# A Review of the Dynamics of Subduction Zone Initiation in the Aegean Region

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**Index Terms**

8104 Continental margins: convergent

**Keywords**

Subduction zones, Hellenic arc, Aegean region, Western Anatolia, Tectonics

**Abstract**

The Hellenic arc, where the African (Nubian) slab subducts beneath the Aegean and Anatolian microplates, has emerged as a type-locality for understanding subduction dynamics, including slab tear, slab fragments, drips, and transfer zones. Based on field evidence and geophysical, tectonics, and geochemical studies, it has been recognized that the subducting African slab is a primary driver for extension in the Aegean and Anatolian microplates and plays a significant role in accommodating present-day westward extrusion of the Anatolian microplate. Thus, understanding the Hellenic arc subduction zone initiation (SZI) age is critical in deciphering ancient mantle flow, how plate tectonics is maintained, and the mechanisms involved in triggering the onset of subduction. The SZI for the Hellenic arc has two disparate ages based on different lines of evidence. A Late Cenozoic (Eocene-Pliocene) SZI is proposed using the analysis of topography combined with estimates of slab age and depth, paleomagnetism, the timing of metamorphism, and volcanic activity, and timing of sedimentation within its accretionary wedge, the Mediterranean Ridge. This age follows an induced-transference SZI model, where a new subduction zone initiates following the jamming of an older subduction zone by buoyant crust due to regional compression, uplift, and underthrusting. A Late Cretaceous-Jurassic SZI age has also been proposed using reconstructions of images of subducted slabs seen using tomography and timing of obducted ophiolite fragments thought to be related to the system. In this case, the induced-transference SZI model fails, and a single subduction zone persists. As a result, continental lithospheric fragments and the ancient oceans between them become incorporated into the overall system without creating a new subduction zone. The presence of a long-lived subduction zone has implications for understanding Earth's mantle dynamics and how plate tectonics operates. This paper describes and summarizes the evidence for both models in the Aegean-Western Anatolia region.

## 1 Introduction

Subduction zones form when two lithospheric plates converge, and one plate abruptly descends beneath the other (Figures 1 and 2) (e.g., White et al., 1970; Hayes, 2018; Stern & Gerya, 2018; Crameri et al., 2020). Large magnitude earthquakes, tsunamis, volcanic eruptions, and landslides occur near and are caused by this specific plate boundary. They are considered exceptional geological environments for recording significant ground-level changes that can trigger tsunamis, impact ground motion, and climate change. Earthquakes that occur in such zones and those triggered by the subduction process far-afield have global consequences. Understanding the dynamics of subduction zones involves diverse and multidisciplinary studies, critical for understanding their associated hazards and how they have influenced the dynamics of plate tectonics over Earth's history (e.g., Stern, 2004; Gerya, 2011; Le Pichon et al., 2019; Crameri et al., 2020).

The Aegean and Anatolian microplates (Figure 1) are significantly impacted by the dynamics of the subducting northern portion of the African (Nubian) plate, which has emerged as the primary driver for extension and the development of metamorphic core complexes in the Aegean region (e.g., Jolivet & Faccenna, 2000; Çemen et al., 2006; Dilek & Sandvol, 2009; van Hinsbergen et al., 2010; van Hinsbergen). The Hellenic and Cyprus arcs are the surface expression of the subducting Nubian Plate and eastern Mediterranean lithosphere beneath the Aegean and Anatolian microplates, respectively (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; Biryol et al., 2011; Royden & Papanikolaou, 2011; Hall et al., 2014; Symeou et al., 2018; Ventouzi et al., 2018).

Constraints regarding the subduction zone initiation (SZI) age of the present-day expression of the Hellenic arc developed from several independent approaches include timing of sedimentation within the intensely folded and faulted rocks of Mediterranean Ridge accretionary prism (Figures 1 and 2), paleomagnetism, the analysis of topography combined with estimates of slab age and depth, reconstructions of subducted slabs using tomography, and the timing of metamorphism and volcanic activity. SZI 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). Some studies suggest a Cenozoic SZI age for Hellenic arc, although estimates vary significantly, from the Eocene-Pliocene (e.g., Meulenkamp et al., 1988; Spakman et al., 1988; Papadopoulos, 1997; Brun & Sokoutis, 2010; Le Pichon et al., 2019) to Mesozoic (Late Cretaceous-Jurassic) (Faccenna et al., 2003; van Hinsbergen et al., 2005; Royden & Papanikolaou, 2011; Jolivet et al., 2013; Malandri et al., 2017; Crameri et al., 2020, van Hinsbergen et al., 2021).

The disparity in the SZI age of onset of Nubian slab subduction along the Hellenic arc is significant as it impacts the tectonic history of the entire Aegean-Anatolian region, one of the most rapidly deforming regions across the Alpine-Himalayan chain. The region has emerged as the type-locality for understanding subduction zone dynamics, including slab tear, slab fragments, drips, and the role of transfer zones triggered by subduction. Understanding its SZI is also critical in deciphering ancient mantle flow, how plate tectonics is maintained, and the mechanisms involved in triggering the onset of subduction, among other factors (e.g., Crameri et al., 2020; van Hinsbergen et al., 2021). This paper aims to summarize the approaches and results of studies that strive to constrain the SZI age of the African (Nubian) slab beneath the Aegean microplate that led to the formation of the Hellenic arc.

# 2 Geometry of the Hellenic arc (Greece to Western Turkey)

## 2.1 Definitions

**The Hellenic subduction system** extends ~1200 km from approximately 37.5°N, 20.0°E offshore the island of Zakynthos to 36.0°N, 29.0°E offshore of the island of Rhodes (Ganas & Parsons, 2009; Le Pichon et al., 2019; Papanikolaou, 2021) (Figures 2 and 3). The system defines the boundary between the northern portion of the **Nubian Plate** and the southern extent of the **Aegean microplate** within the central Mediterranean region (Pearce et al., 2012) and is sometimes referred to as the **Aegean subduction zone** (Wortel et al., 1990; Biryol et al., 2011; Bleier et al., 2007; Polat & Ozel, 2007; Taymaz et al., 2007; Crameri et al., 2020). This boundary between the **Aegean microplate** portion of Eurasia (Nyst & Thatcher, 2004) and the subducting Nubian slab is presently characterized by a strong curvature and fast slab rollback (e.g., Faccena et al., 2013; Evangelidis, 2017). Presently, the African plate advances towards Eurasia NNW at a rate of 5 mm/yr (Fernandes et al., 2003; Ganas & Parsons, 2009), but it subducts northward beneath Crete at a significantly faster rate of 35 mm/yr (McKenzie, 1972; Reilinger et al., 2006). The Aegean area also records the highest deformation rate along the entire Africa/Eurasia convergence zone (McClusky et al., 2000; Kassaras et al., 2005).

The Aegean and Anatolia microplates are sometimes classified as the single **Aegean-Anatolian microplate** (e.g., Jackson, 1994; Oral, 1995; Doutsos & Kokkalas, 2001; Le Pichon et al., 1995) with an Euler pole located north of the Sinai Penninsula (Cianetti et al., 2001). The **Anatolian microplate** itself is a distinct entity that includes over two-thirds of the country of Turkey (Figure 1) (Le Pichon et al., 1995; Oral et al., 1995; Reilinger et al., 1997; Papazachos, 1999). Şengör & Zabcı (2019) consider the whole of Turkey and the Balkan Peninsula within a plate boundary zone.

The Nubian Plate includes the African continent. When Somalia is part of the definition, it is referred to as the **African Plate** or the **Nubia-Somalia Plate** (e.g., McCluksy et al., 2003). The African plate itself is defined by Nubia to the west and Somalia to the east, both of which are separated by a diffuse plate boundary (Chase, 1978, Gordon & Stein 1992, Lemaux et al., 2002; McCluksy et al., 2003; DeMets & Merkouriev, 2016). The African plate is geographically significant and is, at present, the third-largest major tectonic plate (Gaina et al., 2013). A wide range of tectonic interactions characterizes its boundaries and internal dynamics, which were influenced by the collision between the plate and Eurasia (e.g., Meijer & Wortel, 1999; McCluksy et al., 2003; Fernandes et al., 2006; Catalano et al., 2008; DeMets & Merkouriev, 2016). The plate boundary between the Indian and African plates is over 12,000-km-long and may have initiated around 105 Ma in the Early Cretaceous (e.g., van Hinsbergen et al., 2021). Rifting in the Gulf of Aden and the Red Sea signaled the break-off of the Arabian peninsula from Africa at 29-24 Ma (e.g., Bosworth et al., 2005; Wolfenden et al., 2005) fragmentation of the Africa plate and the onset of motion between Somalia and Nubia. Alternatively, the separation may have been more recent (11-10 Ma, Wolfenden et al., 2004; Keranen & Klemperer, 2008; Corti, 2009; see discussion in DeMets & Merkouriev, 2016).

The overriding Aegean microplate (Nyst & Thatcher, 2004) consists mainly of continental lithosphere with a similar thickness as the Anatolian plate (e.g., Zhu et al., 2006; Sodoudi et al., 2006). Large crustal thickness variations exist from western Greece to eastern Anatolia (20–47 km), but within specific regions of the Anatolian plate, thickness appears fairly uniform (Özacar et al., 2010; Karabulut et al., 2019). Moho depth across the Anatolian plate varies between 24 and 48 km, with the thinnest crustal thickness located on the coast of western Turkey and the thickest in Eastern Turkey (Tezel et al., 2013). The Aegean Sea crustal thickness averages ~25 km (Zhu et al., 2006; Tirel et al., 2004; Kind et al., 2015), similar to thickness beneath the central Menderes Massif at 25-30 km (Zhu et al., 2006; Karabulut et al., 2019). Estimates in the eastern Mediterranean overall show the Moho at an average 30 km depth (Marone et al., 2003). The Aegean microplate crustal thickness has thinner southern and central parts (20–22 km) compared to its northern portions (25–28 km) (Karagianni et al., 2005; Sodoudi et al., 2006). An exception to this observation is the crustal thickness beneath western Crete, 32.5-35 km or up to 45-50 km (Bohnhoff et al., 2001; Meier et al., 2004; Snopek et al., 2007). The eastern part of Crete appears thinner (24-26 km) compared to its western portion (32.5 km) (Bohnhoff et al., 2001). Crustal thickness also decreases to 15 km below the central Cretan Sea and to 17 km, approximately 100 km off the southern coast of Crete (Bohnhoff et al., 2001). A crustal thickness of ~40 km is reported for western Greece (Karagianni et al., 2005), north of Antalya Bay (Karabulut et al., 2019), Eastern Anatolia (Kind et al., 2015), and the Anatolian plateau (e.g., Saunders et al., 1998). In Eastern Anatolia, the crustal thickness increases towards the north, from 35 km at the Arabian foreland to 45 km further north (Özacar et al., 2010).

The subducting portion of the African plate has multiple names based on the time frame that was active and differentiate it from other nearby subduction zone systems. In the Aegean and Western Anatolia region, the subducting plate is sometimes referred to as the **Aegean slab** (Wortel & Spakman, 2000; Widiyantoro et al., 2004; Vanderhaeghe et al., 2007; Biryol, 2009; van der Meer et al., 2018; El-Sharkawy et al., 2020), the **African-Mediterranean slab** (Sachpazi et al., 2016), the **Hellenic plate** or **Hellenic slab** (Chang et al., 2010; Vernant et al., 2014; Sachpazi et al., 2016; Confal et al., 2018; Hayes, 2018; Wei et al., 2019; Blom et al., 2020). The Aegean slab usually describes the system during the Late Cretaceous, whereas the other specific names distinguish the Aegean Sea and Western Anatolia subduction system from that beneath Antalya and Cyprus. The subducting slab near Cyprus is referred to as the **Cyprus slab** (Confal et al., 2018), **Cyprus/Tethys slab** (Taylor et al., 2019), **Cyprus–Bitlis slab** (Govers and Fichtner, 2016). The **Antalya slab** is recently defined between the Hellenic and Cyprus arcs (Güvercin et al., 2021).

The western edge of the Hellenic arc is bounded by the **Kephalonia (also Kefalonia, Cephalonia) Transform Zone (KTZ)** (Figure 1) (Bocchini et al., 2018; Hansen et al., 2019). This feature is a dextral strike-slip system that delineates the boundary between the Apulian platform (Adria microplate) and the Eurasian plate or Aegean microplate (Kokinou et al., 2006; Pearce et al., 2012). The KTZ has structural trends and seismicity that extend onshore, and it is the most active zone of shallow seismicity in the broader Aegean region (Kokinou et al., 2006). Over the last 1–4 Myr, the KTZ has linked to the **North Anatolian Fault (NAF)** system via the **Central Hellenic Shear Zone** (Papanikolaou & Royden, 2007; Reilinger et al., 2010; Vassilakis et al., 2011; Royden & Papanikolaou, 2011; Halpaap et al., 2018). The KTZ is considered one of the most seismically active zones in Europe (Pearce et al., 2012; Halpaap et al., 2018) and has been represented as a vertical tear between oceanic and continental lithosphere (Suckale et al., 2009), forming a Subduction-Transform-Edge-Propagator (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 is also proposed between two segments without a tear between, at least at depths shallower than 100 km (Pearce et al., 2012; Halpaap et al., 2018). Nine STEP structures may exist beneath southern Greece, segmenting the subducting African slab and contributing to seismicity and deformation (Sachpazi et al., 2016). Overall, the southwestward expansion and stretching of the Aegean microplate during Plio-Quaternary is accommodated by a northern right lateral tectonic boundary defined by the KTZ and NASZ and a southern left-lateral tectonic boundary, marked by the Pliny and Strabo trenches (Sakellariou et al., 2013).

The active right-lateral NAF and **North Anatolian Shear Zone (NASZ)** extends for ~1200 from the Karlıova triple junction through the Sea of Marmara and Biga Peninsula (Figure 1) (Ketin, 1948; Barka, 1992; Armijo et al., 1999; Şengör & Zabcı, 2019). The NASZ contains the NAF, which 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 & Zabcı, 2019). The fault system disrupts the Aegean extensional domain in its northern section and may have been activated due to subduction rollback processes (e.g., Flerit et al., 2004; Facenna et al., 2006). Current timing estimates suggest the NAF initiated in east Anatolia at 10-12 Ma, and propagated westward through central Anatolia from 7-6 Ma, and was activated in the Aegean region by the Pliocene (Şengör et al., 2005; Facenna et al., 2006). Its western and eastern terminations of the NASZ are poorly defined (Barbot & Weiss, 2021), but the system itself presently accommodates ~24 mm/year of slip along northern Turkey (McClusky et al., 2000; Bulut et al., 2018).

Westward extrusion of the Anatolian plate accommodated by the NASZ and NAF dominates the present-day geodynamics of the Aegean microplate such that active extensional strain has been stated to be absent, except near the Corinth rift, south Viotia, south of Evia, and across the Sperchios-Kammena Vourla rift (Brooks & Ferentinos 1984; Papanikolaou and Royden, 2007; Chousianitis et al., 2013, 2015). However, 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 active thrusting is limited to the central and western sectors of the Hellenic subduction zone and the offshore regions external with respect to it (Maggini & Caputo, 2020). The nature of earthquakes varies with Hellenic arc depth, with normal senses of motion at depths up to 40 km and strike-slip faulting present parallel to strike for events deeper than 40 km (Benetatos et al., 2004). One of the strongest earthquakes of the 20th century, with an Mw 7.7 (or 7.8), occurred in the area of the South Aegean off the coast of the island of Amorgos on 09 July 1956 03:11 UTC (Figure 1) (Okal et al., 2009; Alatza et al., 2020). This event has debated focal mechanisms, as either strike-slip or normal faulting geometries (Okal et al., 2009). An Mw 4.4 earthquake with normal motion recently occurred near its epicenter on 11/27/2018 (23:16 UTC, 36.7565ºN, 25.877 ºE). A normal sense of motion also is found with some recent earthquakes near the NASZ, including 6/12/2017 Mw 6.2 (12:28 UTC, 38.8486ºN, 26.313ºE) and 2017 Mw 5.3 (10:58 UTC, 39.5275 ºN, 26.1373 ºE), likely associated with transtensional motion.

The Hellenic arc itself is separated into a western portion based on well-defined bathymetry. This portion is sometimes termed the **Western Hellenic Subduction Zone or Western Hellenic Arc and Trench (WHA-T) system** and extends ~400 km NW from the central Adriatic Sea to the west coast of Crete (Figures 1 and 3) (Papadopoulos et al., 2010; Pearce et al., 2012; Hansen et al., 2019). This part of the Hellenic Arc is divided into northern and southern sections termed the **Northern Hellenic arc** and **Southern Hellenic arc** (Royden & Papanikolaou, 2011). The Southern Hellenic arc exhibits classical features, but in the north, thick continental crust is subducted beneath northern Greece (Pearce et al., 2012; Halpaap et al., 2018). As a result, this area has significant variations in subduction rate, trench retreat, the occurrence of deep seismicity, and overriding plate extension, likely related to a change in convergence regime from the subduction of oceanic crust in the south to continental crust in the north (Papanikolaou & Royden, 2007; Halpaap et al., 2018). The southern portion of the Hellenic arc near the island of Crete is also affected by the subduction of continental crust (e.g., Meier et al., 2007).

Overall, the Hellenic arc has three distinct regions: an outer compressional non-volcanic arc, a volcanic arc, and an extensional back-arc region covering the broader Aegean Sea region (Figures 1 and 2) (McKenzie 1972; Papazachos, 2019). The Pliocene-Quaternary volcanic arc is located in the Methana Peninsula, along the islands of Milos and Santorini in the Cycladic archipelago, and the island of Nisryos in the Dodecanese (Figure 1) (Le Pichon and Angelier, 1979; McKenzie, 1972; Papazachos and Nolet, 1997, see review in Scoon, 2021). The majority of the subducting Nubian Plate is oceanic, except along the central sector of its southern margin, where the Mediterranean Ridge accretionary complex collides with the African continental margin (Chaumillon & Mascle, 1997; Westbrook & Reston, 2002). High-resolution scattered wave images of the northern portion of the Western Hellenic arc also show the subduction of thick continental crust beneath northern Greece (Pearce et al., 2012; Halpaap et al., 2018) and Crete (e.g., Meier et al., 2007).

The connection of the western portion of the Hellenic arc to its central and eastern sections is diffuse (Beißer et al., 1990; Shaw & Jackson, 2010; Biryol et al., 2011; Özbakır et al., 2013). The eastern portions of the Hellenic arc are sometimes referred to as the **East Hellenic Arc and Trench system (HA-T)** (Papadopoulos et al., 2007). The boundary between the Hellenic and Cyprus arcs in this area is obstructed by up to 300-km wide, 6-10 km-thick section of sediments that comprise the **Mediterranean Ridge** (Figures 1-3; Heezen & Ewing, 1963; Emery et al., 1966; Le Pichon et al., 1982; Kenyon et al., 1982; Kastens et al., 1992; Foucher et al., 1993; Westbrook & Reston, 2002; Kopf et al., 2003). This accretionary complex is the largest structural unit of the Eastern Mediterranean Sea, extending ~2000 km from the Calabrian Rise east of Greece to the Florence Rise (Liminov et al., 1996; Cita et al., 1996). The front of subduction of the Hellenic arc is located south of the Mediterranean Ridge (Figures 1-3) (e.g., Le Pichon et al., 1995; Mascle & Chaumillon, 1997; Mascle et al., 1999; Jost et al., 2002; Westbrook & Reston, 2002; Meier et al., 2007; Jolivet et al., 2013).

The significant size of the Mediterranean Ridge indicates that the Hellenic arc is an accretionary as opposed to an erosive margin (e.g., Von Huene and Scholl, 1991, Von Huene and Scholl, 1993, Cloos and Shreve, 1996; Clift and Vannucchi, 2004). In this case, material is transferred from the subducting plate to the fore-arc wedge and plate boundary zone (e.g., DeFranco et al., 2008; Scholl and Von Huene, 2009). This type of margin is favored when subduction rates are low and seismic moment release rates are high (Le Pichon et al., 1993; Clift and Vannuchi, 2004; DeFranco et al., 2008). However, the Mediterranean Ridge accretionary complex is unusual compared to others worldwide. It formed in a continent-continent collisional setting and has shallow, Messinian-age evaporites (e.g., Cita et al., 1996; Chaumillon & Mascle, 1997). These evaporites influence its deformation and growth rates due to their mechanical properties and effect on fluid flow and pressure (Kastens, 1991; Westbrook & Reston, 2002; Kopf et al., 2003). Extensive mud volcanism associated with the Mediterranean Ridge began in the early Pleistocene 1.75-1.25 Ma (Nikitas et al., 2021).

Understanding the development of the Mediterranean Ridge is critical to determining the Hellenic arc SZI because it is thought to grow via off scraping against a backstop formed by the Alpine nappes of the Hellenic Arc (Kastens, 1991) at a very fast rate (10 km/Myr; Kastens, 1991;

Kopf et al., 2003; Papanikolaou, 2021). The ridge itself varies in geometry along strike (Cita et al., 1996; Chaumillon & Mascle, 1997; Westbrook & Reston, 2002; Kopf et al., 2003; Andronikidis et al., 2017; Papanikolaou, 2021). The wedge accumulates sediments in its western and eastern portions, but the ridge behaves unlike a typical accretionary complex in the central portion between Libya and Crete, where it experienced maximum deformation, and incipient collision had already caused uplift of the area (Chaumillon and Mascle, 1997; Kopf, 2002). Crete is situated on the central forearc of the Hellenic subduction zone and is underlain by material from two microcontinents, leading to larger observed crustal thicknesses (~50 km, e.g., Thomson et al. 1998; Stöckhert 1999; Meier et al., 2004; 2007).

South of Crete, the **Hellenic trenches**, Ptolemy, Pliny, and Strabo (Figure 1) developed between the Mediterranean Ridge and volcanic arc. These trenches are not classical ocean trenches, as earthquakes beneath them originate along low-angle thrusts at 20–40 km (Taymaz et al., 1990; Shaw & Jackson, 2010). Instead, they develop due to back-thrusting beneath the northern edge of the accretionary complex (Galindo-Zaldivar et al., 1996; Westbrook & Reston, 2002) or due to the tearing of the Nubian slab (Özbakır et al., 2013). The Hellenic Trench has been described as the surface expression of a steep (∼30°) reverse fault splaying off the deeper underlying thrust-fault interface of the subduction zone (Shaw et al., 2008; Shaw & Jackson, 2010).

Low-angle thrust faults along the Aegean coast associated with subduction zone tectonics pose significant tsunami hazards (e.g., Tinti et al., 2005; Basili et al., 2013; Howell et al., 2015; Bocchini et al., 2020). Offshore Crete Island is considered one of the most tsunamigenic areas in the entire Mediterranean Sea region (Papadpoulos et al., 2010; Triantafyllou et al., 2019). However, the complexity of the overall Hellenic arc plate boundary combined with its aseismic nature makes earthquake data alone a misleading guide for identifying the likely sources of tsunamigenic earthquakes (Yolsal et al., 2007; England et al., 2015; Howell et al., 2015). Tsunamigenic earthquakes infrequently occur in the eastern Mediterranean (Papazachos and Dimitriu, 1991; Papadopoulos et al., 2007). An evaluation of historical data, including the 1956 Amorgos event that generated the largest of the most recent tsunamis, indicate that a likely trigger of some past tsunamis in the region were submarine landslides generated by earthquakes (e.g., Dominey-Howes, 2002; Okal et al., 2009; Ebeling et al., 2012). Factors that contribute to slope instability across portions of the Hellenic Arc include its sloping bottom, thick accumulations and high rates of recent sedimentation, the presence of closely spaced active faults, active earthquakes, and diapirism (e.g., Ferentinos, 1990; Hooft et al., 2017). The eruption of Santorini (Figure 1) in 1610 BCE generated a tsunami that affected civilizations throughout the eastern Mediterranean (Dominey-Howes, 2004, Friedrich, 2006, Marinatos, 1939, Papadopoulos, 2015; Hooft et al., 2017). Detailed bathymetry across the Mediterranean is critical in understanding tsunami propagation and mitigating its impacts (Figure 1) (e.g., CIESM, 2011).

The **Rhodes Basin** bounds the eastern portion of the Hellenic arc (Figure 1) (Ganas & Parsons, 2009) and is a deep depression lying in continuity with the Pliny and Strabo trenches. The depth of the basin is >4000 m and is one of the deepest portions of the Mediterranean Sea (Woodside et al., 2000). The Rhodes basin may represent an un-subducted portion of the deep Mesozoic Levantine basin (Rotsein & Ben-Avraham, 1985) or a remnant of a former upper-Miocene subduction trench that remained after a shift in the primary convergence zone (Mascle et al., 1986). Deep faults buried beneath the zone may mark the onset of extension (Woodside et al., 2000). Hall et al. (2009) suggest a two-part history of the basin. Following Miocene convergence, the basin experienced middle Pliocene-Quaternary sinistral transpression due to actively curving Hellenic arc and change in the convergence vector of the African plate. Slab tear has been proposed to interpret the presence and structures within the deep Rhodes Basin (Woodside et al., 2000; Faccenna et al., 2014).

A STEP is also suggested for the transition between the Cyprus and Hellenic arcs (Elitez et al., 2016). 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 (Kempler & Ben-Avraham 1987; Zitter et al., 2003; Biryol et al., 2011). This arc became effectively inactive during the onset of westward extrusion of the Anatolian plate (Papanikolaou, 2021).

### 2.2 Geometry of the Hellenic arc subduction zone

To understand when and why the Hellenic subduction zone was established, we must consider its present-day structure. The Hayes (2018) Slab2 model uses active-source seismic data interpretations, receiver functions, local and regional seismicity catalogs, and seismic tomography and models the subducting Nubian slab as uniformly northward dipping to >440 km depths in its northern portion (Figure 3). The Hellenic arc has a well-developed Wadati-Benioff zone at shallower depths but 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). Figure 4A shows the slab clearly defined by earthquake depth vs. latitude across the Hellenic subduction zone. Detailed analysis of the distribution of earthquakes indicate that the western part of the subduction zone dips under 20-30° to the NE and reaches the maximum depth of 180 km, and its eastern section dips under 40° to the NW and reaches a maximum depth of 170 km (Papazachos & Comninakis, 1971; Vaněk et al., 1987; Papazachos et al., 2000; Suckale et al., 2009; Papazachos, 2019). At deeper levels (100–180 km), the Wadati–Benioff zone dips freely (without coupling) at a high angle (~45°) beneath the south Aegean trough and the volcanic arc (Mahatsente et al., 2017; Papazachos, 2019).

Seismic coupling is defined as the ratio between the observed seismic moment release and the rate calculated from plate tectonic velocities (e.g., Ruff and Kanamori, 1983; Scholz and Campos, 2012). The plate interface coupling between the Hellenic trench fault and the Nubia–Aegean is low (<20%) (Papadimitriou and Karakostas, 2008; Vernant et al., 2014) and only 10% in the southwestern portion of the arc (Papadimitriou and Karakostas, 2005). A low seismic coupling scenario tends to produce earthquakes with small rupture zones and high moment release that occur as small isolated patches on the subduction interface (Scholz 1990; Messini et al., 2007). However, in the region near the Ionian islands (Figure 1), complete seismic coupling of the subduction is most consistent with the observed moderate seismic moment release (Laigle et al., 2002; 2004). Higher degrees of seismic coupling are also found in specific regions in the Western Hellenic arc (Jackson and McKenzie, 1988; Messini et al., 2007; Ganas et al., 2020).

Classical ray tomography using body and surface waves is often applied to the Mediterranean region to image the Hellenic arc subduction system (e.g., Piromallo and Morelli, 1997; Spakman et al., 1988; Snieder, 1988; Zielhuis and Nolet, 1994; Piromallo and Morelli, 2003; Marone et al., 2004; Amaru, 2007; Schivardi and Morelli, 2009; Chang et al., 2010; Biryol et al., 2011; Salaün et al., 2012; Legendre et al., 2012; Portner et al., 2018; Blom et al., 2019; Wei et al., 2019; El-Sharkawy et al., 2020). We present a P-wave tomographic image of the Aegean anomaly as an example in Figure 4B. Although surface wave data improves ray coverage and allows the visualization of regions with low seismicity and few stations, body wave tomography affords better resolution yet suffers from vertical smearing and an unknown background velocity model compared to surface waves (Piromallo & Morelli 1997; Kassaras et al., 2005; Schivardi & Morelli, 2009; Salaün et al., 2012). Body wave tomographic images across the Hellenic arc shows the geometry of the subduction zone clearly (e.g., Figure 4B), with the slab sometimes exhibiting a strong velocity contrast (up to 6%) compared to the background model (e.g., Papazachos & Nolet, 1997; Kassaras et al., 2005). Some regional studies show the slab forms an asymmetrical shape, shallowly dipping and larger along the western side and a steeper and smaller in the eastern section (Papazachos & Nolet, 1997; Blom et al., 2019).

Segments of the African slab have long been imaged as subducting into the lower mantle (Figure 4B) (1400±100 km depth; Spakman et al., 1988; Spakman, 1990; 1991; Spakman et al., 1993; Papadopoulos, 1997; Bijwaard et al., 1998; Kárason & van der Hilst, 2000; Wortel & Spakman, 2000; Romanowicz, 2003; Kassaras et al., 2005; Chang et al., 2010; Biryol et al., 2011; Legendre et al., 2012; Govers & Fichtner, 2016; van der Meer et al., 2018; Bocchini et al., 2018; Wei et al., 2019; El-Sharkawy et al., 2020). The width of the subducting slab differs depending on the approach applied to image it. For example, Chang et al. (2010) show the slab widening beneath 660 km in the lower mantle, whereas Primallo & Morelli (2003) has widening within the transition zone. Widening can be associated with a slab avalanche where high viscosity occurs in the lower mantle (Capitanio et al., 2009; Chang et al., 2010). Slab avalanches form when large-volume subducted slabs that temporarily stagnate within the transition zone periodically penetrate the lower mantle (e.g., Solheim and Peltier, 1994; Deschamps and Tackley, 2009; Yang et al., 2018). Slab avalanches are unstable, controlled by mantle thermal instabilities, and will accelerate as slab sinking rates increase with time (e.g., Solheim and Peltier, 1994; Yang et al., 2018). More recent tomographic images show the Hellenic slab extends from the surface to the transition zone in a bent, arcuate shape with a high-velocity structure that flattens from the 410 km discontinuity and is not seen at deeper levels (Blom et al., 2019).

Mantle tomography has also shown that not all slabs in the Mediterranean region are connected to the lithosphere at the surface due to delamination (e.g., Spakman et al., 1988; Dilek & Sandvol, 2009; Wortel & Spakman, 2000). For example, gaps are visible in the slab at depths of 60-100 km just west of the south Hellenides, and slab tear may be visible at the 660 km discontinuity (Wei et al., 2020). Multiple remnant slabs are imaged beneath the North Hellenides (e.g., Wei et al., 2020) and most of Anatolia (e.g., Salaün et al., 2012). Vote Maps, generated from stacking tomography models and identifying where the models agree based on an increasing vote count at a specified depth (Shephard et al. 2017; ), show a slab interpreted to be the African plate between 250-1550 km depths (Crameri et al., 2020).

Although tomographic images are central in some studies of Hellenic arc SZI, their results are challenging to interpret, especially in locating regions of slab tear (e.g., Piromallo & Morelli 2003). The subduction zone's small size, its spatially highly variable nature, and the uneven distribution of its seismic stations make imaging the area more challenging (El-Sharkawy et al., 2020). The presence of slow velocity anomalies throughout the uppermost mantle beneath Anatolia and very slow velocities beneath eastern Anatolia have been explained by innovative ideas regarding how the mantle and lithosphere behave in terms of mantle upwelling, drip-style lithospheric delamination (Göğüş et al., 2017), or rollback (see discussion in Portner et al., 2018). Tomographic images of both the Hellenic and Cyrus arc have transformed ideas regarding how subducting slabs operate as they enter the deep mantle and their impacts on seismicity and the creation of transfer zones (e.g., Meighan et al., 2013), large volume magmatism (e.g., Cocchi et al., 2017), and controlling ore-forming process and mineral deposits (e.g., de Boorder et al., 1998; Rabayrol et al., 2019; Rabayrol & Hart, 2021). Changes in the subducting slab dynamics significantly affect plate dynamics, including plate motion and mantle dynamics (e.g., Gianni et al., 2019).

## 3 Geological background of Aegean-Anatolian Suture Zones

Determining the geological relationship between the present-day Hellenic arc and ancient subduction zones associated with the Aegean and Anatolian microplates is complicated due to the amalgamation of terranes that occurred over its tectonic history. The western portion of the Anatolian microplate exposes ophiolite remnants and other evidence of ancient subduction zones that may have influenced the onset of the present-day Hellenic arc. The Anatolian microplate's development is characterized by continental fragments separated by branches of the Paleo- and Neo-Tethyan oceans that collided and ultimately combined by the Late Cretaceous-Eocene. These exposures of ophiolitic and high-pressure/low-temperature (HP/LT) rock assemblages identify suture zones (Figure 5) (e.g., Şengör & Yılmaz, 1981; Okay, 2008; Moix et al., 2008; Okay & Tuysuz, 1999; Pourteau et al., 2016; Okay et al., 2020). This section describes the evidence of ancient subduction zones that may be linked to or influenced the development of the Hellenic arc.

### 3.1 Intra-Pontide Suture Zone

Western Anatolia contains the Pontide Mountains to the north and the Taurides Mountains to the south (Figure 5) (e.g., Şengör & Yılmaz, 1981; Yilmaz et al., 1997; Okay & Tuysuz, 1999; Pourteau et al., 2016). The Pontide Mountains include a southern Eurasia margin affinity and are characterized by a Pan-African basement with Phanerozoic sedimentary cover units (Yilmaz 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 (Istanbul–Zonguldak Zone, Istanbul Nappe, or Istanbul Zone, Yiğitbaş et al., 2004; Yilmaz et al., 2021). The IPS has been interpreted as 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). Its complex geology, comprised of components from the Istanbul-Zonguldak and/or Sakarya zones, has led to a debate about its presence and utility in paleogeographic reconstructions (Moix et al., 2008). Geochemical data from mafic rocks collected along the IPS show signatures related to a supra-subduction zone that opened in the Middle Jurassic-Early Cretaceous, obducted during the Early-Late Cretaceous boundary, and had closed by the Late Paleocene (Göncüoglu et al., 2014; Sayit et al., 2016).

In Greece, the Vardar suture connects to the IPS (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) or Meliata-Balkan suture (Stampfli, 2000). In Turkey's Biga Penisula, the IPS and Vardar connection may be recorded by an isolated ophiolite-bearing accretionary complex active until the Late Cretaceous (Figure 5) (e.g., Okay et al., 1991). However, this connection is debated by several researchers (Altunkaynak & Genc, 2008; Burchfiel et al., 2008; Şengün et al., 2011). Furthermore, due to the uncertain link between IPS and Izmir-Ankara-Erzincan Suture Zone (IAESZ), the relationship of the Biga Peninsula basement assemblages to those in the Rhodope-Thrace Massif is also debated (e.g., Bonev & Beccaletto, 2007; Elmas, 2012). Both the IPS and IAESZ are thought to mark Late Cretaceous–earliest Tertiary closure of Neo-Tethyan ocean basins (e.g., Pourteau et al., 2010; Akbayram et al., 2016).

### 3.2 Izmir-Ankara-Erzincan Suture Zone (IAESZ)

The IAESZ separates the Pontide's Sakarya Composite Terrane in the north from the Anatolide-Tauride block to the south (Figure 5) (Şengör & Yılmaz, 1981; Okay & Tüysüz, 1999; Tekin et al., 2002; Göncüoğlu, 2010; Yilmaz et al., 2021). 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) (Figure 1) (Channell & Kozur, 1997; Okay & Tuysuz, 1999; Tekin et al., 2002; Moix et al., 2008). However, the feature is not exposed due to the presence of the Aegean Sea (e.g., Burtman, 1994; Stampfli, 2000; Yılmaz et al., 2001; Burchfiel et al., 2008).

In Western Anatolia, blueschist assemblages exposed along the IAESZ are intruded by Suture Zone Granitoids (SZGs) [Topuk, Orhaneli, Tepeldag (Gürgenyayla and Gürgenyayla)] with Paleocene (63.5±2.8 Ma) to Oligocene (31.4±0.6 Ma) ages but primarily characterized by early Eocene emplacement (~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 Tavşanlı Zone is narrow (~50 km) and trends E-W for approximately 250-350 km (Okay & Whitney, 2010; Plunder et al., 2013). Its 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 zone is thought to have formed as a result of northward-dipping subduction and represents the Mesozoic-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 Sivrihisar Massif in the eastern portion of the Tavşanlı has eclogite, blueschist, and Barrovian sequences (Gautier, 1984; Seaton et al., 2009). Paleocene-Eocene ages from the Tavşanlı Zone granites mark the timing of the closure of the IAESZ (e.g., Okay et al., 2020).

The Afyon zone (Afyon–Bolkardag Zone of Okay, 1986; Özdamar et al., 2013 or Ören–Afyon Zone of Pourteau et al., 2013) 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 its southern extent is unclear, the Afyon zone is mapped parallel to the Tavşanlı Zone. 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). Afyon Zone stratigraphy resembles the Tavşanlı Zone and consists of Palaeozoic-Mesozoic Tauride shelf sequences of metasedimentary and metavolcanic rocks that experienced regional greenschist to blueschist facies (Fe–Mg carpholite and glaucophane) metamorphism overlying Pan-African-related basement (Okay, 1984; Candan et al., 2005, Pourteau et al., 2010; Özdamar et al., 2013). HP/LT metamorphism in the Afyon Zone is thought to have occurred coincident with the closure of the Neo-Tethyan Ocean at 70-65 Ma (Pourteau et al., 2010, 2013; Özdamar et al., 2013; Plunder et al., 2013). Portions of the Afyon Zone may have subducted beneath the Tavşanlı Zone during the Late Cretaceous based on zircon ages of granites that intrude Tavşanlı Zone blueschist and altered ophiolitic assemblages (Speciale et al., 2012; Shin et al., 2013).

## 4. Age constraints on the initiation of subduction

### 4.1 Cenozoic estimates

Early estimates for the initiation age of Hellenic arc subduction were Late Miocene to Pliocene (13±3-5 Ma, Le Pichon & Angelier, 1979 and 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 earliest volcanic activity in the South Aegean arc (Pliocene, Pe-Piper & Piper, 2005) and the onset of the KTZ based on geodynamic modeling and GPS data (6-8 Ma, Royden & Papanikolaou, 2011). The Late Miocene time frame is also cited as when the Western Hellenic arc divided into its northern and southern portions along the KTZ (Papanikolaou, 2010). Reconstructions of fault systems in the northern margin of the eastern Mediterranean Sea are also consistent with 15 Ma (Le Pichon et al., 2019).

Interpretations of the Aegean seismic velocity structure, tomography, and seismicity data in the Aegean area suggest Eocene to Late Oligocene onset of the Hellenic arc subduction zone (40-26 Ma; Meulenkamp et al., 1988; Spakman et al., 1988; Papadopoulos, 1997; Brun & Sokoutis, 2010). These ages are consistent with the timing of granite crystallization throughout Western Anatolia and Mediterranean Ridge sedimentation (33-23.6 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 from the Western portion of the Hellenic arc experienced the onset of subduction from 34-27 Ma. In contrast, the Southern Hellenic arc was active at 34 Ma (Royden & Papanikolaou, 2011).

The progressive deceleration in motion of Africa with respect to Europe in the Mediterranean region has been observed to occur 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 may have accelerated from ~0.6 cm/yr during the first 30 m.y. of subduction to 3.2 cm/yr during the past 15 m.y., perhaps due to slab tear during the Middle Miocene-Pliocene (Brun et al., 2017). 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 obliquity of subduction over time, which was accommodated by strain partitioning within the upper Eurasian plate (Faccenna et al., 2013; Philippon et al., 2014; Brun et al., 2016; Cornée et al., 2018). By the Early Oligocene-Eocene, back-arc basin formation was ongoing in the Aegean region as the African plate retreated as it subducted beneath Eurasia (Le Pichon & Angelier, 1981; Royden, 1993; Jolivet & Faccenna, 2000; Agostini et al., 2010; Carminati et al., 2012; Jolivet et al., 2018).

Figures 6 and 7 are N-S temporal cross-sections across Western Anatolia that describe the region's tectonic evolution from the Late Cretaceous to the present. Figure 6A (from Dilek & Altunkaynak, 2009) illustrates the closure of a branch of the Neo-Tethyan Ocean (Izmir-Ankara Ocean), which occurred during the Late Cretaceous when the IPS already marked the division between the Sakarya and Rhodope continents. Break off of the subducting Neo-Tethyan slab occurs ca. 54 Ma along the IAESZ (Figure 6B), concurrent with the development and emplacement of SZG and ophiolite obduction. The Hellenic arc forms due to subduction step-back processes that are completed by the Miocene (~25 Ma) (Figure 6C). Subduction rollback was most likely initiated soon after the complete formation of the Hellenic Arc to start Aegean back-arc extension (Figure 6D). Some variations of this model exist. For example, Shin et al. (2013) proposed an additional subduction system concurrent with the IAESZ but located further south between the IAESZ and the Hellenic arc (Figure 7).

These models are the outcome of large data sets comprised of primarily field observations, geophysical, geochemical, petrological, geochronological, and structural data from multiple rock types, but primarily those affiliated with the upper mantle and continental crust (lithosphere). They provide testable hypotheses for the timing and generation mechanisms of particular magmatic and structural features in the Aegean and western Anatolia to predict the presence of multiple suture zones and subducted slabs. They also delineate the affinity of specific crustal fragments north of the Hellenic arc to be distinct from the African plate in terms of their origin and geological history.

### 4.2 Mesozoic estimates

Significantly older estimates for the onset of the Hellenic subduction exist and suggest the arc has been long-lived since the Jurassic-Late Cretaceous, with stipulations that Cenozoic age estimates are considering only the latest episode of its evolutionary history and disregard its earlier history (Figure 8) (e.g., Jolivet et al., 2018). In this model, the African slab has been part of a single, evolving subduction zone system that was initiated sometime in the Mesozoic during the initiation of the closure of the northern branch of the Neo-Tethys ocean (Faccenna et al., 2003; van Hinsbergen et al., 2005; Royden & Papanikolaou, 2011; Jolivet et al., 2013; Malandri et al., 2017; Jolivet et al., 2018; Crameri et al., 2020; van Hinsbergen et al., 2021). The model has been primarily applied to explain the evolution of the Western Hellenic arc. 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. This process triggered southward slab rollback and migration of the volcanic arc (Figure 8) (van Hinsbergen et al., 2005; Brun & Faccenna 2008; Jolivet & Brun 2010; Jolivet et al., 2013; Cornée et al., 2018). In this model, the subduction zone accumulates tectonic fragments, terrains, and oceanic lithosphere, instead of following the induced transference model of subduction (Figures 6 and 7; e.g., Stern, 2004). In this model, a suture is created, and a new subduction zone develops. Instead, the Hellenic subduction zone front has behaved like an accretionary wedge during its entire development since the Jurassic-Late Cretaceous.

The model also requires a stationary trench from 100-45 Ma to allow for the penetration of the African slab into mantle depths (Bonneau, 1982; Stampfli & Borel, 2002; Jolivet et al., 2003; Capitanio et al., 2010). Penetration to 1200-1400 km depths is estimated to have occurred by ~50 Ma, which triggers the onset of extension in the region by 45 Ma (Jolivet & Faccenna, 2000; Brun & Faccenna, 2008; Capitanio et al., 2009; 2010). The present-day curvature of the Hellenic forearc represents oblique subduction that grew systematically (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 model was developed primarily using interpretations of mantle tomography coupled with estimates of subduction rates over long periods of geological time. Thus, ages found in suture zones that would be entirely unrelated to the Hellenic arc in early SZI models now time the onset of subduction of the African slab. For example, the Late Cretaceous (104 Ma) estimate of Hellenic arc SZI is based on the ages of ophiolites exposed in the Tavsanli Zone of Western Anatolia, part of the IAESZ (Peters et al., 2017; Pourteau et al., 2019; Crameri et al., 2020). The single subduction system also has implications for the affinity of Aegean crustal fragments and terranes in paleogeographical reconstructions. The model requires all the lower plate continental crust to be accreted into the upper plate while subducting continental lithosphere. It also requires the entire Aegean crust from the Vardar (Izmir-Ankara) suture to the Mediterranean ridge to be derived from the lower plate (Figure 6-8). Oceans between the accreted domains were of significant size (500 km in some cases). The process would lead to significant elevation changes and crustal thicknesses that should be recorded in the geological and tomographic record. Critical changes in the zone of subduction transitions also occurred as shear zones developed to accommodate oceanic and continental lithosphere (see discussion in Le Pichon et al., 2019). The model also 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 (van Hinsbergen et al., 2005).

## 5 Discussion

Plate interactions are thought to be most clear at convergent plate boundaries (Silver et al., 1998). However, several outstanding issues remain regarding timing the onset of convergence between the Nubian Plate and Aegean microplate. Several observations indicate that the Aegean microplate does not behave like a typical back-arc basin (Agostini et al., 2010; Doglioni et al., 2002). It lacks an ocean floor and associated ocean crust and is underlain by a thick layer of continental crust (e.g., Makris, 1978). In fact, the Aegean crust is basically comprised of stacked carbonate and oceanic nappes, metamorphic massifs, and core complexes (Jolivet et al. 2013; Papazachos, 2019). Models of deformation along the Hellenic arc intended to capture the large-scale 3-D structure of the Nubian plate suggest the region more closely resembles a continental thrust than a typical oceanic subduction zone (Ganas and Parsons, 2009).

The Aegean microplate is also bound to the north by the active strike-slip systems (NAF, NASZ, KTZ, and Central Hellenic Shear Zone; Brooks & Ferentinos, 1984; Lyberis, 1984; Nyst & Thatcher, 2004; Gürer et al., 2006; Kreemer et al., 2004; Kokkalas et al., 2006; Papanikolaou & Royden, 2007; Reilinger et al., 2010; Royden & Papanikolaou, 2011; Vassilakis et al., 2011; Halpaap et al., 2018). This results in 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). 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). Subduction zones behave chaotically and may retreat advance or remain stationary at different stages, especially if the incoming lithosphere is heterogeneous (e.g., Royden & Husson 2009; Husson et al., 2009). The extent and role of several ancient fault systems, including older Neo-Tethyan suture zones that likely played a significant role in its development, are thus debated.

The Mesozoic SZI age for the Hellenic arc requires all of the lower plate continental crust to be accreted into the upper plate while subducting continental lithosphere. The entire Aegean Crust from the Vardar suture to the Mediterranean ridge would be derived from the lower plate (Figure 8). This process requires a redefinition of components associated with ancient supercontinents and would impact paleogeographic reconstructions of the region. Oceans between the accreted domains were of significant size (500 km in some cases). The process would lead not only to changes in elevation and crustal thicknesses, but also critical changes in the zone of subduction transitions occurring from oceanic to continental shear zones (see discussion in Le Pichon et al., 2019) (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).

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). The requirement of a stationary trench from 100-45 Ma to allow for the African slab to enter mantle depths (Bonneau, 1982; Stampfli & Borel, 2002; Jolivet et al., 2003; Capitanio et al., 2010) seems unusual in that most of the slab today is characterized by significant slab tear. Sites of slab tear vary in scale from regional to local and include the boundary between the Hellenic and Cyprus arcs (Wortel & Spakman, 1992; Biryol et al., 2011), south of Crete at the Pliny–Strabo Shear Zone (Özbakır et al., 2013), the İzmir-Balıkesir Transfer Zone (e.g., Kaya, 1981; Gessner et al., 2013; Cemen et al., 2014; Ersoy et al., 2014), and beneath the Menderes Massif (Biryol et al., 2011; Rabayrol & Hart, 2021), and the eastern Aegean (Jolivet et al., 2015).

At present, it is unclear how the subducting African (Nubian) slab could survive through the Earth's transition zone, suggesting that its composition and convergent rate, and the material properties of the mantle were critical constraints to allow this process to occur (e.g., Kárason & van der Hilst, 2000). Interpretations of the deeper images are debated: the subducted slab may be a single folded body that overturned in the lower mantle (Faccenna et al., 2003) or two different slabs between 2000-1500 km and from 1500 km to the surface (van Hinsbergen et al., 2005; van der Meer et al., 2018). The unhindered slab penetration into lower mantle depths is unexpected due to the relationship between slab morphology and trench migration (e.g., Widiyantoro et al., 2004). However, the process of slabs reaching depths of 1700 km is not uncommon (e.g., Kárason & van der Hilst, 2000).

Numerous examples also suggest significant slab tear across the Aegean and Western Anatolia (e.g., Wortel & Spakman, 1992; Woodside et al., 2000; Biryol et al., 2011; Gessner et al., 2013; Faccenna et al., 2014; Roche et al., 2019; Rabayrol & Hart, 2021). Legrande et al. (2012) note that body wave studies show slab segments are trapped within the mantle transition zone in the western Mediterranean and Alpine region and do not dive deeper as is observed in the Aegean region. Overall, the Mediterranean Sea exhibits lateral variations in crustal and upper mantle structures (Marone et al., 2004a).

Long-lived Aegean subduction would have affected the mantle's thermal and chemical convection and impacted surface conditions, including climate over time (e.g., Sigl et al., 2005). The region contains active and quiescent volcanoes located in the Methana Peninsula (Peloponnese) and islands of the Cycladic archipelago (Figure 1) (see review in Scoon, 2021). Holocene volcanism along the Hellenic arc allows for identifying and characterizing high-frequency climatic changes during the Holocene in particular (e.g., Zanchetta et al., 2011). Fluids released from past and present subducting lithosphere have affected not only volcanism, but also grain mobility and seismic anisotropy throughout the Mediterranean and elsewhere (e.g., Karato, 1995; Van der Meijde et al., 2003; Marone et al., 2004b; Faccenda et al., 2008). This impact over the length of planetary history may have been significant in terms of triggering other subduction zones, controlling sites of seismicity and magmatism, as well as the generation of ore deposits. A Jurassic-age subduction zone that is presently active and capable of generating significant hazards even today has important implications for our understanding of lithospheric strength, the driving force of plate tectonics over geological time, and the tectonic history of the Aegean region.

The single subduction system has its roots in tomography and the interpretation of large low shear-wave velocity anomalies in the lowermost mantle. These compositional anomalies thus appear as crucial as continents are for the surface, yet they are inaccessible and have limited data regarding their thermal histories and chemical compositions. Their role as reservoirs and impact on the evolution of the Hellenic arc over time is unclear. The interpretation that the seismic tomography images reveal a single, long-lived subduction system may be incorrect. The Aegean-Anatolian microplates have long been known to have an extensive and complex orogenic history based on surface geology. With the presence and data from mapped suture zones, it is expected that the region would record multiple subducted slabs from earlier collisional events.

Determining the Hellenic arc SZI is also important because the timing constraints shed light on the mode of initiation (i.e., induced or spontaneous; from compression or extension; Gurnis et al., 2004; Stern, 2004; Gerya, 2011; Stern & Gerya, 2018) based on our understanding of the geology at that time. Subduction initiation requires breaking a zone of weakness in the lithosphere (e.g., Stern & Gerya, 2018; Baes et al., 2020a; Cloetingh et al., 2021). The process may be accommodated along passive margins (e.g., Cloetingh et al., 1984), although the conversion of a passive margin into an active one has not been observed to have occurred during the Cenozoic (Gurnis et al., 2004; Cloetingh et al., 2021). Subduction initiation could also occur due to motion along transform or fracture zones associated with mid-ocean ridges due to large-scale changes in plate motion (e.g., Hall et al., 2003). This process could occur across extinct mid-ocean ridges or within back-arc regions within mature subduction zones, as well as mantle plume interactions with old oceanic lithosphere (Baes et al., 2020a,b; Cloetingh et al., 2021). No evidence or hypothesis exists for a mantle plume as triggering the development of the Hellenic arc. A young SZI initiation age suggests the Hellenic arc developed due to the Late Cenozoic collision of the African plate with Eurasia (e.g., Meulenkamp et al., 1988). An older SZI links its origination to near and parallel to a spreading ridge associated with the western portion of the NeoTethyan ocean (e.g., Maffione and van Hinsbergen, 2018).

Younger estimates for the Hellenic arc are consistent with the observation that nearly half of Earth’s active subduction zones were initiated during the Cenozoic (Gurnis et al., 2004). Crameri et al. (2020) indicate that SZI has happened at least thirteen times during the last hundred million years, with global occurrences of SZI occurring within 55-40 Ma and 16-6 Ma. These times mark higher SZI activity on Earth and are remarkably consistent with Cenozoic estimates of the onset of the Hellenic arc (McKenzie, 1978; Le Pichon & Angelier, 1979; Mercier, 1981; Meulenkamp et al., 1988; Spakman et al., 1988; Papadopoulos, 1997; Brun & Sokoutis, 2010; Royden & Papanikolaou, 2011; Le Pichon et al., 2019). A significant slowdown between Africa’s northward motion relative to surrounding plates has also been attributed to the Eocene-Oligocene collision of Africa and Eurasia, which altered the kinematics and tectonic evolution of not only the Aegean region but the restructuring, reorganization, and fracturing of the African plate elsewhere (e.g., DeMets & Merkouriev, 2016).

## 6 Conclusions

The Hellenic arc results from the convergence of the northern margin of the African (Nubian) plate with the southern portion of the Aegean microplate. The region has emerged as the type locality for subduction zone tectonics, and many novel ideas regarding the development of this particular plate boundary have emerged from the area. Two end-member models for the SZI age of the Hellenic arc evolved using disparate datasets. Seismic tomography images help determine the present-day subduction-related structure beneath the Aegean microplate, and some interpretations suggest a long-lived subduction zone since the Jurassic to Late Cretaceous. However, tomographic images can be limited in their interpretations of deeper depths and our ability to observe slab tears or structures that disrupt the subducting slab. Surface geological data sets suggest a complex, long-lived history of the Aegean region with multiple suture zones. These observations are more consistent with multiple detached slabs, even some unrelated to the African plate. Geological, geochemical, and geochronological constraints from the continental lithosphere affected by subduction help constrain the pathways of rocks affected by subduction dynamics. However, these are limited by the currently available datasets and interpretation within an existing framework with limited exposure. The Hellenic arc is a significant plate boundary that would likely be influenced by differential retreat, inherited lower plate lithospheric heterogeneities, and mantle upwelling. Multidisciplinary studies that couple detailed field mapping of available outcrops in the Western Anatolian and Aegean microplates with geophysical observations, geochemistry, and geochronology are needed to determine its SZI age and which model, a long-lived single subduction zone or multiple subduction zones with the latest SZI in Cenozoic, applies to the region.

## Acknowledgments

No real or perceived financial conflicts of interests exist for the authors. We appreciate the comments from two anonymous reviewers. We thank Jeffrey S. Horowitz for assistance with drafting figures.

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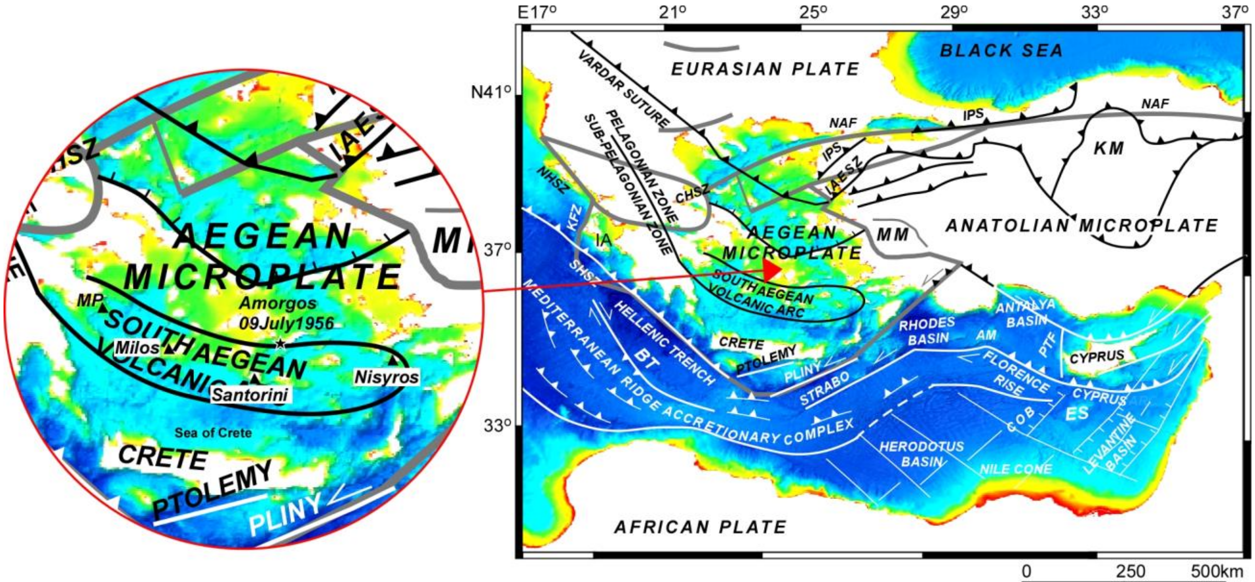



**Figure 1**. EMODnet Digital Bathymetry maps with some structures overlain. The circular inset shows higher magnification of the region as indicated by the arrow. The Aegean and Anatolian microplate boundaries are shown in grey after Nyst & Thatcher (2004). Other structures after Hall et al. (1984) and (2009), Woodside et al. (2002), Peterek & Schwarze (2004), Meier et al. (2007), Kinnaird & Robertson (2012), and Symeou et al. (2018). Abbreviations: AM= Anaximander Mountains; BT= Backthrust; ES = Eratosthenes Seamount; IAESZ = Izmir-Ankara-Erzincan Suture Zone; IPS= Intra-Pontide Suture; KFZ = Kephalonia Fault Zone; KM= Kirşehir Massif; MP = Methana Peninsula; MM= Menderes Massif; NHSZ = North Hellenic Shear Zone; PTF = Paphos Transform Fault; SHSZ = South Hellenic Shear Zone. See supplementary documents for this figure in color.

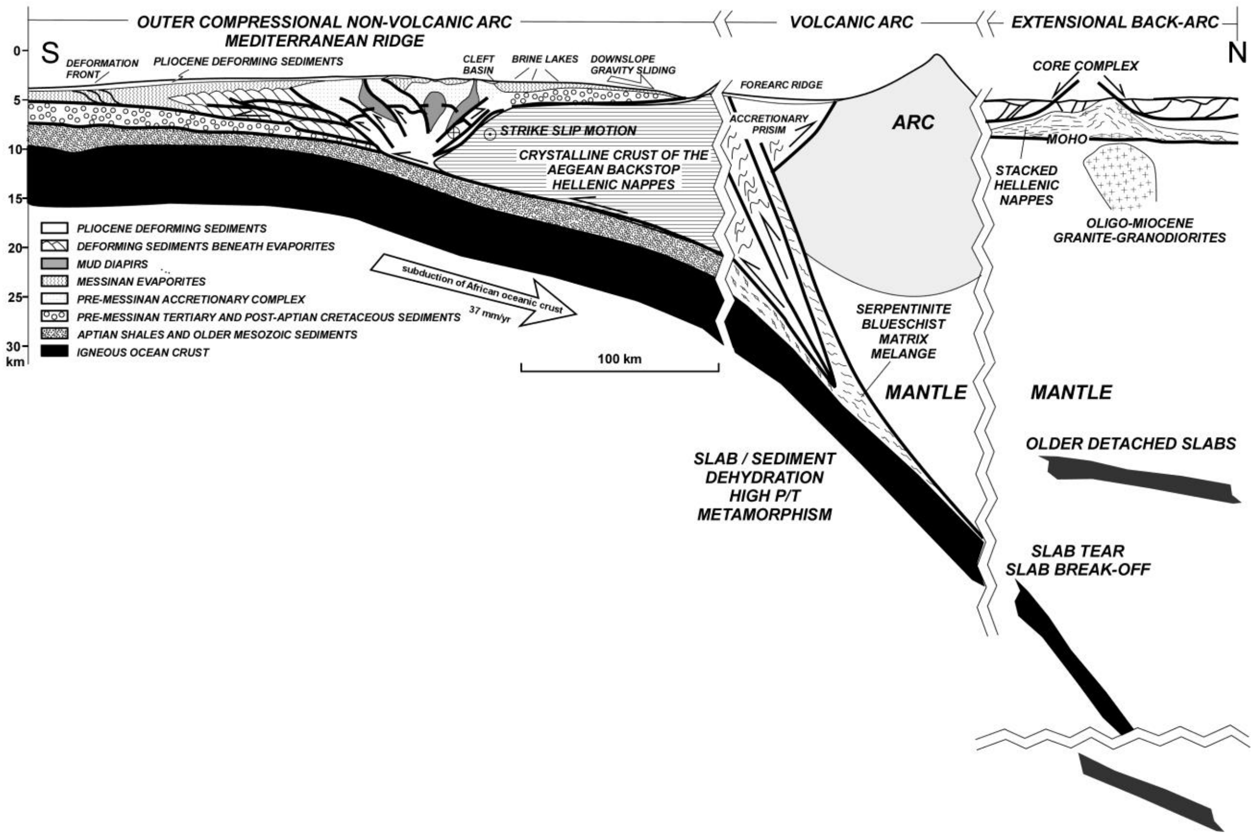
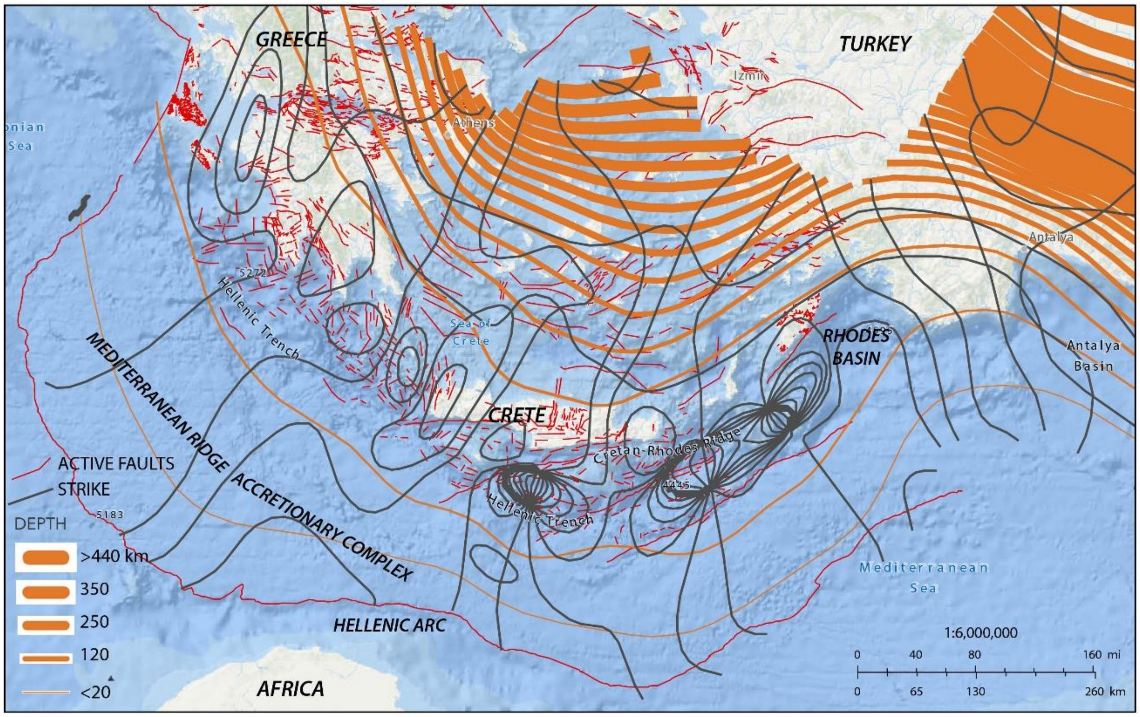



**Figure 2**. North-south generalized cross-section through the Hellenic arc system showing the key structural elements. Map of the Mediterranean Ridge after Westbrook & Reston (2002).




Esri, GEBCO, DeLorme, NaturalVue, Esri, HERE, Garmin, FAO, NOAA,

**Figure 3.** Map of the geometry of the Hellenic and part of the Cyprus subduction zones after Hayes (2018). The figure shows both the depth (up to 440 km) and strike of the subducted African plate (African, Nubian, Aegean, or Hellenic slab) based on the Slab2 model. Overlain in red are fault systems from NOAFaults, a digital database of active faults in Greece (Ganas et al., 2013). Countries and some features are indicated. See supplementary documents for this figure in color.

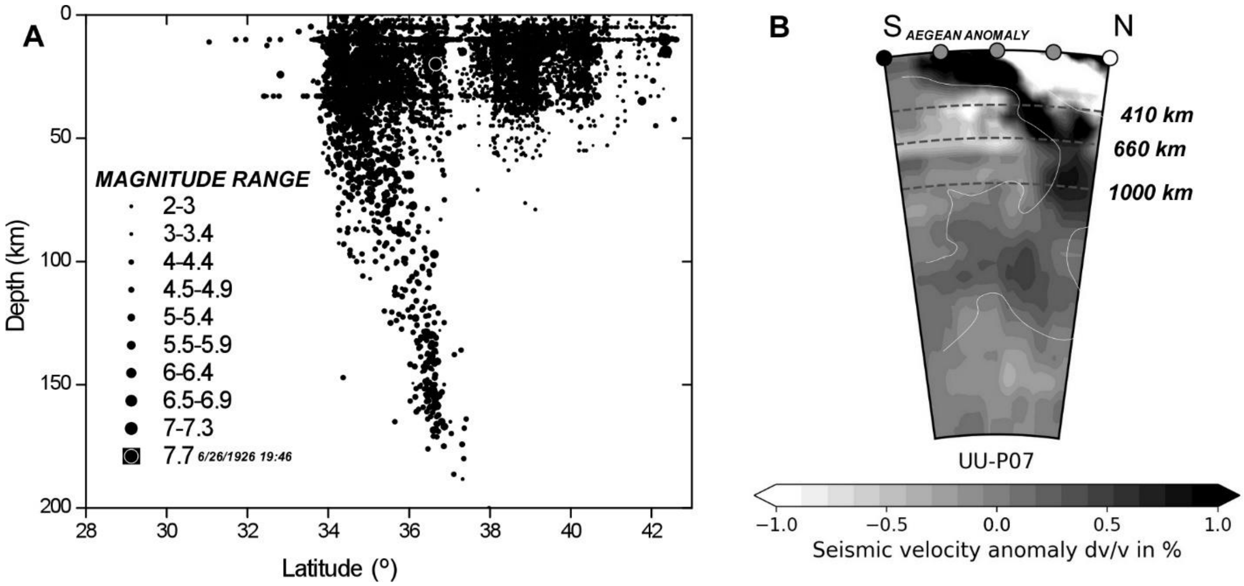



**Figure 4.** (A) Depth vs. latitude of earthquakes taken from a line of the section of 28°-43° and longitude of 24°-28°. Events were extracted from the Turkish Ministry of the Interior, Disaster and Emergency Management Presidency, Earthquake Department Earthquake Catalog (M>=4.0), 1900-20XX (https://deprem.afad.gov.tr/depremkatalogu) from 01/24/1900 to 6/17/2021. We indicate the largest event (6/26/1926, 19:46). The legend shows how the size of the earthquake correlates to the symbol. (B) Cross-section of the Aegean anomaly interpreted as the African slab using the UUP07 P-wave model (Amaru, 2007). 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). See text for references to additional tomographic images.

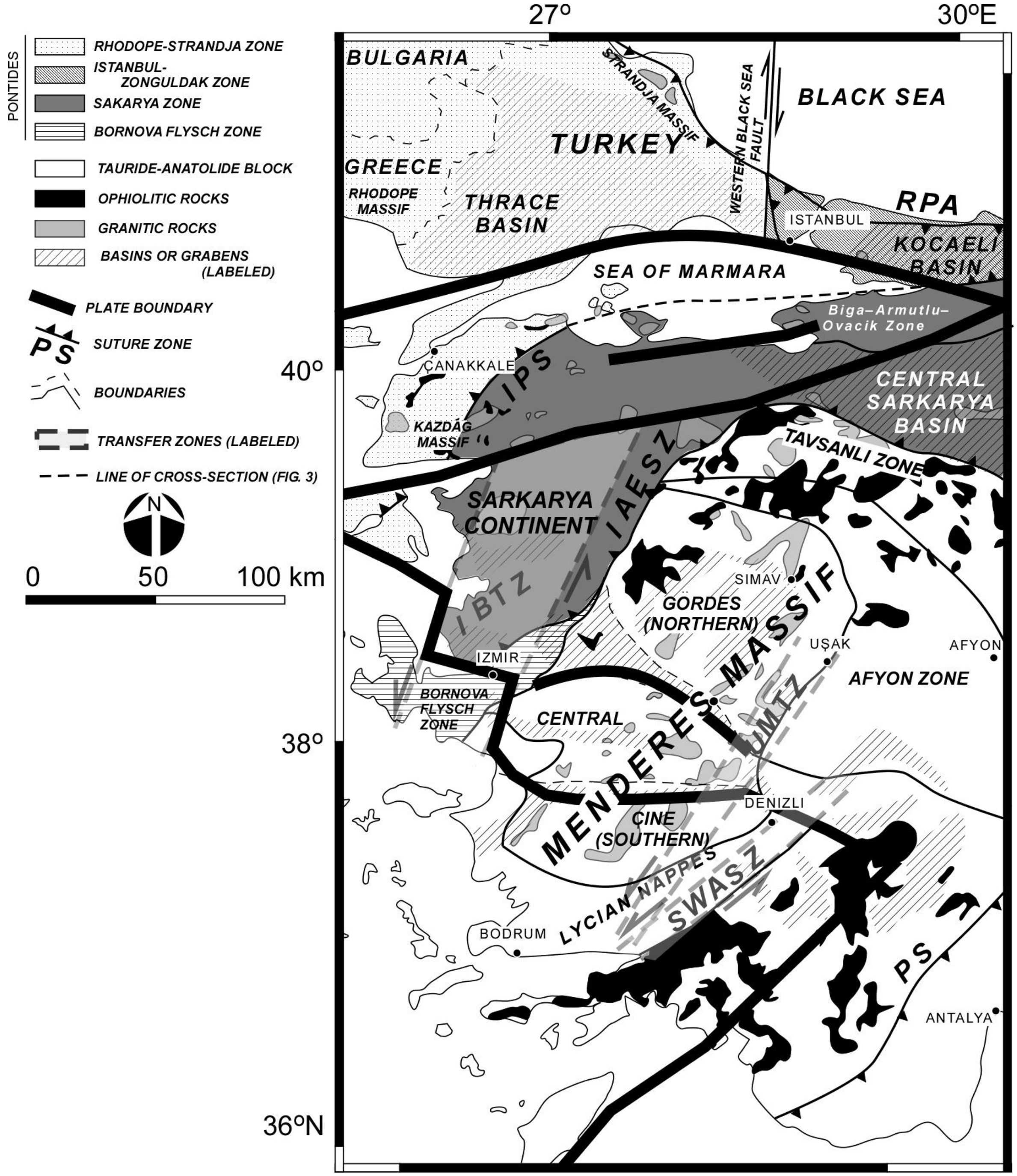



**Figure 5**. Geological map of Western Anatolia focusing on the ophiolite and granite assemblages along the boundary between the Aegean and Anatolia microplates. Plate boundary after Nyst & Thatcher (2004). Terrane boundaries, major fault systems, and transfer zones after Okay (2008), Akbayram et al. (2016), Oner et al. (2010), and Karaoğlu & Helvacı (2014). Abbreviations: RPA= Rhodope -Pontide Arc; İBTZ = 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.

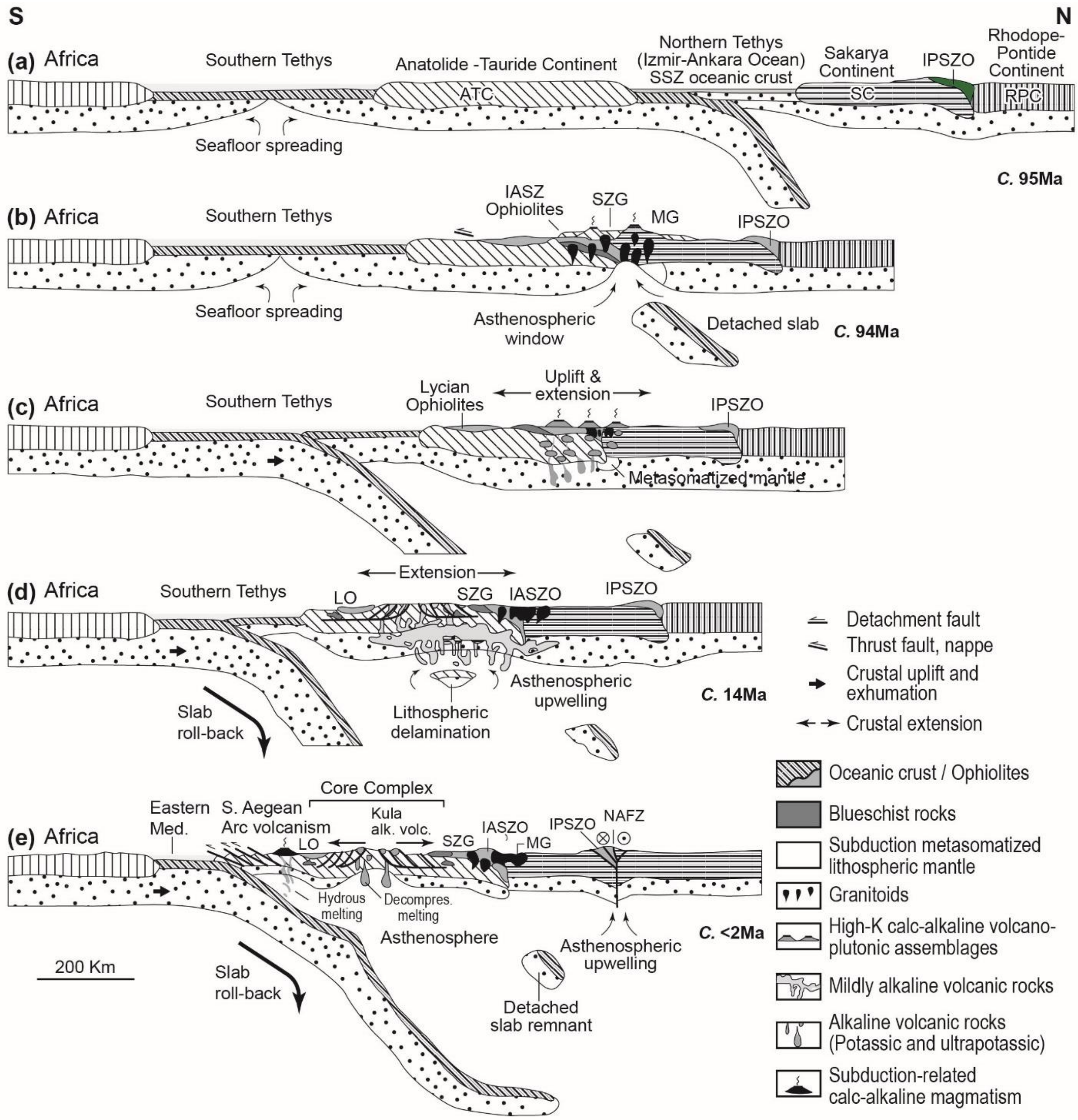



**Figure 6.** North-south cross section across the Aegean region from c. 95 Ma to <2 Ma. In this scenario, the Hellenic arc forms due to the transition from a divergent plate boundary to convergence sometime between 54 Ma and 25 Ma (Dilek & Altunkaynak, 2007). Scale is only a rough approximation. See Dilek & Altunkaynak (2007) for a color image of this figure. Abbreviations: ATC= Anatolide-Tauride Continent; IASZ= Izmir-Ankara Suture Zone; IPSZO= Intra-Pontide Suture Zone ophiolites; MG= Magmatic Granites; NAFZ= North Anatolian Fault Zone; SC = Sakarya Continent; SSZ = supra-subduction zone; SZG = Suture Zone Granitoids; RPC= Rhodope-Pontide Continent.

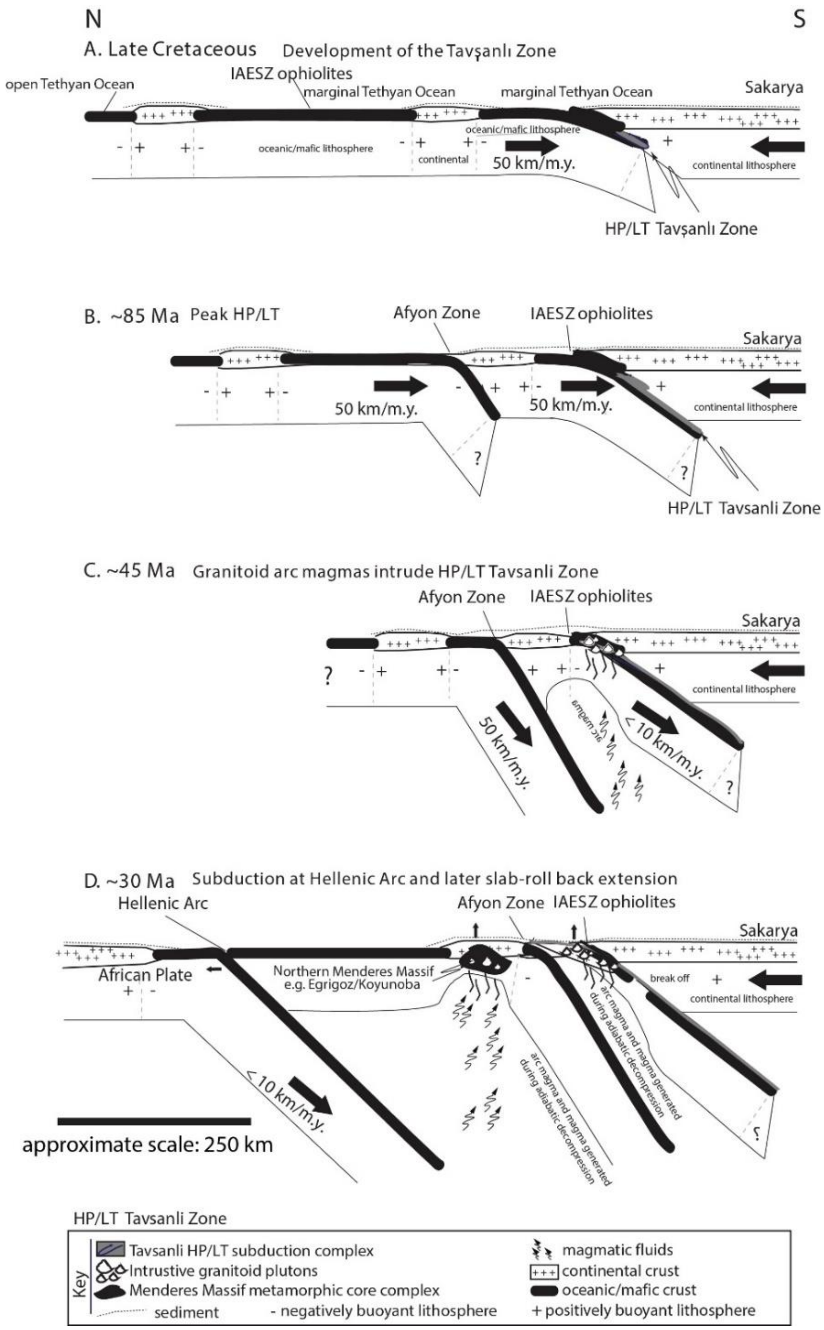



**Figure 7.** Lithospheric-scale cross sections of a model of the Aegean region through Western Anatolia in which the region records and experiences progressively younger subduction zones after Shin et al. (2013). The subduction zone cartoon is adapted from Cloos et al. (2005). Scale is only a rough approximation. Panel (A) shows the development of the IAESZ during the Early to Late Cretaceous. Panel (B) records the peak metamorphism of the zone and the development of the Afyon Zone subduction zone at ~85 Ma. Panel (C) shows granitoid arc magmas from the Afyon Zone intrude the IAESZ at ~45 Ma, emplacing granite bodies that intrude the ophiolite sequences. Panel (D) records the break off of the IAESZ slab and additional magma emplacement at 30 Ma. In this scenario, the Hellenic Arc experiences ongoing subduction.

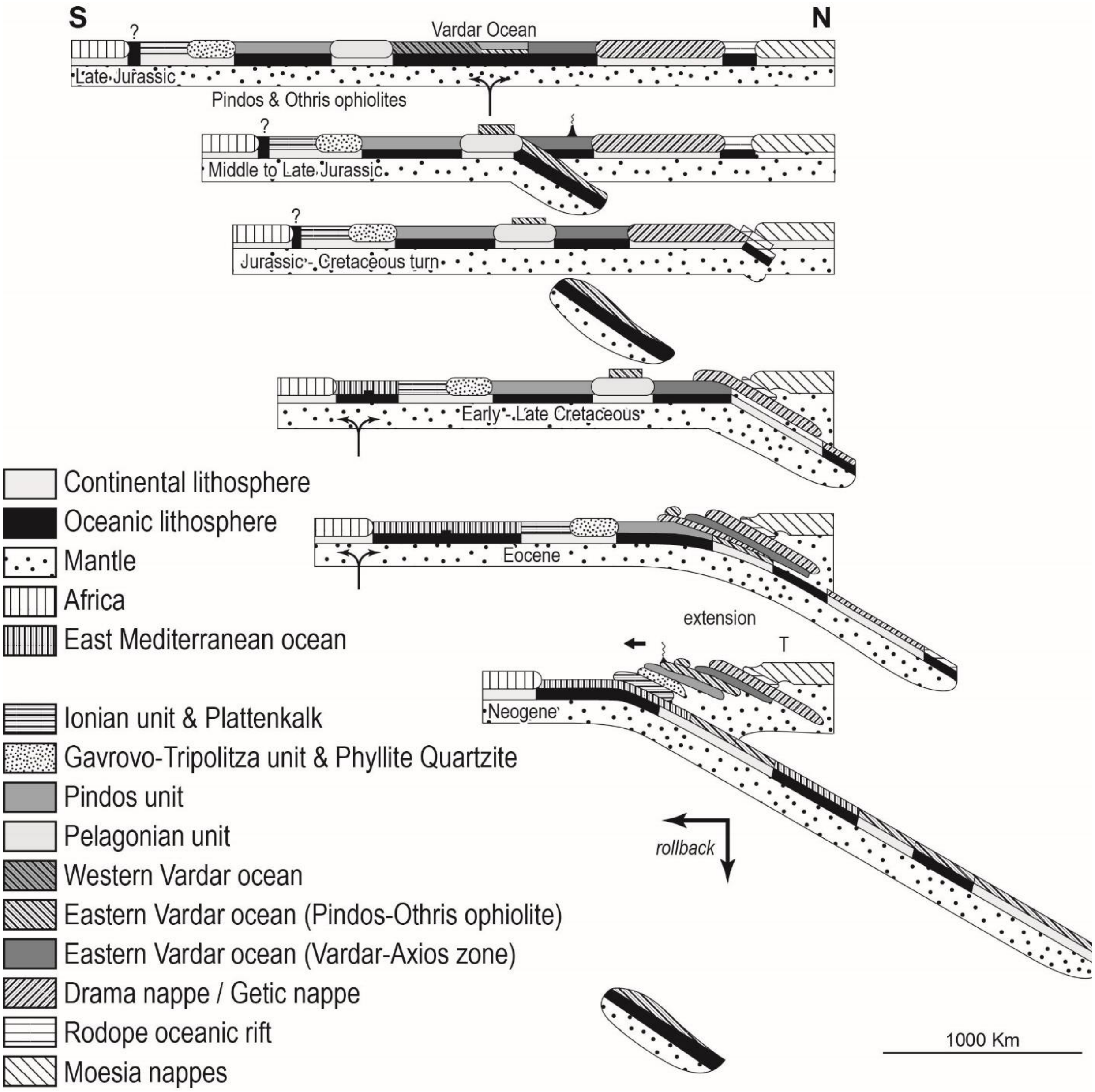



**Figure 8.** Schematic overview of development of nappe stack and subduction during Alpine orogeny in Greece from van Hinsbergen et al. (2005). The cross-section displays the single subduction zone model for the development of the Hellenic arc. See van Hinsbergen et al. (2005) for a color version of this figure.