Fast Pliocene integration of the Central Anatolian Plateau drainage: Evidence, processes, and driving forces

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ABSTRACT

Continental sedimentation was widespread across the Central Anatolian Plateau in Miocene–Pliocene time, during the early stages of plateau uplift. Today, however, most sediment produced on the plateau is dispersed by a well-integrated drainage and released into surrounding marine depocenters. Residual long-term (10⁶–10⁷ yr) sediment storage on the plateau is now restricted to a few closed catchments. Lacustrine sedimentation was widespread in the Miocene–Pliocene depocenters. Today, it is also restricted to the residual closed catchments. The present-day association of closed catchments, long-term sediment storage, and lacustrine sedimentation suggests that the Miocene–Pliocene sedimentation also occurred in closed catchments. The termination of sedimentation across the plateau would therefore mark the opening of these closed catchments, their integration, and the formation of the present-day drainage. By combining newly dated volcanic markers with previously dated sedimentary sequences, we show that this drainage integration occurred remarkably rapidly, within 1.5 m.y., at the turn of the Pliocene.

The evolution of stream incision documented by these markers and newly obtained ⁹Be erosion rates allow us to discriminate the respective contributions of three potential processes to drainage integration, namely, the capture of closed catchments by rivers draining the outer slopes of the plateau, the overflow of closed lakes, and the avulsion of closed catchments. Along the southern plateau margin, rivers draining the southern slope of the Central Anatolian Plateau expanded into the plateau interior; however, only a small amount of drainage integration was achieved by this process. Instead, avulsion and/or overflow between closed catchments achieved most of the integration, and these top-down processes left a distinctive sedimentary signal in the form of terminal lacustrine limestone sequences.

In the absence of substantial regional climate wetting during the early Pliocene, we propose that two major tectonic events triggered drainage integration, separately or in tandem: the uplift of the Central Anatolian Plateau and the tectonic completion of the Anatolian microplate. Higher surface uplift of the eastern Central Anatolian Plateau relative to the western Central Anatolian Plateau promoted more positive water balances in the eastern catchments, higher water discharge, and larger sediment fluxes. Overflow/avulsion in some of the eastern catchments triggered a chain of avulsions and/or overflows, sparking sweeping integration across the plateau. Around 5 Ma, the inception of the full escape of the Anatolian microplate led to the disruption of the plateau surface by normal and strike-slip faults. Fault scarp partitioned large catchments fed by widely averaged sediment and water influxes into smaller catchments with more contrasted water balances and sediment fluxes.

The evolution of the Central Anatolian Plateau shows that top-down processes of integration can outcompete erosion of outer plateau slopes to reintegrate plateau interior drainages, and this is overlooked in current models, in which drainage evolution is dominated by bottom-up integration. Top-down integration has the advantage that it can be driven by more subtle changes in climatic and tectonic boundary conditions than bottom-up integration.

INTRODUCTION

Orogenic plateaus are large areas of subducted relief and high average elevation (Altiplano: 3.8 km, Tibet: 4.5 km) formed in zones of plate convergence. They are thought to form either progressively, by contraction and crustal thickening (e.g., Tibet; England and Houseman, 1989), or more rapidly, by isostatic rebound following delamination of the lithospheric mantle (e.g., Altiplano; Hoke and Garzione, 2008). The margins of these plateaus commonly act as orographic barriers to precipitation, isolating dry interiors from incoming atmospheric moisture. The aridity and tectonic fragmentation of plateau interiors are regarded as responsible for the disintegration of interior plateau drainages into mosaics of closed catchments. The aridity and perched base level of these closed catchments would hinder erosion and favor widespread sedimentation across the plateau interiors (Garcia-Castellanos, 2007; Streecker et al., 2009).

The isolation of closed plateau interiors, however, is not expected to be stable. Drainage reintegration
is expected to start along plateau margins, owing to the presence of opened catchments with lower base levels, steep slopes, and high precipitation. These conditions favor deep dissection of plateau margins, backwearing of outer plateau slopes, migration of the drainage divides inward, toward plateau interiors, and the capture of perched closed catchments and their subsequent dissection (García-Castellanos et al., 2003; Masek et al., 1994; Seeber and Gornitz, 1983; Zeilinger and Schlunegger, 2007).

Tracking drainage evolution accurately enough to test these models is often difficult, owing to the vastness and remoteness of most orogenic plateaus. The Central Anatolian Plateau in Turkey (Fig. 1), however, is an easily accessible and well-studied orogenic plateau small enough to conduct a detailed study of drainage evolution. The plateau used to display traits typical of orogenic plateaus, such as widespread continental sedimentation over the plateau interior. However, today, the plateau interior dominantly drains externally. An important pulse of drainage integration occurred on the plateau between the late Miocene and today, which did not result in substantial dissection of the plateau interior. In this study, we constrained the age, pace, and extent of drainage integration over the southern half of the Central Anatolian Plateau, where geological and geomorphological markers are well preserved. We investigated the mechanisms of integration, as well as the climatic and tectonic events that triggered drainage integration. The work relied on the compilation of many earlier works carried out in various sedimentary basins of the plateau interior, complemented by new lava \(^\text{\textsuperscript{40}}\text{Ar}-\text{\textsuperscript{39}}\text{Ar}\) ages and terrestrial cosmogenic \(^{10}\text{Be}\) erosion rates measured at key locations.

Drainage integration involves catchment capture, catchment avulsion, or lake overflow. Rivers perched on the plateau interior can be captured by rivers that erode the plateau outer slope. Asymmetric erosion across the divide that separates the perched drainages from the outer slope drainages results in the migration of the divide toward the plateau interior, until it intersects with streams of the plateau interior. These later streams are then captured by streams of the outer slope (Masek et al., 1994; Seeber and Gornitz, 1983). River capture is a bottom-up process of integration, as opposed to avulsion and overflow, which are top-down processes (Bishop, 1995). Basin avulsion occurs when a closed catchment is filled with sediment up to the elevation of its watershed, allowing the redirection of rivers into the next catchment downslope (García-Castellanos, 2007). Basin overflow refers to the filling of a terminal lake up to the elevation of the watershed. Water then spills over the watershed into the next downstream catchment. Avulsion and overflow are sensitive to climate and tectonics, as the former affects the water balance of internal lakes and sediment production from intervening ranges, while the latter can affect water and sediment fluxes on an orogenic plateau.

After reviewing the tectonic and geomorphic evolution of the Central Anatolian Plateau, we combine published isotopic data on carbonates, new \(^\text{\textsuperscript{40}}\text{Ar}-\text{\textsuperscript{39}}\text{Ar}\) ages on volcanic rocks, and new terrestrial \(^{10}\text{Be}\) erosion rates to track the evolution of the Central Anatolian Plateau drainage during the rise of the plateau. We first review the evidence for an initial state of disintegrated drainage and then constrain the age and pace of drainage reintegration. We then document associated patterns of erosion and deformation. These data are then used to assess the contributions of basin capture, basin avulsion, and lake overflow to drainage integration. Finally, we analyze how tectonic events at the turn of the Pliocene triggered drainage integration.

GEODYNAMIC, TECTONIC, CLIMATIC, AND TOPOGRAPHIC SETTING

### Present-Day Topography of the Central Anatolian Plateau

The Central Anatolian Plateau has an average elevation of ~1 km. Its surface rises progressively and irregularly from west to east, following a regional trend of eastward increase in elevation, from sea level at the Aegean coast in the west up to ~2 km in the east on the Eastern Anatolian Plateau (Fig. 1). Meridian changes in elevation are sharper. The Pontic Mountains in the north and Taurus Mountains in the south enclose the plateau interior between drainage divides that stand at average elevations of 1.7 and 2.2 km, respectively. Despite these topographic barriers, most of the plateau interior presently drains to surrounding seas across these two mountain ranges. The largest drainages are those of the Kızılırmak River, which flows to the Black Sea, the Ceyhan and Zamanti Rivers, which flow to the Mediterranean Sea, and the Euphrates River, which flows to the Persian Gulf (Fig. 2A). About one quarter of the Central Anatolian Plateau interior, however, is occupied by closed catchments that feed a series of permanent and intermittent lakes, the extent of which fluctuates considerably under the influence of climate oscillations (Fig. 2A; Erol, 1970; Fontugne et al., 1999; Kuzucuoğlu et al., 1999). The Konya closed catchment alone hosts more than half of the endorheic areas of the Central Anatolian Plateau. It feeds two successive permanent lakes (Beuşehir and Suğla). These permanent
lakes drain to Lake Konya, a large, closed terminal lake that expands considerably during wet (“pluvial”) climatic phases and shrinks into a mosaic of shallow residual lakes during the driest times (Fontugne et al., 1999; Karabiyikoğlu et al., 1999; Kuzucuğlu et al., 1999). Lake Konya in turn drains underground to Lake Tuz, another terminal lake that is currently shallow and saline. During the wettest periods of the Pleistocene, Lake Tuz reached depths of up to 100 m (Erol, 1970). Today, most of its water is contributed by slowly migrating (~1 m/yr) groundwater that precipitated over the Taurus Mountains several thousand years ago (Bayari et al., 2009a). Karstic groundwater flow is an important agent of water redistribution during dry periods, particularly in the Taurus Mountains, in which massive Paleozoic and Mesozoic limestones are abundant. Diffuse and channelized gas and groundwater flow through the carbonates has driven extensive hypogenic (Bayari et al., 2009b; Karaisaoğlu et al., 2006), endogenic (Bayari et al., 2009b), and epigenetic karstification. Groundwater flow in the karst affects the water balance of terminal lakes in the closed catchments (Bayari et al., 2009a; Eastwood et al., 2007; Fontugne et al., 1999; Kuzucuğlu et al., 2019). It contributes to lake replenishment (e.g., Lake Konya; Leng et al., 1999) or conversely to lake drainage, for example, from Lake Konya (Leng et al., 1999) and from Lake Beyşehir (Ekmecki, 1993) toward the Mediterranean Sea, feeding large springs on the outer slope of the plateau (Dumanlı spring; Fig. 2A).

**Tectonic Escape of the Anatolian Microplate**

The Anatolian plateau is traditionally divided into two adjacent plateaus (Fig. 1). The higher Eastern Anatolian Plateau lies in a zone of convergence between the Arabian and Eurasian plates, whereas the Central Anatolian Plateau lies on the westward-escaping Anatolian microplate. Surface uplift of both plateaus initiated during the late Miocene (Schildgen et al., 2012; Şengör et al., 2003), together with the final tectonic delineation of the Anatolian microplate. However, the precise unrolling of these events is not well known, owing...
to their tight, almost coeval occurrence. It leaves much room for speculation regarding the potential causal relationships between microplate formation and plateau uplift. The displacement of Anatolia relative to Eurasia is referred to as westward escape, which involves extrusion in the east, away from an area of collision between Arabia and Eurasia (in the Eastern Anatolian Plateau area), and pulling in the west, driven by slab pull and roll back along the Aegean subduction zone (Reilinger et al., 2010). The escape is accommodated by major strike-slip faults on the northern and southeastern boundaries of the Anatolian microplate. The North Anatolian fault nucleated ca. 13–11 Ma near its current eastern termination. Subsequently, the fault propagated westward, completing the northern boundary of the microplate (Facenna et al., 2006) by the turn of the Pliocene (5 Ma; Armijo et al., 1999). The southeastern boundary of the Anatolian microplate, from the eastern end of the North Anatolian fault to the Mediterranean subduction zone, is the East Anatolian fault. It formed rapidly ca. 5 Ma (Yilmaz et al., 2006; Zeilinger and Schlunegger, 2007).

Proto-escape contraction produced fold belts in the Cilicia and Adana basins (Aksu et al., 2005), long-wavelength folding and orogenic bending in the Taurus Mountains (Burton-Ferguson et al., 2005; Fernández-Blanco et al., 2019; Koç et al., 2016; Meijers et al., 2018a), and low-amplitude, long-wavelength folding in the Central Anatolian Plateau interior, manifested by a syntectonic fanning of dips in Miocene–Pliocene strata (in the Sivas, Kangal, and Koazki Basins; Fig. 2B). Distributed contractional proto-escape deformation then gave way to transtensional deformation concentrated along large through-going fault systems (Aksu et al., 2005; Fernández-Blanco et al., 2019).

Full escape of the Central Anatolian Plateau is accommodated by solid block rotation of the Anatolian microplate between the East Anatolian fault and the North Anatolian fault, with little internal deformation (Aktüg et al., 2013; Le Pichon and Kreemer, 2010). This small amount of internal deformation, in turn, is mostly accommodated along the dextral-reverse Sürgü fault, the sinistral-normal Central Anatolian fault zone, and the normal Lake Tuz fault. These faults slip slowly (Lake Tuz fault: 0.03–0.13 mm m.y.\(^{-1}\) [Özsayın et al., 2013], Central Anatolian fault zone: 0.36–1.1 mm yr\(^{-1}\) [Higgins et al., 2015; Sarikaya et al., 2015]), but they are sufficiently fast to affect erosion and sedimentation rates across the Central Anatolian Plateau, and therefore influence the evolution of its drainage.

**Rise of the Anatolian Plateaus**

The rise of the Eastern Anatolian Plateau started ca. 13 Ma, during emergence of its landmass (Keskin, 2003; Şengör et al., 2003). Its subsequent surface uplift trajectory, from sea level to its current average elevation of ~2 km above sea level (asl), is not known. Likewise, the Central Anatolian Plateau started to rise soon before or after the final retreat of the sea from the plateau interior, ca. 11 Ma (Landau et al., 2013; Meijers et al., 2018a; Ocakoğlu, 2002; Poisson et al., 2016). Marine limestones deposited on the southern margin of the plateau support emergence of the plateau before 8–6 Ma (Cosentino et al., 2012; Schlindgen et al., 2012). Basalt flow vesicularity data suggest that the plateau interior rose mostly after 8 Ma, reaching its current average elevation of ~1 km more than 2 m.y. ago (Aydar et al., 2013).

The southern margin of the Central Anatolian Plateau then rose to form the Taurus Mountains, which today include peaks as high as 3–3.5 km. A divide separates the Central Anatolian Plateau interior from its southern outer slope. A decrease in the oxygen isotope (δ18O) values of lacustrine carbonates deposited across the Central Anatolian Plateau interior between 10 and 5 Ma reflects the development of a rain shadow on the lee side of the Taurus Mountains (Meijers et al., 2018a). The δ18O values further suggest that the southern margin had already reached its current average elevation 5 m.y. ago (Meijers et al., 2018a). This chronology is consistent with the age of formation of a large south-dipping monoclinal of uplifted Miocene marine carbonates along the outer slope of the southern plateau margin (Fernández-Blanco et al., 2019), and with the massive delivery of Pliocene detrital sediments to the Cilicia-Adana basin in the Mediterranean Sea (Walsh-Kennedy et al., 2014). Records of conifer pollen in the Central Anatolian Plateau interior indicate that elevations of ~940–2000 m had been reached in the area at the end of the Miocene (Yavuz et al., 2017). Planktonic foraminifer associations of bimodal Miocene and Quaternary ages found at elevations of 1.2–1.5 km (Öğretmen et al., 2018; Yıldız et al., 2003) in carbonates of the middle Miocene Mut formation (1:100,000 geological maps of the General Directorate of Mineral Research and Exploration of Turkey) have recently been interpreted as evidence of more recent surface uplift (Öğretmen et al., 2018; Schlindgen et al., 2012).

**Tectonic Origin of the Anatolian Plateaus**

The origin of the Anatolian plateaus has been widely debated. There is a consensus that the plateaus did not form as a result of crustal thickening, except along their southern (Fernández-Blanco et al., 2019) and northern (Yildirim et al., 2011) margins. Instead, the rise of the Anatolian plateaus likely resulted from some combination of lithospheric delamination and slab break-off, in a suprasubduction setting. The thin lithosphere and elevated heat flux that characterize the Eastern Anatolian Plateau are interpreted to have resulted from the delamination of its lithospheric mantle (Reid et al., 2019; Şengör et al., 2003) and/or from the removal of an oceanic slab on which a thin Eastern Anatolian Plateau, formed as an accretionary wedge, directly rested (Govers et al., 2009; Keskin, 2003). Shear wave velocities indicate that the southern margin of the Central Anatolian Plateau is thinned (40–45 km) relative to the normal crustal thickness (30–35 km) of the Central Anatolian Plateau interior (Abgarmi et al., 2017; Delph et al., 2017). Low-velocity anomalies observed in the underlying mantle are interpreted to have resulted from the partial delamination and foundering of Central Anatolian Plateau lithospheric mantle fragments (Delph et al., 2017; Portner et al., 2018). The surface uplift of the Central Anatolian Plateau has been ascribed to break-off of the descending Arabian-Mediterranean oceanic lithosphere (Schlindgen et al., 2012) and/or associated delamination of the overriding
lithospheric mantle (Bartol and Govers, 2014), as both processes could generate the plateau by isostatic rebound. The slab break-off itself would have resulted from the arrival of Arabian continental crust into the subduction zone. The break-off would have started first in the east under the Eastern Anatolian Plateau and then propagated westward from there (Capitanio, 2016; Faccenna et al., 2006). In support for such a linkage between plateau uplift and slab break-off, several geological events typically assigned to slab break-off have been documented at the time when the plateaus started to form. These events include volcanic flare-up, trenchward migration of volcanism (Keskin, 2003; Schleifarth et al., 2018), changes in volcanic geochemical affinities (Şengör et al., 2008), and a shift from crustal contraction to extension (Göğüş and Pysklywec, 2008). However, seismicity and seismic wave velocities below Anatolia east of Cyprus fail to image remnants of formerly subducted slabs as far down as the transition zone (410–660 km; Berk Biryol et al., 2011; Portner et al., 2018). North of Cyprus, geophysical imaging reveals discontinuous high-velocity seismic anomalies (Berk Biryol et al., 2011; Delph et al., 2017; Portner et al., 2018) generally interpreted as remnants of the Cyprus slab (Berk Biryol et al., 2011; Delph et al., 2017; Portner et al., 2018; Schöllgen et al., 2012). The absence of large slabs in the upper mantle remains surprising, however, considering the young (late Miocene) age of the hypothetical break-off events.

Climate during the Rise of the Central Anatolian Plateau

Today, the climate of Central Anatolia is strongly affected by the Pontic and Taurus Mountains. These orographic barriers isolate the plateau interior from storms tracking from the Black and Mediterranean Seas (Fig. 3; Schemmel et al., 2013). Mean annual precipitation (MAP) decreases from ~1000 to 1500 mm along the sea-facing sides of both ranges down to 300–500 mm over the plateau interior. Mean annual temperature (MAT) along the Black Sea coast is lower (13 °C) than along the Mediterranean Sea coast (20 °C). MATs decrease down to 10–4 °C over the plateau interior as a result of elevation. Inward-draining basins are absent in areas that receive >750 mm of rainfall per year. Closed basins are found in drier areas (500–750 mm/yr), where groundwater flow strongly contributes to the water balance of both closed and integrated catchments (Eastwood et al., 2007; Kuzucuoğlu et al., 2019).

Pollen and spore records suggest that during early to middle Miocene times (ca. 23–15 Ma), Central Anatolia experienced humid, subtropical conditions (Akgün et al., 2007). Floral taxa suggest MATs of 17–21 °C from the late early Miocene to the late middle Miocene (ca. 18–13 Ma), steadily decreasing to 13–17 °C during the early Pliocene, ca. 5 Ma (Kayseri-Özer, 2017). The decrease could have resulted from a combination of global cooling (Zachos et al., 2001) and late Miocene plateau uplift.

In the meantime, MAPs remained steady, fluctuating between 1000 and 1400 mm/yr, with a temporary drop, down to ~900 mm/yr, during the late Miocene (Kayseri-Özer, 2017; Mazzini et al., 2013). Phytolith studies suggest that Miocene Central Anatolia was occupied by open woodlands, grasslands, and forests (Strömberg et al., 2007). Phytoliths, faunal assemblages, and dental ecometrics show that grasslands started spreading more widely during middle Miocene times (Akgün et al., 2007; Denk et al., 2018; Kaya et al., 2018; Strömbberg et al., 2007). Diverse habitats such as mesophytic forests, broad-leaved evergreen forests, xeric open woodlands, and zonal herbs, however, persisted until Pliocene times (Kayseri-Özer, 2017). During the Pliocene, sclerophyll, legume-like, and zonal herb components increased substantially. The increase in the broad-leaved component over the Central Anatolian Plateau during that time may record the surface uplift of the Central Anatolian Plateau relative to Western Anatolia (Kayseri-Özer, 2017).

Drainage Evolution along the Southern Part of the Central Anatolian Plateau

The presence of many well-preserved geomorphic markers in the southern part of the plateau makes it well suited to track major drainage changes throughout the conversion of this suprasubduction margin into the southern Central Anatolian Plateau margin.

Figure 3. Current climate across the Central Anatolian Plateau. (A) Annual rainfall; contour spacing 250 mm/yr. (B) Mean annual temperature; contour spacing 5 °C. Data: World Climate (https://www.worldclim.org).
Steeply Dissected Oligocene–Miocene Topography

Before the rise of the Central and Eastern Anatolian Plateau, Anatolia was a low-elevation accretionary wedge, composed of continental and oceanic ribbons that accreted during successive docking events from the Late Cretaceous to the Eocene (Şengör et al., 2008). After the Eocene, convergence between Arabia-Africa and Eurasia was accommodated by subduction along the Bitlis suture zone, a former trench located south of Anatolia (Fig. 1). There, the Eastern Mediterranean seafloor (Neotethys Ocean) subducted beneath Anatolia. Subduction along this trench continues until today at its western end (along the Aegean and Cyprus trenches), but it came to an end farther east when the Arabian continental crust entered the trench. The age of subduction termination is debated, but it certainly predated the late Miocene (Hüsing et al., 2009).

In Oligocene time, the Anatolian accretionary prism was emplaced; subaerial sedimentation was taking place in fault-bounded intermontane basins, widely scattered across the prism. These included the Ecemiş, Karsanti, Atokprak, Mut, and Adana Basins (Bassant et al., 2005; Jeaffreys and Robertson, 2005; Ünlügenç and Akinci, 2015; Williams et al., 1995). The topography of Central Anatolia started acquiring some of its modern geomorphological traits at the turn of the Miocene, when a phase of massive subsidence above the subduction zone led to the submergence of the southern Anatolian margin (Fig. 4). Submergence started in the east during Aquitanian time (e.g., in the Sivas [Poisson et al., 2016], Elbistan [Yusufoğlu, 2013], Darende [Booth et al., 2013], Malatya [Kaymakçı et al., 2008], Adana [Burton-Ferguson et al., 2005], and Kahramanmaraş [Gül et al., 2011; Hüsing et al., 2009]) basins, progressing westward during middle Miocene times (e.g., Dikime [Oçakoğlu, 2002] and Mut [Eriş et al., 2005] basins). The forearc region located closest to the trench has since remained submerged (e.g., Adana [Burton-Ferguson et al., 2005] and Cilicia [Aksu et al., 2005] basins), while the more internal forearc regions, farther from the trench, started re-emerging during a phase of contraction in middle Miocene time, forming the outer slope of the future southern plateau margin (Fernández-Blanco et al., 2019; Walsh-Kennedy et al., 2014).

The amplitude of the early Miocene marine transgression was much larger than the amplitude of the global eustatic early Miocene transgression (<100 m; Miller et al., 2005). In the distal forearc region, valleys were filled with fluvial sediments, submerged, and buried under marine sediments (e.g., Mut [Eriş et al., 2005]) and Adana [Derman and Gürbüz, 2007] basins. They define a basal Miocene unconformity with a local relief reaching several hundreds of meters in the Mut Basin (Eriş et al., 2005) and up to 1500 m in the Adana Basin (Burton-Ferguson et al., 2005). Farther up the continental slope, the margin remained emerged and dissected by steep gorges hundreds of meters deep. These valleys were backfilled with hundreds of meters of coarse fluvial sediments (>1300 m in the Dikme Basin [Oçakoğlu, 2002]), >600 m along the paleo-Göksu valley [personal observation], and >500 m in the paleoappearances of Tekir and Doksanlar [Gül et al., 2011]). These fluvial sediments thus preserve the lowest parts of a highly dissected and steep eroding mountain range, which occupied most of southern Central Anatolia during the early Miocene. The presence of this range, however, did not significantly affect the δ18O record of continental carbonates deposited in the region now corresponding to the Central Anatolian Plateau interior (Lüdecke et al., 2013; Mazzini et al., 2013; Meijers et al., 2016). The range did not form a continuous orographic barrier to precipitation, but mountain-derived florals in Oligocene and Miocene pollen assemblages (Akgün et al., 2007) indicate that it locally reached substantial elevations.

Miocene Relief Reduction and Drainage Disintegration

Late Miocene to Pliocene continental sediments rest on a shallowly dipping erosional surface that bevels folded middle Miocene sediments in places (see Supplemental Material Methods S1). Unlike the Oligocene–Miocene paleovalleys, this basal

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unconformity exhibits little local relief. It rises at shallow angle (2%–12%) away from Miocene–Pliocene depocenters, projecting above surrounding relief, over which it connects to perched, low-relief surfaces. The unconformity and the perched surfaces are the remnants of a once more extensive, regional low-relief surface (see Supplemental Material Methods S1). The topographic differences between the early and late Miocene unconformities reveal that Central Anatolia underwent substantial local relief reduction during the middle or early-late Miocene. Relief reduction allowed late Miocene–early Pliocene sediments to spread widely across the Central Anatolian Plateau.

The late Miocene–early Pliocene sediments (m3pl and pl of 1:100,000 geological quadrangles Central Anatolia) exhibit substantial local relief reduction during the middle or early-late Miocene. Relief reduction allowed late Miocene–early Pliocene sediments to spread widely across the Central Anatolian Plateau.

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Flow direction indicators and sediment provenance are the remnants of a once more extensive, regional low-relief surface (see Supplemental Material Methods S1). The topographic differences between the early and late Miocene unconformities reveal that Central Anatolia underwent substantial local relief reduction during the middle or early-late Miocene. Relief reduction allowed late Miocene–early Pliocene sediments to spread widely across the Central Anatolian Plateau.

The late Miocene–early Pliocene sediments (m3pl and pl of 1:100,000 geological quadrangles of the General Directorate of Mineral Research and Exploration of Turkey, https://www.mta.gov.tr/eng/; Atabey et al., 2000) were deposited in coalescent depocenters more extensive than the Oligocene and early Miocene depocenters. Mio-Pliocene sediments came to cover ~70% of the plateau interior. Their depocenters were separated by narrow mountain ranges and isolated hills. Sedimentation was dominated by fine grained and terrigenous. Flow direction indicators and sediment provenance studies (Jaffey and Robertson, 2005) demonstrated depositional architectures compatible with centripetal drainages. Covariance analyses of δ18O and δ13C values in lacustrine carbonates further indicated that some of these drainages fed chains of open and closed lakes (Meijers et al., 2018a). Such widespread sedimentation, with the presence of playas and lakes, is common on orogenic plateaus, where sedimentation commonly takes place in hydrologically closed catchments. Today, long-term (10^6–10^7 yr) sediment storage over the Central Anatolian Plateau interior takes place in closed catchments (such as those of lakes Tuz, Konya, and Yay; Fig. 2A). Open catchments instead disperse their sediments farther downstream, into the Black Sea, the Mediterranean Sea, and the Persian Gulf. It is therefore highly likely that the end of sediment retention within the Central Anatolian Plateau interior corresponded to the time at which now-opened catchments became connected to these marine sinks. Therefore, the retention of sediments in the now-opened catchments also supports the view that Miocene–Pliocene sedimentation took place in closed catchments.

■ METHODS

Mapping of Geomorphic and Stratigraphic Markers

We used the distribution of stratigraphic and geomorphic markers to track the evolution of the drainage during the rise of the Central Anatolian Plateau. High-resolution panchromatic (40-cm-resolution), and multispectral (2-m-resolution) satellite images (from satellites Ikonos, GeoEye 1, QuickBird 2, WorldView 1, WorldView 2) provided by the Polar Geospatial Center (University of Minnesota) were processed by principal component analysis. Resulting images were pan-sharpened using ArcGIS, and these were used to laterally extrapolate lithological contacts observed in the field. The resulting correlations were used to establish superposition relationships between dated volcanic layers and lacustrine carbonates. Correlations were ground-proofed across the study area during bi-annual field campaigns from 2011 to 2014.

Extraction and Interpolation of the Basal Late Miocene Unconformity

Along the southern plateau margin, we reconstructed the extent of a subdued topography that was lapped onto by Upper Miocene marine and continental sediments (Fig. 5A). This topography is preserved below Miocene–Pliocene depocenters. Everywhere else, the surface has been variously eroded, dissected, and folded at broad wavelengths (see Supplemental Material Methods S1 [footnote 1]). It remains best preserved next to the late Miocene–Pliocene basins, where it connects to their basal late Miocene unconformity. The surface can be traced over limestones affected by karstification.
at characteristic lengths of 10 m, but it is harder to track over extensively dissected Oligocene–Miocene sedimentary basins, ophiolites, as well as on ranges affected by glacial erosion. To obtain a first-order approximation of this subdued topography, we modeled it as a simple plane affected by long-wavelength deformation. Its topographic remnants were extracted from the present-day topography, and the intervening, eroded surface was interpolated. Criteria for surface extraction were local slope and residual relief values. Filtering parameters were adjusted to bedrock type. A moving 750-m-to 3000-m-wide circular window was passed over the 3° Shuttle Radar Topography Mission (SRTM) digital elevation model within the identified surface remnants, selecting highest elevations to eliminate the effects of dissection and dissolution (Brocard et al., 2011). A cubic spline interpolation was then passed over these elevation points to reconstruct the paleosurface over completely dissected areas. Summits without preserved remnants in highly dissected areas, but located within 200 m of the interpolated surface, were then added to the input data set. A second interpolation was performed, allowing the surface to break freely across identified tectonic discontinuities (hull function in ArcGIS). The resulting envelope (Fig. 5B) is regarded as accurate within ±100 m on average over the area of concern. The modern topography was finally subtracted from the envelope to reveal the distribution of dissection across the plateau margin.

Ar/Ar Dating of Volcanic Markers

New 40Ar/39Ar ages from 10 basalts and three ignimbrites (Fig. 6; Table S2-1; Fig. S2-1 [see footnote 1]) were combined with previously published 40Ar/39Ar ages (Meijers et al., 2018a, 2018b; Özsayın et al., 2013) to establish the chronology of drainage integration. To achieve this, we also relied on stratigraphic logs and descriptions (Meijers et al., 2020), as well as a number of radiometric, biostratigraphic, and magnetostatigraphic ages ascribed to continental fluvial and lacustrine sedimentary rocks and interlayered volcanic rocks (Meijers et al., 2018a, 2018b). The 13 new samples for 40Ar/39Ar dating were prepared and measured at the U.S. Geological Survey, Denver, Colorado. Technical details on sample preparation and measurement are provided in Supplemental Material Methods S2 (see footnote 1).

Constraining the Age of Terminal Depositional Surfaces

The end of sediment retention within the Central Anatolian Plateau interior corresponds to the time at which closed catchments integrated to form the current drainage. Knowing the age of termination of sedimentation in each basin therefore provides a way to track drainage integration across the plateau interior through space and time. The dating of Miocene–Pliocene sediments relies on mammal stratigraphy, magnetostratigraphy, and radiometric dating of volcanic deposits. Interbedded chronostratigraphic markers provide a maximum bound to the age of termination of sedimentation in each closed basin. Overlying markers provide a minimum age of termination. An estimate of the most likely age of termination was calculated assuming that sedimentation rates measured down section remained constant until final sedimentation (Meijers et al., 2020, 2018b). In the absence of site-specific sedimentation rates, an average of the regional sedimentation rate was used (75 ± 0.5 cm/k.y.; Meijers et al., 2018a).

Calculation of Long-Term Erosion Rates

Valley deepening rates (Table S5-1) were calculated using the difference in the elevation between present-day valley floors, volcanic deposits perched on valley sides, and the terminal depositional surfaces of Pliocene depocenters. Erosion rates below the late Miocene low-relief surface were determined assuming an age of emergence of 8 Ma along the southern margin.

Detrital Terrestrial 10Be Erosion Rates

We measured the concentration of terrestrial 10Be produced in quartz from the sandy bed-load fraction (250–500 μm) of two rivers and from the gravelly fraction of a third river (because quartz was too scarce in its sandy bed-load fraction). Riverborne quartz provides a spatially integrated average denudation rate of the feeding catchment. This rate is temporally averaged over the residence time of quartz in the production layer of soil, which amounts here to a few thousands of years (Brown et al., 1995; Granger et al., 1996). We targeted the few catchments that drain quartz-bearing lithologies along the southern plateau margin (Fig. 6). In the same area, we sampled two bedrock outcrops and two large boulders on ancient pediments that hang above the Ceyhan River gorge. They were used to obtain site-specific erosion rates on these surfaces. In situ–produced 10Be in quartz was extracted at the PennCIL lab of the Department of Earth and Environmental Science at the University of Pennsylvania, Philadelphia, Pennsylvania, and 10Be concentration was measured at PRIME lab, Purdue University, West Lafayette, Indiana, using accelerated mass spectrometry. Technical details on sample preparation and measurements are provided in Supplemental Material Methods S3 (see footnote 1).
Modeling of Isostatic Rebound

The terminal sedimentary sequences deposited in the Miocene–Pliocene sedimentary basins of the Central Anatolian Plateau are shallowly tilted at long wavelengths. Tilting occurred coeval with and/or subsequent to the initiation of the full-escape tectonic regime. Initiation of full escape was accompanied by an almost complete cessation of contraction across the plateau interior and the growth of large transtensional structures. In this context, the persistence of long-wavelength folding under the full-escape regime is intriguing. Several possible explanations can be put forward, but we tested here whether folding could result from flexural isostatic unloading of the plateau surface, in response to the drainage of the lakes and to the incision of the Miocene–Pliocene basins following drainage integration. We conducted a simulation of the amount of flexural upwarping produced by the dissection of sediments around Sivas (Fig. 2B). This area was the best suited for this simulation owing to the abundance of lacustrine cap carbonates preserved along the margins of the Sivas Basin. To conduct the modeling, we first reconstructed the topography of the Pliocene floor before its dissection, by passing a topographic envelope over the remnants of the terminal Pliocene depositional surfaces. We then calculated the eroded volume by subtracting the present-day topography from the reconstructed surface (Fig. 7A). Last, we used gFlex (Wickert, 2016) to simulate the elastic response of the underlying lithosphere to erosional unloading for five different elastic thicknesses (0, 5, 10, 15, 20, and 25 km), considering a Young’s modulus of \( E = 65 \times 10^9 \) Pa with null vertical displacement and slope values along the model boundaries (“0Displacement0Slope”).

We employed the finite difference solution from gFlex based on the approach proposed by Van Wees and Cloetingh (1994). The model produces a map of the amplitude of the vertical rebound produced, under these different elastic thicknesses, by the removal of the eroded volume. Lateral variations in the amount of vertical rebound result in a tilting of the terminal sedimentary sequences, which varies with the modeled elastic thickness.

Finally, we compared the predicted tilting values for the terminal sedimentary sequences to those observed along the basin margins. Because the isostatic rebound lifts up the original surface basin, a more comprehensive model would include an iterative erosion of the upwarped floor of the basin, which would result in a larger eroded volume than the volume used in this calculation. The modeling therefore only provides a minimum estimate of the amount of rebound and elastic thickness necessary to produce the observed tilting of the sedimentary sequences.
RESULTS: GEOMORPHIC EVOLUTION OF THE SOUTHERN HALF OF THE CENTRAL ANATOLIAN PLATEAU

Age of Drainage Reintegration

Sedimentation ended within a remarkably short time span (1.5 m.y.) across the Central Anatolian Plateau, during the early Pliocene, from 5.5 to 3.0 Ma (Fig. 8; Table S4-1). Outliers are found in the Kozaklı Basin (7–6 Ma) and along the southern margin of the Lake Tuz Basin. In this latter case, growth of the Taurus arch initiated a retreat of sedimentation northward, away from the arch axis. Sedimentation ended at 11 Ma close to the axis (Meijers et al., 2020). In the basin of Lake Konya, 10 km south of the current limit of sedimentation (Fig. S1-1C), lacustrine carbonates postdate a 6.9 ± 0.2 Ma ignimbrite (12), Table S2-1) and a 6.2 ± 0.1 Ma tuff (Meijers et al., 2018a), but they predate overlying outliers of the 5.0 ± 0.3 Ma Kızılkaya ignimbrite. Carbonates in a shallow anticline located between the current depocenters of Lake Konya and Lake Tuz are even younger, concealing the 5.0 ± 0.3 Ma Kızılkaya ignimbrite. Because the cessation of sedimentation in that region tracks the formation of the arch rather than drainage integration, it was excluded from the following analysis.

Patterns of Erosion Associated with Drainage Integration

We used several data sets to assess the distribution of erosion rates over the Central Anatolian Plateau interior and its southern margin during and following drainage integration. We combined measurements of in situ-produced $^{10}$Be erosion rates averaged over a few thousands of years (Table S3-1) to valley deepening rates (Table S5-1) averaged over the past millions of years. The detrital terrestrial $^{10}$Be rates were obtained from the sand fraction of three tributaries of the Ceyhan River (Fig. 6). They document average catchment-wide denudation rates of 38 to 54 ± 4 m/m.y. over quartz-feeding areas covering 4–30 km$^2$. The $^{10}$Be of the gravel fraction documents denudation rates of 36 ± 3 m/m.y. over 43 km$^2$. Schist and pegmatite outcrops located in the Nurhak Mountains, high on the flanks of the Ceyhan River gorge, erode at 32–69 ± 5 m/m.y. A resistant, 42-cm-high quartzite boulder resting on a pediment yielded an exposure age >260 ± 25 ka, corresponding to an exhumation rate <1.6 ± 0.2 m/m.y. All $^{10}$Be measurements consistently documented erosion rates <100 m/m.y.

Rivers of the Central Anatolian Plateau interior similarly incise at <100 m/m.y. over several millions of years (Fig. 9). Some incision rates are as low as 20 m/m.y. Conversely, incision rates along the Central Anatolian Plateau outer slope tend to exceed 100 m/m.y.; and peak at 250 m/m.y. in middle Miocene sedimentary rocks (Mut Basin and Ecemış fault corridor). Intermediate incision rates of 70–160 m/m.y. are found in the canyons of the Ceyhan and Göksu Rivers. River incision rates of 600 m/m.y. were documented at the base of the Central Anatolian Plateau outer slope, along the Mut River (Schildgen et al., 2012).

The Kızılrmak River locally incises the plateau surface at a faster pace than surrounding rivers. In the Sivas Basin, the river incises three times faster than the nearby Zamantı River. Further downstream, the Kızılrmak River has incised a gorge in the footwall of the Kayseri releasing bend at 55–118 m/m.y., mostly between 5 and 3 Ma. Farther downstream in Cappadocia, incision rates of 80–110 m/m.y. have been sustained over the past 2 m.y. (Doğan, 2011).

The spatial distribution of erosion is provided by the volume eroded below the late Miocene subdued surface (Fig. 9). Incision increases from a few tens of meters on the plateau interior to 2 km on the outer slope. A line drawn between shallowly and deeply dissected areas almost coincides with the former divide between the closed catchments of the

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**Figure 8.** (Top panel) Temporal constraints on the termination of sedimentation in various depocenters of the Central Anatolian Plateau interior along an E-W transect (locations on Fig. 2B). Dashed line—continuous sedimentation; dotted line—intervening phase of erosion or no deposition. (Lower panel) Elevation of the topmost sediments at each depocenter.
on the Central Anatolian Plateau interior and the southern outer slope. This divide tracked the Taurus Mountains arch, south of the Lake Tuz and Cappadocia basins, and possibly the Nurhak Mountains, south of Elbistan Basin (Figs. 5B and 9). Erosion crosscuts the divide where it has proceeded the deepest into the southern outer slope (Ecemis fault corridor and Mut Basin).

**Disruption of the Central Anatolian Plateau Surface by Tectonic Faults during the Early Pliocene**

The integration of the Central Anatolian Plateau drainage from ca. 5.5 to 3.0 Ma (see “Age of Drainage Reintegration” section) occurred during the development of the full-escape tectonic regime (ca. 5 Ma; see “Tectonic Escape of the Anatolian Microplate” section). The development of full escape triggered the growth of prominent fault scarps across the Central Anatolian Plateau interior. These fault scarps disrupted formerly topographically continuous depositional surfaces that linked the contiguous Miocene–Pliocene depocenters. In doing so, fault scarps altered the locations of the watersheds of contiguous closed catchments. The new $^{39}\text{Ar}/^{39}\text{Ar}$ ages were combined with field observations to constrain the relative timing of the rise of fault scarps, the cessation of sedimentation in the closed catchment, and the development of the modern drainage. An emphasis was put on key locations, where faulting impeded or promoted water and sediment exchanges between depocenters.

**Discontinuation of Surface Fluxes across the Lake Tuz Fault Scarp**

The Lake Tuz fault scarp (Fig. 10) is the most prominent active fault scarp of the Central Anatolian Plateau. It roughly separates the region concentrating the remaining closed catchments from the region of opened catchments (Fig. 2A). Slip on the fault before ca. 6.8 Ma included strike-slip and reverse components. Since Pliocene times, slip has been mostly normal with a minor strike-slip component (Krystopowicz et al., 2020; Kürçer and Göktten, 2012; Özsayın et al., 2013). Lake Tuz, the terminal lake of the Konya closed catchment, abuts its scarp. Before the scarp formed, the sedimentary basin of Lake Tuz extended farther to the east, connecting to depocenters located along the Kızılırmak River and in western Cappadocia, but the directionality of interbasin fluxes is not known. Growth of the fault scarp was instrumental in separating the Konya closed catchment from its now-open neighbors. Indeed, there is no topographic evidence (such as wind gaps or underfit valleys) that the Konya closed catchment ever overflowed the scarp, even during the highest stages of Lake Tuz (Erol, 1970; Özsayın et al., 2013).

**Discontinuation of Surface Fluxes across the Volcano-Tectonic Province of Cappadocia**

Large volcano-tectonic depressions formed in Cappadocia during the eruption of rhyolitic ignimbrites from 10 to 5 Ma (Aydar et al., 2012; Le Pennec et al., 2014. B—Basin. Green dashed lines—Miocene–Pliocene divides of the Taurus Mountains (TM) and Nurhak Mountains (NM). Black line—landward-migrating front of dissection. Sediment thickness contours spacing = 500 m (thick contours) and 100 m (thin contours). Incision rates from Table S5-1 are reported in m/m.y. (see text footnote 1), with the measurement interval between the time in brackets and the present-day (unless specified otherwise). *Data from Schildgen et al. (2012).
The absence of large unconformities among the (e.g., Derinkuyu fault [Toprak, 1998]; Lake Tuz fault [Krystopowicz et al., 2020]), creating more expansive regions of the Central Anatolian Plateau interior. Post-5 Ma tectonism and magmatism have therefore contributed to the reduction of surface fluxes between the western (Konya closed catchment) and eastern (open catchments) regions of the Central Anatolian Plateau interior.

**Formation of the Kayseri Releasing Bend: Age of Inception of Escape Tectonics**

The Kayseri releasing bend (Fig. 10) is a prominent topographic feature of the discontinuous Central Anatolian fault zone (Koçyiğit and Beyhan, 1998). The area is rich in volcanic markers, as a result of its proximity to the province of Cappadocia. It is therefore a key area in which to finely constrain the inception of full-escape tectonics. Growth of normal fault scarps around the bend contributed to cut communications between the still-closed and now-open catchments. The formation of the bend was thought to postdate the 2.7-m.y.-old eruption of the Valibaba Tepe–Incesu ignimbrite (Dirik, 2001; Koçyiğit and Erol, 2001), because the ignimbrite is offset by the faults of the bend and seals a nondissected Pliocene terminal surface. It was therefore surmised that these surfaces underwent sediment deposition when the Valibaba Tepe–Incesu ignimbrite was erupted. Because these surfaces are now perched and incised around the releasing bend, their incision was thought to have started with the formation of the bend. In fact, well-developed soils are invariably found between the Valibaba Tepe–Incesu ignimbrite (Mues-Schumacher et al., 2004) and the underlying sediments, indicating a protracted absence of sediment aggradation before the emplacement of the ignimbrite. Sediments were already channeled in nearby valleys, as evidenced by patches of ignimbrite found along the flanks of the Kızılırmak, Zamantı, and Karasu River valleys (Fig. 11, section f-f*; Fig. 12D). They indicate that rivers had already incised their valleys within 75%–100% of their current depth by 2.7 Ma. Upon its eruption, the ignimbrite overran various topographic obstructions and was deflected by others (Mues-Schumacher et al., 2004; Şen et al., 2003), such as the bounding scarps of the Kayseri releasing bend near its NE end (Fig. 11D). The releasing bend therefore does not postdate the ignimbrite.

The new $^{40}$Ar/$^{39}$Ar ages help to constrain the age of the bend. The end of Pliocene sedimentation in the depocenters of Kayseri and Emmiler (Fig. 12D) is constrained by basalt flows [5] and [3] (Fig. 11, sections e-e*, f-f*), emplaced at 4.9 and 4.2 Ma near the top of the Pliocene series. Lacustrine carbonates lap onto a 4.9 Ma basalt in the Kayseri Basin (Fig. 11, section f-f*, km 10–20), which lies near the top of the Miocene–Pliocene sequence (Fig. 11, section f-f*, km 30–35). A 4.2 Ma basalt [3] (Fig. 11, section e-e*) also lies at the top of the depositional sequence 25 km to the NE (Fig. 12D). A 4.2 Ma basalt [4], located 5 km farther south, could belong to the same lava flow, but it was emplaced in a 100-m-deep, N-S–striking valley. This valley is reportedly incised into Pliocene sediments that interfere laterally with Cappadocian ignimbrites as young as 5.8 Ma (Fig. 11, section e-e*; Fig. 12D). When the 4.2 Ma basalt [4] was emplaced, it flowed to the NW along the valley

**Figure 10. Distribution of the cap carbonates and potential extent of lakes at 5 Ma, with indication of the fault scarps that started to form in Pliocene times. Dashed lines—anticline axes. Green areas—Miocene–Pliocene carbonates of the Konya closed catchment (see text). VTI—Valibaba Tepe–Incesu. Basins: AB—Altınıy, TB—Tufanbeyli. Volcanic obstructions: D—Develi, E—Erciyes, ER—Erenler, KC—Karasadac, KD—Kepez Dağ, ME—Melendiz, PH—paleo-Hasan, YD—Yamadag, Faults: DF—Derinkuyu fault, CAFZ—Central Anatolian fault zone, EAF—East Anatolian fault, KF—Karanfilda fault, SF—Salanda fault. Topographic obstructions: AD—Ada Dağlar, BK—Bolkar Dağlar, ND—Nurhak Dağlar, TD—Tahalti Dağ. Areal extent of ignimbrites are modified from Le Pennec et al. (1994) and Şen et al. (2003). E and W represent east and west sections of each basin. Where figure space allows, basin names have been duplicated, otherwise lines have been drawn, linking East and West to basin name.
Figure 11. Cross sections of key areas (locations on Fig. 2B). Lower Zamantı: b-b*: Güzelce-Akmezar (GA), c-c*: Yayılaçık-Taşçı-Hoşça (YTH), d-d*: Karaköy-Zamantı-Develi volcano (KZD). Kızılırmak River (Emmiler Basin): e-e*: Karakaya-Amarat-Özvatan (KAO), f-f*: Emmilier-Erkilet-Boğaz köprü (EEB), g-g*: Çiftlikköy-Kabaktepe (CK) in the Ceyhan River gorge. Numbers above sections #/#—changes in strike along cross-section, in °N clockwise, going from one end of the section to the other (from # to # *). Faults: CAFZ—Central Anatolian fault zone, KF—Karanfildağ fault. Rock formations: βc—basaltic breccia, β—basalt flow, cc—Pliocene lacustrine cap carbonates, Cs—Cambrian sandstones, EoFO—Eocene flysch and olistostrome, JKg—Jurassic–Cretaceous granites to gabbros, JKΘ—Jurassic–Cretaceous ophiolite, JKΘσ—Jurassic–Cretaceous ophiolitic mélange, JKΘπ—Jurassic–Cretaceous ophiolitic serpentinite, JKc—Jurassic–Cretaceous carbonates, mα—middle Miocene andesite, mβ—middle Miocene basalt, m2—middle Miocene terrestrial sediments, m3pl—late Miocene–Pliocene fluvial and lacustrine sediments, mξ—mesozoic metamorphic marble, Ol—Oligocene terrestrial sediments; Tc—Triassic carbonates, Pc—Permian carbonates, Pgs—Paleogene sandy turbidites, PQ—Pliocene–Quaternary pyroclastics, Pzζ—Paleozoic schists, VTI—Valibaba Tepe–Incesu ignimbrite, ξc—Paleozoic–Mesozoic marble. Underlined section names (Meijers et al., 2018a, 2020): AZ—Akmezar, BK—Büyükkünye, OV—Ozvatan, TS—Taşhan. Ages (in Ma): [#] 40Ar-39Ar, this paper; (a, b, d) Meijers et al. (2018a), (c) Jaffey et al. (2004). [b] 10Be catchment-averaged surface denudation rate.
floor, away from the scarp, but also over a short distance to the SE, eventually cascading more than 200 m down the fault scarp (Fig. 11, section e-e*; Fig. 12D). Onset of scarp growth therefore occurred between 5.8 Ma (age of the incised sediment fill) and 4.2 Ma (minimum age of the valley). This timing is consistent with the cessation of lacustrine carbonate deposition over the 4.9 Ma basalt (5), 25 km farther to the SW (Fig. 12D). Collectively, these data indicate that the Kayseri releasing bend started to form ca. 5 Ma.

The two newly dated 4.2 Ma basalts provide a paleogeographical snapshot of the area. They indicate incison of the Pliocene fill close to the releasing bend scarp coeval to deposition farther away into the footwall. They evidence, therefore, synsedimentary tilting over the present-day footwall. This tilting has two possible, non–mutually exclusive origins. First, the area of sedimentation (the Emmiler Basin) lies in structural continuity with similar Miocene–Pliocene basins located farther to the NE, along the northern margin of the Sivas Basin. The tilting and uplift of the Emmiler Basin SE margin, near the releasing bend, could therefore have resulted, like in the Sivas Basin, from the long-wavelength anticlinal folding of the Sivas Basin (Fig. 10). Like in the Sivas Basin, the anticline would have here been drained by a river, transverse to the axis of arching, flowing from the anticlinal axis toward the Emmiler Basin. That river would have incised its early Pliocene valley into the Miocene–Pliocene sediments that were affected by folding. This interpretation is supported by the thinning and complete disappearance of the Miocene–Pliocene cover toward the fold axis, where Oligocene–Lower Miocene sediments are directly exposed to erosion. This axial region is now down-faulted and lies on the floor of the Kayseri releasing bend. The tectonic activity of the Kayseri releasing bend therefore disrupted a preexisting broad anticline, which represents the continuation of the Sivas anticline. The NW dip of the Miocene–Pliocene fill "surface" (Fig. 12D), crosswise to the strike of the bounding scarp, suggests that tilting resulted from a combination of tilting along the NW limb of the Sivas Basin anticline and of local tilting in the footwall of the Kayseri releasing bend bounding scarp.

Figure 12. Detailed drainage maps along the southern CAP (locations on Fig. 2B): (A) Central Anatolian Plateau southern outer slope drainage, (B) Ceyhan River drainage, (C) zone of connection between the Zamantı River and the southern outer slope, and (D) middle reaches of the Kızılırmak River near the Kayseri releasing bend. In D, contour line spacing at the surface of the Pliocene basin = 50 m. Red dashed line in A—Taurus arch axis. B—Basin; F—fault. Q—Quaternary; CAFZ—Central Anatolian fault zone, VTI—Valibaba Tepe–İncesu ignimbrite. BK—Büyükkünye section (Meijers et al., 2018a, 2020). Ages (in Ma): [#] $^{40}$Ar/$^{39}$Ar, this paper; (#) formerly published. See reference list in caption of Figure 11.
**Funneling of Surface Fluxes within the Kayseri Releasing Bend**

While interbasinal flow is impeded by the long, continuous scarp of the Tuz Gölü fault, interbasinal flow is maintained across the scarps that bound the releasing bend of Kayseri, owing to two rivers that flow through the scarps. These rivers are the Kızılırmak River, which tracks from the Sivas Basin, and the Karasu River (Fig. 12D), which only drains the interior of the releasing bend. The Kızılırmak River has established its course along the northern side of the Sivas Basin (Fig. 7A), along a belt of Miocene–Pliocene continental deposits that are bounded by basement ranges in the NE and by the NW limb of the broad Sivas anticline to the SE. The Kızılırmak River collects surface flow along this former series of depocenters until it reaches the Kayseri releasing bend. The river collects, in particular, all the streams that drain the northern limb of the Sivas anticline. This pattern used to continue farther west, across the area now disrupted by the Kayseri releasing bend. There, the Kızılırmak River keeps collecting Miocene–Pliocene basins (Emmiler and northern Cappadocia; Figs. 10 and 12D), bound to the north by basement ranges. The shallow anticline of Sivas also continues as a broad arc in eastern Cappadocia, and beyond Cappadocia, between the two Pliocene–Quaternary depocenters of the Tuz and Konya lakes (Fig. 10). The early Pliocene valley incised into the Miocene–Pliocene sediments, and the valley of the Karasu River both drained the northern limb of the anticline toward the Kızılırmak River. However, while the Karasu River has maintained its course across the growing scarps of the bend, the river that carved the Pliocene valley was unable to maintain its flow as the scarps formed, and it was rerouted along the scarps, toward either the Karasu or the Kızılırmak River.

**Rise of Fault Scars along the Drainage Divide in the Tahtalı Mountains**

The Karanfıdağ fault bounds the Dikme and Zamanti depocenters to the SE (Figs. 10 and 11, section c-c*, km 2, and section d-d*, km 2). Vertical displacements during Miocene times on this strike-slip fault resulted in the down-faulting of these basins relative to the Tahtalı Mountains (Ocakoğlu, 2004), which carry the drainage divide between the plateau interior and the southern outer slope. Vertical displacements have been described as continuing into Quaternary times (Ocakoğlu, 2004), but we did not find any unambiguous evidence of deformation after 6 Ma. An array of antithetic normal faults at Taççı accommodates 7 m of net southeastward down-faulting in Miocene conglomerates. The faults are sealed by 6.12 ± 0.06 Ma basaltic phreatomagmatic breccia (9; Figs. 6, 11, section c-c*, and 12C). Normal faults in the 2.7 ± 0.2 Ma Valibaba Tepe–İncesu ignimbrite (at f. on Fig. 6) may record a continuation of tectonic activity into the late Pliocene, but they may have been produced by gravitational spreading toward the newly incised Zamanti River valley. It seems likely therefore that the fault was not active during drainage integration.

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The Karanfıdağ fault bounds the Dikme and Zamanti depocenters to the SE (Figs. 10 and 11, section c-c*, km 2, and section d-d*, km 2). Vertical displacements during Miocene times on this strike-slip fault resulted in the down-faulting of these basins relative to the Tahtalı Mountains (Ocakoğlu, 2004), which carry the drainage divide between the plateau interior and the southern outer slope. Vertical displacements have been described as continuing into Quaternary times (Ocakoğlu, 2004), but we did not find any unambiguous evidence of deformation after 6 Ma. An array of antithetic normal faults at Taççı accommodates 7 m of net southeastward down-faulting in Miocene conglomerates. The faults are sealed by 6.12 ± 0.06 Ma basaltic phreatomagmatic breccia (9; Figs. 6, 11, section c-c*, and 12C). Normal faults in the 2.7 ± 0.2 Ma Valibaba Tepe–İncesu ignimbrite (at f. on Fig. 6) may record a continuation of tectonic activity into the late Pliocene, but they may have been produced by gravitational spreading toward the newly incised Zamanti River valley. It seems likely therefore that the fault was not active during drainage integration.

**Faults Isolating the Central Anatolian Plateau Interior from Its Southern Margin**

While the drainage divide between the Central Anatolian Plateau interior and its southern outer slope is pinned to the axis of the Taurus Mountains arch south of the remaining closed catchments of the plateau interior, its position farther east follows narrow mountain ranges (the Tahtalı and Nurhak Mountains) that are bounded by discreet tectonic dislocations on their flanks facing the plateau interior. The rise of fault scarps along these flanks may have promoted and/or maintained the isolation of the closed catchments of the plateau interior respective to the southern outer slope. Here, we review the chronological evidence regarding the activity of these faults and nearby basin development.

**Rise of the Sürgü Fault Scarp and Formation of the Elbistan Basin**

Dextral-reverse slip along the Sürgü fault resulted in the rise of the Nurhak Mountains relative to the Elbistan Basin (Fig. 12B). The age of initiation of the scarp is not known, but because the activity of the Sürgü fault is closely related to that of the East Anatolian fault zone (Koç and Kaymakçı, 2013), it is probable that the scarp started forming together with the East Anatolian fault during the early Pliocene. Wind gaps and abandoned valleys across the Nurhak Mountains document the tectonic defeat of several rivers that used to cross the mountains (Fig. 12B), and the formation of a drainage divide along the mountain range. The defeated rivers are now collected along the fault scarp, joining to form the Ceyhan River, the only river that has maintained its course across the scarp.

The Ceyhan River drains the Elbistan Basin, which is currently bypassed by sediments. The basin was a net sediment trap until at least 4.7 ± 0.2 Ma, when sedimentation in swamps or shallow lakes led to the deposition of lacustrine carbonates and alkaline peat (Yusufoglu, 2013). Sediment trapping and water ponding upstream could have been a transient response of the Ceyhan River to uplift on the footwall of the Sürgü fault (Brocard et al., 2012; van der Beek et al., 2002). It is unclear whether sediment trapping was associated with a full hydrologic closure of the basin for some time. This would imply a discontinuation of flow along the Ceyhan River across the Nurhak Mountains. Its gorge is currently >1.5 km deep, and it comprises two nested valleys (Fig. 11, section g-g*): a broad, upper valley to which most incoming tributaries are graded, and a lower, 500-m-deep inner gorge. The formation of the inner gorge may have resulted from the resumption of (or increase in) outflow from the Elbistan Basin, or from the upstream propagation of a wave of incision nucleating along the outer plateau slope. The activity of the Sürgü fault therefore overlapped with drainage integration. However, its activity would have favored catchment closure rather than opening. The nearby Göksu River displays a distribution of sedimentation and deposition similar to that of the Ceyhan River, and yet it has not faced the rise of a mountain range across its course. Pliocene lacustrine lignites similar to those of the Elbistan Basin were deposited in the Tufanbeyli Basin (Fig. 10), in the headwaters of the Göksu River, and the river
flows into a deep canyon incised farther downstream (Fig. S1-1B). These similarities suggest that local faulting did not control the timing of sedimentation and erosion during Pliocene time, and that the evolution of the drainage was controlled by a more regional driver.

**DISCUSSION 1: MECHANISMS AND MARKERS OF DRAINAGE INTEGRATION**

The rapid pulse of drainage integration between 5.5 and 3 Ma may represent the response of surface drainage to a large and sudden modification of the tectonic and/or climatic environment. It may also represent a large response to more subtle changes in boundary conditions that were amplified by chain reactions and runaway processes. The involved mechanisms and evidence for each are reviewed next.

**Bottom-Up Capture by the Outer Slope Drainage**

Erosion is expected to slow down over plateau interiors during the rise of orogenic plateaus as a result of decreasing atmospheric moisture delivery and the establishment of closed catchments with high base levels. Conversely, erosion is expected to increase on the outer slopes, where rivers benefit from increasing topographic gradients and the focusing of orographic precipitation on windward slopes (Garcia-Castellanos, 2007; Masek et al., 1994; Whipple and Gasparini, 2014). Ultimately, the backwearing of the outer slopes promotes inward migration of the drainage divides that separate the outer slopes from the plateau interiors (Fig. 13B-1). Outer slope drainages are therefore expected to intersect and capture drainages of the plateau interiors (Fig. 13B-2; Garcia-Castellanos et al., 2003; Zeiling and Schlunegger, 2007; Zhang et al., 2014). Upon capture, the captured drainage adjusts to the outer slope base level, and incision spreads across the captured catchment (Brocard et al., 2011). Incision of the captured catchments in turn increases the topographic gradient between the captured catchments and the still-closed surrounding catchments. Surface lowering in the captured catchments also promotes the penetration of atmospheric moisture into the captured regions (Brocard et al., 2012). The penetration of such outer-slope characteristics into a captured catchment promotes the capture of other closed catchments located further inside the plateau (Fig. 13B-3). The process is expected to repeat itself, leading to a piecemeal reintegration of interior drainages. The diagnostic characteristics of such bottom-up integration include (1) a substantial inward migration of the drainage divide, (2) deep incision following capture, and (3) the development of substantial topographic gradients before each capture.

The distribution of erosion rates across the southern Central Anatolian Plateau (Fig. 9) shows that the outer slope erodes faster than the Central Anatolian Plateau interior. In the west, the water divide between currently closed catchments and the outer slope (Fig. 2A) coincides with the axis of maximum surface uplift. The outer slope is drained by series of remarkably parallel rivers (Fig. 12A). Parallel networks are the first drainages to grow on newly exposed slanting surfaces with no obstacle to free flow (Phillips and Schumm, 1987). The parallel drainage was established during the emergence of the southern slope of the plateau. These rivers first incised the marine carbonates that used to mantle the southern slope, and are now only preserved on interfluve positions. Incision of the outer slope led to the entrenchment of the rivers into these marine carbonates and then into underlying meta-carbonates and ophiolites, onto which the rivers are superimposed. The outer slope drainage shape has evolved into a less primary layout in the two areas where the incision is the deepest.

First, in the Mut Basin, river incision has proceeded deeply into erodible early–middle Miocene sedimentary rocks (Fig. 9). Excavation has contributed large sediment volumes to the Mediterranean Sea since the Pliocene, trapped in proximal sinks (Cilia and Iskenderun basins; Fig. 9; Walsh-Kennedy et al., 2014). There, the Taurus Mountains divide has retreated 30–50 km inland (Fig. 9) with respect to the axis of uplift, now reaching middle Miocene lacustrine carbonates dipping toward the plateau interior. Second, incision has proceeded –2 km into the Oligocene–Miocene infill of a narrow basin developed along the Ecemiş fault corridor (Jaffey and Robertson, 2005), and into the Oligocene–Miocene Aktoprak Basin. The superimposed margin-transverse, parallel drainage network crosses the Ecemiş fault corridor and the tectonic structures of the southern Aktoprak Basin (Figs. 12A and 12C). The parallelism of the rivers vanishes in the headwaters, beyond the former position of the drainage divide. There, the headwaters occupy an area were rivers of the outer slope (such as the Çakıt River; Fig. 13A) have penetrated 25–35 km into the plateau interior and dissected the northern limb of the Taurus Mountains arch. This evolution is confirmed by provenance analysis of Miocene–Pliocene fluvial sedimentary rocks (Radwany et al., 2017). The coincidence of deep incision, divide breaching, and substantial incision of the formerly closed regions makes a case for drainage integration by capture along the outer slope. However, the penetration depth of these drainages is small compared to the size of the Central Anatolian Plateau interior.

Divide breaching and incision are decoupled farther east. The Zamantı River, which drains part of the plateau interior, flows close to the divide (Tahtalı Mountains) for some distance before veering south and crossing the divide (Fig. 12A). This turn looks like an elbow of capture in map view. However, this change of direction occurs in an area where there is evidence neither for a divide migration nor for a former continuation of the river over the plateau interior. A continuation of the river farther west would have been prevented by the presence of the Develli volcano since late Miocene time (Fig. 12C; Fig. 11, section c-c*). Besides, at odds with events expected to follow capture, incision along the Zamantı River did not increase, but instead decreased during the Pliocene (Fig. 13A). Indeed, the 2.8 ± 0.2 Ma Valibaba Tepe–İncesu ignimbrite (Fig. 11, section b-b*, km 21, and section c-c*, km 14 and 27) spread across a Zamantı River valley that had already reached its current depth (Meijers et al., 2020). The Zamantı River also flows across a series of coalescent Pliocene depocenters that share comparable elevations and low erosion rates. Stream captures are therefore highly unlikely.
to have contributed to drainage integration along the Zamantı River, and to the connection of the river with the drainage of the outer slope.

Farther east, the drainage of the outer slope formed following early–middle Miocene emergence; it may therefore predate plateau uplift. The location of the divide over time in this area is unclear: It may have been permanently located in the Tahtalı Mountains. Following the rule that sediment deposition and the development of lacustrine conditions on the Central Anatolian Plateau imply closed-catchment conditions, the divide may have momentarily shifted south into the Nurhak Mountains during lacustrine deposition in the Elbistan and Tufanbeyli Basins. Reintegration along the same drainage lines, however, is more readily accounted for by the processes described in the following section.

To sum up, clear cases of capture by the outer slope drainage are rare along the southern plateau margin. Their contribution to drainage integration is restricted to their immediate surroundings, such that their overall contribution to the integration of the Central Anatolian Plateau is minimal.

**Top-Down Integration by Overflow or Avulsion**

The fact that the modern lakes of the Central Anatolian Plateau are located in closed catchments and not in open catchments cannot be accounted for by obvious systematic differences that would be less conducive to the formation of lakes in open

![Diagram](http://pubs.geoscienceworld.org/gsa/geosphere/article-pdf/doi/10.1130/GES02247.1/5271392/ges02247.pdf)
catchments. Lakes in open catchments have simply a shorter lifetime because they fill up with sediments, or drain by outlet incision at a faster pace than those located in the closed catchments. This is because specific water and sediment discharge are higher in open catchments. Such higher fluxes probably likewise explain how formerly closed catchments became open in the first place: They experienced an increase in sediment and/or water discharge that led to their opening, by basin avulsion or overflow. Basin avulsion occurs when an inward-draining depocenter is filled with sediment up to its lowest rim, allowing for its rivers to pour either into the next closed catchment (Fig. 13C-1) or directly onto the outer slope (Fig. 13C-2). It does not require the establishment of a positive water balance. On the other hand, a more positive water balance, through either an increase in water delivery or a decrease in water loss (by evaporation or infiltration), will drive the formation/expansion of the terminal lake, until water losses and input equilibrate. Such equilibrium may not be reached before the lake starts overflowing the closed catchment watershed (Fig. 13B-2), allowing its opening by overflow.

A pulse of incision, akin to the one produced by capture, may follow avulsion or overflow, owing to the establishment of a lower-elevation base level. However, its amplitude is commensurate to the amount of the drop in base level and can therefore be quite small (Fig. 13C-2). Avulsion and overflow are therefore compatible with the low amount of incision observed across the Central Anatolian Plateau interior (Fig. 13A). In areas occupied by contiguous closed catchments, avulsion and overflow are powerful agents of drainage integration because integration of one catchment into the next modifies the water balance in the receiving catchment. This modification potentially triggers its avulsion or overflow (Fig. 13B-2). Successions of closed catchments can thus integrate into one single drainage within a few 10⁶ yr (Enzel et al., 2003), and this chain reaction can spark widespread integration without substantial external forcing. Avulsion and overflow therefore better account for rapid (<1.5 m.y.) drainage integration across the Central Anatolian Plateau as compared to slower processes of successive captures, which have only achieved integration over limited areas next to the outer plateau slope. Incision of the Pliocene depocenters is limited from a few tens of meters to at most 300 m along the Kızılrmak River (Fig. 13A). This contrasts with the deep dissection observed on the southern plateau margin (Fig. 9), which has not propagated into the plateau, as would be expected if integration were driven by a bottom-up process. Along the rivers that drain to the outer slope of the plateau, the propagation of incision over the past 4–5 m.y. has been very limited. This can be explained by the fact that the interior drainages, in this phase following integration, contribute little additional water and sediment to the river drainages that they integrate, preventing a rapid headward migration of knickpoints and therefore thwarting the transmission of the signal of a new base level to the integrated plateau interior. The shallow slopes and limited run-off over the plateau interior limit the production of sediment load necessary to achieve incision. The modest breaching of the former divide restricts the penetration of atmospheric moisture into the plateau interior, limiting the increase in runoff, stream power, and sediment discharge, and therefore the ability of knickpoints along the Ceyhan, Göksu, and Zamanti Rivers (Figs. 12A, 12B, and 13A) to migrate upstream (Fig. 13C-3).

Discriminating avulsion or overflow is more difficult, because both processes can act in concert: The protracted filling of a closed basin, which lifts up the bed of its terminal lake (García-Castellanos et al., 2003), can end up with the overflow of the terminal lake during a wet phase. However, avulsion brings the floor of a closed catchment to the elevation of its lowest rim, while the formation of large and deep lakes can result in the formation of conspicuous perched shorelines. Such markers can help differentiate avulsion from lake overflow. However, the dissection of the terminal depositional surfaces, the continued growth of alluvial fans along their margins, and finally their long-wave-length deformation by differential compaction, erosional rebound, and tectonics have modified the differences in elevation between basin floors and their rims. No shorelines older than the late Pleistocene have been found thus far (Erol, 1970; Karabiyikoğlu et al., 1999). The only major distinctive marker of drainage integration is the deposition of final lacustrine limestones in many of the formerly closed catchments at the time of integration.

Cap Limestones: Prominent Markers of Drainage Integration

The lacustrine limestones that cap the Pliocene series (known as the Kişladağ Limestone) are distinctively less terrigenous than lacustrine carbonates deposited earlier. These well-cemented limestones stand out in the landscape, forming dramatic mesas throughout the Central Anatolian Plateau that systematically cap the entire Pliocene sequences. Because the cap limestones do not share the same age from one basin to the next, it is unlikely that a regional climatic event drove their deposition. Instead, because they were deposited just before the closed catchments integrated, their formation seems genetically related to the opening of the closed catchments. It can be objected that, because cap limestones are resistant to erosion, they could simply represent the first resistant layers encountered by erosion during the removal of the topmost Pliocene layers, and not true terminal sequences. Several lines of evidence, however, indicate that cap limestones are indeed terminal sequences. First, even in areas where the carbonates are protected by the overlying 2.7 Ma Valibaba Tepe–İncesu ignimbrite, little to no sediment is found between the carbonates and the ignimbrite. Instead, the Valibaba Tepe–İncesu ignimbrite seals deep paleosols formed by protracted exposure and weathering of the carbonates. Second, near the front of mountain ranges, the lacustrine limestones interfinger with fanglomerates (e.g., western Sivas Basin; Meijers et al., 2020; Fig. 11, section e-e*, km 8), the surface of which grades downslope directly to the top of the carbonates, and not to higher and now-eroded depositional surfaces.

The distribution of these cap carbonates may have been more widespread than today before the dissection of the Pliocene basins started. However, they are distributed in restricted clusters along the margins of the basins, suggesting either that they
were emplaced in lakes of limited extent, or that they were restricted to marginal areas in shallow lakes (Della Porta, 2015), away from the main detrital influxes, and close to sources of dissolved carbonates. Many of these lacustrine limestones are indeed found in close proximity to active or ancient travertine and/or tufa mounds (Fig. 10). They interfinger with Pliocene travertine or tufa at several locations, in particular along the southern border of the Konya/Tuz Basin, and along the Zamantı River. The travertine and tufa were formed by dissolution of Paleozoic–Mesozoic limestones under the effects of deep metamorphic outgassing (Karaisağlı and Orhan, 2018).

**Cap Limestones: Precursors or Products of Drainage Integration?**

We explored the depositional environment, specifically associated to drainage integration, that may have generated these carbonates. Deposition of these limestones after the deposition of more terrigenous lacustrine sediments requires either a decrease in the terrigenous supply or an increase in the biogenic supply. A decrease in the terrigenous fraction would occur by dilution of terrigenous inputs in lakes that became larger and finally overflowed. In this case, the cap carbonates would be markers of drainage integration by overflow, and they would slightly predate drainage integration. This hypothesis, however, is not supported by the evolution of δ^{18}O and δ^{13}C values in the Pliocene lacustrine carbonates during the millions of years that predate drainage integration. These values should show a systematic drift toward more open hydrological conditions over time, which is not observed (Meijers et al., 2020). In addition, the various cap carbonates that have been studied do not follow similar isotopic trends (Meijers et al., 2020), suggesting that they were deposited in small, separate lakes.

Alternately, the carbonates could have been deposited in restricted lakes, as a result of drainage integration, immediately after the beginning of drainage integration. In this case, the terrigenous fraction would have been funneled along the newly formed valleys that dissected the center of the Pliocene basins, away from these marginal lakes. The coexistence of incision and deposition within a single basin is illustrated today by Lake Seyfe (Fig. 14A). The lake lies in a location sheltered from detrital inputs and from incision, in a corner of the Kozaklı Basin. The floor of the Kozaklı Basin has been incising over at least the past 6.3 ± 0.4 m.y., but the area of Lake Seyfe has been spared. The lake has a small catchment, which provides a small inflow of water and sediment. The lake is currently hydrologically closed (Onalgil et al., 2015; Reis and Yilmaz, 2008) and separated from the incised floor of the Kozaklı Basin by a shallow sill composed of shallowly dipping Miocene–Pliocene sediments. The lake is the successor of a much larger lake in which extensive lacustrine carbonates were deposited. These carbonates are found 30 m below the bed of Lake Seyfe, and they are overlain by 30 m of lacustrine sediments that include gypsiferous clays (Çelik et al., 2008). These sediments indicate that the lake has remained closed, despite large size fluctuations (Reis and Yilmaz, 2008). Restricted surface outflow has prevented incision of the lake outlet and the draining of Lake Seyfe over the past 6 m.y. The Pliocene cap limestones, likewise, tend to occur along the margins of the Miocene–Pliocene depocenters (Fig. 10). The lakes in which they formed were shorter-lived features, probably because their water balance was more positive, as suggested by the isotopic composition of the limestones (Meijers et al., 2018a). This more positive balance promoted more perennial surface outflow, faster outlet incision, and earlier draining of the lakes.

The concentration of these lakes along the margins of the Pliocene depocenters would have resulted from the ponding of small streams draining the surrounding mountains upon their arrival in the depocenters, due to the upwarping of the basin floors by processes described in the following section. The sediments underlying Lake Seyfe and the cap carbonates dip at shallow angles away from the depocenters and toward the surrounding ranges (Fig. 14). Synsedimentary tilting is evidenced by an increase in the dip of the Pliocene strata with depth. This is observed, for example, on both sides of the Sivas Basin anticline (Fig. 7B), and to the north of Lake Seyfe (Fig. 14A). Tilting continued after the draining of the marginal lakes, as evidenced by the tilting of the cap carbonates. The lakes were replaced by streams that flowed along the margins of the depocenters, parallel to the strike of the tilted beds, for some distance before joining the main rivers (e.g., Figs. 14C and 14D).

**Cap Limestones as Markers of Tectonic Changes**

Tilting is a common isostatic flexural response to tectonic denudation in the footwall of normal faults (King and Ellis, 1990). Cap carbonate deposition over the footwall of the Lake Tuz fault (Fig. 14D) occurred when uplift and tilting in the footwall of the Lake Tuz fault isolated the footwall from the main detrital sources of the Lake Tuz basin. A lake formed between the mountain range that borders the basin and the tilted basin floor, in which lacustrine carbonates were deposited. The lake eventually drained and was replaced by streams that flow for some distance along the front of the range, in the back of the tilted basin floor, before veering and crossing the fault scarp (Fig. 14D). Similar isolation by back tilting over the footwall of the Lake Tuz fault may account for the deposition of cap carbonates farther to the SE in western Cappadocia (Fig. 10), as well as over the footwall of the Salanda fault (Fig. 10), and over the tectonic shoulders of the Kayseri releasing bend in eastern Cappadocia (Fig. 10). In all these cases, carbonate deposition was a direct consequence of the development of full-escape tectonics.

By contrast, the centrifugal tilting of the cap limestones in the Sivas Basin (Figs. 14B and 14C) involved shallow elongate doming along the axis of the Sivas Basin (Fig. 7A). Episodic contractional deformation took place in the Sivas Basin from the Eocene to the middle Miocene (Yilmaz and Yilmaz, 2008). It could have continued up to at least the time of dissection of the basin in the early Pliocene, if not until today. Contraction could therefore have contributed to the observed doming. The left-lateral Central Anatolian fault moved in Pliocene time in a central position in the basin (Koçyiğit and Beyhan, 2006). It could have continued up to at least the time of dissection of the basin in the early Pliocene, if not until today. Contraction could therefore have contributed to the observed doming. The left-lateral Central Anatolian fault moved in Pliocene time in a central position in the basin (Koçyiğit and Beyhan, 2006).
1998) during the development of the full-escape tectonic regime. Across the Sivas Basin, the Central Anatolian fault zone consists of discontinuous right-lateral strands (Akyuz et al., 2013) subjected to north-south compression (Kaymakci et al., 2010). The strike-slip segments almost align with the axis of the dome (Fig. 7A), suggesting that contraction across the Central Anatolian fault zone could contribute to the doming. Alternatively, doming could represent an isostatic flexural response to the erosion of the Sivas sedimentary basin since drainage integration, which has generated 150–300 m of incision along the Kızılırmak River (Fig. 7A). The formation of lakes by flexural isostatic rebound at the periphery of captured catchments has also been proposed in eastern Tibet, to explain the deposition of lacustrine sequences during the incision of the depocenters (Zhang et al., 2014).

The tilting produced by flexural isostatic rebound was calculated for various elastic thicknesses and compared to the tilting of the cap carbonates (Fig. 7B). The modeled flexural rebound is able to reproduce the 0.15% ± 0.07% surface dip of the carbonates in the northern Sivas Basin, and the 1.0% ± 0.2% surface dip of the carbonates in the Kangal Basin at elastic thicknesses of 15–20 km. This elastic thickness is comparable to some values reported for the Altiplano and Puna orogenic plateaus (Pérez-Gussinyé et al., 2008), although values as low as 10 km for these plateaus have also been proposed (McNab et al., 2018). Flexural rebound could therefore fully account for the doming of the Kangal Basin, under certain values of flexural rigidity. Because some of the tilting predated incision, it is likely that at least some of the tilting also resulted from contraction. The contribution of each process is uncertain but possibly marked by a shift from contraction-driven to unloading-driven upwarping during the establishment of full-escape tectonics.

In the Kozaklı Basin, Lake Seyfe lies south of a deeply dissected shallow anticline that inverts an Oligocene depocenter (Figs. 10 and 14A). The tilting of the Pliocene beds, below and north of Lake Seyfe, may likewise have resulted from gentle Miocene–Pliocene contraction, with an added contribution of flexural rebound, following the excavation of Oligocene sediments. At a larger scale, the cap

Figure 14. Drainage of some lacustrine carbonates located in the areas of Lake Seyfe (A), eastern Sivas Basin (B), central Sivas Basin (C), and Tuz Lake basin (D). Locations are shown on Figure 2B. Cap carbonate scarp retreat (gray arrows) occurs by landsliding (C) and headward stream erosion (D). Bedding dip in ‰.
carbonates of Kozaklı can be viewed as having been produced by the far-field doming of the Sivas Basin combined with a far-field flexural response to the deep dissection of the Oligocene sediments along the Delice River (Fig. 2A).

The cap carbonates are associated with drainage integration across the Central Anatolian Plateau. They reflect either a dilution of detrital input in deepening lakes or the channeling of the largest detrital sources along rivers partially draining depocenters. In this latter case, they would result from the ponding of water along the margins of the depocenters, the centers of which would be upwarped and already eroding.

**DISCUSSION 2: REGIONAL DRIVERS OF DRAINAGE INTEGRATION**

In the absence of significant changes in the regional climate during the rise of the Central Anatolian Plateau, we review how the two most conspicuous tectonic events of the Pliocene could have affected the drainage of the Central Anatolian Plateau, namely, the rise of the Central Anatolian Plateau from 10 to 5 Ma, and the full tectonic escape of Anatolia after 5 Ma. We also address the contributions of other phenomena that could have affected drainage evolution on the Central Anatolian Plateau, namely, volcanism and karstification.

**Effect of the Rise of the Central Anatolian Plateau on Drainage Integration**

Uplift of the plateau interior decreased surface air temperatures and evapotranspiration, while the rise of the plateau margins decreased precipitation over the plateau interior. The rise of an arid but cooling plateau interior leads to a decrease in evapotranspiration. If such decrease is not compensated by a decrease in precipitation, driven by a commensurate uplift of the plateau margins, and therefore by an intensification of the rain shadows that they cast over the plateau interior, then the uplift of the plateau interior may result in a more positive water balance, which promotes overflow and avulsion. The fact that Pliocene integration affected the eastern part of the plateau, which have risen to higher elevations, rather than the western part (Fig. 1), where all closed catchments are located (Fig. 2A), supports this possibility. Top-down integration could have initiated in the east, in some of the most elevated catchments, and progressed westward, finally reaching the outer slopes (Fig. 15B). The perpetuation of closed conditions in the Konya catchment would then have been a matter of chance, the catchment being simply located off the trajectory of top-down integrating drainages.
Effect of Escape Tectonics on Drainage Integration

Are there genetic links explaining the temporal coincidence of drainage integration and escape tectonics? The incision of the Pliocene depocenters led to the removal of rock volumes from the plateau surface. These volumes were too modest to modify body forces enough such as to trigger the transition from proto-escape to full escape. On the other hand, the extensive deposition of sediments across the Central Anatolian Plateau may have concealed fault growth before drainage integration. The rise of topographic scarps would then have resulted from a decrease in sedimentation rates, as a result of drainage integration. Alternatively, the inception, or acceleration, of fault slip during the full escape could have generated topographic scarps that altered water and sediment routing, spurring drainage integration. Both processes are not mutually independent, because the rise of a fault scarp at one place alters the routing of sediments to other fault scarps.

The Lake Tuz fault is the only fault for which enough data have been collected to evaluate the ratio of fault slip to sedimentation rates. At the turn of the Pliocene, the Lake Tuz basin could have received terrigenous contributions from newly opened contiguous catchments (see “Disruption of the Central Anatolian Plateau Surface by Tectonic Faults during the Early Pliocene” section) that are currently collected by the Kızılırmak River. The Lake Tuz fault scarp started to grow ca. 3.6 ± 0.3 Ma (Table S4-1) and is currently 400–500 m tall at its highest point (Krystopowicz et al., 2020; Özasyn et al., 2013; Yıldırım, 2014). The total vertical separation across the fault is larger as a result of sedimentation over the hanging wall, and scarp growth may slightly predate the 3.6 ± 0.3 Ma age of deposition of cap limestones over the footwall (which appears to have been a consequence of the formation of the scarp; see “Cap Limestones as Markers of Tectonic Changes” section). This allows us to calculate a lower-estimate fault scarp growth rate of 140 ± 10 m/m.y., which is significantly larger than the sedimentation rates of 80 m/m.y. over the hanging wall from 7 Ma to 5 Ma (Meijers et al., 2018a). Scarp growth therefore was most likely triggered by an increase in fault slip, rather than by a decrease in sedimentation rate.

The rise of the Lake Tuz fault, and of other faults farther east (see “Disruption of the Central Anatolian Plateau Surface by Tectonic Faults during the Early Pliocene” section), impeded east-west fluxes between open catchments and still-closed catchments. They deflected and funneled these fluxes toward the northwest, away from central Cappadocia and the Lake Tuz Basin. They could have therefore reduced the delivery of terrigenous sediments to Cappadocia, allowing the deposition of cap limestones in Cappadocia from 4.6 Ma to 4.2 Ma.

In the meantime, the rise of the scarps partitioned vast, unconfined areas of sedimentation (Fig. 15A) into smaller, fault-bounded subbasins. This led to more contrasted water and sediment fluxes in each subbasin compared to earlier, larger depocenters. Subcatchments with more negative water balances (e.g., Lake Tuz, Konya closed catchment) remained closed and became drier, hosting evaporites (Fig. 15C). In contrast, less evaporative catchments (e.g., Sivas) would have overflowed and integrated the Kızılırmak River drainage.

Influence of Volcanism on Drainage Integration

Volcano emplacement is responsible for the formation of large lakes on the eastern Anatolian and Iranian plateaus (Lake Van next to the Nemrut volcano; Lake Sevan [Armenia] next to the Ghegam Ridge). On the Central Anatolian Plateau, it contributed to the formation of Lake Sugla, next to the Eremli volcano (Figs. 2A and 10), and Lake Yay, next to the Erciyes volcano (Figs. 2A and 10). However, drainage integration does not coincide with a peak in volcanic activity, and no large volcanic obstructions formed at the time of drainage integration. Preexisting volcanic obstructions, however, locally had a decisive influence on the shaping of the modern drainage, and on the distribution of open and closed catchments. The Develi volcano (Figs. 10 and 12C) grew during the late Miocene in a bottleneck between the eastern Pliocene depocenters (Zamantı, Kangal, Sivas) and the western depocenters (Lake Tuz, Cappadocia, Kızılırmak). The Pliocene Zamanti River established its course between the Sivas Basin anticline and the mountains bordering the basin in the south (Tahtalı Mountains). The Develi volcano prevented westward integration of the river toward the still-closed catchment of Konya and forced the Zamanti River to exit the plateau and flow down the plateau outer slope (Fig. 11, section c-c’; Figs. 12A, 12C, and 15D).

A major change in volcanic activity happened, however, at the time of drainage integration. From 10 Ma to 5 Ma (Le Pennec et al., 2005; Piper et al., 2013), extensive ignimbrites were produced in Cappadocia (Fig. 10), with some of the individual ignimbrites spreading over areas as large as 180 x 120 km (Le Pennec et al., 1994). This ignimbritic flare-up ceased suddenly at 5 Ma (Aydar et al., 2012). Our dating shows that this shift coincided with the inception of the Kayseri releasing bend, and it may have been contemporary with the initiation of the Lake Tuz fault scarp south of Cappadocia. These latter areas have since been privileged sites for the formation of stratovolcanoes. The development of full-escape tectonics initiated a major change in the type of volcanic activity. The ignimbrite sheets deposited during the millions of years preceding drainage integration were quickly buried in the Pliocene depocenters (Kadir et al., 2013) and did not obstruct surface flow and sediment fluxes. However, owing to their substantial extent, they influenced the geochemistry of lacustrine carbonates (Ersel et al., 2014) and modified the permeability of the ground surface. Cappadocian ignimbrites are poorly welded, highly porous, and permeable, and they probably promoted groundwater flow through the depocenters. Conversely, the strongly welded Valibaba Tepe–İnceuşu ignimbrite prevented infiltration, increased surface runoff, and promoted overflows and avulsion. Our field analysis shows that the Valibaba Tepe–İnceüşu ignimbrite was emplaced over an already incised landscape (see “Formation of the Kayseri Releasing Bend: Age of Inception of Escape Tectonics” section), precluding the possibility that it influenced drainage integration.
Influence of Karstification on Drainage Integration

The increase of the topographic gradient between the Central Anatolian Plateau interior and the Mediterranean Sea led to the establishment of subterranean pathways across the permeable Taurus Mountains (see “Present-Day Topography of the Central Anatolian Plateau” section). Karstification is instrumental in the development and persistence of closed conditions in many tectonic-karstic depressions located west of the Central Anatolian Plateau (Kuzucuoğlu et al., 2019). By promoting water leakage away from the Central Anatolian Plateau interior, it may have delayed drainage integration on the eastern part of the Central Anatolian Plateau, and it may have contributed to the persistence of closed conditions on its western part.

Karstification does not systematically inhibit drainage integration. It can also accelerate drainage integration (Authemayou et al., 2018), most notably by accelerating the breaching of topographic divides by groundwater-assisted capture (Brocard et al., 2011, 2012). Such a process was involved in the final stage of breaching of the plateau marginal divide by the Zamanti River (Fig. 12A). There, to connect the Zamanti Basin on the Central Anatolian Plateau interior to the middle Miocene Dikme Basin on the upper outer slope (Fig. 12C), the Zamanti River did not follow the shortest route along the base of the Tahtalı Mountains (Fig. 12C), but instead incised a 10-km-long gorge into limestones. The gorge crosses the local surface geomorphologic grain produced by flat-floored karstic valleys (poljes; Fig. 11, section d-d’). Avulsion was not involved in the process, because the elevation of the floor of the Zamanti Basin remains 150–300 m below the dried valleys crosscut by the Zamanti River gorge. A terminal lake, ponded against the Develi volcano, would have followed preexisting topographic lows, but the gorge is randomly incised across the local topographic grain (Fig. 12C). From one end to the other, the gorge is cut into limestones that are occasionally concealed beneath volcanic rocks of the Develi volcano (Fig. 11, section c-c’). The formation of the gorge is therefore best explained by the collapse of karstic conduits that once connected the Zamanti and Dikme Basins. Drainage integration at the scale of the entire Zamanti River was dominated by top-down processes of avulsion or overflow, but its final diversion over the plateau margin (Fig. 15E) involved confinement behind the Develi volcano and groundwater-assisted capture.

CONCLUSIONS

Significant relief reduction during the middle to late Miocene and drainage disintegration during the late Miocene led to widespread deposition of continental sediments across the rising Central Anatolian Plateau.

Drainage reintegration occurred during the early Pliocene. It was rapid (<1.5 m.y.) and led to the reconnection of the interior drainage to the drainage of the southern outer slope, establishing the present-day drainage. Since then, long-term sediment storage over the plateau interior has been confined to a few remaining closed catchments.

Patterns of incision across the plateau surface (obtained using 10Be and new 40Ar-39Ar ages) reveal that the rivers that drain the outer slope have not penetrated far into the plateau interior.

Most of the drainage integration was driven by top-down processes of lake overflow and/or overfilling and avulsion of closed basins, rather by bottom-up capture.

Drainage integration did not trigger dramatic waves of accelerated river incision, owing to the limited amount of water and sediment contributed by the plateau interior.

Drainage integration entrained the deposition of distinctive lacustrine limestones. Rather than being deposited in vast, overflowing lakes before drainage integration, these limestones were deposited during drainage integration along the fringes of Pliocene depocenters, due to the upwarping of the basin floors. This upwarping was produced, in part or in whole, by isostatic rebound driven by tectonic denudation in the footwall of normal faults, and by erosional unloading within the most deeply incised basins.

The rise of the Central Anatolian Plateau and the development of the full tectonic escape regime triggered the integration of the Central Anatolian Plateau drainage. The higher-elevation eastern Central Anatolian Plateau hosts most of the catchments that opened during this phase of integration. Higher elevations favored overflow and avulsion. Escape tectonics, on the other hand, partitioned formerly expansive and unconfined depocenters into smaller, fault-bounded basins, sharpening the contrasts in water balance between subbasins. This intensified evaporative trends in some catchments while promoting overflow and avulsion in others.

Unlike most orogenic plateaus, where widespread volcanic derangement of drainage networks occurs, volcanism on the Central Anatolian Plateau did not produce enough topographic obstructions at the time of the integration to contribute much to drainage integration. Karstification inhibited the opening of some closed catchments but assisted the breaching of some watersheds during drainage integration.

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REFERENCES CITED

Abgarmi, B., Delph, J.R., Ozçar, A.A., Beck, S.L., Zandt, G., Sandvol, E., Turkelli, N., and Biryol, C.B., 2017, Structure of the crust and African slab beneath the Central Anatolian Plateau from receiver functions: New insights on isostatic compensation and slab dynamics: Geosphere, v. 13, no. 6, p. 1774-1787, https://doi.org/10.1130/GES01509.1.

Akgün, F., Kayseri, M.S., and Akkiraz, M.S., 2007, Palaeoclimatic evolution and vegetational changes during the late Oligocene-Miocene period in Western and Central Anatolia (Turkey): Palaeogeography, Palaeoclimatology, Palaeoecology, v. 253, no. 1–2, p. 56-90, https://doi.org/10.1016/j.paleo.2007.03.034.

Aksu, A., Calon, T., Hall, J., Mansfield, S., and Yapar, D., 2005, The Cilicia–Adana basin complex, Eastern Mediterranean: Neogene evolution of an active fore–arc basin in an obliquely convergent margin: Marine Geology, v. 221, no. 1–4, p. 121–159, https://doi.org/10.1016/j.margeo.2005.03.011.

Downloaded from http://pubs.geoscienceworld.org/gsa/geosphere/article-pdf/doi/10.1130/GES02247.1/5271392/ges02247.pdf by guest
Earth and Planetary Science Letters, v. 257, no. 3, p. 372–380, https://doi.org/10.1016/j.epsl.2007.02.039.

García-Flores, D., Gómez-Mondragón, D., Ocampo, J., and Escobar, J., 2004, Integration of the Central Anatolian Plateau drainage, Turkey, with paleontologic, radiochronologic, geochemical and paleomagnetic data: Journal of Volcanology and Geothermal Research, v. 141, no. 1, p. 45–64, https://doi.org/10.1016/j.jvolgeores.2004.09.004.

Lepetit, P., Virecque, L., Piperno, J.D., Sudo, M., Gürel, A., Çopuroğlu, I., and Starostenko, V., eds., Sedimentary Basin Tectonics from Cenozoic to Recent: Implications for deformation in the interior of the Central Anatolian Plateau: Geosphere, v. 16, n. 4, p. 1107–1124, https://doi.org/10.1130/GES01275.1.

Kürçer, A., and Gökten, Y.E., 2012, Paleoseismological Three Dimensional Virtual Photography Method: A Case Study: Baglakayas–2010 Trench, Tuz Gölü Fault Zone, Central Anatolia, Turkey, in Sharkov, E., ed., Tectonics—Recent Advances: London, UK, IntechOpen, p. 201–228, http://dx.doi.org/10.5727/48194.

Kuzcuoğlu, C., Cinta, J., Black, S., Denelle, F., Fontugne, M., Karabiyikoglu, M., Kuzcuoglu, C., Mastani, K., Limodin-Loouet, N., Moulard, D., andOrth, P., 1999, Reconstruction of climatic changes during the late Pleistocene, based on sediment records from the Konya Basin [Central Anatolia, Turkey]: Geophysical Research Letters, v. 34, no. 1, p. 175–198, https://doi.org/10.1029/2008GL034191.026.

Kuzcuoğlu, C., Çiner, A., and Kazanc, N., 2019, The geomorphological regions of Turkey, in Kuzcuoglu C., Ciner A., and Kazanc N., eds., Landscapes and Landforms of Turkey: World Geomorphological Landscapes: Cham, Switzerland, Springer, https://doi.org/10.1007/978-3-030-03515-0_4.

Lepetit, P., Viereck, L., Piper, J.D., Sudo, M., Gürel, A., Çopuroğlu, I., Bertaux, J., Black, S., Denelle, F., Fontugne, M., Karabiyikoglu, M., Kuzcuoglu, C., and Starostenko, V., eds., Sedimentary Basin Tectonics from Cenozoic to Recent: Implications for deformation in the interior of the Central Anatolian Plateau: Geosphere, v. 16, n. 4, p. 1107–1124, https://doi.org/10.1130/GES01275.1.

Lepetit, P., Virecque, L., Piperno, J.D., Sudo, M., Gürel, A., Çopuroğlu, I., Gruber, M., Mayer, B., Koch, M., and Tataz, O., 2014, Ar/Ar dating of ignimbrites and Plinian ash-fall layers from Cappadocia, central Turkey: Reconciling field constraints with paleontologic, radiochronologic, geochemical and paleomagnetic data: Journal of Volcanology and Geothermal Research, v. 141, no. 1, p. 45–64, https://doi.org/10.1016/j.jvolgeeres.2004.09.004.

Lepetit, P., Virecque, L., Piperno, J.D., Sudo, M., Gürel, A., Çopuroğlu, I., Gruber, M., Mayer, B., Koch, M., and Tataz, O., 2014, Ar/Ar dating of ignimbrites and Plinian ash-fall layers from Cappadocia, central Turkey: Implications for chronostatigraphic and Eastern Mediterranean paleoenvironmental record: Chemie der Erde (Geochemistry), v. 74, no. 3, p. 471–488, https://doi.org/10.1130/GES01432349.

Kroh, A., and Garzione, C.N., 2008, Paleosurfaces, paleoelevation, and the mechanisms for the late Miocene topographic relief: A record of terminal eastward subduction in southern Turkey: Geological Research, v. 63, no. 1, p. 59–87, https://doi.org/10.1023/A:1008024127346.

Lenné, J.-L., Bourdier, J.-L., and Temel, A., 2004, Neogene ignimbrites of the Nevşehir Plateau (central Turkey): Stratigraphy, distribution and source constraints: Journal of Volcanology and Geothermal Research, v. 63, no. 1, p. 59–87, https://doi.org/10.1016/S0377-0273(03)00019-3.

Kroh, A., and Garzione, C.N., 2008, Paleosurfaces, paleoelevation, and the mechanisms for the late Miocene topographic relief: A record of terminal eastward subduction in southern Turkey: Geological Research, v. 63, no. 1, p. 59–87, https://doi.org/10.1023/A:1008024127346.

Lenné, J.-L., Temel, A., and Bourdier, J.-L., 2005, Stratigraphy and age of the Cappadocia ignimbrites, Turkey: Reconciling field constraints with paleontologic, radiocarbon, geochemical and paleomagnetic data: Journal of Volcanology and Geothermal Research, v. 141, no. 1, p. 45–64, https://doi.org/10.1016/j.jvolgeeres.2004.09.004.
International Journal of Earth Sciences, v. 93, no. 2, p. 314–328, https://doi.org/10.1007/s00531-004-0390-v.

Örgümen, N., Giudici, P., Baran, P., Faranda, C., Karanika, K., Giozi, E., Radef, G., and Cosenzino, D., 2018, Evidence for 1.5 km of uplift of the Central Anatolian Plateau’s southern margin in the last 450 kyr and implications for its multi-phased uplift history: Tectonics, v. 37, no. 1, p. 359–390, https://doi.org/10.1002/2017TC004805..

Onal, M., Helvaci, C., and Ceyhan, F., 2004, Geology and tectonic potential of the middle Miocene Görün (Sivas) basin, Central Anatolia: Geosphere, v. 10, no. 2, p. 118, https://doi.org/10.1130/GF017B475.

Onaligil, N., Kadir, S., Kılıç, T., Eren, M., and Gürel, A., 2015, Mineralogy, geochemistry and genesis of the modern sediments and sediments from the vicinity of Mount Erciyes, Cappadocia, Turkey: Journal of African Earth Sciences, p. 102, https://doi.org/10.1016/j.jafrearsci.2014.10.020.

Ozsoy, E., Çiner, A., Rojas, B., Dirik, K., Melnick, D., Fernandez Blanxart, D., Bottini, G., Schildgen, T., García, Y. and Strecke, M.R., 2013, Plio-Quaternary extensional tectonics of the Central Anatolian Plateau: A case study from the Tuz Gölü Basin, Turkey: Turkish Journal of Earth Sciences, v. 22, no. 5, p. 691–714.

Powes-Guisinger, M., Lowry, A., Phipps Morgan, J., and Tassara, A., 2008, Effective elastic thickness variations along the Andean margin and their relationship to subduction geometry: Geophysics Geosciences Geosystems, v. 9, no. 2, p. 249–267, https://doi.org/10.1016/j.g猴子.2007.07.005.

Phillips, L.F., and Schumm, S., 1987, Effect of regional slope on drainage networks: Geology, v. 15, no. 9, p. 813–816, https://doi.org/10.1130/0091-7613(1987)015<0813:ESORS>2.0.CO;2.

Piper, J.A.D., 1974, The Chemical Weathering of Rocks and Continents: John Wiley & Sons, New York, 419 p.

Pirie, W.C., Ritz, C., Guglielmo, A., Kappen, J., Litt, F., Roberts, A., and Alkner, Z., 2013, Paleolandscape of the Cappadocian volcanic succession, central Turkey: Major ignimbrite emplacement during two short (Miocene) tectonothermal episodes and Neogene tectonics of the Anatolian collage: Journal of Volcanology and Geothermal Research, v. 262, p. 47–67, https://doi.org/10.1016/j.jvolgeores.2013.06.008.

Poison, A., Vrielynck, B., Wernli, R., Negri, A., Bassetti, M.-A., Büyükmücev, M., Özer, S., Guillou, H., Kavak, S., and Temiz, H., 2016, Miocene transgression in the central and eastern parts of the Sivas Basin (Central Anatolia, Turkey) and the Cenozoic palaeogeographical evolution: International Journal of Earth Sciences, 105, no. 1, p. 339–368, https://doi.org/10.1007/s00531-015-1248-1.

Portner, D.E., Delph, R., Biryol, C.B., Beck, S.L., Zandt, G., Ozaar, A.A., Sandur, E., and Turrell, N., 2018, Subduction termination through progressive slab slab displacement across Eastern Mediterranean subduction zones from updated P-wave tomography beneath Anatolia: Geosphere, v. 14, no. 3, p. 907–925, https://doi.org/10.1130/GES01617.1.

Radwany, M., Whitney, D., Brocard, G., Umhoefer, P.J., and Teyssier, C., 2017, Ophiolite gabbro from source to sink: A record of tectonic and surface processes in Central Anatolia: Geosphere, v. 13, no. 5, p. 1328–1358, https://doi.org/10.1130/GES01469.1.

Reilinger, R., McClusky, S., Paradissis, D., Ergintav, S., and Ver- rant, P., 2016, Geodetic constraints on the tectonic evolution of the Aegean region and strain accumulation along the Hellenic subduction zone: Tectonophysics, v. 488, no. 1, p. 22–30, https://doi.org/10.1016/j.tecto.2008.05.027.

Reis, S., and Yilmaz, H.M., 2008, Temporal monitoring of water level changes in Seyfe Lake using remote sensing: Hydrological Processes: International Journal (Toronto, Ontario), v. 22, no. 22, p. 4448–4454.

Robertson, A., Unluenguc, Ü.C., Inan, T., and Tast, K., 2004, The Misis-Andırın Complex: A Mid-Tertiary mélangé related to the late-stage subduction of the southern Nectethys in S Turkey: Journal of Asian Earth Sciences, v. 22, no. 5, p. 413–453, https://doi.org/10.1016/S0921-1577(02)00062-2.

Senkaya, M.A., Yıldırım, C., and Çiner, A., 2015, Late Quaternary alluvial fans of Emi Valley in the Ecmig fault zone, south central Turkey: Insights from cosmogenic nuclides: Geomorphology, v. 226, p. 512–525, https://doi.org/10.1016/j.geomorph.2014.10.008.

Schenneng, R., Mikes, T., Rojaj, B., and Mulch, A., 2013, The impact of topography on isotopes in precipitation across the Central Anatolian Plateau (Turkey): American Journal of Science, v. 313, no. 2, p. 61–80, https://doi.org/10.2475/12.2013.35.

Schildgen, T., Cosenzino, D., Bookhagen, B., Niedermann, S., Yıldırım, C., Echter, H., Wittmann, H., and Strecke, M.R., 2012, Multi-phased uplift of the southern margin of the Central Anatolian Plateau, Turkey: A record of tectonic and upper mantle processes: Earth and Planetary Science Letters, v. 317, p. 85–95, https://doi.org/10.1016/j.epsl.2011.12.003.

Schleifarth, W., Darin, M., Reid, M., and Umhoefer, F.J., 2018, Dynamics of episodic Late Cretaceous-Cenozoic magmatism across Central Anatolia: New insights from an extensive geochronology compilation: Geosphere, v. 14, no. 5, p. 1990–2008, https://doi.org/10.1130/GES01647.1.

Seebauer, L., and Gornitz, V., 1983, River profiles along the Himalayan arc as indicators of active tectonics: Tectonophysics, v. 92, no. 4, p. 335–367, https://doi.org/10.1016/0040-1951(83)90201-9.

Şen, E., Küçükgöz, B., Aydar, E., Gougaraud, A., and Vincent, P.M., 2003, Volcanological evolution of Mount Erciyes strato-volcano and origin of the Valibaba Tepe ignimbrite (Central Anatolia, Turkey): Journal of Volcanology and Geothermal Research, v. 125, no. 3–4, p. 225–246, https://doi.org/10.1016/S0377-0273(03)00110-0.

Şengör, A.M.C., Özeren, S., Genç, T., and Zor, E., 2003, East Anatolian high plateau as a mantle-supported, north-south shortened domal structure: Geophysical Research Letters, v. 30, no. 24, 8946, https://doi.org/10.1029/2002GL017958.

Şengör, A.M.C., Özeren, M.S., Keskin, M., Sakınç, M., Özbakır, M.A., and Sobel, E.R., 2009, Does the topographic distribution of the central Andean Puna Plateau result from climatic or geodynamic processes? Geology, v. 37, no. 7, p. 643–646, https://doi.org/10.1130/G2654A.1.
Strömberg, C.A., Werdelin, L., Friis, E.M., and Saraç, G., 2007, The spread of grass-dominated habitats in Turkey and surrounding areas during the Cenozoic: Phytolith evidence: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 250, no. 1–4, p. 18–49, https://doi.org/10.1016/j.palaeo.2007.02.012.

Toprak, V., 1998, Vent distribution and its relation to regional tectonics, Cappadocian volcanics, Turkey: Journal of Volcanology and Geothermal Research, v. 85, no. 1–4, p. 55–67, https://doi.org/10.1016/S0377-0273(98)00049-3.

Ünlüenç, U.C., and Akıncı, A.C., 2015, Sedimentary development of the Oligocene Karsante Basin, southern Turkey, in its regional tectonic setting: Journal of Asian Earth Sciences, v. 105, p. 173–191, https://doi.org/10.1016/j.jseaes.2015.03.034.

van der Beek, P., Champel, B., and Mugnier, J.L., 2002, Control of detachment dip on drainage development in regions of active fault-propagation folding: Geology, v. 30, no. 5, p. 471–474, https://doi.org/10.1130/0091-7613(2002)030<0471:CODDDO>2.0.CO;2.

Van Wees, J., and Cloetingh, S., 1994, A finite-difference technique to incorporate spatial variations in rigidity and planar faults into 3-D models for lithospheric flexure: Geophysical Journal International, v. 117, no. 1, p. 179–196, https://doi.org/10.1111/j.1365-246X.1994.tb03311.x.

Walsh-Kennedy, S., Aksu, A., Hall, J., Hiscott, R., Yalçırak, C., and Cilti, G., 2014, Source to sink: The development of the latest Messinian to Pliocene-Quaternary Cilicia and Adana Basins and their linkages with the inland Mut Basin, Eastern Mediterranean: Tectonophysics, v. 622, p. 1–21, https://doi.org/10.1016/j.tecto.2014.01.019.

Whipple, K., and Gasparini, N., 2014, Tectonic control of topography, rainfall patterns, and erosion during rapid post-12 Ma uplift of the Bolivian Andes: Lithosphere, v. 6, no. 4, p. 251–268, https://doi.org/10.1130/L325.1.

Wickert, A., 2016, Open-source modular solutions for flexural isostasy: gFlex v1.0: Geoscientific Model Development, v. 9, no. 3, p. 997–1017, https://doi.org/10.5194/gmd-9-997-2016.

Williams, G., Ünlüenç, U.C., Kelling, G., and Demirkol, C., 1995, Tectonic controls on stratigraphic evolution of the Adana Basin, Turkey: Journal of the Geological Society [London], v. 152, no. 5, p. 873–882, https://doi.org/10.1144/gsjgs.152.5.0873.

Yavuz, N., Culha, G., Demir, Ş.S., Utescher, T., and Aydin, A., 2017, Pollen, ostracod and stable isotope records of palaeoenvironment and climate: Upper Miocene and Pliocene of the Çankırı Basin (Central Anatolia, Turkey): Palaeogeography, Palaeoclimatology, Palaeoecology, v. 467, p. 149–165, https://doi.org/10.1016/j.palaeo.2016.04.023.

Yıldırım, C., 2014, Relative tectonic activity assessment of the Tuz Gölü fault zone, Central Anatolia, Turkey: Tectonophysics, v. 630, p. 183–192, https://doi.org/10.1016/j.tecto.2014.05.023.

Yıldırım, C., Sildigen, T.F., Echter, H., Melnick, D., and Streeker, M.R., 2011, Late Neogene and active orogenic uplift in the Central Pontides associated with the North Anatolian fault: Implications for the northern margin of the Central Anatolian Plateau, Turkey: Tectonics, v. 30, no. 5, TC5005, https://doi.org/10.1029/2010TC002756.

Yildiz, A.L., Toker, V., Demircan, H., and Sevim, S., 2003, Paleoenvironmental interpretation and findings of Pliocene-Pleistocene nano-plankton, planktic foraminifera, trace fossil in the Mut Basin [Mut Havzası Pliyosen-Pleyistosen nano-plankton, planlık foraminifera, iz fosil bulguları ve paleoortam yorumu]: Yerbilimleri, v. 28, p. 123–144.

Yılmaz, A., and Yılmaz, H., 2006, Characteristic features and structural evolution of a post collisional basin: The Sivas Basin, Central Anatolia, Turkey: Journal of Asian Earth Sciences, v. 27, no. 2, p. 164–176, https://doi.org/10.1016/j.jseaes.2005.02.006.

Yılmaz, H., Over, S., and Ozden, S., 2006, Kinematics of the East Anatolian fault zone between Turkgöllü (Kahramanmaras) and Celikhan (Adıyaman), eastern Turkey: Earth, Planets, and Space, v. 58, no. 11, p. 1463–1473, https://doi.org/10.1186/BF03352645.

Yusufoglu, H., 2013, An intramontane pull-apart basin in tectonic escape deformation: Elbistan Basin, Eastern Taurides, Turkey: Journal of Geodynamics, v. 65, p. 308–329, https://doi.org/10.1016/j.jog.2012.05.012.

Zachos, J., Pagani, M., Sloan, L., Thomas, E., and Billups, K., 2001, Trends, rhythms, and aberrations in global climate 65 Ma to present: Science, v. 292, no. 5517, p. 686–693, https://doi.org/10.1126/science.1059412.

Zeilinger, G., and Schlunegger, F., 2007, Possible flexural accommodation on the eastern edge of the Altiplano in relation to focused erosion in the Rio La Paz drainage system: Terra Nova, v. 19, no. 5, p. 373–380, https://doi.org/10.1111/j.1365-3121.2007.00762.x.

Zhang, H., Zhang, P., Champagnac, J.-O., Mohor, P., Anderson, R.S., Kirby, E., Craddock, W.H., and Liu, S., 2014, Pleistocene drainage reorganization driven by the isostatic response to deep incision into the northeastern Tibetan Plateau: Geology, v. 42, no. 4, p. 303–306, https://doi.org/10.1130/G35115.1.