnardot et al., 2008; Husson et al., 2012; Chen et al., 2017).

The Mut Basin, which was initially uplifted at c. 8 – 5 Ma, contains no evidence for upper crustal shortening (Fernandez-Blanco, 2014; Fernandez-Blanco et al., 2019), and so three sub-upper crustal causes of uplift have been proposed, each relating to the Miocene evolution of the Cyprus slab. Fernandez-Blanco (2014) suggested that if deformation of the overriding plate occurred above a region of thermal weakening at the base of the overriding plate (e.g., Fuller et al., 2006; Fernandez-Blanco et al., 2020), then surface-breaking thrusts would not form. They inferred inflation of the Mut

Basin monocline by underthrusting and accretion of subducted sediments during oceanic subduction. Schildgen et al. (2014), however, demonstrated that given slow rates of Neogene convergence, uplift by accretion of subducted sediments would take tens of millions of years.

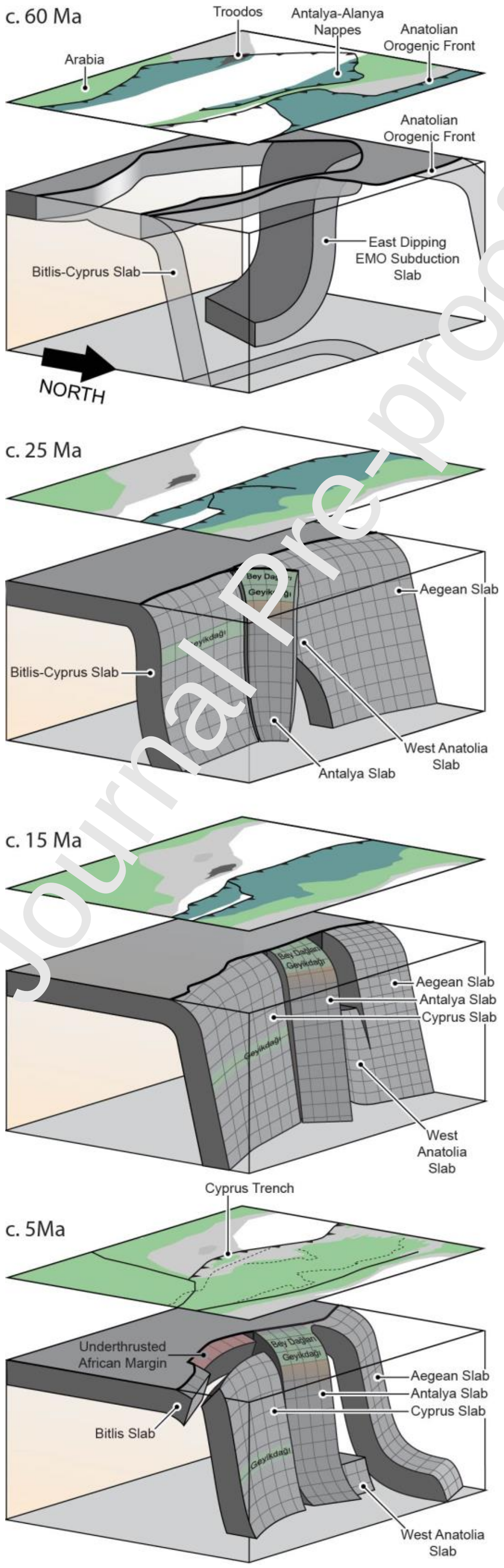

Figure 10: *Interpreted evolution of upper mantle below Central Anatolia combined with plate reconstructions shown in Figure 4. 60 Ma: East-dipping slab subducts the Triassic and older oceanic lithosphere of the EMO. North and north-northeast-dipping slabs related to the Anatolian orogen have been omitted. 25 Ma: The Antalya slab has been abandoned in the upper plate, and subduction continues on the Bitlis-Cyprus, West Anatolian, and Aegean slabs. 15 Ma: The Antalya Slab fills a gap formed by West Anatolian slab break-off. 5 Ma: The Bitlis and West Anatolian slabs have entered the lower mantle. The Cyprus slab has broken off, and the African Margin lithosphere is subducting. The Antalya Slab continues to steepen and slowly delaminate the western Central Taurides.*

As an alternative, deep underplating of the African distal continental margin following continental collision to the south has been proposed (Delph et al., 2017; Meijers et al., 2018; McPhee & van Hinsbergen, 2019). The onset of the Tauride collision with the North African continental margin likely occurred around c. 11 – 7 Ma, based on the age of thrusting on the distal, previously obducted African margin exposed on northern Cyprus (McPhee & van Hinsbergen, 2019). Reconstructed Neogene Africa-Eurasia convergence of 8 km/Myr (e.g., van Hinsbergen et al., 2020), could have feasibly brought the North African continental margin below the Mut Basin by c. 0.45 Ma, and may therefore, have caused or contributed to 1.6 km of late Pleistocene uplift documented by Öğretmen et al. (2018). This continental collision may have caused uplift by replacing an oceanic foreland with a thicker and more buoyant continental foreland, and/or by accretion and duplexing of rocks deep below the Taurides. Miocene and younger underthrusting would cause short-wavelength flexural uplift of the southern plateau and could explain coeval subsidence of the Cilicia Basin (Walsh-Kennedy et al., 2014) (Figure 2). This mechanism, however, cannot account for any earlier uplift, because the African margin was located too far south at c. 8 – 5 Ma.

Slab break-off has been invoked as a cause of c. 8 – 5 Ma uplift of the southern Central Taurides and southern plateau interior (e.g., Abgarmi et al., 2017; Meijers et al., 2018; Öğretmen et al., 2018; Portner et al., 2018; Schildgen et al., 2012b). In this scenario, uplift would be driven by rebound after removal of the slab load, and inflow of asthenosphere into the gap created (Wortel & Spakman, 1992; Davies & von Blanckenburg, 1995; Buiter et al., 2001). Numerical modelling results suggesting long-term uplift rates of around 0.1 – 0.8 mm/yr, peaking at 2 mm/yr (Duretz et al., 2011; 2014) - comparable with post c. 8 – 5 Ma rates observed in the southern plateau margin. Tomographic models are unequivocal as to whether the Cyprus slab has broken off or not (Biryol et al., 2011; van der Meer et al., 2018; Portner et al., 2018), but if slab break-off has occurred, it probably did so geologically recently, as the gap in the slab is narrow in the tomography (Figure 7C). A lack of seismicity at depths exceeding 50 km (Figure 7E) supports the possibility of break-off below Cyprus and south Central Anatolia. It is also possible that slab break-off was contemporaneous with and perhaps linked to underthrusting of the African margin. It is, however, difficult to discriminate between the effects that the two mechanisms may have had on uplift. In any case, our analysis suggests that mild Pliocene draw-down and Pleistocene uplift (Öğretmen et al., 2018) are straightforwardly explained as the combined effects of recent slab break-off and the collision of Anatolia and the African margin on Cyprus.

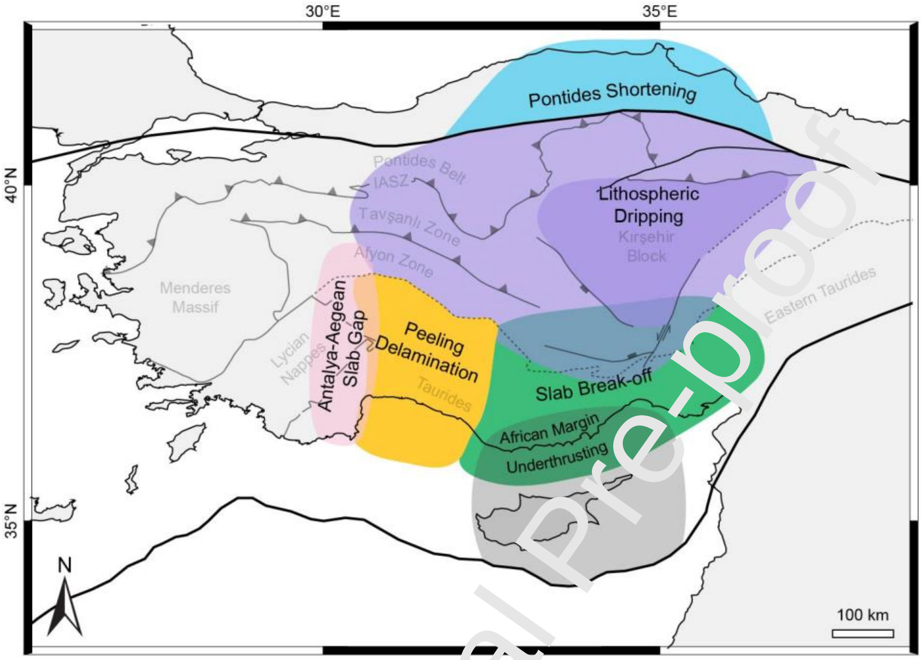

Figure 11: Interpreted footprint of the processes we have discussed in this section.

5.6 Outlook for Other Plateau Regions

The discussion above illustrates the use of the Central Anatolia for studying the geodynamic causes of plateau rise. Where in more evolved plateaus such as the Tibetan, Colorado, or Altiplano-Puna plateaus, discerning the various contributions of crustal thickening, slab evolution, and lithosphere dynamics may be challenging, the young and well-resolved Anatolian case has already revealed some of its great potential to contribute fundamentally to our understanding of the relationship between orogeny, geodynamics, and the formation of high topography. Studying the plateau crustal evolution and kinematic history provides essential constraints on the geodynamic conditions that preceded and changed to cause plateau rise (Figure 10). Our analysis shows that orogenic architecture and evolution itself may not explain the formation of high topography – Central Anatolian plateau rise clearly postdates most of the crustal accretion and deformation. Later subduction dynamics have played a crucial role by modifying the plateau margin via collision, delamination, and break-off - and likely to a lesser extent by modifying mantle flow (Figure 11).

6. Conclusions

We have evaluated potential causes of Central Anatolian Plateau rise using the long-term kinematic record of Anatolian orogenesis that, when combined with a history of Africa-Europe convergence, reveals subduction evolution. We combined this review with constraints on the spatial and temporal evolution of uplift, as well as geophysical data sets that illuminated mantle (and crustal) structure, and found that:

- Neogene crustal thickening is not a viable contributor to plateau-scale uplift since c. 8 – 5 Ma because there is no corresponding record of Neogene crustal shortening. Instead, plateau-scale uplift was driven by the modification or removal of the lithospheric mantle.
- Miocene plateau-scale peeling delamination of a flat slab is inconsistent with Late Cretaceous-Eocene slab retreat that widely exhumed high-grade metamorphic rocks. Global tomographic models of the mantle also show no evidence for plateau-scale slab retreat below Central Anatolia.
- A lithospheric instability may have formed by Late Cretaceous arc magmatism and 40% shortening of the Kırşehir Block. Lithospheric dripping in the plateau interior is, therefore, a viable mechanism of lithospheric mantle removal and uplift.
- Ingress of hot asthenosphere through slab gaps was likely active for millions of years before rapid plateau uplift, given the timing of gap formation. This phenomenon may have contributed to the long-wavelength uplift of the plateau interior but was unlikely to have been its sole cause.
- Early to middle Miocene uplift of the western Central Taurides, and coeval subsidence and subsequent Pliocene uplift of the Antalya Basin signalled retreat and small-scale peeling delamination of the Antalya slab. Subsidence associated with this process may have also contributed to the coeval translation of the Lycian Nappes and oroclinal bending of the western Central Taurides.
- Underthrusting of the African continental margin may be a kinematically-viable cause of post 0.45 Ma Mut Basin uplift but cannot explain any earlier uplift. Shallow break-off of the Cyprus slab is consistent with most seismic tomography models and an absence of deep seismicity below Cyprus: this is a plausible driver of c. 8 – 5 Ma Mut Basin uplift.

- The nascent nature of the Central Anatolian plateau allows an assessment of the relative timing and importance of different processes during plateau formation, and it may, therefore, inspire the analysis of the growth of mature high orogenic plateaus elsewhere.

Acknowledgements

D.J.J.v.H acknowledges NWO Vici grant 865.17.001. We thank Laurent Jolivet and Oğuz Göğüş for their constructive reviews.

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Credit Author Statement

P.J. McPhee: Conceptualisation, Writing, Visualisation; **A. Koç:** Conceptualisation; D.J.J. **van Hinsbergen:** Supervision, Editing, Conceptualisation, Funding Acquisition

Declaration of interests

☒ The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

☐The authors declare the following financial interests/personal relationships which may be considered as potential competing interests:

Highlights

- Lithospheric dripping is compatible with the orogenic evolution since the Cretaceous
- Cyprus slab break-off at c. 8-5 Ma related to initial south plateau margin uplift
- Subduction of African margin a viable cause of 0.45 Ma south plateau margin uplift
- Small-scale peeling delamination a driver of Antalya Basin subsidence and uplift