Extensional Tectonics in Western Anatolia, Turkey: Eastward continuation of the Aegean Extension

Elizabeth Catlos¹, Thomas Etzel², and Ibrahim Çemen³

¹The University of Texas at Austin
²ExxonMobil, 22777 Springwoods Village, Parkway Spring, TX 77389, USA
³University of Alabama, Dept. of Geological Sciences, Tuscaloosa, Alabama, 35487-0338, USA

November 22, 2022

Abstract

Western Anatolia is located at the boundary between the Aegean and Anatolian microplates. It is considered a type-location for marking a significant transition between compressional and extensional tectonics across the Alpine-Himalayan chain. The onset of lateral extrusion in Western Anatolia and the Aegean during the Eocene is only one of its transitional episodes. The region has a geological history marked by diverse tectonic events starting from the Paleoproterozoic through the Cambrian, Devonian, and Late Cretaceous, as recorded by its suture zones, metamorphic history, and intrusions of igneous assemblages. Extension in Western Anatolia initiated in a complex lithospheric tectonic collage of multiple sutured crustal fragments from ancient orogens. This history can be traced to the Aegean microplate, and today both regions are transitioning or have transitioned to a stress regime dominated by strike-slip tectonics. The control for extension in Western Anatolia is widely accepted as the rollback of the African (Nubian) slab along the Hellenic arc, and several outstanding questions remain regarding subduction dynamics. These include the timing and geometry of the Hellenic arc and its connections to other subduction systems along strike. Slab tear is proposed for many regions across the Anatolian and Aegean microplates, either trench-parallel or perpendicular, and varies in scale from regional to local. The role of magma in driving and facilitating extension in Western Anatolia and where and why switches in stress regimes occurred along the Anatolia and Aegean microplates are still under consideration. The correlation between Aegean and Anatolian tectonic events requires a better understanding of the detailed metamorphic history recorded in Western Anatolia rocks, possible now with advances in garnet-based themobarometric approaches. Slab tear and ultimate delamination impact lithospheric dynamics, including generating economic and energy deposits, facilitating lithospheric thinning, and influencing the onset of transfer zones that accommodate deformation and provide conduits for magmatism.
Extensional Tectonics in Western Anatolia, Turkey: Eastward continuation of the Aegean Extension

E. J. Catlos¹, T. M. Etzel¹,², and I. Çemen³

¹The University of Texas at Austin, Jackson School of Geosciences, Dept. of Geological Sciences, Austin TX 78712-1722, USA
²ExxonMobil, 22777 Springwoods Village, Parkway Spring, TX 77389, USA
³University of Alabama, Dept. of Geological Sciences, Tuscaloosa, Alabama, 35487-0338, USA

Corresponding author: Elizabeth Catlos (ejcatlos@jsg.utexas.edu)

Index terms

8110 Continental tectonics: general (0905)
8109 Continental tectonics: extensional (0905)
3618 Magma chamber processes (1036)
3651 Thermobarometry
3652 Pressure-temperature-time paths

Keywords

Aegean, Anatolia, extension, Western Turkey, tectonics

Abstract

Western Anatolia is located at the boundary between the Aegean and Anatolian microplates. It is considered a type-location for marking a significant transition between compressional and extensional tectonics across the Alpine-Himalayan chain. The onset of lateral extrusion in Western Anatolia and the Aegean during the Eocene is only one of its transitional episodes. The region has a geological history marked by diverse tectonic events starting from the Paleoproterozoic through the Cambrian, Devonian, and Late Cretaceous, as recorded by its suture zones, metamorphic history, and intrusions of igneous assemblages. Extension in Western Anatolia initiated in a complex lithospheric tectonic collage of multiple sutured crustal fragments from ancient orogens. This history can be traced to the Aegean microplate, and today both regions are transitioning or have transitioned to a stress regime dominated by strike-slip tectonics. The control for extension in Western Anatolia is widely accepted as the rollback of the African (Nubian) slab along the Hellenic arc, and several outstanding questions remain regarding subduction dynamics. These include the timing and geometry of the Hellenic arc and its connections to other subduction systems along strike. Slab tear is proposed for many regions across the Anatolian and Aegean microplates, either trench-parallel or perpendicular, and varies in scale from regional to local. The role of magma in driving and facilitating extension in Western Anatolia and where and why switches in stress regimes occurred along the Anatolia and Aegean microplates are still under consideration. The correlation between Aegean and Anatolian tectonic events requires a better understanding of the detailed metamorphic history recorded in Western Anatolia rocks, possible now with advances in garnet-based thermobarometric approaches. Slab tear and ultimate delamination impact lithospheric dynamics, including generating economic and energy deposits, facilitating lithospheric thinning, and influencing the onset of transfer zones that accommodate deformation and provide conduits for magmatism.
1 Introduction

The Aegean and eastern Mediterranean are considered the most rapidly deforming regions across the Alpine-Himalayan chain (Figure 1) (e.g., Papazachos & Delibasis 1969; Papazachos & Comninakis, 1971; McKenzie, 1972; Şengör et al., 1985; Taymaz et al., 1991; Jackson, 1994; Reilinger et al., 1997; Nyst & Thatcher, 2004; Le Pichon et al., 2019; Meng et al., 2021). The Aegean and Anatolia microplates, sometimes classified as the single Aegean-Anatolian microplate, are a complex amalgamation of a series of terranes that today experience over two-thirds of the country of Turkey (Le Pichon et al., 1995; Oral et al., 1995; Reilinger et al., 1997; Papazachos, 1999). It is the westernmost protrusion of the Asian continent, with a pole of rotation located in the northern Sinai Peninsula (e.g., Reilinger et al., 2010). The Black Sea bounds it to the north and the Mediterranean Sea to the south. The Aegean microplate is largely comprised of continental crust and sediments obscured by the Aegean Sea (Le Pichon & Angelier, 1981; Jolivet & Patriat, 1999; Makris et al., 2013). The Sea of Marmara connects the Black and Aegean Seas through the Bosphorus and Dardanelles straits and separates a fragment of Eurasia’s microplate (Nyst & Thatcher, 2004).

Deciphering the assembly of the Aegean and Anatolian microplates and their past and present-day deformation drivers impacts our understanding of continental tectonics, subduction zone processes, lithospheric deformation, ore generation process, and hazards (e.g., Jackson, 1994; Meng et al., 2021; Rabayrol & Hart, 2021). The borders of the Aegean and Anatolian microplates coincide with fault systems that played vital roles in triggering changes in their tectonic nature (e.g., McKenzie, 1972; 1978). The microplates share some borders, including the right-lateral strike-slip North Anatolian transform fault and the Western Anatolian Extensional Province (WTEP) (Figure 1) (e.g., Ketin, 1948; Şengör et al., 1985; Barka, 1992; Armijo et al., 1999; Çemen et al., 2006; Barka et al., 2000; McClusky et al., 2000; Chousianitis et al., 2015). The subducting Hellenic and Cyprus arcs and the complex dynamics coinciding with the Florence Rise make up their southern borders (e.g., Le Pichon & Angelier, 1979; Angelier et al., 1982; Anastasakis & Kelling 1991; Papazachos et al., 2000; Ergün et al., 2005; Chousianitis et al. 2015; Haddad et al. 2020). The southern boundary is separated from the Anatolia plate by the WTEP., a zone of N-S extension (Figure 1) (McClusky et al., 2000; Chousianitis et al., 2015). Although many of its bounding fault systems are presently active, both the Anatolian and Aegean microplates contain internal structures, including transfer zones (Figure 1 and Figure 2) (e.g., Nyst & Thatcher, 2004; Çemen et al., 2006; Oner et al., 2010; Aktuğ et al., 2013; Özkaymak et al., 2013; Uzel et al., 2013; Seghedi et al., 2015; Barbot & Weiss, 2021).

Several tectonic models applied to the Aegean and Anatolian microplates have transformed our ideas about the lithosphere’s response to extensional, strike-slip, and compressional forces (see review in Aktuğ et al., 2013). Advances in tomography and GPS technology have contributed to our understanding of its present-day dynamics (e.g., Barka & Reilinger, 1997; McClusky et al., 2000; Ganas & Parsons, 2009; Komut et al., 2012; Aktuğ et al., 2013; Jolivet et al., 2015; Ventouzi et al., 2018). The deformation, metamorphism, and igneous
activity exposed in the upper portions of the microplate’s lithosphere provide constraints on processes that operated in its lower lithosphere over long periods of geological time (e.g., Jackson, 1994; Komut et al., 2012).

This review paper is divided into two primary parts. The first section reviews some of the chronology and tectonic history of the juncture between the Aegean and Anatolian microplates from data available in Western Anatolia (Figure 1 and Figure 2). The goal is to outline how the boundary results from an accumulation of a series of tectonic processes that record stress transitions in the geological past. The second part of the paper aims to present outstanding questions that remain in unraveling its complex dynamics. This particular area of the Anatolian microplate has been the focus of attention for almost fifty years (e.g., McKenzie 1972) and has become the type-locality for understanding subduction zone dynamics, a focus of diverse and multi-disciplinary studies.

2 Geological Background

2.1 Assembly of key components (Paleoproterozoic–Eocene)

The Anatolian microplate is comprised of multiple continental fragments separated by oceans that collided and ultimately combined by the Late Cretaceous–Eocene, with exposures of ophiolitic and high-pressure/low-temperature rock assemblages that mark the suture zones (Figure 2, Figure 3, Figure 4) (e.g., Şengör & Yılmaz, 1981; Okay, 2008; Moix et al., 2008; Okay & Tuysuz, 1999; Pourteau et al., 2016; Okay et al., 2020). Western Anatolia is explicitly defined by the amalgamation of two terranes: the Pontides to the north and the Anatolides-Taurides to the south (e.g., Şengör & Yılmaz, 1981; Yılmaz et al., 1997; Okay & Tuysuz, 1999; Pourteau et al., 2016). The Pontides extends across northern Turkey and is comprised mainly of Pan-African basement blocks and Phanerozoic sedimentary cover units that may have originated from the southern Eurasia margin before back-arc extension initiated and created the Black Sea (Yılmaz et al., 1997; Moix et al., 2008; Pourteau et al., 2010; Okay et al., 2013).

The Intra-Pontide suture zone (IPS) is mapped within the Pontide zone between the Sakarya continental zones and Istanbul-Zonguldak Unit (also Istanbul–Zonguldak Zone, Istanbul Nappe, or Istanbul Zone, see Yiğitbaş et al., 2004) (Figure 2, Figure 3, Figure 4). The Istanbul portion of the unit exists in the west (Istanbul, Gebze, south Camdağ regions) and the Zonguldak to the east (north Camdağ, Zonguldak, Safranbolu regions), both being Gondwanan fragments (e.g., Okay et al., 2006; Bozkaya et al., 2012). The IPS has varying interpretations, including an accretionary complex, a suprasubduction zone, and a remnant of a former ocean basin that may have extended into eastern Europe (e.g., Okay et al., 1996; Robertson & Ustaömer, 2004; Göncüoğlu et al., 2012; 2014; Marroni et al., 2014; Akbayram et al., 2016; Sayit et al., 2016; Frassi et al., 2018). Geological units within the IPS may also be from components from the Istanbul-Zonguldak or Sakarya zones, which has led to a debate about its presence and utility of the IPS in paleogeographic reconstructions (Moix et al., 2008).

Magmatic assemblages help us understand the tectonic processes involved in Western Anatolia, so we present a summary of some available time constraints for several key granite bodies dispersed throughout this region in Tables 1-8 and Figure 4. Zircon ages extracted from metagranites and quartzite units indicate that the Istanbul-Zonguldak Unit has a Precambrian basement with Gondwanan units (Chen et al., 2002; Yiğitbaş et al., 2004; Ustaömer et al., 2005; 2011) and stratigraphic similarities with Paleozoic rocks from the southern margin of Laurasia (Görür et al., 1997; Kaldova et al., 2003). Some of the oldest Neoproterozoic granites in Western
Anatolia are found in the Istanbul Zone (Table 1, Karadere or Karabuk metagranite, Figure 4; Chen et al., 2002; Ustaömer et al., 2016; Di Rosa et al. 2019), although zircons from the Karacabey (Tamsali) and Karaburun plutons in the Western Sakarya Zone and the Çine Massif in the southern portion of the Menderes Massif also yield Paleoproterozoic and Neoproterozoic ages (Tables 5 and 8; Loos & Reischmann 1999; Aysal et al., 2012; Ustaömer et al., 2012). The Triassic ages from granites that intrude the Istanbul-Zonguldak Unit are thought to time partial closure of the Paleotethyan Ocean (Table 1, e.g., Ustaömer et al., 2016). Some of the youngest mineral ages from Istanbul-Zonguldak granites are Late Cretaceous (40\(\text{Ar}/39\text{Ar}\) ages, 93.3±2.0 Ma, 86.1±2.0 Ma, Delaloye & Bingöl, 2000), which are similar to estimates for the activity within the subduction-accretion complex associated with the Izmir-Ankara-Erzincan Suture Zone (IAESZ) (Figure 2, Figure 3, Figure 4) (Okay et al., 2020).

The IAESZ separates the Pontide’s Sakarya Composite Terrane in the north from the Anatolide-Tauride block to the south (Figure 2 and Figure 4) (Şengör & Yılmaz, 1981; Okay & Tüysüz, 1999; Tekin et al., 2002; Göncüoğlu, 2010). Both the IPS and IAESZ mark late Cretaceous–earliest Tertiary closure of Neo-Tethyan ocean basins (e.g., Pourteau et al., 2010; Akbayram et al., 2016). In the Aegean microplate, the IAESZ is thought to record the closure of the Vardar ocean and link with the Vardar ophiolite (or Axios-Vardar suture zone) (Channell & Kozur, 1997; Okay & Tuysuz, 1999; Tekin et al., 2002; Moix et al., 2008), but its exposure beneath the Aegean Sea is masked (e.g., Burtman, 1994; Stampfli, 2000; Yılmaz et al., 2001; Burchfiel et al., 2008). The Vardar suture may also connect to the IPS that separates the Sakarya Zone from the Istanbul Zone (Şengör & Yılmaz, 1981; Okay & Satir, 2000; Okay et al., 2001; Beccaletto & Jenny, 2004; Okay et al., 2010; d’Atri et al., 2012; Di Rosa et al. 2019), and may connect to the Meliata-Balkan suture of Greece (Stampfli, 2000). The IPS and Vardar connection may be evidenced in the Biga Peninsula by an isolated ophiolite-bearing accretionary complex that was active until the Late Cretaceous (Figure 2 and Figure 4) (e.g., Okay et al., 1991). Some disagree and do not map any major suture within the Biga Peninsula (Altunkaynak & Genc, 2008; Burchfiel et al., 2008; Sengun et al., 2011). Because of the uncertain link between the sutures, the relationship of the basement of the Biga Peninsula to that in the Rhodope-Thrace Massif is debated (Bonev & Beccaletto, 2007; Elmas, 2012). In Western Anatolia, the Pamphylian Suture (Figure 2) may connect to the Alanya and Bitlis suture zones further to the east (Centikaplan et al., 2016) and beneath the Lycian nappes to the Cycladic domain to the west (Stampfli & Kozur, 2006).

In Western Anatolia, blueschist assemblages exposed along the IAESZ are intruded by Suture Zone Granitoids (SZGs) [Topuk, Orhaneli, Tepeldag (Gürgenyayla and Gürgenyayla), Table 2; Figure 4]. These granitoids have Paleocene (63.5±2.8 Ma) to Oligocene (31.4±0.6 Ma) ages but are largely thought to have crystallized in the early Eocene (~45-47 Ma, Okay & Satir, 2006; Altunkaynak, 2007). The SGZs intrude the western portion of the Tavşanlı Zone, a blueschist sequence overlain by a Cretaceous accretionary complex and ophiolitic sheet. The zone formed as a result of northward-dipping subduction and represents the Mesozoic to Eocene closing of the northern branch of the Neo-Tethyan Ocean (Okay, 1986; 2008; Okay & Kelley, 1994; Sherlock et al., 1999; Moix et al., 2008, Shin et al., 2013; Plunder et al., 2013; Fornash & Whitney, 2020). The Tavşanlı Zone is narrow (~50 km) and trends E-W for approximately 250-350 km (Okay & Whitney, 2010; Plunder et al., 2013). The western and central portions contain blueschist facies metavolcanic and metasedimentary rocks with rare metabasalts (Okay, 1980a, 1980b, 1982; Okay & Kelley, 1994, see Seaton et al., 2009).
The Sivrihisar Massif further to the east is the only portion of the Tavşanlı Zone to contain eclogite and blueschist and Barrovian sequences (Figure 4 and Figure 5) (Gautier, 1984; Seaton et al., 2009). Rb-Sr and $^{40}$Ar/$^{39}$Ar phengite ages from the Sivrihisar Massif constrain high-pressure/low-temperature (HP/LT) metamorphism to ~88–80 Ma (Sherlock et al., 1999; Seaton et al., 2009; Whitney et al., 2011; Pourteau et al., 2013; Shin et al., 2013). Older ages from the HP/LT assemblages reported from the western portion of the Tavşanlı Zone may suffer from excess argon (see review in Shin et al., 2013). Barrovian-metamorphosed marble from the Sivrihisar massif contains ~59 Ma muscovite ($^{40}$Ar/$^{39}$Ar), timing their exhumation (Seaton et al., 2009). Late Cretaceous and early Paleocene ages are also reported from eastern Tavşanlı Zone granitoids, which are medium to high K., calc-alkaline, metaluminous, I-type, and post-collisional [Kaymaz, Sivrihisar, Sarıkavak (Topkaya), Günyüzü (Karacaören, Tekoren, Dînek, Kadinicik bodies) Figure 4 and Figure 5, Table 2] (e.g., Shin et al., 2013; Demirbilek et al., 2018). However, these results are interpreted as inheritance (Shin et al., 2013; Demirbilek et al., 2018). The Sivrihisar granite’s age is often cited to be 53±3 Ma, based on a hornblende $^{40}$Ar/$^{39}$Ar age clearly affected by excess argon (Sherlock et al., 1999) (Figure 5B and C). However, the Sivrihisar granite contains zircon that is 78.4±8.5 Ma (likely inherited) to 41.9±2.3 Ma (U-Pb, ±1σ, Shin et al., 2013). Figure 5B and C show the K-feldspar $^{40}$Ar/$^{39}$Ar age from the same sample, which yields a plateau age of 46.02±0.21 Ma (MSWD 4.21), similar to those reported for the Sivrihisar and nearby Kaymaz granite and SZGs (Table 2). The flat $^{40}$Ar/$^{39}$Ar age spectrum is consistent with rapid cooling during exhumation (Figure 5D). Paleocene-Eocene ages from the Tavşanlı Zone granites mark the timing of the closure of the IAESZ (e.g., Okay et al., 2020).

The Tavşanlı zone is one component of the larger Anatolide-Tauride block, a microcontinent that rifted away from the northern margin of Gondwana beginning in the early Permian (Figure 2, Figure 3, and Figure 4) (Stampfli & Kozur, 2006) or Triassic (e.g., Şengör & Yılmaz, 1981; Şengör et al., 1984; Okay & Tuysuz, 1999; Robertson & Ustaömer, 2009a, 2009b). The Taurides comprise the southern portion of the Anatolide-Tauride block and is Neoproterozoic-Early Cambrian (Infracambrian) basement overlain by Cambrian to Eocene marine sediments (e.g., Gutnic et al., 1979; Özgül, 1997; Candan et al., 2016). The Anatolide terrane is the metamorphic equivalent of the Taurides and is subdivided into zones based on lithologies and the type and age of metamorphism (see review in Bozkurt & Oberhansli, 2001; Candan et al., 2016; Moix et al., 2008). These include the Tavşanlı Zone, Afyon Zone, Menderes Massif, and Lycian nappes (Figure 2, Figure 3, and Figure 4). The Tavşanlı and Afyon zones are sometimes considered as part of a single Kütahya–Bolkardağ Belt (Özcan et al., 1988; Göncüoğlu et al., 1997; 2012).

Note that a series of granite bodies intrude the IPS between the Sakarya and Istanbul Zones also ages that resemble the SZGs and eastern portions of the Tavşanlı Zone. These Middle Eocene Magmatic Rocks (MEMR), also known as the South Marmara Granitoids [Şevketiye, İlyasdağ tonalite (Marmara Island), Karabiga (Lapeski), Fistikli (Armutlu–Yalova), Kapıdağ, and Avsa Island; Figure 4, Table 3] are located in close association with the IPS and range in age from the Late Cretaceous (71.9±1.8 Ma) to Late Eocene (34.3±0.9 Ma). The MEMR are unique in these ages, as further east, along strike of the IPS and into the central portion of the Sakarya Zone, some of the oldest plutons in Western Anatolia are exposed (Pamukova, Gemlik, Inhisar, Gevyke, Bilecik, Söğüt, Figure 4, Table 4). Some of these intrusions are associated with economically important kaolinite deposits (e.g., Kadir & Kart, 2009). The Cambrian Gemlik granite body is located in the vicinity of the MEMR granites (Figure 4). Its age is more
consistent with Cadomian Orogeny (650–550 Ma) granites further north in the Istanbul-Zonguldak and Strandja zones (e.g., Şahin et al., 2014) and similar-age rocks from the basement or core of the Afyon Zone and Menderes Massif (e.g., Dannat, 1997; Loos & Reichmann, 1999; Şahin et al., 2014; Hetzel & Reischmann, 1996). Western Anatolian granites with Cambrian ages are termed the Late Pan-African Granitoids or Cadomian Granitoids and are associated with tectonic events along the northern margin of Gondwana (Gürsu & Göncüoğlu, 2006; Şahin et al., 2014). We identify some of these granites in their particular zones in Figure 4 and distinct sections of Tables 1, 4, and 8. Note that the entire core of the Menderes Massif is considered Pan-African (primarily late Neoproterozoic to Cambrian) basement (see review in Oberhänsli et al., 2010).

Proterozoic zircon ages are found in the Pontides zone, but some of its central and western granite assemblages also record Silurian-Devonian ages [Saricakaya, Table 4; Karaburun, Güveylerobasi (Çamlık-related), Karacabey (Tamsali), Eybek (Çamlık), Güveylerobasi, Table 5; Figure 4]. These ages are linked to the amalgamation of a fragment of Avalonia terrane in a subduction-zone type setting (Aysal et al., 2012; Sunal, 2012; Topuz et al., 2020). Variscan-age (Carboniferous) granitoids are also reported for granites in the Central and Western Sakarya Zone and Afyon Zone (Tables 4 and 6; Figure 4). Some of these results could represent inherited cores or xenocrystic grains from the surrounding metamorphic assemblages. For example, the Miocene-age Alaçam granite in the Afyon Zone has reported Carboniferous ages, but the older ages were likely entrained from its basement units (Hasözbek et al. 2010; Candan et al. 2016).

The Afyon zone is considered the southward palaeogeographic extension of the Tavşanlı zone (Candan et al., 2005; Pourteau et al., 2010; Akal, 2013; Özdamar et al., 2013). Although it is often mapped as closely and narrowly paralleling the Tavşanlı Zone, the southern extent of the Afyon Zone is unclear, and a portion may also be exposed between the southern Menderes Massif and Lycian Nappes (Okay, 1986; Candan et al., 2005; Pourteau et al., 2013; Ustaömer et al., 2020). The zone has also been termed the Afyon–Bolkardag Zone (Okay, 1986; Özdamar et al., 2013) and Ören–Afyon Zone (Pourteau et al., 2013). The zone consists of Pan-African-related basement underlying shelf-type Palaeozoic-Mesozoic sequence of the Taurides and metasedimentary and metavolcanic rocks, portions of which have undergone regional greenschist to blueschist facies (Fe–Mg carpholite and glaucophane) metamorphism (Figure 3) (Okay, 1984; Candan et al., 2005, Pourteau et al., 2010; Özdamar et al., 2013). In this sense, its stratigraphy resembles that of the Tavşanlı Zone (Candan et al., 2005). Rhyolitic volcanic assemblages contain zircon that crystallized in the Late Triassic time extension along the northern margin of Gondwana as the Neo-Tethyan Ocean developed (230±2 Ma and 229±2 Ma; Özdamar et al., 2013). Triassic ages reported for granitic assemblages found within the Istanbul-Zonguldak zone, central and western Sarkarya, and Menderes Massif are also attributed to this event (Figure 4; Okay et al., 2020). LT/HP metamorphism in the Afyon Zone is thought to have occurred at 70-65 Ma coincident with the closure of the Neo-Tethyan Ocean (Pourteau et al., 2010, 2013; Özdamar et al., 2013; Plunder et al., 2013). Based on zircon ages from granites intruding Tavşanlı Zone blueschist and altered ophiolitic assemblages, portions of the Afyon Zone may have subducted beneath the Tavşanlı Zone during the Late Cretaceous (Speciale et al., 2012; Shin et al., 2013). Upper Palaeocene-Lower Eocene sedimentary rocks overlie the metamorphic rocks of the Afyon Zone (Candan et al., 2005).
The Menderes Massif is considered the metamorphic basement on which the rocks of the Afyon Zone were deposited before regional metamorphism (Okay, 1984). The Menderes Massif exposes ~40,000 km² of metamorphic and igneous rocks, and its stratigraphy was originally described as a gneiss ‘core’ and Paleozoic schist envelope with overlying Mesozoic-Cenozoic marble ‘cover’ (e.g., Schuiling, 1962; Durr, 1975; Şengör et al., 1984). The massif has also been mapped as a large-scale recumbent fold (Okay, 2001; Gessner et al., 2002), a series of nappes stacked during south-directed thrusting (Ring et al., 1999; 2001; Gessner et al., 2001), or north-directed thrusting (Hetzel et al., 1995a,b) (see Gessner et al., 2013). In the nappe model, the core is represented by the Çine and Bozdağ nappes, whereas the cover would be the Bayındır and Selimiye nappes (Ring et al., 2001), although all nappes may be part of the Menderes Massif core series stacked during Eocene out-of-sequence thrusting (Régnier et al., 2007). Timeframes recorded by the massif begin in the Archean and Neoproterozoic based on zircons extracted from metagranites and orthogneisses with geochemical signatures dominated by reworking of old crust (Oberhansli et al., 2010; Zlatkin et al., 2013). During this time, the Menderes Massif was part of a collage of terranes associated with NE Africa and Arabia (Şengör et al., 1984; von Raumer et al., 2015). Some Neoproterozoic zircons (ca. 570 Ma) have an older crust signature, but others suggest a proximal juvenile source resembling the Arabian-Nubian shield (Zlatkin et al., 2013).

Cambrian metagranites, orthogneisses, granulites, and ecologites, mica schists are exposed throughout the massif (Hetzel & Reischmann, 1996; Dannat, 1997; Loos & Reichmann, 1999; Neubauer, 2002; Oberhansli et al., 2010; Zlatkin et al., 2013; Koralay, 2015). Cambrian-Ordovician monazite and zircon inclusions are found in Menderes Massif garnets (Catlos & Çemen, 2005; et al., 2019). During this time, the Menderes Massif was affected by events related to the Cadomian Orogeny, and its core units were intruded by Pan African S- and I-type granites followed by metamorphism (Neubauer, 2002). Note that other terranes within Western Anatolia likewise have a Cadomian signature (Figure 4, e.g., Kozur & Gönçüoğlu, 1998).

Granulite-facies metamorphism in the Menderes Massif was suggested to have occurred at 580.0±5.7 Ma to 660±61 Ma by (U-Pb monazite ages, Oelsner et al., 1997; U-Pb zircon ages, Korolay et al., 2006). Middle-Triassic zircons in metagranites are found in its central portions (Figure 4; Dannat, 1997; Koralay et al., 2001).

The timing of Menderes Massif nappe stacking is largely thought to have occurred during the Eocene-Oligocene, or sometime after the Late Cretaceous (Main Menderes Metamorphism, MMM., e.g., Satir & Friedrichsen, 1986; Konak et al., 1987; Dora et al., 1995; Bozkurt & Park, 1999; Bozkurt & Satir, 2000; Bozkurt & Oberhansli, 2001; Candan et al., 2001; Lips et al., 2001; Gessner et al., 2011). Gessner et al. (2001) report that the Bayındır nappe deformed once during the Eocene related to MMM., whereas the Bozdağ, Çine, and Selimiye nappes record pre-MMM and MMM events. Figure 6 shows a paleogeographic reconstruction of the possible setting of the fragments comprising Western Anatolia during the closure of the IAESZ during the Eocene. This paleogeographic timeframe is critical for understanding the complex tectonic scenario that set the scene before the onset of extension.

The Aegean Orogeny (Searle & Lamont, 2020a) is proposed for the tectonic history further to the west of the Menderes Massif, including the Cycladic Metamorphic Core Complexes but may mirror its development. In this scenario, subduction and a continent-continent collision occur between the Eurasian and Adria-Apulia/Cyclades plates as marked by ophiolite obduction at 74 Ma (Lamont et al., 2020a) and HP eclogite and blueschist facies
metamorphism at 57 Ma–46.5 Ma (Tomaschek et al., 2003; Lagos et al., 2007; Bulle et al., 2010; Dragovic et al., 2012). The HP metamorphism (P = 11-12 kbar) is documented by ophiolitic melanges that may record a cycle of Alpine collisional thickening followed by extension and overprinting via extension (Papanikolau, 1987, Okrusch & Bröcker, 1990; Avigad & Garfunkel, 1991; Katzir et al., 2000; Parra et al., 2002; Laurent et al., 2018; Lamont et al., 2020b). HP metamorphism is recognized as part of a NE-trending subduction-exhumation channel (e.g., Xypolias & Alsop, 2014; Laurent et al., 2018; Gerogiannis et al., 2019). Crustal thickening and regional kyanite– sillimanite grade Barrovian-type metamorphism occur from 22–14 Ma, followed by orogenic collapse. The island of Naxos exemplifies the process with structural data that suggest it is the result of the gravitational collapse of the Aegean orogenic wedge (Vanderhaeghe, 2004). This model emphasizes the role of compression in forming Aegean metamorphic core complexes (e.g., Coney and Harms, 1984; Searle and Lamont, 2020a,b), which is an alternative to the perspective of solely extensionally-driven core complexes discussed in the next section.

2.2 Extensional history (Oligocene-Miocene)

Following the final amalgamation of the various terranes as described in the previous section, Western Anatolia experienced a switch from the dynamics of collision to extension and extrusion (e.g., Berckhemer, 1977; Le Pichon & Angelier, 1979; 1981; Şengör & Yılmaz, 1981; Şengör et al., 1985; Meulenkamp et al., 1988; Buick, 1991; Jolivet et al., 1994; Seyitoglu & Scott, 1996; Okay & Satir, 2000; Bozkurt, 2001; Çemen et al., 2006). A sequence of partial melting, Barrovian metamorphism, and granitoid emplacement has been cited for providing evidence of a change from crustal shortening to extensional tectonism (e.g., Keay et al., 2001; Altunkaynak, 2007; Dilek & Altunkaynak, 2007; Altunkaynak et al., 2012; Rossetti et al., 2017). The process may be recorded by numerous Oligocene to Miocene-age granites (Figure 4, Tables 5-8) and linked to the development of metamorphic core complexes located from northeastern Greece and southern Bulgaria through the Aegean Sea and western Turkey.

In continental orogenic domains, metamorphic core complexes are deep crustal domes exhumed and deformed during extension and are commonly surrounded by sedimentary and volcanic rocks, which may be partly deposited during their exhumation (Tirel et al., 2008). Core complexes in western Turkey and the Aegean region include the Rhodope, Kazdağ, Uludağ, Cyclades, Menderes, and Crete massifs (Figure 1, Figure 2, and Figure 4) (Sokoutis et al., 1993; Hetzel et al., 1995a,b; Burg et al., 1996; Lips et al., 1999; Bozkurt & Oberhänslı, 2001; Candan et al., 2001; Lips et al., 2001; Ring et al., 2003; Bozkurt & Sözbilir, 2004; Duru et al., 2004; Vanderhaeghe, 2004; Catlos & Çemen, 2005; Brun & Sokoutis, 2007; Okay et al., 2008; Cavazza et al., 2009; Kruckenberg et al., 2011; Gessner et al., 2013; Baran et al., 2017).

In Western Anatolia specifically, the Menderes, Kazdağ, and Uludağ massifs are central locations for studying post-collision extensional tectonics (Figures 1, Figure 2, and Figure 4) (e.g., Şengör et al., 1984, Bozkurt & Park, 1994; Hetzel et al., 1995a,b; Yılmaz et al., 2001; Işık & Tekeli, 2001; Çemen et al., 2006; Topuz & Okay, 2017). The Menderes Massif has global importance due to its role as the largest zone of active continental extension (e.g., Jolivet & Facenna, 2000; Çemen et al., 2006). The region has long attracted the attention of those seeking to understand the driving forces of extension from a variety of perspectives (e.g., Lister et al., 1984, Thomson & Ring, 2006; Régnier et al., 2007; Gessner et al., 2013; Uzel et al., 2015). Both low-angle detachment faults and high-angle normal faults bound sedimentary basins and separate the Menderes Massif into northern (Gördes), central (Ödemis), and southern (Çine) submassifs.
Manuscript accepted to AGU Books

(Figure 2). In the central Menderes Massif, Miocene-age granites are cut by the low-angle Alasehir detachment, helping to constrain the timing of extension (Alasehir, Salihli, Turgutlu, Table 8). The Kazdağ Massif is smaller in scale compared to the Menderes Massif and is a NE-SW oriented structural dome or tectonic window flanked by detachment structures (Figure 2 and Figure 4) (Okay et al., 1991; Okay & Satir, 2000; Duru et al., 2004; Bonev et al., 2009; Cavazza et al., 2009). This massif’s Evciler (Kazdağ) pluton routinely yields Oligo-Miocene crystallization ages from a range of chronometers (Table 5). The Uludağ Massif is NW-SE trending and has high-grade metamorphic and intrusive Eocene-Miocene age granitic rocks (Figure 4, Table 5, Okay et al., 2008). Large Neogene basins bind the northern and southern sections of the Uludağ Massif, and late-stage exhumation is largely thought to have occurred during the Early Miocene (e.g., Topuz & Okay, 2017).

Besides these localities, Miocene ages have been reported for granites in the eastern Tavşanlı zone (Table 2) [Kaymaz and Tekoren granodiorite (Günyüzü); Shin et al., 2013; Demirbilek et al., 2018]. These ages likely represent metamorphism and subsequent alteration associated with the large-scale extension/exhumation affecting Western Anatolia during this time. Early Miocene ages also characterize granites closely associated with the Menderes, Kazdağ, and Uludağ metamorphic core complexes. For example, Miocene ages are reported for a group of granites near the Kazdağ Massif, extensively exposed in the Biga Peninsula and western Pontides [Kozak, Eybek, Katrandag, Cataldag (Bozenkoy, Cataltepe, Tufaldag, Balicikhisar), Kuscayir, and Kestanbol (Ezine), Figure 4, Table 5] and from a series of plutons grouped as the Younger South Marmara Granitoids (Yenice, Ilica, Kizildam, Danisment, Sarioluk, Davutlar, and Yeniköy; Figure 4, Table 5; Karacik et al., 2008). North of the Menderes Massif, Miocene-age plutons also intrude the Afyon Zone, in close association with the Simav fault system, which includes the lower angle Simav Detachment Fault (SDF) and higher-angle Simav Fault further south (Koyunoba, Alaçam, and Egrigöz, Figure 2, Figure 4 and Figure 7, Table 7; Isik et al., 2003).

The Simav structures are at the boundary between two dynamically distinct regions in western Turkey: a northern component dominated by the NAFZ that accommodates the lateral extrusion of the Anatolian block and a southern zone of large-scale crustal extension (Seyitoğlu, 1997; Ersoy et al., 2010). The Simav Fault is a distinct, a high-angle (~45-60º) system that extends ~150 km between the towns of Banaz in the east and Sındırgı in the west (Figure 7) (Ambraseys & Tchalenko, 1972; Seyitoğlu, 1997; Ersoy et al., 2010; Hetzel et al., 2013). The structure near the town of Simav has >200 m of relief between the top of the hanging-wall and footwall, and dips steeply to the north, roughly perpendicular to the current extension direction (Tekeli et al., 2001; Işık et al., 2003). This fault is thought to have formed during the Pliocene and is currently active (Seyitoğlu, 1997; Ring & Collins, 2005). Deciphering the sense of motion of the Simav Fault has implications for the understanding of the neotectonic regime of Turkey and is discussed further in the section regarding outstanding questions in Aegean tectonics.

Estimates of timing core complex exhumation and extension in Western Anatolia have relied on calc-alkaline magmatism, widespread continental sedimentation, and mineral chronometers (Sokoutis et al., 1993; Gautier et al., 1999; Catlos & Çemen, 2005; Altunkaynak & Genç, 2008; Brun & Sokoutis 2010; Brun et al., 2016). In some locations, the complexes record progression of magmatism from earlier Eocene-age mantle melts and input from asthenosphere upwelling to later Oligocene to Late Miocene crustal contamination and subduction signatures, with emplacement ages that young to the south (e.g., Delaloye & Bingöl, 2000; Altunkaynak &
Dilek, 2006; Dilek & Altunkaynak, 2007; Altunkaynak, 2007; Altunkaynak & Genç, 2008; Dilek & Altunkaynak, 2009; Altunkaynak et al., 2012; Karaoğlu & Helvacı, 2014). However, this simple scenario of melt origin and emplacement can be complicated, as the melts are influenced by varied protoliths of varying sources, ages, and degrees of crustal anatexis (Pe-piper, 2000; Stouraiti et al., 2010; 2018).

Late Cenozoic (since ~32 Ma) plutonic rocks are also widespread in the Aegean (e.g., Altherr et al., 1982; Henjes-Kunst et al. 1988; Pe-piper, 2000; Keay et al., 2001; Brichau et al., 2007; 2008). The origin of the granites is linked to subduction migration along the Hellenic arc (e.g., Fytikas et al., 1984; Schaarschmidt et al., 2021) or regional, widespread extensional deformation (e.g., Boztuğ et al., 2009). Barrovian metamorphism on Naxos is thought to have influenced the development of fluid-fluxed melts at ca. 8–10 kbar between 18.5 Ma and 17 Ma (Lamont et al., 2019; Searle and Lamont, 2020b). Peak metamorphism is thought to have occurred at 20.7-16.7 Ma (Keay et al., 2001). In some locations, coeval mafic and felsic melts were emplaced (Seyitoğlu & Scott, 1996; Aldanmaz et al., 2000; Okay & Satir, 2000; Pe-Piper & Piper 2001; Ozgenç & Ilbeyli, 2008). Magma compositions were influenced by a range of factors, including inflowing mantle at the site of melting, the nature of the subduction component and the degree of interaction between mantle and subduction components, as well as the melting of fluid-rich mantle and the assimilation/crystallization history of the resulting hydrous magma (e.g., Pearce & Stern, 2006). Extensive geochemical and isotopic studies of Miocene I-type granitoid plutons of the central Aegean Sea show little evidence for a significant contribution of mantle-derived magmas (Altherr & Siebel, 2002).

Cenozoic magmatism in the Anatolian microplate consists of three distinct, continuous geochemical phases (Innocenti et al., 2005; Dilek & Altunkaynak, 2007; Altunkaynak & Genç, 2008; Akay, 2009; Altunkaynak et al., 2012). Magmatic rocks represent a Late Eocene-Middle Miocene phase with orogenic character and a petrological affinity ranging from calc-alkaline to dominant high-K calc-alkaline to shoshonites. During the Late Miocene–Early Pliocene, alkaline volcanic rocks appear. The third phase is characterized by Pliocene–Quaternary Na-enriched alkali basalts with an oceanic island basalt (OIB) signature (Aldanmaz, 2012). The first volcanic activity in the South Aegean Active Volcanic Arc occurred between 5 and 2 Ma (e.g., Müller et al., 1979; Fytikas et al., 1984; Matsuda et al., 1999; Elburg & Smet, 2020). The driver of extension is widely thought to be the rollback of a subducting African slab (Figure 8, Figure 9, and Figure 10) (e.g., Jolivet & Faccenna, 2000; Çemen et al., 2006; van Hinsbergen, 2010; Royden 1993; Faccenna et al. 2003, 2014; Brun & Faccenna 2008). We discuss the slab and arc dynamics, geometry, and age in the section regarding outstanding questions in Aegean tectonics.

2.3 Strike-slip History (Late Miocene, Pliocene-present)

The Aegean and Anatolian microplates have emerged to be type-localities for the model of tectonic escape based on GPS vectors (Reilinger et al., 2006). In this scenario, the Anatolian plate moves westward in response to the collision of Arabia and Eurasia (e.g., Şengör & Yılmaz, 1981; Şengör et al., 1985; Bozkurt, 2001). The North and East Anatolian transform fault systems accommodate extrusion, and rollback along the Hellenic arc is suggested to provide space to accommodate the escaping plate (McKenzie, 1972; Dewey & Şengör, 1979; Le Pichon & Angelier, 1979; Jackson & McKenzie, 1984; Barka & Kadinsky-Cade, 1988; Taymaz et al., 1991; Reilinger et al., 1997; McClusky et al., 2000; Tatar et al., 2013). Philippon et al. (2014) suggest a two-stage evolution of the arc. At 30 Ma, extension was only driven by the southward retreat of the Hellenic trench at a rate lower than 1 cm/yr, but since the last 13 Ma, the
interaction of trench retreat with Anatolia escape accelerated the rate of trench retreat in the southwest direction at a rate of up to 3 cm/yr.

In western Turkey, extrusion tectonics is dominated by the active right-lateral North Anatolian strike-slip fault (NAF) and North Anatolian Shear Zone (NASZ), which extends for ~1200 km from the Karlıova triple junction through the Sea of Marmara and Biga Peninsula (Figure 1) (Ketin, 1948; Barka, 1992; Armijo et al., 1999; Şengör & Zabci, 2019). The NASZ contains the NAF and is speculated to have accommodated from 25 to 110 km of displacement, depending on location since the late Miocene (Westaway 1994; Yoshioka 1996; Armijo et al., 1999; Hubert-Ferrari et al. 2002; Şengör & Zabci, 2019). The structure accommodates ~24 mm/year of slip along northern Turkey (McClusky et al., 2000; Bulut et al., 2018). The geometries of its western and eastern terminations are poorly defined (Barbot & Weiss, 2021).

The NAF splits into three strands as it trends westward into Western Anatolia and the Aegean Sea (Figure 1) (e.g., Emre et al., 1998; Kürcer et al., 2008; Beniest et al., 2016; Şengör & Zabci, 2019). Each segment is comprised of several en échelon fragments (Emre et al., 1998; Kürcer et al., 2008). The northernmost E-W striking segments within the Sea of Marmara change strike in the Northern Aegean Sea towards a NE-SW orientation in the North Aegean Trough, maintaining its right-lateral strike-slip character but splits across three basins and two transpressional ridges (Bulut et al., 2018). A branch between the northern and central segments originates southeast of Sapanca Lake (Kürcer et al., 2008) and terminates at the western end of the North Aegean Trough (Ferentinos et al., 2018). This structure enters the Aegean Sea and trends into the Northern Skyros Basin. Strands of the NAF have also been linked to the KTZ through the transtensional Central Hellenic Shear Zone (Royden & Papanikolaou, 2011; Evangelidis, 2017). In the Western Anatolia-Marmara region, the NAF may have been active since the Pliocene (e.g., Ünay et al., 2001).

Sakellariou et al. (2013) suggest that the southwestward expansion and stretching of the Aegean microplate during Plio-Quaternary is accommodated by a northern right lateral tectonic boundary marked by the KTZ and NASZ, and a southern left-lateral tectonic boundary, marked by the Pliny and Strabo trenches (Figure 9). Papanikolaou and Royden (2007) note that regional extension has a much-reduced role in the dynamics of the Aegean microplate and that, in fact, no active extensional strain is present, except for a small southeastern domain (Figure 1) (Corinth rift, south Viotia, south of Evia, and across the Sperchios-Kammena Vourla rift; Brooks & Ferentinos 1980; Chousianitis et al., 2013, 2015). Maggini & Caputo (2020) report that seismogenic faults in the internal Aegean domain associated with the Hellenic subduction arc are characterized by pure normal and strike-slip kinematics or by a combination and that active thrusting is limited to the central and western sectors of the Hellenic subduction zone and the offshore regions external to it.

Figure 11 shows the focal mechanisms for some recent earthquakes (2010-2020) that appear along the Aegean-Anatolian microplate boundary. Recent earthquakes with focal mechanisms consistent with reverse faulting have occurred south of Crete, including those associated with an Mw 6.4 earthquake on 5/2/2020. These earthquakes occurred at relatively shallow depths (6.5-9.6 km, Table 10) and may be associated with a plate interface zone defined by the upper plate and splay-thrust faults (Saltogianni et al., 2020). Observations and modeling of historical and recent earthquakes have shown that uplift along the Hellenic arc margin offshore of Crete is controlled by reverse fault motion with little contribution from plate-interface slip (e.g., Mouslopoulou et al., 2015).
Extrusion and deformation in Western Anatolia are also accommodated by transfer zones, where strain is transferred from one structural element to another and displacement changes between individual fault and basin segments (e.g., Gawthorpe & Hurst, 1993; Barbot & Weiss, 2021). Some examples of these zones include the NE-SW trending strike-slip dominated Izmir–Balıkesir transfer zone (İBTZ), Uşak-Mugla Transfer Zone (UMTZ), and Southwestern Anatolian Shear Zone (SWASZ) (Figure 1 and Figure 2) (Çemen et al., 2006; Oner et al., 2010; Sözbilir et al., 2011; Gessner et al., 2013; Özkaymak et al., 2013; Uzel et al., 2013; Karaoğlu & Helvacı, 2014; Seghedi et al., 2015). These transfer zones have been considered as significant portions of the larger Western Anatolian Shear Zone (WASZ) or Western Anatolian Extensional Province (Figure 1) and may have developed due to mantle processes related to the subduction of the Aegean slab (e.g., Gessner et al., 2013; Uzel et al., 2020). Some transfer zones trend into other fault systems. For example, the İBTZ is speculated to connect to the Mid-Cycladic Lineament (MCL) in central Greece and the NASZ in northern Turkey (Figure 1, Figure 11) (Uzel et al., 2013; Seghedi et al., 2015; Westerweel et al., 2020). The MCL is a strike-slip structure that may be the result of the reactivation of the Vardar suture zone, evidenced by the North Cycladic Detachment (Figure 11), to accommodate westward extrusion of Anatolia in the Late Miocene (e.g., Philippon et al., 2014). These transfer zones have been used to illustrate that the Aegean and Anatolian microplates experienced or are currently transitioning from a stress regime dominated by extension to transform tectonics (Papanikolaou & Royden, 2007; Cavazza et al., 2009).

Presently, normal fault motion exists within the İBTZ as illustrated by focal mechanisms from a 2020 Mw 6.6 earthquake and 2018 Mw 4.5 earthquake within the zone. An Mw 4.4 earthquake with normal motion occurred off the coast of Amorgos near the 1956 Mw 7.7 (or 7.8) earthquake, one of the strongest earthquakes of the 20th century in the area of the South Aegean (Okal et al., 2009; Alatzà et al., 2020). The 1956 event has debated focal mechanisms, as either strike-slip or normal faulting geometries (see Okal et al., 2009). A normal sense of motion also is found with some recent earthquakes near the NASZ, including 2017 Mw 6.2 and 2017 Mw 5.3 earthquakes (Figure 11, Table 10). These events are likely associated with transtensional motion.

3. Outstanding Questions in Aegean Tectonics

As outlined in the previous section, significant contributions have been made regarding the fundamental tectonics and geological history recorded by rocks throughout the Western Anatolian microplate. However, outstanding questions remain to be addressed regarding the boundary between the Aegean and Anatolian microplates that affect our understanding of the mechanisms that drive extension in the Earth’s lithosphere. Most of these questions center on how upper lithospheric and crustal deformation are linked and are related to lower lithosphere and mantle processes.

3.1. Slab dynamics

3.1.1 African slab geometry and connections to other subduction systems

Based on several geophysical, tectonic, and geochemical developments, the subducting African (Nubian or Aegean) slab has emerged as the primary driver for extension in the Aegean and Anatolian microplates and the development of their metamorphic core complexes (Figure 1, Figure 2, Figure 8, Figure 9, and Figure 10) (e.g., Jolivet et al. 2013; Jolivet & Faccenna, 2000; Çemen et al., 2006; Dilek & Sandvol, 2009; van Hinsbergen et al., 2010; van Hinsbergen & Schmid 2012; Salatín et al., 2012; Faccenna et al., 2014; El-Sharkawy et al., 2020; Barbot &
Weiss, 2021). The Hellenic and Cyprus arcs are the surface expression of the subducting African plate and eastern Mediterranean lithosphere beneath the Anatolian and Aegean microplates (e.g., Le Pichon & Angelier, 1979; Angelier et al., 1982; Anastasakis & Kelling, 1991; Papazachos et al., 2000; Ergün et al., 2005; Ganas & Parsons, 2009; Hall et al., 2009; Royden & Papanikolaou, 2011; Hall et al., 2014; Symeou et al., 2018; Ventouzi et al. 2018).

Although it has a well-developed Wadati-Benioff zone dipping ~30° from 20-100 km depth and ~45° from 100-150 km depths (Figure 10B) (e.g., Papazachos & Comninakis, 1971; Papazachos et al., 2000; Sukale et al., 2009; Hayes, 2018), it has a debated slab geometry at intermediate depths (150-250 km, Suckale et al., 2009; Agostini et al., 2010; see review in Hansen et al., 2019; El-Sharkawy et al., 2020). Seismic body wave tomography shows it extends into the upper and lower mantle to 1400±10 km depth (Figure 10A) (e.g., Spakman et al., 1988; Bijwaard et al., 1998; van der Meer et al., 2018; see review in Bocchini et al., 2018). However, the slab may be a single folded body that overturned in the lower mantle (Faccenna et al., 2003), or two slabs, located between 2000-1500 km and from 1500 km to the surface (van Hinsbergen et al., 2005; van der Meer et al., 2018). Mantle tomography has shown multiple subducted slabs beneath the Aegean and Anatolian microplates (e.g., Spakman et al., 1988; Spakman, 1990; 1991; Wortel & Spakman, 2000; Govers & Fichtner, 2016; van der Meer et al., 2018; Wei et al., 2019). Blom et al. (2019) show the Hellenic slab, visible in both S and P velocity, extending from the surface to the transition zone in a bent, arcuate shape. A high-velocity structure exists beneath the Hellenic arc and the Aegean Sea that flattens from the 410 km discontinuity and is not seen at deeper levels. Wei et al. (2020) show a gap in the subducting slab at depths of 60-100 km just west of the south Hellenides. In the South Hellenides, slab tear may be visible at the 660 km discontinuity, whereas four slabs are imaged beneath the North Hellenides.

Interpretations of these tomographic images have indicated that more slab is imaged than is reflected by seismicity (e.g., Spakman et al., 1988; Papadopoulos, 1997; Bijwaard et al., 1998), and that a variation of slab exists thickness across the Aegean Sea (e.g., Karagianni et al., 2002). Mantle tomography has also shown that not all slabs in the Mediterranean region are connected to the lithosphere at the surface, consistent with past delamination (e.g., Spakman et al., 1988; Dilek & Sandvol, 2009; Wortel & Spakman, 2000). Challenges in imaging the subduction zone include its small size, its spatially highly variable nature, and the uneven distribution of its seismic stations (El-Sharkawy et al., 2020).

The Hellenic subduction system is comprised of three regions: an outer compressional non-volcanic arc, a volcanic arc, and an extensional back-arc region that makes up the broader Aegean Sea region (Figure 8) (McKenzie 1972; Papazachos, 2019). Although the Western Hellenic Arc (also termed the North and Southern Hellenic arc, Royden & Papanikolaou, 2011) has a well-defined topography, trench, sedimentation, and strain pattern (Stanley et al., 1978; Papadopoulos et al., 1988; Hatzfeld et al., 1990; Cocard et al., 1999), the central and eastern portions of the Hellenic arc are more difficult to characterize as the boundary becomes diffuse (Beißer et al., 1990; Shaw & Jackson, 2010; Özbakır et al., 2013). The Hellenic arc’s connection with the Cyprus arc and even the nature of plate motion along strike of the Cyprus arc has been debated (Anastasakis & Kelling, 1991; Woodside et al., 2002; Ergün et al., 2005; Hall et al., 2009; Harrison et al., 2012; Kinnaird & Robertson, 2012; Symeou et al., 2018). The surface morphology of the southern and eastern portions of the Hellenic arc and its connection to the Cyprus arc is obstructed by up to 300-km wide, 6-10 km-thick section of sediments that comprise the Mediterranean Ridge (Figure 8 and Figure 9; Heezen & Ewing, 1963; Emery et al.,
The ridge is a giant accretionary complex, extending ~2000 km from the Calabrian Rise east of Greece to the Florence Rise, and is the largest structural unit of the Eastern Mediterranean Sea (Liminov et al., 1996; Cita et al., 1996). The front of subduction of the Hellenic arc is located south of the Mediterranean Ridge (e.g., Jost et al., 2002; Westbrook & Reston, 2002; Jolivet et al., 2013). The majority of the subducting African plate beneath the ridge is oceanic, except along the central sector of its southern margin, where the accretionary complex collides with the African continental margin (Chaumillon & Mascle, 1997; Westbrook & Reston, 2002). The ridge may be the fastest outward growing wedge in most recent Earth history, with a rate of up to 10 km/Myr (Kastens, 1991; Kopf et al., 2003). It has been speculated to grow by off scraping against a backstop formed by the Alpine nappes of the Hellenic Arc (Kastens, 1991).

The intensively folded and faulted rocks of the Mediterranean Ridge vary in geometry along strike (Cita et al., 1996; Chaumillon & Mascle, 1997; Westbrook & Reston, 2002; Kopf et al., 2003). In its western and eastern portions, the wedge accumulates sediments, but in its central portion between Libya and Crete, the ridge behaves unlike a typical accretionary complex. In this area, a trench system (the Hellenic trenches; Ptolemy, Pliny, and Strabo; Figure 9) developed in between the accretionary complex and volcanic arc, likely as a result of back-thrusting beneath the northern edge of the complex (Galindo-Zaldívar et al., 1996; Westbrook & Reston, 2002).

The accretionary complex is unusual compared to others worldwide, not only because of these back thrusts but also because it appears to have formed in a continent-continent collisional setting and contains shallow, Messinian-age evaporites (e.g., Cita et al., 1996; Chaumillon & Mascle, 1997). These evaporites influence its deformation and fast growth rate due to their mechanical properties and effect upon fluid flow and pressure (Kastens, 1991; Westbrook & Reston, 2002; Kopf et al., 2003). Understanding the development of the Mediterranean Ridge is critical to determining the initiation age of the Hellenic arc, as described in the next section.

### 3.1.2 The age of subduction of the African slab

The Subduction Zone Initiation (SZI) age is defined as the onset of downward plate motion forming a new slab, which later evolves into a self-sustaining subduction zone (Crameri et al., 2020). Constraints regarding SZI age of the present-day expression of the Hellenic arc developed from several independent approaches, including timing sedimentation within the Mediterranean Ridge (Kastens, 1991; Kopf et al., 2003), analysis of topography combined with estimates of slab age and depth (McKenzie, 1978; Le Pichon et al., 2019), reconstructions of subducted slabs using tomography (e.g., Spakman et al., 1988), palaeomagnetism (Savostin et al., 1986; Marsellos et al., 2010), and the timing of metamorphism and volcanic activity (e.g., Fytikas et al., 1984). Early estimates for the initiation of Hellenic arc subduction are 13±3-5 Ma (Le Pichon & Angelier, 1979) to 5-10 Ma (McKenzie, 1978; Mercier, 1981) based on interpretations of seismic activity coupled with assumptions regarding the age of subducted lithosphere and subduction depths. These ages are similar to the onset of the KTZ based on geodynamic modeling and GPS data (Figure 1) (6-8 Ma, Royden & Papanikolaou, 2011) and the timing of the earliest volcanic activity in the South Aegean arc (Pliocene, Pe-Piper & Piper, 2005). Reconstructions of fault systems in the northern margin of the eastern Mediterranean Sea are consistent with estimates of 15 Ma (Le Pichon et al., 2019).

However, interpretations of the Aegean seismic velocity structure, tomography, and seismicity data in the Aegean area suggest older estimates (26-40 Ma; Meulenkamp et al., 1988;
Spakman et al., 1988; Papadopoulos, 1997; Brun & Sokoutis, 2010). These ages are more consistent with the ages of granitic intrusions found throughout Western Anatolia (Figure 4, Tables 5-8) and the timing of the onset of sedimentation associated with the Mediterranean Ridge at 23.6-33 Ma (Fytikas et al., 1984; Kastens, 1991). Younger estimates from the ridge are also reported (~19 Ma, Kopf et al., 2003). Plate reconstructions suggest that the Northern Hellenic trench experienced the onset of subduction from 27-34 Ma, whereas the southern Hellenic segment was active at 34 Ma (Royden & Papanikolaou, 2011).

If the incoming lithosphere is heterogeneous in terms of thickness and compositions, subduction zones may behave chaotically, in that they may, over time, retreat, advance, or remain stationary at different stages (e.g., Royden & Husson 2009; Husson et al., 2009). The progressive deceleration in motion of Africa with respect to Europe in the Mediterranean region is observed to have occurred since 35 Ma, and in the eastern Mediterranean from 35 Ma to 10 Ma to a convergence rate of a few mm/yr (Savostin et al., 1986; Marsellos et al., 2010). The rate of trench retreat is estimated to have accelerated from ~0.6 cm/y during the first 30 M.y. of subduction to 3.2 cm/yr during the past 15 m.y., perhaps due to Middle Miocene-Pliocene slab tear (Brun et al., 2017). Differences in the timing of initiation and rate of subduction exist between segments along the Western Hellenic Arc and should also be expected to occur along other portions of the Hellenic and Cyprus arcs (Royden & Papanikolaou, 2011; Pearce et al., 2012). The timing of interpreted ductile ‘extensional’ shear fabrics in metamorphic rocks can also be complicated as these may record extrusion instead of processes associated with slab rollback (see Searle and Lamont, 2020b).

These Late Cenozoic estimates are difficult to reconcile with the model in which the Hellenic arc is a single, evolving subduction zone system that initiated in the Mesozoic (Jurassic) (Facenna et al., 2003; van Hinsbergen, 2005; Royden & Papanikolaou, 2011; Jolivet et al., 2013; Malandri et al., 2017). In this scenario, the Vardar suture in Greece, equivalent to the IAESZ (Channell & Kozur, 1997; Okay & Tuysuz, 1999; Moix et al., 2008), and Pindos suture zone, equivalent to units within the Antalya domain and Dilek peninsula (Stampfli & Kozur, 2006) had buoyant microcontinents that entered and locked subduction, triggering southward slab rollback and migration of the volcanic arc (van Hinsbergen et al. 2005; Brun & Facenna 2008; Jolivet & Brun 2010; Jolivet et al., 2013; Cornée et al., 2018). The model eliminates the need for multiple sutures and subducted slabs to be present beneath western Turkey and the Aegean and simplifies the evolution of the Aegean microplate to a single evolving, long-lived subduction system. The present-day curvature of the Hellenic forearc thus represents oblique subduction and a plate-boundary expression that grew systematically over long periods of geological time (Huchon et al., 1982; Le Pichon et al., 1995; ten Veen & Kleinspehn, 2003; Gautier et al. 1999; Le Pichon et al., 2002; Wallace et al., 2005, 2008; van Hinsbergen & Schmid, 2012; Philippon et al., 2014; Cornée et al., 2018).

The single subduction system requires all the lower plate continental crust to be accreted into the upper plate while subducting continental lithosphere and requires the entire Aegean Crust from the Vardar suture to the Mediterranean ridge was derived from the lower plate (e.g., Figure 2 in van Hinsbergen et al. 2005). Oceans between the accreted domains were of significant size (500 km in some cases), and the process would lead to significant elevation changes, crustal thicknesses, and critical changes in the zone of subduction transitions occurred from oceanic to continental shear zones (see discussion in Le Pichon et al., 2019). Not all units
record blueschist facies conditions, and some experienced Barrovian prograde (burial) P-T paths, such as on the island of Naxos (e.g., Lamont et al., 2019).

Currently, the Hellenic arc is migrating SW faster than the counterclockwise rotation of Anatolia (ten Veen & Kleinspehn, 2003), and the rate of convergence between Africa and Eurasia is 4 cm/yr (Reilinger et al., 1997; Kahle et al., 2000; McClusky et al., 2000; Hollenstein et al., 2008). Timing constraints on Aegean forearc curvature, due to opposite rotations, clockwise in the west and counterclockwise in the east, are Eocene and Middle Miocene (Morris & Robertson 1993; Cornée et al., 2018). Trench bending and rollback increased subduction obliquity over time, which has been accommodated by strain partitioning within the upper Eurasian plate (Philippon et al. 2014; Brun et al. 2016; Cornée et al., 2018). Subduction zones with limited trench-parallel lengths on the order of the Hellenic arc (600-800 km) and narrow slabs (<1,500 km) typically have rapid retreat rates (Schellart et al., 2007; Bolhar et al., 2010).

3.1.3 The number, location, and impacts of slab detachments and tears

An additional key focus of study has been identifying the location, depth, and relationship of ancient and present-day active subducting slabs and their detachment mechanisms beneath the Aegean and Anatolian microplates (see review in Hansen et al., 2019; El-Sharkawy et al., 2020). Several locations across the Aegean and Anatolian microplates have been suggested to be affected by slab tear, either trench parallel or perpendicular (Figure 1). The tearing process in the near term can lead to intermediate-depth seismicity (e.g., Meighan et al., 2013) and explain earthquakes that appear inconsistent with a coherent subducting slab (e.g., Clark et al., 2008). Tears can lead to large volume magmatism (e.g., Cocchi et al., 2017), changes in igneous geochemistry, and facilitate the ore-forming process and mineral deposits (e.g., de Boorder et al., 1998; Rabayrol et al., 2019; Rabayrol and Hart, 2021). The process leads to asthenosphere upwelling and changes in thermal and fluid regimes (e.g., Roche et al., 2018; Gessner et al., 2018). Slab tear has been related to present and past geothermal activity in Western Anatolia and the generation of a late Eocene-Miocene metallogenic period (Pb-Zn- followed by Au-rich) (Menant et al., 2018; Gessner et al., 2018; Rabayrol & Hart, 2021). Their presence significantly affects plate dynamics, including subduction rates, plate motion, and mantle dynamics (e.g., Gianni et al., 2019).

These sites vary in scale from regional to local and include the boundary between the Hellenic and Cyprus arcs (Wortel & Spakman, 1992; Biryol et al., 2011), at the Anaximander Mountains (Woodside et al., 1992), south of Crete at the Pliny–Strabo Shear Zone (Özbakır et al., 2013), the İBTZ transfer zone (e.g., Kaya, 1981; Gessner et al., 2013), and beneath the Menderes Massif itself (Biryol et al., 2011; Rabayrol & Hart, 2021). A tear is speculated to generate a ~200 km-depth low-velocity anomaly below western Turkey (Roche et al., 2019). Slab tear has been used to interpret the deep Rhodes Basin (Faccenna et al., 2014; Woodside et al., 2000) and tectonic activity within southwest Anatolia (Biryol et al., 2011; Roche et al., 2019).

Trench-parallel tear affects the subducting African lithosphere between northern Greece and the Gulf of Corinth along the Western Hellenic Arc (Hansen et al., 2019). Trench-perpendicular tear may accommodate the region between the Hellenic and Cyprian arcs, which differ in subduction steepness and material subducted (Dilek & Sandvol, 2009). The Cyprian arc involves shallower subduction dynamics with the Eratosthenes seamount and Anixamander Mountains (mud volcanoes; Lykousis et al., 2009) impinging on the trench (Figure 9) (Kempler
The back thrusts and tectonic geometry of the Mediterranean Ridge has led to speculation that the African slab detached in the region between Libya and Crete (Kopf et al., 2003). Alternatively, a Subduction Transform Edge Propagator (STEP, a tear fault or a hinge fault, Govers & Wortel, 2005) may exist in this region (Özbakır et al., 2013). Nine of these structures have been proposed to exist beneath southern Greece, segmenting the subducting African slab and contributing to seismicity and deformation (Sachpazi et al., 2016). A STEP is also proposed for the transition between the Cyprus and Hellenic arcs (e.g., Salaün et al., 2012; Elitez et al., 2016; Portner et al., 2018).

The KTZ (Figure 1 and Figure 9) has been a particular subject of the debate regarding slab tear (see Bocchini et al., 2018; Hansen et al., 2019). The structure is part of the Western Hellenic Subduction Zone, considered one of the most seismically active areas in Europe (Pearce et al., 2012; Halpaap et al., 2018). The KTZ may represent a vertical tear along oceanic and continental lithosphere (Suckale et al., 2009), forming the KTZ as a STEP-fault (Govers & Wortel, 2005). The STEP fault may be in its initial stages of forming (Evangelidis, 2017; Özbakır et al., 2020), or the slab may have entirely detached (Wortel & Spakman, 2000). A smooth transition has also been proposed between two segments, without the presence of a tear between, at least at depths shallower than 100 km (Pearce et al., 2012; Halpaap et al., 2018).

Despite the fragmentation of the subducting African lithosphere, the thickness of the Aegean and Anatolian crust is remarkably similar (Zhu et al., 2005; Karabulut et al., 2019). Estimates from the central Menderes Massif are 28–30 km (Zhu et al., 2005), whereas the thickness beneath the Aegean Sea averages ~25 km (Zhu et al., 2005; Tirel et al., 2004; Kind et al., 2015). The crustal thickness in the southern and central parts of the Aegean is reported to be thinner (20–22 km), whereas the northern Aegean Sea shows a relatively thicker crust (25–28 km) (Karagianni et al., 2005; Sodoudi et al., 2006). Depending on the model used, the crustal thickness beneath western Crete could be 32.5–35 km or up to 45 km (Snopek et al., 2007). Karabulut et al. (2019) demonstrates large crustal thickness variations (20–47 km) from western Greece to eastern Anatolia but shows that these are fairly uniform within specific regions. In Western Anatolia, the crustal thicknesses are 25–30 km, increasing slightly to the north, whereas in southern Anatolia, crustal thicknesses decrease from 35 to 25 km in the Mediterranean Sea, except north of Antalya Bay, where the thickness locally reaches 40 km. A thickness of 40 km is in line with estimates of Eastern Anatolia (Kind et al., 2015), western Greece (Karagianni et al., 2005), and the Anatolian plateau (Saunders et al., 1998).

These thickness estimates seem at odds with large-scale back-arc thinning typically seen in subduction zone settings (e.g., Saunders & Tarney 1984). The Aegean is not a typical back-arc basin (Agostini et al., 2010; Doglioni et al., 2002) because it is underlain by a thick layer of continental crust and lacks an ocean floor (e.g., Makris, 1978), is disrupted by the active North Anatolian Shear Zone (NASZ) in its northern portion (e.g., Brooks & Ferentinos, 1980; Gürer et al., 2006; Kokkalas et al., 2006; Kreemer et al., 2004; Lyberis, 1984). The region displays a complex tensional regime where crustal stretching is inconsistent with the geometry and direction of the subducting Hellenic slab (e.g., Mantovani et al., 1997; Agostini et al., 2010). The premise of extrusion tectonics driven by convergence in the west requires a free lateral boundary in the east. However, the Aegean plate is constructed mainly of continental lithosphere and has a similar thickness as the Anatolian plate, as seen in both bathymetry (Figure 9) and seismic reflection (e.g., Zhu et al., 2005; Sodoudi et al., 2006). However, slab ruptures associated with the differential retreat, inherited lower plate lithospheric heterogeneities, and mantle upwelling...
would provide accommodation for the microplates to extrude (Agostini et al., 2010; Govers & Fichtner et al., 2016; Karabulut et al., 2019). The onset of the NASZ may be the result of slab deformation and detachment beneath the Bitlis–Hellenic subduction zone, which accelerated slab retreat in the west and indentation of the continent along the Bitlis–Zagros suture zone (Figure 1) (Faccenna et al., 2006; Schildgen et al., 2014).

### 3.2 Timing, number, and geometry of transfer zones

Transfer zones play a significant role in accommodating tectonic escape (Barbot & Weiss, 2021), and despite their importance in accommodating the present-day subduction dynamics, when, how, and why specific transfer zones occur across Western Anatolia is debated. For example, the İBTZ is a deep crustal transform fault zone consisting of NE-trending active strike-slip dominated faults and accommodates differential deformation between the Cycladic and Menderes core complexes (Uzel et al., 2013; 2020). The İBTZ is also mapped as the Western Anatolian Transfer Zone (WATZ, Gessner et al., 2013; 2017). The zone may be the surface expression of a tear in the subducting African slab (Gessner et al., 2013; Uzel et al., 2015; Sümer et al., 2018) or a transition between extensional and strike-slip dynamics due to the southward rotation rollback of the subduction zone (Ersoy & Palmer, 2013; Özkaymak et al., 2013; Ersoy et al., 2014; Ersoy et al., 2017; Uzel et al., 2020). Based on a compilation of data from igneous rocks throughout Western Anatolia, Uzel et al. (2020) suggest that volcanic activity in the region is always associated with the İBTZ as recorded by the positions of the eruption centers that follow the trend of the transfer zone. A lack of $^{40}$Ar/$^{39}$Ar ages from igneous assemblages between 15.97 and 13.82 Ma is attributed to a pulse of core complex exhumation and a change in partitioning extension between the Cyclades and Menderes Massif. Geochemical compositions of Miocene-age (17.48–14.94 Ma) volcanoes within the transfer zone indicate their origins are decompression melting of the upper mantle/lower crust, consistent with the outcome of regional transtensional movements in a post-collisional setting (Seghedi et al., 2015). Slab-tear typically results in asthenosphere-derived (Ocean-Island Basalt, OIB-like) Na-alkaline basalts, which are only exposed in the region within the northern Menderes Massif (Kula volcanics) (Holness & Bunbry, 2006; Ersoy et al., 2017).

The İBTZ may trend further south into the MCL, an extensional fault exposed near or on the island of Paros that records orogen-parallel extension or transform fault motion (Figure 11) (Morris & Anderson, 1996; Avigad et al., 1998; Walcott & White, 1998; Pe-Piper et al., 2002; Tirel et al., 2009; Gessner et al., 2013; Philippin et al., 2014; Beniest et al., 2016; Malandri et al., 2017). Besides the İBTZ, the SWASZ and UMTZ are located near each other on the border of the Menderes Massif, but their influence on each other is presently unclear (Figure 2).

### 3.3 Magmatic influence in driving extension

Throughout Western Anatolia, magmatic pulses are exposed as geochemically variable extrusive and intrusive igneous rocks (Tables 1-9; Figure 4; e.g., Rossetti et al., 2017). Metamorphic core complexes with their associated post-collisional magmatic suites offer insights into the tectonic processes controlling crustal extension (e.g., Perkins et al., 2018). Extensional systems cut igneous intrusions in Western Anatolia metamorphic core complexes, and their ages are critical for timing events that facilitated their emplacement. Geochemical data regarding the depths of granite formation lends additional insight into how the mantle processes operated in the past. The picture, however, is complicated by the influence of the collisional dynamics that characterized the earlier assembly of the microplate (see Assembly section).
Granite crystallization ages provide information regarding how extension during the Eocene to Miocene migrated through Western Anatolia and the Aegean region in the past (e.g., Delaloye & Bingöl, 2000; Pe-Piper, 2000; Altunkaynak & Dilek, 2006; Altunkaynak et al., 2012).

Magma bodies can drive extension through the conductive transfer of heat from upwelling of hot, asthenospheric mantle beneath significantly extended crust, and small volume partial melts can exploit crustal pathways developed during extensional deformation (e.g., McKenzie & Bickle 1988; von Blanckenburg & Davies 1995; Perkins et al., 2018). Volatiles facilitate additional crustal deformation and metamorphism, resulting in feedbacks between decompression and mantle upwelling and driving additional lithospheric melting (Teyssier & Whitney, 2002; Kendall et al., 2005; Whitney et al., 2013; Platt et al., 2015; Perkins et al., 2018).

The Menderes Massif of western Turkey is suggested to be a key area to study feedback relationships between magma generation/emplacement, rheological weakening, activation of extensional detachment tectonics (Rossetti et al., 2017). The island of Naxos likewise illustrates the interplays between lower crustal flow and upper crustal extension and between buoyancy- and isostasy-driven controls in developing migmatite domes (Kruckenberg et al., 2011). The connections between detachment faulting and magma emplacement have also been explored in the Cyclades (e.g., Rabillard et al., 2018).

To determine the role between magma generation and extension requires understanding intrusive rock relationships to fault structures. In Western Anatolia, maps of the same pluton are commonly inconsistent in terms of the locations of structures that may have affected or result from exhumation. For example, the northern boundary of the Kozak pluton (Figure 4) is shown by some as an intrusive contact (Akal & Helvacı, 1999) but by others as fault-bounded (Altunkaynak & Yılmaz, 1998; 1999; Yılmaz et al., 2001). The Eğrigöz, Koyunoba, and Alaçam plutons (Figure 4) have been the focus of many field-based, geochemical and geochronological studies, but conflicting ideas exist regarding their relationship to the SDF (Figure 7) (see Catlos et al., 2012). For example, Işık and Tekeli (2001) map the SDF only along the northern portion of the Eğrigöz pluton, whereas Ring and Collins (2005) and Işık et al. (2004) indicate the SDF is exposed along the western edge of both the northern Eğrigöz and Koyonba plutons. Seyitoğlu et al. (2004) place the SDF within the central portion of the Eğrigöz pluton, whereas Ersoy et al. (2010) mark the structure as following the outer boundaries of the Eğrigöz and Koyunoba bodies. Thomson and Ring (2006) place the detachment prominently along the northern edge and central portion of the Eğrigöz granite and along the eastern edge of the Koyunoba body. Recent gravity measurements suggest an igneous intrusion at depth near the Simav Fault (Toker et al., 2018, 2019). The 12-15 km-thick intrusion is located in the NE margin of the Simav graben at 2.5-3 km below the surface and has been suggested to be a primary driver of recent-day seismicity. Developing links between magmatism and extensional dynamics requires a critical structural understanding of the granite petrology, structures, and clear delineation between how it appears affected by fault systems (e.g., Kruckenberg et al., 2011; Rabillard et al., 2018).

In Western Anatolia, many published maps also do not distinguish different granite types or textural orientations (Karacik & Yılmaz, 1998; Akal & Helvacı, 1999; Şahin et al., 2010). Mineral lineations and solid-state or magmatic fabrics associated with faulting or shearing are rarely reported. Besides the standard structural and petrographic analyses, cathodoluminescence (CL) images of extensional-related Western Anatolia granites (Salihli and Turgutlu, Catlos et al., 2010; Eğrigöz, Koyunoba, and Alaçam, Catlos et al., 2012; Figure 4) help document mineral zoning, deformation, and fluid alteration (e.g., Ramseyer et al., 1992; Catlos et al., 2016).
Western Anatolia granites share many similar microtextural characteristics in CL, including evidence for fluid interactions and multiple generations of microcracks. The samples show secondary alteration textures, mineral growth generations, and evidence for fluid migration. The generations of microfractures, microcracks, and microfaults seen in CL document that these granites experienced brittle deformation multiple times, both at depth and at lower temperatures near the surface (Catlos et al., 2010; 2012). CL imagery is a powerful tool for identifying mineral textural relationships, growth histories, and deformation structures of Western Anatolia granite assemblages.

3.4 Timing the switches in the stress regimes in Western Anatolia

The Simav Fault system illustrates another outstanding question regarding deciphering stress regimes within Western Anatolia (Figure 7). On 19 May 2011, a magnitude 5.7 (Mww, USGS and Turkish Ministry of the Interior, Disaster and Emergency Management Presidency, Earthquake Department, AFAD) earthquake occurred near the town of Simav. The epicenter was located ~53 km NNW of Uşak and ~82 km WSW from Kütahya in western Turkey at 20:15:23.4 GMT. The estimated depth of the earthquake varies (Doğangün et al., 2013). Table 9 reports the 24.46 km result from AFAD, although the USGS Earthquake Catalog suggests a shallower 7.0 km depth. Görgün (2014) estimated a best-fit hypocenter depth of 10 km and 6.0 magnitude (Mw). Karasözen et al. (2016) indicate that the centroid depth was 7–9 km, but the hypocenters of the mainshock and largest aftershocks were located systematically deeper at depths of 10–22 km. In approximately the same location, an Mw ~5.1 event preceded the mainshock on 17 February 2009, and an Mw 4.4 foreshock occurred 15 min before the mainshock (e.g., Karasözen et al., 2016).

The Simav region is considered to be one of the most seismically active portions of Western Anatolia (Inel et al., 2013; Görgün, 2014), and the 19 May 2011 Simav (Kütahya) earthquake was the largest felt in the region since the destructive 1969 Demirci and 1970 Gediz earthquake sequences (e.g., İlhan, 1971; Ambroseys & Tchalenko, 1972; Eyidoğan & Jackson, 1985). All of these earthquakes involved dominant normal faulting with nucleation zones from 6-10 km depth and dips of 30-50° (Eyidoğan & Jackson, 1985; Emre & Duman, 2011; Görgün, 2014; Karasözen et al., 2016). However, a strike-slip component is recorded by some of the aftershocks of the 1969 and 1970 earthquakes and the 2011 Simav event (Figure 7B) (Ambroseys & Tchalenko, 1972; Eyidoğan & Jackson, 1985; Emre & Duman, 2011). In addition, Figure 7B shows that some earthquakes in the Simav region after the 2011 event also yield fault plane solutions that include some or a significant strike-slip component.

The epicenters of these earthquakes occurred near the Simav Fault (Figure 7) (Seyitoğlu, 1997; Ersoy et al., 2010). The fault extends ~150 km between the towns of Banaz in the east and Sındırgı in the west (Ambroseys & Tchalenko, 1972; Seyitoğlu, 1997; Ersoy et al., 2010). It may be part of a larger extensional Akşehir-Simav Fault System (Koçyiğit & Deveci, 2007), which extends >250 km from the town of Akşehir in south-central Turkey and includes the Sultandağ Fault in the east (Aksari et al., 2010). Or, it may be part of the Sındırgı-Sincanlı Fault Zone (SSFZ) between the towns of Soma and Afyon (Doğan & Emre, 2006). The Simav Fault may also connect to the Muratdağ Fault near the town of Gediz in an en echelon pattern, which lends support for a right-lateral system (Ambroseys & Tchalenko, 1972). Where the Akşehir-Simav Fault System is located between the cities of Uşak and Afyon is unclear (e.g., Karasözen et al. 2016).
The Simav Fault is assigned as an active right-lateral strike-slip fault in active tectonic maps of Turkey (Şaroğlu et al., 1992; Emre et al., 2011). This sense of motion is based on offsets of metamorphic zones east of Simav (Konak, 1982; Seyitoğlu, 1997) and its relationship to the formation of the NAFF (Konak, 1982; Doğan & Emre, 2006; Emre & Duman, 2011). The strike-slip motion is also consistent with uniform (magnitude and orientation) GPS plate velocity vectors that show the region is extruding through an SW motion from 30–40 mm/yr (McClusky et al., 2000; Reilinger et al., 2006, 2010). However, the detailed analysis of the Simav fault mechanisms consistently indicates a normal mechanism (Görgün, 2014; Yolsal-Çevikbilen et al., 2014; Demirci et al., 2015; Karasözen et al., 2016; Bello et al., 2017; Mutlu, 2020). This origin is linked to subduction-related extension along the Hellenic and Cyprus arcs (e.g., Seyitoğlu, 1997; İşık et al., 2003; Ersay et al., 2010; Görgün, 2014; Yolsal-Çevikbilen et al., 2014; Demirci et al., 2015; Karasözen et al. 2016; Bello et al., 2017).

If the Simav Fault was initiated as a strike-slip system but switched to extension sometime after the Late Miocene is possible (Oygür & Erler, 2000). Strike-slip motion has also been speculated to predate subsidence currently experienced by Western Anatolia and may be related to Eocene to Oligocene compression (Oygür & Erler, 2000). Based on an analysis of the available data from the 19 May 2011 event, Görgün (2014) indicate that the hypocenter distribution is consistent with the activation of two nearly parallel faults: one northern one with a fault plane trending mainly E–W and dipping towards SE and a southern fault plane trending NW–SE and dipping towards SE. The strike-slip mechanisms are delegated to smaller fault segments that experience a stress change after the mainshock and more minor secondary faults in the region with different mechanisms. Karasözen et al. (2016) suggest the potential involvement of structures inherited from earlier deformation phases of shortening and extension in evaluating the nature of motion along the structure.

The Simav E-W trending-graben hosts one of Turkey’s most important geothermal systems (Bello et al., 2017). Based on a study of geothermal activity, soil radon gas release, and regional seismicity patterns, İnan et al. (2012) suggests that the epicentral area of the 19 May 2011 Simav earthquake is located within a block that is tectonically separated from Aegean Extensional Province and the Marmara Region. The observation is also supported by geodetic data that show a region surrounding the event behaves distinctly from the Aegean Extensional Province (Tiryakioğlu, 2011). Yolsal-Çevikbilen et al. (2014) suggest the magnitude of the stress drop associated with the 19 May 2011 event (62 bars) is more consistent with an intraplate earthquake compared to those associated with Aegean plate boundaries (3-11 bars).

3.5 Relating geological units and events across boundaries
As noted in the Geological Background section, several units and structures can be correlated from Western Anatolia to the Aegean region. For example, the Cyclades Blueschist Unit (CBU) from the southern portion of the Menderes Massif (Figure 12A) is often matched to outcrops exposed in the Cyclades (Ring et al., 1999; Roche et al., 2018; Çetinkaplan et al., 2020; Barbot and Weiss, 2021), but distinguishing structures developed during subduction-related burial and prograde metamorphism from those that formed due to decompression and retrogression is problematic (e.g., Rosenbaum et al., 2002; Xypolias et al., 2012; Çetinkaplan et al., 2020). The CBU experienced multiple phases of deformation and mineralogical transformations (e.g., Seman et al., 2017; Gerogiannis et al., 2019). Identifying local internal structures from those that would correlate as significant deformation zone poses a challenge.
Çetinkaplan et al. (2020) suggest that the contact between the Menderes Massif and the CBU, now defined by a ductile thrust fault, was originally a lithosphere-scale transform fault zone.

The timing of detachment systems in the Menderes Massif are similar to those estimated in the Cyclades. Three major Aegean microplate detachment systems include the North Cycladic Detachment on Andros, Tinos, and Mykonos (Figure 11) (e.g., Jolivet et al., 2010), the Naxos-Paros Detachment on Naxos and Paros (Buick, 1991; John & Howard, 1995; Cao et al., 2013), and the West Cycladic Detachment on Serifos (Grasemann et al., 2012). The North Cycladic Detachment may have initiated activity in the Oligocene until the Late Miocene (e.g., Jolivet et al., 2010). The Naxos-Paros Detachment records retrogression associated with its latest activity in the Late Miocene (e.g., Cao et al., 2017). These time frames are similar to constraints estimated for the activity of detachment faulting in the central Menderes Massif (Hetzel et al., 1995a, Hetzel et al., 1995b, Işık et al., 2003, Glodny & Hetzel, 2007; Catlos et al., 2010). The Cyclades Detachments cross-cut blueschist-amphibolite facies fabrics and post-date HP metamorphism and peak Barrovian metamorphism (Searle and Lamont, 2020a).

Another correlation links the lithologies, conditions, and metamorphic history of Menderes Massif nappes to those in the Cyclades (e.g., Robertson et al., 1991; Ring et al., 1999; Stampfli, 2000; Çetinkaplan et al., 2020). Menderes Massif nappes have zoned garnets useful for generating P-T conditions and paths (e.g., Figure 12B and Figure 13). These paths are often developed by connecting peak metamorphic conditions of individual rocks, inferences from mineral assemblages, pseudosections, or Gibbs method thermodynamic modeling (e.g., Ashworth & Evirgen, 1984; 1985a,b; Ring et al., 2001; Whitney & Bozkurt, 2002; Çenki-Tok et al., 2016; Etzel et al., 2019; 2020). Despite these studies, the number and timing of garnet-growth events recorded in the rocks remain unclear. Some Çine nappe rocks experienced two stages of garnet growth (Ring et al., 2001), whereas other samples are consistent with one episode (Régnier et al., 2007). Pan-African garnet growth is recorded in the Menderes Massif, and conditions could reflect events unrelated to MMM (Ring et al., 2004; Catlos & Çemen, 2005). Gessner et al. (2001) report that the Bayındır nappe deformed once during the Eocene related to MMM, whereas the Bozdag, Çine, and Selimiye nappes record pre-MMM and MMM events. This contradicts Oberhaensli et al. (1997), who suggest the cover sequence records deformation during the Eocene, but structurally lower units record pre-MMM events. Studies of Bozdag nappe rocks show prograde burial, but conditions decrease downward by ~40°C/kbar per km of structural section (inverted metamorphism, Ring et al., 2001). Selimiye nappe rocks record exhumation and retrogression (Régnier et al., 2007). Paths in Figure 12B were generated by connecting peak metamorphic conditions of individual rocks, inferring from mineral assemblages, pseudosections, or Gibbs method thermodynamic modeling. P-T paths that decrease in pressure or temperature suggest the potential for tectonic switching as unloading and refrigeration occur when the thrust reverses and experiences extension.

Challenges for generating P-T conditions and paths include a prior garnet-producing history and retrograde fluid-induced alteration and overprinting as the core complex formed (e.g., Satir & Taubald, 2001; Régnier et al., 2003; Catlos & Çemen, 2005; Baker et al., 2008; Candan et al., 2011). Menderes Massif rocks are known to yield problematic P-T estimates based on evidence of disequilibrium among phases and the application of barometers to inappropriate (uncalibrated) mineral compositions (Ashworth & Evirgen, 1984; 1985a,b). In some cases, calculated conditions appear at odds with observed mineral assemblages and structural data (Ring et al., 2001; Whitney & Bozkurt, 2002). Pressure estimates using conventional approaches
are challenging to obtain due to the lack of appropriate mineral assemblages (Iredale et al., 2013). Problems may arise if the chosen mineral compositions for thermobarometric calculations are associated with retrogression instead of the desired prograde conditions. P-T paths that only rely on core and rim measurements are also limited in their ability to test models developed regarding lithospheric response to perturbations, including motion within fault zones.

One promising avenue to address this issue is the application of isochemical phase equilibria modeling. Figure 13 shows this approach applied to garnets from the Menderes Massif’s Çine, Selimiye, and Bayindir nappe from Etzel et al. (2019) and Etzel et al. (2020) and a sample from the Northern Menderes Massif from Cenki-Tok et al. (2016). The researchers report petrological details, X-ray element maps, and geochemical data from the rocks. They compositionally analyzed micaschists with a mineral assemblage of garnet + biotite + plagioclase + muscovite + quartz + rutile ± ilmenite ± pyrite ± zircon ± monazite. The sample from the Northern Menderes Massif contains kyanite and small porphyroblasts of staurolite. Using data reported in the papers, isochemical phase diagrams were created using rock bulk compositions, the software package TheriaK-Domino (de Capitani & Brown, 1987; de Capitani & Petrakakis, 2010) with the Holland and Powell (1998; 2010) thermodynamic data set, and appropriate mixing models in the system MnO–Na2O–CaO–K2O–FeO–MgO–Al2O3–SiO2–H2O–TiO2. Isoptes of ±0.01 mole fraction spessartine, almandine, pyrope, and grossular corresponding with the garnet core composition, are plotted on the phase diagram. This portion of the diagram with intersecting isopleths approximates the chemical system at the time garnet began growth. This diagram also tests if the thermodynamic data set and mixing models used in the modeling are appropriate for these particular samples, as expected mineral assemblages appeared in the phase diagrams with intersecting isopleths.

After the garnet core conditions are estimated, a Matlab script was applied to each step along a garnet compositional traverse from core to rim to yield both an estimate of the P-T conditions of incremental growth and a new effective bulk rock composition, ultimately culminating in a high-resolution P-T path. High-resolution P-T paths are defined as those derived from fractionated equilibrium phase diagram modeling and the resolution is an outcome of the number of garnet fractionated steps. Garnets with complex zoning profiles, modified by diffusion, or rocks that experienced major changes in bulk composition over their growth history are not candidates (e.g., Catlos et al., 2018). However, even these types of samples may provide clues by exploring the reason for their failure (e.g., Catlos et al., 2018; Etzel et al., 2020). Ideal samples are those with garnets that preserve prograde, gradational core-to-rim zoning profiles. Garnets from the Selimiye and Bayindir nappes of the Southern and Central Menderes Massif, respectively, show similar trajectories. However, the Çine nappe garnet yields an N-shape path and a significantly different metamorphic history.

Either tectonically-driven extension may have created the N-shaped P-T path during orogenesis or the result of erosional exhumation during pulses of thrust motion (Etzel et al., 2019). Etzel et al. (2019) developed two thermal models: erosional denudation followed by fault reactivation (Figure 14A) and tectonic switching (Figure 14B), which are briefly summarized here. Figure 14A and Figure 14B show an upper equilibrium thermal grid (depth vs. horizontal distance) before faulting with the position of fault (grey line) arbitrarily selected at 30°. Fault displacement varies linearly across shear zones. The grid includes reflecting side boundaries and top and bottom maintained at 25°C and 700°C and an initial geothermal gradient at 25°C/km indicated by shaded zones. A hatched area shows the position of the Selimiye samples, and the
grey bar represents the approximate initial location of the Çine nappe garnet with the N-shaped P-T path. This position is also represented by point 1 in the P-T path insets. In Figure 14C and Figure 14D, the fault is active. A finite-difference solution to the diffusion-advection equation is used to examine the P-T variations in the hanging wall and footwall due to its motion. The rock sample experiences the point 1 to 2 in the P-T path insets. Fault motion stops and denudation occurs in Figure 14E and, whereas extension occurs in Figure 14F. This process is based on the mid-rim lower pressure portion of the garnet P-T path and is represented by points 2 to 3 on the P-T path insets. Although the end, the surface geometry in the denudation phase (Figure 13E) and extensional phase (Figure 14F) are similar, the shape of the isotherms is different and leads to the development of a decrease in temperature in the P-T loop observed in the tectonic switching model. Finally, the fault is reactivated, represented by Figure 14G to Figure 14H. The decrease in pressure with increasing temperature is related to an episode of denudation (model 1) rather than a tectonic switch from compression to extension (Etzel et al., 2019).

The P-T paths reported in Figure 13 approximate how a garnet with specific compositional zoning would behave in a closed system of a known bulk composition as it evolves during increasing T. A critical assumption is that the minerals in a sample experienced equilibrium, which can never be proven for any rock system (e.g., Spear & Peacock, 1989; Lanari & Duesterhoeft, 2019). Closed system behavior also requires the original compositions of the mineral phases, and the bulk rock has not changed significantly since metamorphism (e.g., Lanari and Engi, 2017). Multiple sources of error are inherent, including uncertainty in the accuracy of end-member reactions, electron microprobe analyses, calibration errors, variations in activity models, compositional heterogeneity, and uncertainty associated with the thermodynamic properties inherent in the choice of internally consistent database (e.g., Kohn & Spear, 1991; White et al., 2014; Palin et al., 2016; Lanari & Duesterhoeft, 2019). Garnets with significant changes in composition over short distances from core to the rim and those affected by diffusion cannot be modeled. Garnets in samples that experienced significant changes in bulk composition or multiple deformation episodes resulting in modification of composition are also unsuitable.

A significant value of the high-resolution P-T path and isopleth approaches is that a user can detect when systems stray from the equilibrium and closed system assumptions. Confidence in paths and conditions increases when minerals assemblages agree with rock observations and if the P-T paths reproduce trends in garnet zoning. Samples collected from the same outcrop or nearby should yield similar P-T conditions and paths. In addition, a user can gauge the extent of overlapping mineral isopleths in P-T space, including if matrix mineral compositions overlap the garnet rim conditions. These paths are the first steps in developing critical insights into the metamorphic history of the assembly of the Menderes Massif and, combined with age information from the garnet itself or matrix or mineral inclusions, can be used to test models for the development of Western Anatolia.

4. Conclusions

This paper is divided into two major sections. The first outlines, as much as is possible, our present-day understanding of the geological history of Western Anatolia from its assembly through its extensional and strike-slip history. We aim to illustrate the complex tectonic scenario before the onset of large-scale extension and emphasize the present-day change in stress regime towards strike-slip tectonics. The transitions are also comparable in duration and timing to those experienced by the Aegean microplate.
The second part highlights some outstanding questions that remain to be addressed. These include issues regarding the dynamics of the African slab along the Hellenic arc, the arc's geometry and connections to other subduction systems, and reconciling the Jurassic initiation age of subduction with Late Cenozoic sedimentation, magmatic, and paleogeographic data that are consistent with younger initiation. In addition, a large number of regions of slab tear are proposed throughout the African slab, and their influence on accommodating extrusion, creating economic resources, and driving lithospheric thinning and magmatism should be explored. Other questions include investigating the influence of transfer zones in accommodating deformation and the role of magma in driving extension in Western Anatolia.

The interface between Western Anatolia and the Aegean region exemplifies tectonic transitions and how the interplay between large-scale tectonics influences smaller-scale processes. The Aegean and western Turkey contain helpful assemblages that can be exploited to time these processes that shape the lithosphere and are critical in understanding the region's hazards and mineralizations. Extracting high-resolution P-T paths from Western Anatolia garnet-bearing rocks is a promising approach to evaluate tectonic models and correlate and compare metamorphic histories of nearby assemblages and from those across long distances.

**Data Availability Statement**

Data supporting the conclusions of this paper and color figures are publically available from Texas Data Repository Dataverse (https://doi.org/10.18738/T8/ER3WQV).

**Acknowledgements**

We appreciate constructive reviews of the original manuscript from Spyros Pavlides (Aristotle University of Thessaloniki, Greece), Thomas Lamont (University of Bristol) and three anonymous reviewers. We appreciate analytical assistance with the $^{40}$Ar/$^{39}$Ar analysis from the Australia National University Argon Facility.

**References**


http://dspace.library.uu.nl/bitstream/handle/1874/19338/index.htm;jsessionid=88B6AA4941C5E76F6D65034D89D4065E0?sequence=17


https://doi.org/10.1130/0091-7613(1999)027<0267:WPOTNA>2.3.CO;2


https://doi.org/10.1007/s00410-010-0504-4


https://doi.org/10.1016/j.lithos.2009.02.010


https://doi.org/10.1016/j.tecto.2018.08.003


Işık, V., Tekeli, O., & Seyitoğlu, G. (2004). The \(^{40}\text{Ar}/^{39}\text{Ar}\) age of extensional ductile deformation and granitoid intrusion in the northern Menderes core complex: implications for the initiation of extensional tectonics in western Turkey. Journal of Asian Earth Sciences, 23, 555–566.


Kaldova, J., Leichmann, J., Babek, O., & Melichar, R. (2003). Brunovistulian terrane (Central Europe) and Istanbul zone (NW Turkey): Late Proterozoic and Paleozoic tectonostratigraphic development and paleogeography. Geologica Carpathica, 54(3), 139-152.


Menant, A., Jolivet, L., & Vrielynck, B. (2016). Kinematic reconstructions and magmatic evolution illuminating crustal and mantle dynamics of the eastern Mediterranean region since the late Cretaceous. Tectonophysics, 675, 103-140. https://doi.org/10.1016/j.tecto.2016.03.007


Pearce, F.D., Rondenay, S., Sachpazi, M., Charalampakis, M., & Royden, L. H. (2012). Seismic investigation of the transition from continental to oceanic subduction along the western


Turkey and the Cycladic Massif in the Aegean; do they really correlate? Journal of the

Cretaceous-Tertiary collisional belt; regionally symmetric cooling patterns and their relation
to extensional faults in the Anatolide Belt of western Turkey. Geological Magazine, 140(4),
421-441.

temperature paths; an example from the Menderes Nappes of western Turkey. American
Journal of Science, 301(10), 912-944.


https://doi.org/10.1144/GSL.SP.1984.017.01.02

in the Armutlu Peninsula, NW Turkey. Tectonophysics, 381(1–4), 175-209.

https://doi.org/10.1016/j.tecto.2002.06.002

and comparable units in southern Turkey by subduction-accretion processes: Implications for
the tectonic development of Tethys in the Eastern Mediterranean region. Tectonophysics,

the closure of Palaeotethys: Evidence from the Chios Melange (E Greece), the Karaburun
Melange (W Turkey), and the Teke Dere Unit (SW Turkey). Sedimentary Geology, 220(1-2),
29-59. https://doi.org/10.1016/j.sedgeo.2009.06.005

paleotectonic evolution of eastern Mediterranean Neotethys. Palaeogeography,

fragmentation beneath the Aegean/Anatolia transition zone: Insights from the tectonic and
metamorphic evolution of the Eastern Aegean region, Tectonophysics, 754, 101-129.

https://doi.org/10.1016/j.tecto.2019.01.016

(Dodecanese, Greece) and correlations between the Aegean Domain and the Menderes

ff10.1144/jgs2018-028ff.ffinsu-01795049

Roche, V., Sternai, P., Guillou-Frottier, L., Menant, A., Jolivet, L., Bouchot, V., & Gerya, T.
(2018). Emplacement of metamorphic core complexes and associated geothermal systems

https://doi.org/10.1016/j.epsl.2018.06.043

Rosenbaum, G., Avigad, D., Sánchez-Gómez, M (2002). Coaxial flattening at deep levels of
orogenic belts: evidence from blueschists and eclogites on Syros and Sifnos (Cyclades,
Greece). Journal of Structural Geology, 24(9), 1451-1462. https://doi.org/10.1016/S0191-
8141(01)00143-2.

Magmatism and crustal extension: Constraining activation of the ductile shearing along the


https://doi.org/10.1046/j.1365-246x.1998.00571.x


Sherlock, S., Kelley, S., Inger, S., Harris, N., & Okay, A. (1999). $^{40}$Ar-$^{39}$Ar and Rb-Sr geochronology of high-pressure metamorphism and exhumation history of the Tavşanlı Zone, NW Turkey. Contributions to Mineralogy and Petrology, 137, 46–58.


Table 1. Brief summary of some available ages from granitic assemblages that intrude the Istanbul-Zonguldak Zone.

<table>
<thead>
<tr>
<th>Granite</th>
<th>Locationa</th>
<th>Approach</th>
<th>Age</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Late Pan-African Granitoids or Cadomian Granitoids</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Karadere (Karabuk metagranite)</td>
<td>1</td>
<td>U-Pb zrn</td>
<td>924±4</td>
<td>Chen et al. (2002)</td>
</tr>
<tr>
<td>Karadere (Karabuk metatonalite)</td>
<td>1</td>
<td>U-Pb zrn</td>
<td>668±7</td>
<td>Chen et al. (2002)</td>
</tr>
<tr>
<td>Bolu (Tüllükiris)</td>
<td>2</td>
<td>U-Pb zrn</td>
<td>576±6</td>
<td>Ustaömer et al. (2005)</td>
</tr>
<tr>
<td>Bolu (Kapıkaya)</td>
<td>2</td>
<td>U-Pb zrn</td>
<td>565.3±1.9</td>
<td>Ustaömer et al. (2005)</td>
</tr>
<tr>
<td>Karadere (Karabuk)</td>
<td>1</td>
<td>Sm-Nd grt + wr</td>
<td>559±8</td>
<td>Chen et al. (2002)</td>
</tr>
<tr>
<td><strong>Devonian</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Bolu</td>
<td>2</td>
<td></td>
<td>389, 200</td>
<td>Ustaömer et al. (2012)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>273-255</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>229.6±4.2/2.3</td>
<td></td>
</tr>
<tr>
<td>Bolu</td>
<td>2</td>
<td>40Ar/39Ar or + hbl</td>
<td>381.1±7.1</td>
<td>Delaloye and Bingöl (2000)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>93.3±2.0</td>
<td></td>
</tr>
<tr>
<td><strong>Permo-Triassic</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Bolu (Sınınicë Group)</td>
<td>2</td>
<td>207Pb/206Pb zrn</td>
<td>262±19</td>
<td>Ustaömer et al. (2005)</td>
</tr>
<tr>
<td>Sancaktepe</td>
<td>3</td>
<td>U-Pb zrn</td>
<td>257.3±1.5</td>
<td>Aysal et al. (2018)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>253.7±1.8</td>
<td></td>
</tr>
<tr>
<td>Akyazi</td>
<td>4</td>
<td>40Ar/39Ar or + chl</td>
<td>240.4±4.9</td>
<td>Delaloye and Bingöl (2000)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>86.1±2.0</td>
<td></td>
</tr>
</tbody>
</table>

a See Figure 4 for locations of these granite bodies.

b Abbreviations after Whitney and Evans (2010), wr= whole rock.
Table 2. Brief summary of some available ages from granitic assemblages that intrude the Tavşanlı Zone.

<table>
<thead>
<tr>
<th>Granite</th>
<th>Location</th>
<th>Approach</th>
<th>Age</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Western Tavşanlı Zone: Suture Zone Granitoids</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Topuk</td>
<td>5</td>
<td>$^{40}$Ar/$^{39}$Ar bt+kfs</td>
<td>63.5±2.8</td>
<td>Delaloye and Bingöl (2000)</td>
</tr>
<tr>
<td>Orhaneli</td>
<td>6</td>
<td>$^{40}$Ar/$^{39}$Ar bt+hbl</td>
<td>57.9±1.2</td>
<td>Delaloye and Bingöl (2000)</td>
</tr>
<tr>
<td>Orhaneli</td>
<td>6</td>
<td>$^{40}$Ar/$^{39}$Ar bt+hbl</td>
<td>52.6±0.4</td>
<td>Harris et al. (1994)</td>
</tr>
<tr>
<td>Topuk</td>
<td>5</td>
<td>$^{40}$Ar/$^{39}$Ar bt+hbl</td>
<td>47.8±0.4</td>
<td>Harris et al. (1994)</td>
</tr>
<tr>
<td>Tepeldag</td>
<td>7</td>
<td>U-Pb zrn</td>
<td>44.9±0.2</td>
<td>Okay and Satir (2006)</td>
</tr>
<tr>
<td>(Gürgenyayla)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Tepeldag</td>
<td>7</td>
<td>Rb-Sr bt</td>
<td>44.7±0.4</td>
<td>Okay and Satir (2006)</td>
</tr>
<tr>
<td><strong>Eastern Tavşanlı Granitoids</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Kaymaz</td>
<td>9</td>
<td>U-Pb zrn</td>
<td>84.98±6.27</td>
<td>Gautier (1984)</td>
</tr>
<tr>
<td>Sivrihisar</td>
<td>10</td>
<td>U-Pb zrn</td>
<td>79.9±8.6</td>
<td>Shin et al. (2013)</td>
</tr>
<tr>
<td>Sankavak (Topkaya)</td>
<td>11</td>
<td>U-Pb zrn</td>
<td>65.9±3.8</td>
<td>Gautier (1984)</td>
</tr>
<tr>
<td>Sivrihisar</td>
<td>10</td>
<td>$^{40}$Ar/$^{39}$Ar bt+hbl</td>
<td>62.9±1.3</td>
<td>Delaloye and Bingöl (2000)</td>
</tr>
<tr>
<td>Karacaören</td>
<td>10</td>
<td>$^{40}$Ar/$^{39}$Ar hbl+bt</td>
<td>59.3±3.0</td>
<td>Demirbilek et al. (2018)</td>
</tr>
<tr>
<td>Tekoren granodiorite (Günyüzü)</td>
<td>10</td>
<td>$^{40}$Ar/$^{39}$Ar hbl+bt</td>
<td>57.8±2.3</td>
<td>Demirbilek et al. (2018)</td>
</tr>
<tr>
<td></td>
<td>10</td>
<td>$^{40}$Ar/$^{39}$Ar hbl+kfs</td>
<td>55.9±2.7</td>
<td>Demirbilek et al. (2018)</td>
</tr>
<tr>
<td></td>
<td>10</td>
<td>$^{40}$Ar/$^{39}$Ar kfs</td>
<td>54.0±2.1</td>
<td>Demirbilek et al. (2018)</td>
</tr>
<tr>
<td>Kaymaz</td>
<td>9</td>
<td>$^{40}$Ar/$^{39}$Ar kfs</td>
<td>52.1±2.0</td>
<td>Demirbilek et al. (2018)</td>
</tr>
<tr>
<td>Sivrihisar</td>
<td>10</td>
<td>$^{40}$Ar/$^{39}$Ar hbl</td>
<td>53.2±2.1</td>
<td>Demirbilek et al. (2018)</td>
</tr>
<tr>
<td>Kadinicik (Günyüzü)</td>
<td>10</td>
<td>$^{40}$Ar/$^{39}$Ar hbl+wr</td>
<td>52.8±2.4</td>
<td>Demirbilek et al. (2018)</td>
</tr>
<tr>
<td>Kaymaz</td>
<td>9</td>
<td>U-Pb zrn</td>
<td>44.3±4.9</td>
<td>Shin et al. (2013)</td>
</tr>
<tr>
<td>Sivrihisar</td>
<td>10</td>
<td>Rb-Sr kfs+bt</td>
<td>47.0±1.6</td>
<td>Bağcı et al. (2012)</td>
</tr>
<tr>
<td>Sivrihisar</td>
<td>10</td>
<td>$^{40}$Ar/$^{39}$Ar kfs</td>
<td>46.02±0.21</td>
<td>This study</td>
</tr>
<tr>
<td>Sivrihisar</td>
<td>10</td>
<td>Rb-Sr kfs+bt</td>
<td>40.8±3.0</td>
<td>Bağcı et al. (2012)</td>
</tr>
</tbody>
</table>

*See Figure 4 for locations of these granite bodies.*
Table 3. Brief summary of some available ages from granitic assemblages associated with rocks between the Sakarya and Istanbul Zones.

<table>
<thead>
<tr>
<th>Granite</th>
<th>Location</th>
<th>Approach</th>
<th>Age</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Şevketiye</td>
<td>Middle Eocene Magmatic Rocks (South Marmara Granitoids)</td>
<td>$^{40}\text{Ar}/^{39}\text{Ar}$ ms</td>
<td>71.9±1.8</td>
<td>Delaloye and Bingöl (2000)</td>
</tr>
<tr>
<td>İlyasdağ</td>
<td>Marmara Island</td>
<td>U-Pb zrn</td>
<td>56.7±0.8</td>
<td>Ustaömer et al. (2009)</td>
</tr>
<tr>
<td>Karabiga (Lapeski)</td>
<td></td>
<td>U-Pb xtm</td>
<td>52.7±1.9</td>
<td>Beccaletto et al. (2007)</td>
</tr>
<tr>
<td>Fistikli (Armutlu–Yalova)</td>
<td></td>
<td>$^{40}\text{Ar}/^{39}\text{Ar}$ bt+ms</td>
<td>48.2±1.0</td>
<td>Delaloye and Bingöl (2000)</td>
</tr>
<tr>
<td>Karabiga (Lapeski)</td>
<td></td>
<td>$^{40}\text{Ar}/^{39}\text{Ar}$ bt</td>
<td>45.3±0.9</td>
<td>Delaloye and Bingöl (2000)</td>
</tr>
<tr>
<td>Kapıdağ</td>
<td></td>
<td>$^{40}\text{Ar}/^{39}\text{Ar}$ hbl+bt</td>
<td>42.2±1.0</td>
<td>Delaloye and Bingöl (2000)</td>
</tr>
<tr>
<td>Avsa Island</td>
<td></td>
<td>K-Ar bt</td>
<td>40.9±1.1</td>
<td>Karacık et al. (2008)</td>
</tr>
</tbody>
</table>

See Figure 4 for locations of these granite bodies.

Abbreviations after Whitney and Evans (2010), wr= whole rock.
### Table 4. Brief summary of some available ages from granitic assemblages that intrude the Central Sakarya Zone.

<table>
<thead>
<tr>
<th>Granite Location</th>
<th>Age</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Late Pan-African Grantoids or Cadomian Granitoids</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Pamukova</td>
<td>582.0±9.1</td>
<td>Okay et al. (2008)</td>
</tr>
<tr>
<td></td>
<td>446.0±3.8</td>
<td></td>
</tr>
<tr>
<td>Gemlik</td>
<td>575.5±3.6</td>
<td>Okay et al. (2008)</td>
</tr>
<tr>
<td></td>
<td>438.9±4.5</td>
<td></td>
</tr>
<tr>
<td><strong>Silurian-Devonian</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Saricakaya</td>
<td>419±6</td>
<td>Topuz et al. (2020)</td>
</tr>
<tr>
<td></td>
<td>434±7</td>
<td></td>
</tr>
<tr>
<td></td>
<td>319±5 Ma</td>
<td></td>
</tr>
<tr>
<td><strong>Carboniferous</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Inhisar</td>
<td>348.5±6.6</td>
<td>Delaloye and Bingöl</td>
</tr>
<tr>
<td></td>
<td>213.5±4.4</td>
<td>(2000)</td>
</tr>
<tr>
<td>Gevyke</td>
<td>327±12</td>
<td>Ustaömer et al. (2016)</td>
</tr>
<tr>
<td>Söğüt granite (Saricakaya, Çaltı)</td>
<td>327.2±1.9</td>
<td>Ustaömer et al. (2012)</td>
</tr>
<tr>
<td>Söğüt granite (Saricakaya, Küplü)</td>
<td>324.3±1.3</td>
<td>Ustaömer et al. (2012)</td>
</tr>
<tr>
<td>Söğüt granite (Saricakaya, Borçak)</td>
<td>319.5±1.1</td>
<td>Ustaömer et al. (2012)</td>
</tr>
<tr>
<td>Bilecik</td>
<td>312.1±6.0</td>
<td>Delaloye and Bingöl</td>
</tr>
<tr>
<td></td>
<td>233.5±4.8</td>
<td>(2000)</td>
</tr>
<tr>
<td>** Permian**</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Söğüt granite</td>
<td>290±4.8</td>
<td>Okay et al. (2002)</td>
</tr>
<tr>
<td><strong>Jurassic to Late Cretaceous</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Pamukova</td>
<td>168.2±3.5</td>
<td>Delaloye and Bingöl</td>
</tr>
<tr>
<td></td>
<td>123.0±2.8</td>
<td>(2000)</td>
</tr>
<tr>
<td>Beypazari</td>
<td>95.4±4.2</td>
<td>Speciale et al. (2012)</td>
</tr>
<tr>
<td></td>
<td>70.5±3.4</td>
<td></td>
</tr>
<tr>
<td>Beypazari</td>
<td>80.1±1.4</td>
<td>Okay et al. (2020)</td>
</tr>
<tr>
<td></td>
<td>79.2±0.9</td>
<td></td>
</tr>
<tr>
<td>Beypazari</td>
<td>74.8±0.4</td>
<td>Okay et al. (2020)</td>
</tr>
<tr>
<td></td>
<td>73.2±1.4</td>
<td></td>
</tr>
<tr>
<td>Beypazari</td>
<td>82.9±1.8</td>
<td>Delaloye and Bingöl</td>
</tr>
<tr>
<td></td>
<td>77.7±4.5</td>
<td>(2000)</td>
</tr>
</tbody>
</table>

*a See Figure 4 for locations of these granite bodies.

*b Abbreviations after Whitney and Evans (2010), wr= whole rock.
Table 5. Brief summary of some available ages from granitic assemblages that intrude the Western Pontides Zone.

<table>
<thead>
<tr>
<th>Granite</th>
<th>Location</th>
<th>Approach</th>
<th>Age</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Proterozoic</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Karacabey (Tamsali)</td>
<td>23</td>
<td>U-Pb zrn</td>
<td>1961.9±16.4</td>
<td>Aysal et al. (2012)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>(inherited cores)</td>
<td>804±10.5</td>
<td></td>
</tr>
<tr>
<td>Evciler (Kazdağ)</td>
<td>24</td>
<td>U-Pb zrn</td>
<td>805, 286</td>
<td>Ustaomer et al. (2012)</td>
</tr>
<tr>
<td>Karaburun</td>
<td>25</td>
<td>U-Pb zrn</td>
<td>1800, 960, 380, 297</td>
<td>Ustaomer et al. (2012)</td>
</tr>
<tr>
<td><strong>Devonian</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Güveylerobası</td>
<td>26</td>
<td>U-Pb zrn</td>
<td>401.5±4.8</td>
<td>Aysal et al. (2012)</td>
</tr>
<tr>
<td>(Çamlık-related)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Karacabey (Tamsali)</td>
<td>23</td>
<td>U-Pb zrn</td>
<td>400.3±1.4</td>
<td>Aysal et al. (2012)</td>
</tr>
<tr>
<td>Eybek (Çamlık)</td>
<td>27</td>
<td>U-Pb zrn</td>
<td>397.5±1.4</td>
<td>Okay et al. (2006)</td>
</tr>
<tr>
<td>Karacabey (Tamsali)</td>
<td>23</td>
<td>Pb-Pb zrn</td>
<td>395.9±4.1</td>
<td>Sunal (2012)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>393.8±2.7</td>
<td></td>
</tr>
<tr>
<td>Güveylerobası</td>
<td>26</td>
<td>U-Pb zrn</td>
<td>371.2±2.3</td>
<td>Ustaömer et al. (2016)</td>
</tr>
<tr>
<td><strong>Permo-Triassic</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Karacabey (Tamsali)</td>
<td>23</td>
<td>40Ar/39Ar bt</td>
<td>298.3±5.8</td>
<td>Delaloye and Bingöl (2000)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>199.4±4.0</td>
<td></td>
</tr>
<tr>
<td>Karacabey (Tamsali)</td>
<td>23</td>
<td>40Ar/39Ar bt</td>
<td>304.5±3.7</td>
<td>Sunal (2012)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>223.0±7.5</td>
<td></td>
</tr>
<tr>
<td>Kozak</td>
<td>28</td>
<td>U-Pb zrn</td>
<td>280.2±18.2</td>
<td>Black et al. (2013)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>259.1±13.8</td>
<td></td>
</tr>
<tr>
<td>Karaburun</td>
<td>25</td>
<td>U-Pb zrn</td>
<td>244.4±1.5</td>
<td>Ustaomer et al. (2012)</td>
</tr>
<tr>
<td>Evciler</td>
<td>24</td>
<td>U-Pb zrn</td>
<td>229.6±0.6</td>
<td>Ustaomer et al. (2012)</td>
</tr>
<tr>
<td>Karacabey (Tamsali)</td>
<td>23</td>
<td>(U/Th)-He zrn</td>
<td>93.0±6.9</td>
<td>Sunal (2012)</td>
</tr>
<tr>
<td><strong>Late Eocene-Oligocene-Miocene</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Kozak</td>
<td>28</td>
<td>40Ar/39Ar or +bt</td>
<td>37.6±3.3</td>
<td>Delaloye and Bingöl (2000)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>19.5±0.4</td>
<td></td>
</tr>
<tr>
<td>Kozak</td>
<td>28</td>
<td>U-Pb zrn</td>
<td>36.5±6.6</td>
<td>Black et al. (2013)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>17.1±0.7</td>
<td></td>
</tr>
<tr>
<td>Evciler (Kazdağ)</td>
<td>24</td>
<td>40Ar/39Ar chl+bt</td>
<td>36.0±1.4</td>
<td>Delaloye and Bingöl (2000)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>26.4±0.6</td>
<td></td>
</tr>
<tr>
<td>Evciler (Kazdağ)</td>
<td>24</td>
<td>U-Pb zrn</td>
<td>24.8±4.6</td>
<td>Erdoğan et al. (2013)</td>
</tr>
<tr>
<td>Evciler (Kazdağ)</td>
<td>24</td>
<td>207Pb-206Pb zrn</td>
<td>28.2±4.1</td>
<td>Erdoğan et al. (2013)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>26.0±5.6</td>
<td></td>
</tr>
<tr>
<td>Uludağ</td>
<td>29</td>
<td>U-Pb zrn</td>
<td>34.7±0.34</td>
<td>Topuz and Okay (2017)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>28.2±0.39</td>
<td></td>
</tr>
<tr>
<td>Eybek</td>
<td>27</td>
<td>U-Pb zrn</td>
<td>32.5±3.0</td>
<td>Black et al. (2013)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>21.0±1.2</td>
<td></td>
</tr>
<tr>
<td>Katrandag</td>
<td>30</td>
<td>40Ar/39Ar hbl+chl</td>
<td>27.6±0.6</td>
<td>Delaloye and Bingöl (2000)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>24.7±0.6</td>
<td></td>
</tr>
<tr>
<td>Location</td>
<td>Sample Code</td>
<td>Technique</td>
<td>40Ar/39Ar Age 1 (Ma)</td>
<td>40Ar/39Ar Age 2 (Ma)</td>
</tr>
<tr>
<td>------------------------</td>
<td>-------------</td>
<td>-----------</td>
<td>---------------------</td>
<td>---------------------</td>
</tr>
<tr>
<td>Uludag</td>
<td>29</td>
<td>40Ar/39Ar bt</td>
<td>26.8±0.8</td>
<td>24.7±0.7</td>
</tr>
<tr>
<td>Eybek</td>
<td>27</td>
<td>40Ar/39Ar bt</td>
<td>26.6±0.8</td>
<td>21.1±0.4</td>
</tr>
<tr>
<td>Cataldag (Bozenkoy)</td>
<td>31</td>
<td>K-Ar bt+hbl</td>
<td>25.9±0.5</td>
<td>21.27±0.44</td>
</tr>
<tr>
<td>Evciler (Kazdağ)</td>
<td>24</td>
<td>Rb-Sr</td>
<td>25.0±0.3</td>
<td></td>
</tr>
<tr>
<td>Kozak</td>
<td>28</td>
<td>K-Ar bt+hbl</td>
<td>23.0±3.8</td>
<td>14.6±1.0</td>
</tr>
<tr>
<td>Cataldag (Cataltepe)</td>
<td>31</td>
<td>K-Ar bt</td>
<td>22.0±0.3</td>
<td>21.7±0.1</td>
</tr>
<tr>
<td>Cataldag (Turfaldag)</td>
<td>31</td>
<td>K-Ar bt</td>
<td>21.9±0.6</td>
<td>21.2±0.6</td>
</tr>
<tr>
<td>Cataldag (Balicikhisar)</td>
<td>31</td>
<td>40Ar/39Ar bt</td>
<td>20.8±0.4</td>
<td></td>
</tr>
<tr>
<td>Evciler (Kazdağ)</td>
<td>24</td>
<td>Rb-Sr</td>
<td>20.7±0.2</td>
<td>20.5±0.2</td>
</tr>
</tbody>
</table>

**Younger South Marmara Granitoid Bodies**

<table>
<thead>
<tr>
<th>Location</th>
<th>Sample Code</th>
<th>Technique</th>
<th>40Ar/39Ar Age 1 (Ma)</th>
<th>40Ar/39Ar Age 2 (Ma)</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Yenice</td>
<td>32</td>
<td>40Ar/39Ar hbl</td>
<td>47.6±1.4</td>
<td>20.1±1.1</td>
<td>Delaloye and Bingöl (2000)</td>
</tr>
<tr>
<td>Ilica</td>
<td>33</td>
<td>K-Ar hbl</td>
<td>37.9±0.1</td>
<td>25.6±1.9</td>
<td>Boztuğ et al. (2009)</td>
</tr>
<tr>
<td>Kizildam</td>
<td>34</td>
<td>K-Ar wr+bt</td>
<td>23.9±0.6</td>
<td>20.7±0.8</td>
<td>Karacık et al. (2008)</td>
</tr>
<tr>
<td>Danisment</td>
<td>35</td>
<td>K-Ar wr+bt</td>
<td>23.2±1.1</td>
<td>22.1±0.6</td>
<td>Karacık et al. (2008)</td>
</tr>
<tr>
<td>Ilica</td>
<td>33</td>
<td>K-Ar wr+bt</td>
<td>22.8±0.5</td>
<td>18.4±2.2</td>
<td>Karacık et al. (2008)</td>
</tr>
<tr>
<td>Sarioluk</td>
<td>36</td>
<td>K-Ar hbl</td>
<td>22.6±0.8</td>
<td></td>
<td>Karacık et al. (2008)</td>
</tr>
<tr>
<td>Yenice</td>
<td>32</td>
<td>K-Ar wr+bt</td>
<td>21.9±1.1</td>
<td>18.8±1.3</td>
<td>Karacık et al. (2008)</td>
</tr>
<tr>
<td>Davutlar</td>
<td>37</td>
<td>K-Ar wr+bt</td>
<td>21.6±0.6</td>
<td>18.4±1.1</td>
<td>Karacık et al. (2008)</td>
</tr>
<tr>
<td>Yeniköy</td>
<td>36</td>
<td>K-Ar wr</td>
<td>20.1±1.0</td>
<td></td>
<td>Karacık et al. (2008)</td>
</tr>
</tbody>
</table>

---

*See Figure 4 for locations of these granite bodies.*

*Abbreviations after Whitney and Evans (2010), wr= whole rock.*
Table 6. Brief summary of some available ages from granitic assemblages that intrude the Rhodope-Strandja Zone (Biga Peninsula only).

<table>
<thead>
<tr>
<th>Granite</th>
<th>Location</th>
<th>Approach</th>
<th>Age</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Kuscayir</td>
<td>38</td>
<td>(^{40})Ar(^{39})Ar hbl</td>
<td>39.4±0.8, 35.7±0.8</td>
<td>Delaloye and Bingöl (2000)</td>
</tr>
<tr>
<td>Kestanbol (Ezine)</td>
<td>39</td>
<td>U-Pb zrn</td>
<td>26.2±2.0, 18.8±1.0</td>
<td>Black et al. (2013)</td>
</tr>
<tr>
<td>Kestanbol (Ezine)</td>
<td>39</td>
<td>(^{40})Ar(^{39})Ar</td>
<td>22.21±0.07, 21.22±0.09</td>
<td>Akal (2013)</td>
</tr>
<tr>
<td>Kestanbol (Ezine)</td>
<td>39</td>
<td>(^{40})Ar(^{39})Ar hbl</td>
<td>20.5±0.6</td>
<td>Delaloye and Bingöl (2000)</td>
</tr>
</tbody>
</table>

\(^a\) See Figure 4 for locations of these granite bodies.

\(^b\) Abbreviations after Whitney and Evans (2010), wr= whole rock.
Table 7. Brief summary of some available ages from granitic assemblages that intrude the Afyon Zone.

<table>
<thead>
<tr>
<th>Granite</th>
<th>Location</th>
<th>Approach</th>
<th>Age</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Paleozoic Granitoids</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Sandıklı</td>
<td>39</td>
<td>U-Pb zrn</td>
<td>541±9</td>
<td>Gürsu et al. (2004)</td>
</tr>
<tr>
<td>Alaçam (basement)</td>
<td>41</td>
<td>U-Pb zrn</td>
<td>331.3±1.7</td>
<td>Candan et al. (2016)</td>
</tr>
<tr>
<td>Alaçam (basement)</td>
<td>41</td>
<td>U-Pb zrn</td>
<td>314.3±4.8</td>
<td>Hasözbek et al. (2010)</td>
</tr>
<tr>
<td><strong>Late Eocene-Oligocene-Miocene</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Balkan (Muratdag)</td>
<td>40</td>
<td>40Ar/39Ar or</td>
<td>35.5±3.0</td>
<td>Delaloye and Bingöl (2000)</td>
</tr>
<tr>
<td>Koyunoba</td>
<td>42</td>
<td>U-Pb zrn</td>
<td>30.0±3.9</td>
<td>Catlos et al. (2012)</td>
</tr>
<tr>
<td>Alaçam</td>
<td>41</td>
<td>40Ar/39Ar or</td>
<td>27.1±1.0</td>
<td>Delaloye and Bingöl (2000)</td>
</tr>
<tr>
<td>Alaçam</td>
<td>41</td>
<td>U-Pb zrn</td>
<td>25.3±1.5</td>
<td>Catlos et al. (2012)</td>
</tr>
<tr>
<td>Egrigöz</td>
<td>43</td>
<td>40Ar/39Ar bt+or</td>
<td>24.6±1.4</td>
<td>Delaloye and Bingöl (2000)</td>
</tr>
<tr>
<td>Egrigöz</td>
<td>43</td>
<td>U-Pb zrn</td>
<td>24.1±1.3</td>
<td>Catlos et al. (2012)</td>
</tr>
<tr>
<td>Egrigöz</td>
<td>43</td>
<td>U-Pb zrn</td>
<td>20.7±0.6</td>
<td>Ring and Collins (2005)</td>
</tr>
<tr>
<td>Koyunoba</td>
<td>42</td>
<td>40Ar/39Ar kfs</td>
<td>20.37±0.03</td>
<td>Etzel et al. (2020)</td>
</tr>
<tr>
<td>Alaçam</td>
<td>41</td>
<td>Rb-Sr bt</td>
<td>20.17±0.20</td>
<td>Hasözbek et al. (2010)</td>
</tr>
<tr>
<td>Alaçam</td>
<td>41</td>
<td>40Ar/39Ar ms</td>
<td>20.2±0.3</td>
<td>Işık et al. (2004)</td>
</tr>
<tr>
<td>Egrigöz</td>
<td>43</td>
<td>40Ar/39Ar kfs</td>
<td>20.02±0.03</td>
<td>Etzel et al. (2020)</td>
</tr>
<tr>
<td>Alaçam</td>
<td>41</td>
<td>U-Pb zrn</td>
<td>20.0±1.4</td>
<td>Hasözbek et al. (2010)</td>
</tr>
<tr>
<td>Baklan</td>
<td>40</td>
<td>40Ar/39Ar wr</td>
<td>19.3±0.9</td>
<td>Aydoğan et al. (2008)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>17.8±0.7</td>
<td></td>
</tr>
</tbody>
</table>

a See Figure 4 for locations of these granite bodies.
b Abbreviations after Whitney and Evans (2010), wr= whole rock.
<table>
<thead>
<tr>
<th>Granite</th>
<th>Location</th>
<th>Approach</th>
<th>Age</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Late Pan-African Granitoids or Cadomian Granitoids</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Çine Massif</td>
<td>north of Milas</td>
<td>U-Pb zrn</td>
<td>662±3</td>
<td>Loos and Reichmann (1999)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>207Pb/206Pb zrn</td>
<td>537.2 ± 2.4</td>
<td>Dannat (1997)</td>
</tr>
<tr>
<td>Demirci–Gördes</td>
<td></td>
<td>207Pb/206Pb zrn</td>
<td>544.1 ± 4.3</td>
<td></td>
</tr>
<tr>
<td>Ödemiş–Kiraz</td>
<td></td>
<td>207Pb/206Pb zrn</td>
<td>528.0 ± 4.3</td>
<td>Dannat (1997)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>570 ± 5</td>
<td></td>
</tr>
<tr>
<td>Çine Massif</td>
<td></td>
<td>207Pb/206Pb zrn</td>
<td>546.0 ± 1.6</td>
<td>Hetzel and Reischmann (1996)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>546.4 ± 0.8</td>
<td></td>
</tr>
<tr>
<td>Bafa Lake-Çine Massif</td>
<td>U-Pb zrn</td>
<td>521 ± 5</td>
<td>Loos and Reichmann (1999)</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>572 ± 7</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Yatağan</td>
<td>207Pb/206Pb zrn</td>
<td>555.5 ± 6.2</td>
<td>Dora et al. (2005)</td>
<td></td>
</tr>
<tr>
<td>North of Yatağan</td>
<td>U/Pb zrn</td>
<td>549 ± 26</td>
<td>Dora et al. (2005)</td>
<td></td>
</tr>
<tr>
<td>Triassic</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Alasehir</td>
<td>44</td>
<td>U-Pb zrn</td>
<td>222.9 ± 1.1</td>
<td>Ustaömer et al. (2016)</td>
</tr>
<tr>
<td>Late Eocene-Oligocene-Miocene</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Alasehir</td>
<td>44</td>
<td>40Ar/39Ar bt</td>
<td>36.4 ± 2.2</td>
<td>Delaloye and Bingöl (2000)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>16.6 ± 0.3</td>
<td>36.4 ± 2.2</td>
<td></td>
</tr>
<tr>
<td>Gordes</td>
<td>45</td>
<td>40Ar/39Ar ms</td>
<td>28.8 ± 0.6</td>
<td>Delaloye and Bingöl (2000)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>19.4 ± 0.7</td>
<td>28.8 ± 0.6</td>
<td></td>
</tr>
<tr>
<td>Salihli</td>
<td>46</td>
<td>Th-Pb mnz</td>
<td>21.7 ± 4.5</td>
<td>Catlos et al. (2010)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>9.6 ± 1.6</td>
<td>21.7 ± 4.5</td>
<td></td>
</tr>
<tr>
<td>Turgutlu</td>
<td>47</td>
<td>Th-Pb mnz</td>
<td>19.2 ± 5.1</td>
<td>Catlos et al. (2010)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>11.5 ± 0.8</td>
<td>19.2 ± 5.1</td>
<td></td>
</tr>
<tr>
<td>Salihli</td>
<td>46</td>
<td>U-Pb ttn</td>
<td>17.07 ± 0.2</td>
<td>Rossetti et al. (2017)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>14.36 ± 0.3</td>
<td>17.07 ± 0.2</td>
<td></td>
</tr>
<tr>
<td>Turgutlu</td>
<td>47</td>
<td>U-Pb mnz</td>
<td>16.1 ± 0.2</td>
<td>Glony and Hetzel (2007)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>14.06 ± 0.03</td>
<td>16.1 ± 0.2</td>
<td></td>
</tr>
<tr>
<td>Salihli</td>
<td>46</td>
<td>U-Pb aln</td>
<td>15.0 ± 0.3</td>
<td>Glony and Hetzel (2007)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>5.05 ± 0.02</td>
<td>15.0 ± 0.3</td>
<td></td>
</tr>
</tbody>
</table>

\[a\] See Figure 4 for locations of these granite bodies.

\[b\] Abbreviations after Whitney and Evans (2010), wr= whole rock.
Table 9. List of selected earthquake events along the Simav Fault and associated fault systems.

<table>
<thead>
<tr>
<th>No. a</th>
<th>Event-ID b</th>
<th>Time (UTC)</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Depth (km)</th>
<th>Rms c</th>
<th>Mag d</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>465625</td>
<td>2/18/2020 16:09</td>
<td>39.1015</td>
<td>27.8453</td>
<td>14.68</td>
<td>0.45</td>
<td>5.2</td>
</tr>
<tr>
<td>2</td>
<td>150860</td>
<td>12/10/2011 5:15</td>
<td>38.8625</td>
<td>30.1883</td>
<td>13.44</td>
<td>0.96</td>
<td>4.2</td>
</tr>
<tr>
<td>3</td>
<td>319040</td>
<td>12/2/2015 15:52</td>
<td>39.1495</td>
<td>28.154</td>
<td>10.85</td>
<td>0.4</td>
<td>4.0</td>
</tr>
<tr>
<td>4</td>
<td>132605</td>
<td>6/10/2011 22:47</td>
<td>39.0975</td>
<td>28.3405</td>
<td>34.38</td>
<td>0.85</td>
<td>4.7</td>
</tr>
<tr>
<td>5</td>
<td>160143</td>
<td>3/29/2012 10:13</td>
<td>38.6035</td>
<td>30.004</td>
<td>12.77</td>
<td>0.73</td>
<td>4.2</td>
</tr>
<tr>
<td>6</td>
<td>367059</td>
<td>3/29/2017 18:10</td>
<td>38.2003</td>
<td>31.0575</td>
<td>14.87</td>
<td>0.36</td>
<td>4.0</td>
</tr>
<tr>
<td>7</td>
<td>495401</td>
<td>2/9/2021 15:51</td>
<td>38.5965</td>
<td>31.6318</td>
<td>7.01</td>
<td>0.49</td>
<td>4.7</td>
</tr>
<tr>
<td>8</td>
<td>495403</td>
<td>2/9/2021 15:53</td>
<td>38.59</td>
<td>31.6495</td>
<td>4.61</td>
<td>0.41</td>
<td>4.1</td>
</tr>
<tr>
<td>9</td>
<td>367501</td>
<td>4/3/2017 9:05</td>
<td>38.4801</td>
<td>31.7975</td>
<td>13.84</td>
<td>0.25</td>
<td>4.0</td>
</tr>
<tr>
<td>10</td>
<td>136512</td>
<td>7/27/2011 9:58</td>
<td>38.3278</td>
<td>31.8802</td>
<td>17.79</td>
<td>0.33</td>
<td>4.8</td>
</tr>
<tr>
<td>11</td>
<td>128573</td>
<td>5/19/2011 20:15</td>
<td>39.1328</td>
<td>29.082</td>
<td>24.46</td>
<td>0.49</td>
<td>5.7</td>
</tr>
<tr>
<td>12</td>
<td>128577</td>
<td>5/19/2011 20:25</td>
<td>39.1442</td>
<td>29.1078</td>
<td>7.00</td>
<td>0.44</td>
<td>4.6</td>
</tr>
<tr>
<td>13</td>
<td>128603</td>
<td>5/19/2011 21:12</td>
<td>39.113</td>
<td>29.0377</td>
<td>7.74</td>
<td>0.57</td>
<td>4.8</td>
</tr>
<tr>
<td>14</td>
<td>128672</td>
<td>5/20/2011 0:13</td>
<td>39.1413</td>
<td>29.1065</td>
<td>16.92</td>
<td>0.62</td>
<td>4.1</td>
</tr>
<tr>
<td>15</td>
<td>128701</td>
<td>5/20/2011 0:58</td>
<td>39.1147</td>
<td>29.0837</td>
<td>17.38</td>
<td>0.78</td>
<td>4.3</td>
</tr>
<tr>
<td>16</td>
<td>129252</td>
<td>5/21/2011 21:43</td>
<td>39.1037</td>
<td>29.0513</td>
<td>7.00</td>
<td>0.11</td>
<td>4.0</td>
</tr>
<tr>
<td>17</td>
<td>129791</td>
<td>5/24/2011 2:55</td>
<td>39.1013</td>
<td>29.0217</td>
<td>16.80</td>
<td>0.45</td>
<td>4.2</td>
</tr>
<tr>
<td>18</td>
<td>131192</td>
<td>5/30/2011 22:03</td>
<td>39.1567</td>
<td>29.0112</td>
<td>15.29</td>
<td>0.85</td>
<td>4.0</td>
</tr>
<tr>
<td>19</td>
<td>132022</td>
<td>6/5/2011 21:29</td>
<td>39.143</td>
<td>29.095</td>
<td>6.98</td>
<td>0.55</td>
<td>4.0</td>
</tr>
<tr>
<td>20</td>
<td>134386</td>
<td>6/29/2011 11:40</td>
<td>39.1232</td>
<td>29.0032</td>
<td>9.28</td>
<td>0.75</td>
<td>4.0</td>
</tr>
<tr>
<td>21</td>
<td>135896</td>
<td>7/19/2011 21:16</td>
<td>39.1048</td>
<td>29.093</td>
<td>17.66</td>
<td>0.67</td>
<td>4.1</td>
</tr>
<tr>
<td>22</td>
<td>138300</td>
<td>8/25/2011 4:19</td>
<td>39.139</td>
<td>29.0957</td>
<td>22.54</td>
<td>0.77</td>
<td>4.3</td>
</tr>
<tr>
<td>23</td>
<td>161414</td>
<td>4/16/2012 10:10</td>
<td>39.1227</td>
<td>29.1222</td>
<td>6.90</td>
<td>0.5</td>
<td>4.7</td>
</tr>
<tr>
<td>25</td>
<td>161902</td>
<td>4/20/2012 16:39</td>
<td>39.1525</td>
<td>29.0975</td>
<td>20.59</td>
<td>0.81</td>
<td>4.4</td>
</tr>
<tr>
<td>26</td>
<td>177315</td>
<td>10/30/2012 0:12</td>
<td>39.1385</td>
<td>29.1787</td>
<td>21.35</td>
<td>0.76</td>
<td>4.1</td>
</tr>
<tr>
<td>27</td>
<td>188611</td>
<td>3/12/2013 20:47</td>
<td>39.1203</td>
<td>29.0583</td>
<td>12.81</td>
<td>0.52</td>
<td>4.1</td>
</tr>
<tr>
<td>28</td>
<td>197002</td>
<td>6/9/2013 14:18</td>
<td>39.1392</td>
<td>29.022</td>
<td>15.61</td>
<td>0.68</td>
<td>4.1</td>
</tr>
<tr>
<td>29</td>
<td>234353</td>
<td>7/15/2014 12:25</td>
<td>39.13</td>
<td>29.0041</td>
<td>9.92</td>
<td>0.32</td>
<td>4.1</td>
</tr>
<tr>
<td>30</td>
<td>309933</td>
<td>9/3/2015 8:23</td>
<td>39.1226</td>
<td>29.1225</td>
<td>10.24</td>
<td>0.49</td>
<td>4.1</td>
</tr>
</tbody>
</table>

a. See Figure 7A for events 1-10 and Figure 7B for events 11-30.

b. Parameters were extracted from https://depem.afad.gov.tr/depemkatalogu 1900-20XX Earthquake Catalog (M>4.0), Turkish Ministry of the Interior, Disaster and Emergency Management Presidency, Earthquake Department (AFAD).

c. Rms= root-mean-square (RMS) travel time residual in seconds.

d. All magnitudes are ML (original magnitude relationship defined for local earthquakes), except events 1, 3, 6, 7, 9, 29, and 30, which are moment magnitudes (Mw).
Table 10. List of selected earthquake events along the Aegean-Anatolian plate boundary.

<table>
<thead>
<tr>
<th>No.</th>
<th>Event-ID</th>
<th>Time (UTC)</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Depth (km)</th>
<th>Rms&lt;sup&gt;c&lt;/sup&gt;</th>
<th>Mag&lt;sup&gt;d&lt;/sup&gt;</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>199626</td>
<td>7/12/2013 0:36</td>
<td>40.3738</td>
<td>25.946</td>
<td>27.85</td>
<td>0.45</td>
<td>4.3</td>
</tr>
<tr>
<td>2</td>
<td>201060</td>
<td>7/30/2013 5:33</td>
<td>40.3028</td>
<td>25.7902</td>
<td>20.01</td>
<td>0.52</td>
<td>5.3</td>
</tr>
<tr>
<td>3</td>
<td>184151</td>
<td>1/19/2013 19:26</td>
<td>39.6382</td>
<td>25.6795</td>
<td>20.91</td>
<td>0.5</td>
<td>4.2</td>
</tr>
<tr>
<td>4</td>
<td>360268</td>
<td>2/6/2017 10:58</td>
<td>39.5275</td>
<td>26.1373</td>
<td>9.83</td>
<td>0.21</td>
<td>5.3</td>
</tr>
<tr>
<td>5</td>
<td>183497</td>
<td>1/12/2013 13:47</td>
<td>39.6447</td>
<td>25.6733</td>
<td>6.89</td>
<td>0.8</td>
<td>4.0</td>
</tr>
<tr>
<td>6</td>
<td>155511</td>
<td>1/29/2012 21:03</td>
<td>38.7387</td>
<td>26.0447</td>
<td>32.39</td>
<td>0.68</td>
<td>4.2</td>
</tr>
<tr>
<td>7</td>
<td>101387</td>
<td>3/26/2010 18:35</td>
<td>38.1457</td>
<td>26.177</td>
<td>24.26</td>
<td>0.2</td>
<td>4.7</td>
</tr>
<tr>
<td>8</td>
<td>426091</td>
<td>11/27/2018 23:16</td>
<td>36.7565</td>
<td>25.877</td>
<td>16.15</td>
<td>0.62</td>
<td>4.4</td>
</tr>
<tr>
<td>9</td>
<td>426096</td>
<td>11/27/2018 23:46</td>
<td>36.6493</td>
<td>25.4535</td>
<td>5.95</td>
<td>0.75</td>
<td>4.1</td>
</tr>
<tr>
<td>10</td>
<td>418888</td>
<td>8/19/2018 5:46</td>
<td>35.8861</td>
<td>26.0695</td>
<td>28.49</td>
<td>0.77</td>
<td>4.9</td>
</tr>
<tr>
<td>11</td>
<td>309516</td>
<td>8/27/2015 0:25</td>
<td>34.7751</td>
<td>25.8068</td>
<td>7.06</td>
<td>0.52</td>
<td>4.5</td>
</tr>
<tr>
<td>12</td>
<td>472843</td>
<td>5/2/2020 16:44</td>
<td>34.5521</td>
<td>25.8181</td>
<td>6.76</td>
<td>0.56</td>
<td>5.1</td>
</tr>
<tr>
<td>13</td>
<td>472824</td>
<td>5/2/2020 13:45</td>
<td>34.2973</td>
<td>25.7371</td>
<td>9.63</td>
<td>0.98</td>
<td>5.2</td>
</tr>
<tr>
<td>14</td>
<td>472819</td>
<td>5/2/2020 12:51</td>
<td>34.2226</td>
<td>25.8253</td>
<td>6.65</td>
<td>0.98</td>
<td>6.4</td>
</tr>
<tr>
<td>15</td>
<td>472825</td>
<td>5/2/2020 13:33</td>
<td>33.9548</td>
<td>26.0141</td>
<td>6.5</td>
<td>0.96</td>
<td>4.6</td>
</tr>
<tr>
<td>16</td>
<td>472827</td>
<td>5/2/2020 14:21</td>
<td>34.2123</td>
<td>26.232</td>
<td>5.86</td>
<td>0.93</td>
<td>4.8</td>
</tr>
<tr>
<td>17</td>
<td>294406</td>
<td>4/16/2015 18:07</td>
<td>34.8434</td>
<td>26.7275</td>
<td>12.34</td>
<td>0.62</td>
<td>5.9</td>
</tr>
<tr>
<td>18</td>
<td>169403</td>
<td>7/4/2012 23:46</td>
<td>35.1613</td>
<td>26.9993</td>
<td>34.09</td>
<td>0.35</td>
<td>5.0</td>
</tr>
<tr>
<td>19</td>
<td>293183</td>
<td>3/27/2015 23:34</td>
<td>35.7295</td>
<td>26.576</td>
<td>56.13</td>
<td>0.47</td>
<td>5.0</td>
</tr>
<tr>
<td>20</td>
<td>507881</td>
<td>8/1/2021 4:31</td>
<td>36.3843</td>
<td>27.0805</td>
<td>10.86</td>
<td>0.17</td>
<td>5.5</td>
</tr>
<tr>
<td>21</td>
<td>187555</td>
<td>2/27/2013 22:05</td>
<td>36.7298</td>
<td>26.5115</td>
<td>140.27</td>
<td>0.43</td>
<td>4.1</td>
</tr>
<tr>
<td>22</td>
<td>417483</td>
<td>7/26/2018 8:17</td>
<td>37.6546</td>
<td>26.696</td>
<td>4.5</td>
<td>0.52</td>
<td>4.5</td>
</tr>
<tr>
<td>23</td>
<td>483762</td>
<td>10/30/2020 11:51</td>
<td>37.879</td>
<td>26.703</td>
<td>14.9</td>
<td>1</td>
<td>6.6</td>
</tr>
<tr>
<td>24</td>
<td>375576</td>
<td>6/12/2017 12:28</td>
<td>38.8486</td>
<td>26.313</td>
<td>15.96</td>
<td>0.28</td>
<td>6.2</td>
</tr>
<tr>
<td>25</td>
<td>431610</td>
<td>2/20/2019 18:23</td>
<td>39.6011</td>
<td>26.4261</td>
<td>5.8</td>
<td>0.37</td>
<td>5.0</td>
</tr>
<tr>
<td>26</td>
<td>411695</td>
<td>5/3/2018 2:04</td>
<td>39.967</td>
<td>26.8993</td>
<td>10.39</td>
<td>0.35</td>
<td>4.3</td>
</tr>
<tr>
<td>27</td>
<td>284923</td>
<td>12/16/2014 9:02</td>
<td>40.1298</td>
<td>27.0845</td>
<td>17.35</td>
<td>0.29</td>
<td>4.3</td>
</tr>
<tr>
<td>28</td>
<td>115792</td>
<td>11/3/2010 2:51</td>
<td>40.3997</td>
<td>26.3147</td>
<td>28.9</td>
<td>0.59</td>
<td>5.1</td>
</tr>
<tr>
<td>29</td>
<td>199626</td>
<td>7/12/2013 0:36</td>
<td>40.3738</td>
<td>25.946</td>
<td>27.85</td>
<td>0.45</td>
<td>4.3</td>
</tr>
</tbody>
</table>

a. See Figure 11 for events.
b. Parameters were extracted from https://deprem.afad.gov.tr/depremkatalogu 1900-20XX Earthquake Catalog (M>=4.0), Turkish Ministry of the Interior, Disaster and Emergency Management Presidency, Earthquake Department (AFAD).
c. Rms= root-mean-square (RMS) travel time residual in seconds.
d. All magnitudes are ML (original magnitude relationship defined for local earthquakes), except events 4, 8-14, 16, 17, 19, 20, 23-28, which are moment magnitudes (Mw).
Figure 1. Tectonic map of the Aegean and Anatolian microplates. Plate boundaries after McClusky et al. (2000), Nyst & Thatcher (2004), Piper et al. (2010), Harrison et al. (2012), and Tan (2013). Only some major fault systems are labeled. NAF= North Anatolian Fault, EAFZ= East Anatolian Fault Zone, CATB = Central Anatolian Thrust Belt, DSF = Dead Sea Fault; KTZ = Kephalonia Transform Zone; MCL = Mid-Cycladic lineament; İBTZ = İzmir–Balıkesir transfer zone; NAT= North Aegean Trough; NAF = North Anatolian Fault (N-, northern, C- central, and S- southern segments); KM= Kazdağ Massif. Region of slab tear in western Turkey and the Aegean after Jolivet et al. (2015), near Crete (Özbakır et al., 2013), Cyprus (Woodside et al., 2002), and between the Aegean domain and the Menderes Massif (Roche et al., 2019). Boundaries between Central and Western Anatolia after Şengör et al. (1985).
Figure 2. Geological map of Western Anatolia focusing on the ophiolite and granite assemblages along the boundary between the Aegean and Anatolia microplates. Plate boundary after Nyst & Thatcher (2004). Terrane boundaries, major fault systems, and transfer zones after Okay (2008), Akbayram et al. (2016), Oner et al. (2010), and Karaoğlu & Helvacı (2014). Abbreviations: RPA = Rhodope-Pontide Arc; IBTZ = Izmir–Balıkesir Transfer Zone (also sometimes referred to as the Western Anatolia Transfer Zone, Gessner et al., 2013; 2017); SWASZ = South West Anatolian Shear Zone; IPS = Intra-Pontide suture zone; IAESZ = Izmir-Ankara-Erzincan suture zone; PS = Pamphylian suture zone; UMTZ = Uşak-Mugla Transfer Zone.
Figure 3. North–south generalized cross-section across western Turkey after Okay (1986) and Shin et al. (2013). IAESZ=İzmir-Ankara-Erzincan Suture Zone. See Figure 2 for the approximate line of section on the geological map.
Figure 4. Geological map showing structures and locations of Western Anatolia granite bodies. Base map after Delaloye & Bingöl (2000), Senel & Aydal (2002), and Okay (2008). See Tables 1-7 for the granite names that correspond to the numbers in this figure. Abbreviations: IPS = Intrapontide Suture Zone, IAESZ = Izmir-Ankara-Erzincan Suture Zone, SDF= Simav Detachment Fault, AD= Alasehir Detachment. Locations of Figures 5 and 7 are indicated.
Figure 5. (A) Simplified geologic map of the Sivrihisar Massif (eastern Taşlanmış Zone) overlain on a hillshade raster. Map after Senel & Aydal (2002), Özsayin & Dirik (2007), and Shin et al. (2013). See the data repository for the color figure. (B) Sivrihisar granite K-feldspar age spectra for sample TA04. The upper profile by Sherlock et al. (1999) and the lower are our results. (C) $^{36}\text{Ar}/^{40}\text{Ar}$ vs. $^{39}\text{Ar}/^{40}\text{Ar}$ plot comparing our data to Sherlock et al. (1999). Our results show mixing between a radiogenic and atmospheric component of argon with four lower points from initial isothermal steps. Sherlock et al. (1999) data is affected by excess argon (ArE). (D) One possible thermal history path for the Sivrihisar granite based on the rapidly cooled K-feldspar ages, zircon ages, and zircon saturation temperature from Shin et al. (2013).
Figure 6. Paleogeographic reconstruction of Western Anatolia (center box) and the surrounding region at 46 Ma prior to the onset of extension (after Stampfli & Kozur, 2006). Abbreviations relevant to Western Anatolia are Mn=Menderes Massif, Kb=Karaburun, Dg=Denizgören ophiolite, Sk=Sakarya Is=Istanbul, Zo=Zonguldak, BS=Black Sea, Er=Eratosthenes seamount. For other abbreviations, please see Stampfli & Kozur (2006).
**Figure 7.** (A) Map of the Simav Fault and associated structures. Small dots are extracted from the USGS Earthquake Catalog magnitude 2.5+ (http://earthquake.usgs.gov/earthquakes/search) of events from 1952-2021. Location of fault strands after Konak (2002). Inset shows the location near the town on Simav in panel (B). (B) Map of the surrounding area of Simav with earthquakes plotted. In this map, events were extracted from the Turkish Ministry of the Interior, Disaster and Emergency Management Presidency, Earthquake Department Earthquake Catalog (M>=4.0), 1900-20XX (https://deprem.afad.gov.tr/depremkatalogu). The size of the circle represents magnitude. The figure highlights 2011 earthquakes by additional solid dots. Base maps in both panels are from ESRI. Focal mechanism solutions in both panels were extracted from the Turkish catalog. See Table 9 for details of the events. For locations of faults in panel (B), see Mutlu (2020).
Figure 8. North-south generalized cross-section through the Hellenic arc system showing the key structural elements. Map of the Mediterranean Ridge after Westbrook & Reston (2002).
Figure 9. (A) EMODnet Digital Bathymetry map with some structures overlain. The Aegean and Anatolian microplate boundaries are shown in grey after Nyst and Thatcher (2004). Other structures after Hall et al. (1984) and (2009), Woodside et al. (2002), Peterek & Schwarze (2004), Meier et al. (2007); Kinnaird & Robertson (2012), and Symeou et al. (2018). Abbreviations: BT= Backthrust; KFZ = Kephalonia Fault Zone; IAESZ = Izmir-Ankara-Erzincan Suture Zone; KM= Kirşehir Massif, AM= Anaximander Mountains; PTF = Paphos Transform Fault, ES = Eratosthenes Seamount. (B) Profiles along the lines of section shown in panel (A). Abbreviations: CT= CFH = LN= Lycian Nappes, MR= Mediterranean Ridge Accretionary Complex, HB = Herodotus Basin, HSZ= Hellenic Shear Zone, NAF= North Anatolian Fault; AM = Anaximander Mountains; CA= Cyprus Arc. Hashed regions in panel (B) indicate area speculated to be affected by slab tear (e.g., Woodside et al., 2002; Özbakır et al., 2013; Jolivet et al., 2015). See supplementary files for the color figure.
Figure 10. (A) Cross-section of the Aegean anomaly interpreted as the African slab using the UUP07 P-wave model (Amaru, 2007). The line of section used latitude of 28°-43° and longitude of 24°-28°. For more detailed views of the anomaly, see van der Meer et al. (2018), Wei et al. (2019), Blom et al. (2020), and El-Sharkawy et al. (2021). The depths of the dashed lines are 410, 660, 1000 km from the surface. Interpretations of the geology below 1000 are debated and discussed in the text. Image created using Hosseini et al. (2018). (B) Depth vs. estimated earthquake depth for the same latitude and longitude as seen in panel (A). In this map, events were extracted from the Turkish Ministry of the Interior, Disaster and Emergency Management Presidency, Earthquake Department Earthquake Catalog (M>=4.0), 1900-20XX (https://deprem.afad.gov.tr/depremkatalogu). Events are from 01/24/1900 to 6/17/2021. We indicate the largest event (6/26/1926, 19:46).
Figure 11. Map of plate boundaries between the Aegean and Anatolian microplates with some faults indicated (after Nyst & Thatcher, 2004; Uzel et al., 2013; Pe-Piper et al., 2002; Menant et al., 2016). Focal mechanisms are from the Turkish Ministry of the Interior, Disaster and Emergency Management Presidency, Earthquake Department Earthquake Catalog (M>=4.0), 1900-20XX (https://deprem.afad.gov.tr/depremkatalogu). Events are only from 2010-2020 and details are presented in Table 10. The size of the circle represents magnitude. The 9 July 1956 Amorgos earthquake epicenter is also indicated after Alatza et al. (2020). See Okal et al. (2009) for discussions regarding the focal mechanism of this event. The base map is from ESRI. The abbreviations IBTZ = Izmir–Balıkesir transfer zone; NCSD= North Cyclades Detachment System; MCL= Mid-Cycladic Lineament.
Figure 12. (A) Interpretative thrust sequence during the formation of Anatolide belt after Gessner et al. (2013). CBU= Cyclades Blueschist Unit; CMT= Cyclades Menderes Thrust; SSZ= Selimiye Shear Zone, BT= Bozdag Thrust. (B) P-T paths from Menderes Massif nappes (Ring et al., 2001; Whitney & Bozkurt, 2002; Rimmelé et al., 2005; Régnier et al., 2007).
Figure 13. Isochemical phase diagrams with overlapping garnet core compositional isopleths for garnet-bearing samples from the (A) Çine nappe, (B) Selimiye nappe after Etzel et al. (2019), (C) Northern Menderes Massif using data from Cenki-Tok et al. (2016), and (D) the Central Menderes Massif (Etzel et al., 2020). Mineral abbreviations after de Capitani and Brown (1987) and de Capitani and Petrakakis (2010). Labeled stripes are compositional isopleths of ±0.1 mole fraction for endmember garnet core compositional contents, except for panel (D), which overlies ±0.2 mole fraction and is for the reported representative composition for that garnet by Cenki-Tok et al. (2016). The grey polygon in each diagram represents the conditions estimated for garnet growth in the samples. High-resolution P-T paths for the samples are shown in panels (A), (B), and (D). See supplementary figures for this figure in color.
Figure 14. Snapshots of thermal models of the Çine nappe for the (left) fault reactivation and (right) tectonic switching model after Etzel et al. (2019). (A) and (B) are the upper equilibrium thermal grid (depth vs. horizontal distance) before faulting with the position of fault (grey line) arbitrarily selected at 30°. Fault displacement varies linearly. The grid includes reflecting side boundaries and top and bottom maintained at 25°C and 700°C and an initial geothermal gradient at 25°C/km indicated by shaded bars. The position of the Selimiye samples is inferred by a hatch area, and the grey bar represents the approximate initial location of the Çine nappe garnet with the N-shaped P-T path. This is also represented by point 1 in P-T path insets. In panels (C) and (D), the fault is activated and a finite-difference solution to the diffusion-advection equation is used to examine the P-T variations in the hanging wall and footwall as a result of motion. The rock sample experiences the path from 1 to 2 on the P-T path insets. In panels (E) and (F), motion stops. In panel (E), extension occurs, whereas denudation occurs in panel (F). This is modeled based on the mid-rim lower pressure portion of the garnet P-T path and is represented by points 2 to 3 on the P-T path insets. In panels (G) and (H), the fault is reactivated, represented by points 3 to 4 on the P-T path insets.