

Sedimentary petrology and geochemistry of the siliciclastic rocks from the Isparta Angle area (SW Türkiye): Implications for provenance and tectonic settings

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Abstract The Antalya Complex within the Isparta Angle (IA) in the southwest of Türkiye includes various rock types formed in different geodynamic settings and contains important data that shed light on the geodynamic evolution of the Eastern Mediterranean region. In this study, source rocks and tectonic settings of ancient sedimentary basins within the IA have been interpreted using geochemical contents of siliciclastic rocks in different ages. In this regard, whole-rock major, trace and rare earth element (REE) analyses have been conducted on the Cenozoic turbiditic sandstones (from the Eocene Isparta Formation and the Miocene Güneyce Formation in the north of the IA) and integrated with the available data of the Triassic shales from the east and Upper Cretaceous sandstones from the southwest of the IA. The results indicate that Cenozoic sandstone samples are calcium-rich (CaO, 30.54–43.87 wt%, average 37.63 wt%), comparable to those of Upper Cretaceous sandstones (CaO, average 24.92 wt%) and Late Triassic (CaO, range between 11.27–26.71 wt%) shales.

They demonstrate an excellent carbonate dilution effect on the majority of major oxides. In contrast to the Late Triassic shales with close affinity to Post-Archean Average Archean Shale (PAAS), Cenozoic and Upper Cretaceous ophiolite-related clastic sediments display strike depletions in Nb–Ta, Th, Rb, Zr and Hf and enrichments in Ni, Co and Cr, suggesting mafic and ultramafic rocks in a source region. Notable enrichments of Cr (average 80.5 ppm), V (average 105.5 ppm), Ni (average 42.3 ppm) and Co (average 13.9 ppm) in Late Triassic shales can be attributed to the influences of Triassic OIB-like basaltic rift magmatism in their source regions. In general, their igneous source rocks were derived from mixed sources, e.g., (1) predominantly continental felsic and lesser rift-related basaltic mafic source rocks for Late Triassic shales, (2) mainly ophiolitic mafic-ultramafic and lesser continental felsic source rocks for Upper Cretaceous sandstones and (3) primarily continental felsic and lesser ophiolitic mafic-ultramafic source rocks for Cenozoic sandstones. These inferences also imply that the distinct parent rock compositions of the Late Triassic, Upper Cretaceous and Cenozoic aged siliciclastic rock samples reflect their distinct tectonic settings such as the passive rift setting of Late Triassic shales, an SSZ setting for the Upper Cretaceous sandstones and a collisional setting of the Cenozoic sandstones, respectively.

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1 Introduction

The southwestern part of Türkiye, characterised by autochthonous and ophiolite-bearing allochthonous rock assemblages, is an extremely complicated area with the structural

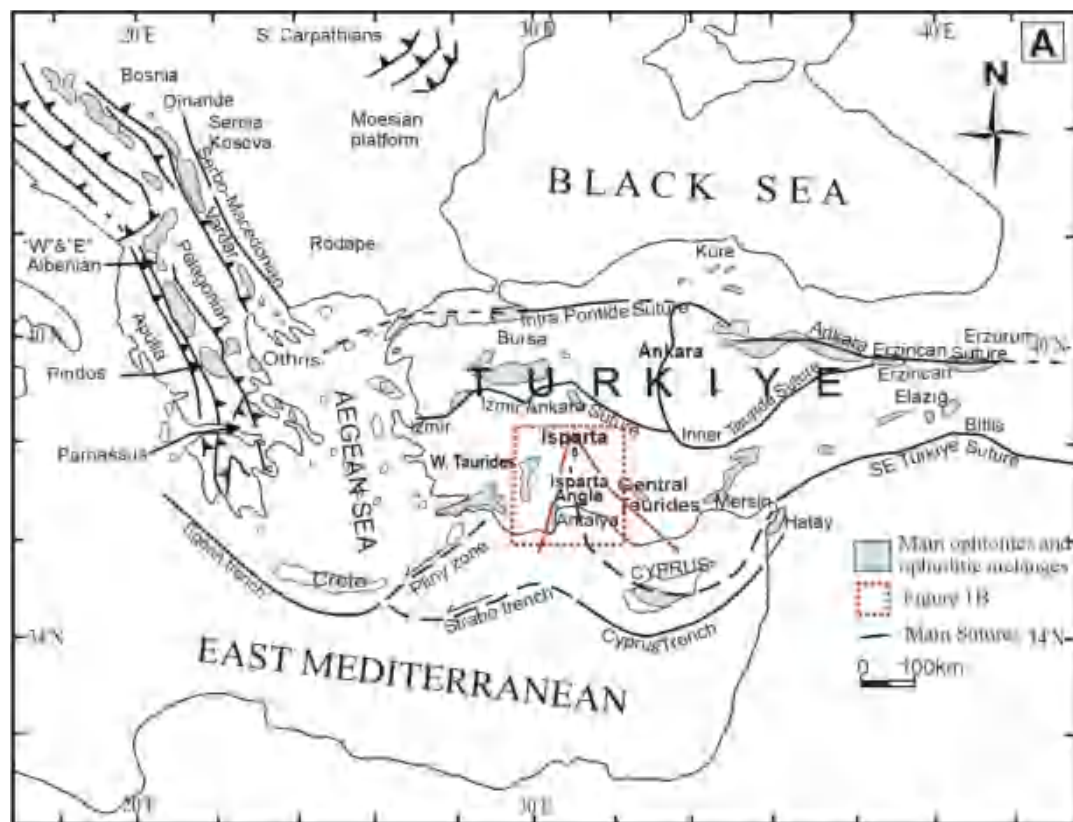
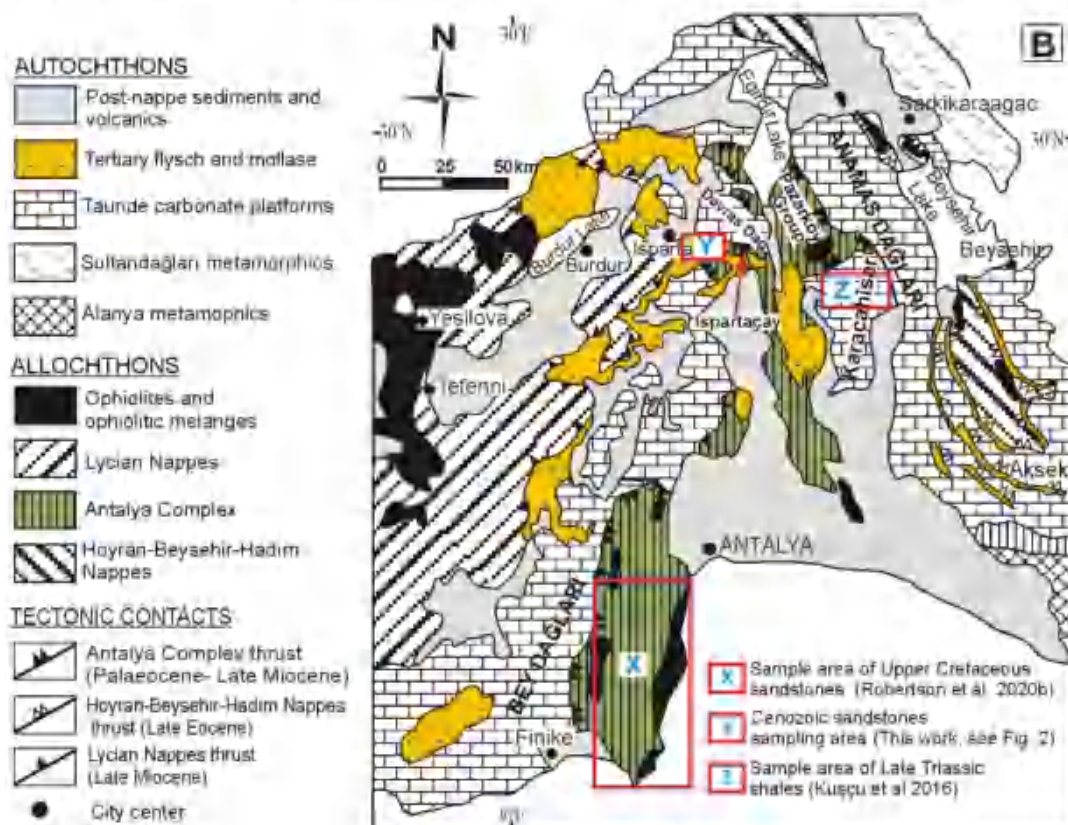


Fig. 1 **A** Tectonic map of the East Mediterranean showing distribution of main ophiolites, major suture and tectonic zones in Türkiye (remodified from Robertson et al. 2006 and Dilek and Sandvol 2009, references therein). **B** Simplified geological map of the Isparta Angle (remodified from Robertson, 1993, references therein). Sample areas of Upper Cretaceous sandstones (Robertson et al. 2020b) (X), Cenozoic (Eocene to Miocene) sandstones (this work) (Y) and Late Triassic shales (Kuşçu et al., 2016) (Z) are also shown.

complexity of the allochthonous units, i.e., the western Lycian Nappes, the Antalya Complex (AC) and the eastern Beyşehir-Hoyran-Hadim Nappes (Koçyiğit et al. 2024). The Lycian and the Beyşehir-Hoyran-Hadim Nappes are separated from the Antalya Complex by the western part of the Tauride carbonate axis, which forms a north-pointing cusp in the north of the Gulf of Antalya, the so-called Isparta Angle (IA). The IA area is one of the key regions for understanding the paleotectonic evolution of the eastern Mediterranean region. The rock assemblages in the IA area formed as a result of different tectonic processes, including continental rifting, passive margin evolution, ophiolite formation, subduction, accretion and collision (Robertson and Mountrakis 2006). It is widely accepted that the ophiolites are supra-subduction zone ophiolites formed within the Southern Neotethys during the Late Cretaceous (Bağcı and Parlak 2009; Caran et al. 2010; Elitok 2012; Güneş et al. 2021). Due to the complexity of the region, various paleotectonic reconstructions have been suggested and remain a matter of debate. The present Mediterranean Sea in the south of Türkiye is a remnant of the Neotethys, and the IA is located at the intersection of the Aegean and Cyprus arcs (Fig. 1A), which also control the active tectonics and related earthquakes, especially in the central and western part of Türkiye with the Tauride Belt (Çiçek and Koçyiğit 2024). Numerous studies related to IA, e.g., on stratigraphy and sedimentological evolution (Akbulut 1980; Hayward 1982; Robertson and Woodcock 1981a, b, c, 1982; Waldron 1982; Robertson and Woodcock 1984; Lagabriele et al. 1986; Flecker et al. 1998; Ilgar et al. 2024a, b; Nemek and Özaksoy 2024; Larsen et al. 2024a, b, c; Wathne et al. 2024), structural geology and tectonic evolution (Marcoux et al. 1989; Glover and Robertson 1998; Şengör et al. 2019; Koçyiğit et al. 2024), fossil findings (Marcoux and Baud 1986a, b; Yurtsever et al. 2003; Angiolini et al. 2007), palaeomagnetism (Kissel et al. 1993; Morris and Robertson 1993; van Hinsbergen et al. 2010), petrology (Bozcu et al. 2003) and ophiolite geology and petrology (Çelik and Delaloye 2003; Bağcı et al. 2006; Bağcı and Parlak 2009; Elitok 2012), have been carried out for the region. Nevertheless, petrological studies on syn- and post-ophiolite emplacement sedimentary basin deposits are scarce.

It is widely known that mineralogical, chemical and textural characteristics of epiclastic sedimentary rocks are influenced by the following factors: (1) the source

rocks and their geotectonic position; (2) the weathering conditions, such as climate, relief and exposure time to the weathering process; (3) the mode and length of transport to the sediment basin; (4) the processes that occur after sediment deposition, such as diagenesis and hydrothermal alteration (Bilic and Garasic 2021). Besides, the chemical compositions of clastic rocks give important information about the plate tectonic settings of their source area(s) and depositional basins (e.g., Maravelis et al. 2015, 2021, 2023; Kovani et al. 2023; Botziolis et al. 2023). In other words, clastic rocks from different tectonic settings have distinctive geochemical characteristics for a particular tectonic setting (Roser and Korsch 1986). There is a remarkable relationship between the mineralogical-geochemical compositions of the clastic sediments and the distinct tectonic settings in which they formed. In this regard, the IA, hosting from Triassic to Cenozoic sediments, is a natural laboratory to test sediment geochemistry applications in determining the tectonic settings of ancient sedimentary terranes. In this study, we evaluated geochemical data on Mesozoic (Triassic and Upper Cretaceous) and Cenozoic (Eocene and Miocene) siliciclastic rocks within the IA, including the Antalya Complex, to discuss their correspondence with proposed tectonic settings based on geological data and the geochemistry of igneous rocks.

The IA plays a critical role in understanding the tectonic evolution of the Mesozoic-Cenozoic Neotethys in the Eastern Mediterranean region (Koçyiğit et al. 2024). It is characterised by para-autochthonous units dominated by Mesozoic platform carbonates and allochthonous assemblages (from west to east the Lycian Nappes, also known as the Teke Nappes), the Antalya Complex (also known as the Antalya Nappes), Alanya Nappes and the Beyşehir-Hoyran-Hadim Nappes. Mesozoic sedimentary, igneous and minor metamorphic rocks of mostly microcontinental and oceanic origin constitute the Antalya complex (Robertson and Woodcock 1982; Robertson 1993), which dominated the core area of the IA. All the lithologies exposed in the region reflect the imprints of paleotectonic evolution from rifting to subduction-accretion and collision processes. Although there are debates on the origin and emplacement of the complex, e.g., northern or southern origin, the most accepted model is that the Antalya Complex units derived from the Southern Neotethys, located to the south of the platform carbonates (Beydağları and Anamas-Akseki) according to the paleogeographic situation at that time.

Considering the historical geodynamic reconstruction of the region, the main tectonic events that shaped the East Mediterranean domain are Cadomian arc magmatism formed by the southward subduction of the Proto-Tethyan ocean beneath North Gondwana in a back-arc rift setting (based on detrital zircon geochronology of a mafic-intermediate

dyke suits in the Karacahisar dome, West-Central Taurides, Dedeoğlu Yıldız et al. (Deniz 2022)), pre-, syn- and post-rift sedimentary-tectonic development of N Gondwana, related to opening and closure of S Neotethys in the E Mediterranean region (based on detrital zircon geochronology of the Antalya Complex, Antalya Nappes, Chen et al. 2022), Permian and Triassic rifting (Poisson 1977; Robertson 1993; Şahin et al. 2012; Robertson et al. 2020a) and the Late Triassic to Early Jurassic opening of the southern branch of the Neo-Tethys (Şengör and Yılmaz 1981; Şengör et al. 1984; Robertson et al. 1996; Garfunkel 1998, 2004; Robertson 2007, 2022; Frizon de Lamotte et al. 2011; Tugend et al. 2019; van Hinsbergen et al. 2020). Several lines of evidence suggested that by the Late Triassic, a continental margin rifting had already been developed in the IA (e.g., Garfunkel 2004). Important indications of Triassic intra-continental rifting in the IA area also record its separation from Gondwana (Robertson 1993, 2000; Robertson et al. 2024). The presence of intraplate Late Triassic basaltic lavas in the Antalya Complex suggests the onset of an oceanic basin (Robertson and Waldron 1990). This basin was connected with the concomitant oceanic basin recorded in Cyprus. Late Triassic OIB-like extrusives in the Antalya Complex (Robertson and Waldron 1990; Maury et al. 2008; Bağcı and Parlak 2009; Elitok 2012; Bağcı et al. 2020; Karaoğlu et al. 2022) were formed in a tensional regime at the continent-ocean transition zone during rifting of the southern Neotethyan Ocean. Accordingly, all ophiolites and their associated metamorphic soles, isolated dikes and volcanics in the Antalya Complex are invariably Upper Cretaceous in age (~95–89 Ma, Çelik et al. 2006; van Hinsbergen et al. 2016; Bozcu et al. 2003; Bağcı and Parlak 2009; Parlak 2016; Parlak et al. 2019) and show a supra-subduction zone-type (SSZ-type) signature (e.g., Dilek and Furnes 2011; Elitok 2012; Parlak 2016); they are commonly interpreted as evidence of commencing of an intra-oceanic subduction zone formed within the Neotethyan Ocean before ~95 Ma (Bağcı et al. 2006; Çelik et al. 2006; Elitok 2012; Maffione et al. 2017; Menant et al. 2016; Plunder et al. 2013, 2016; Robertson et al. 2012; van Hinsbergen et al. 2015, 2016). Late Cretaceous-Palaeocene subduction-collision-exhumation of a microcontinent along the northern, active margin of Southern Neotethys is evidenced from the Alanya Massif and the adjacent Antalya Complex (S Türkiye) (Robertson and Parlak 2020). Following the Late Cretaceous-Early Paleocene, first emplacement of the Antalya Complex (Robertson and Parlak 2020) and Late Paleocene-Middle Eocene ophiolite obduction and collision continued until the last emplacement of Early Pliocene collision (Aksu phase, Poisson et al. 2003b; Robertson et al. 2012; McPhee and van Hinsbergen 2019; Searle and Lamont 2022). Robertson (2000) concluded that the intervening southerly Neotethyan oceanic

basin remained partly open in the Early Tertiary, finally closing by diachronous collision in the Eocene-Oligocene, followed by further overthrusts in the Miocene. The Antalya Complex is transgressively overlain by Neogene sedimentary basins, and deposition in Eocene-Oligocene marine basins was controlled mainly by the emplacement of the Lycian Nappes in the region at the time (Poisson et al. 2003b; Tokar et al. 2012). The emplacement of the Lycian Nappes in SW Anatolia continued until the end of the Late Miocene. The IA has been under the influence of the tensional Neotectonic regime since the Mio-Pliocene (Koçyiğit 1983, 1984; Glover and Robertson 1998; Görmüş et al. 2001; Poisson et al. 2003a, 2003b; Robertson et al. 2003; Flecker et al. 2005; Koçyiğit and Deveci 2007; Koçyiğit et al. 2012, 2024). Collision of the Arabian and Anatolian plates was followed by westward extrusion of the Anatolian plate and southward expansion in the Aegean Sea. In general, the geological evolution of Türkiye is characterised by closure of the Neo-Tethys, rotation and amalgamation of smaller continental fragments and collision of Anatolide-Tauride block with the southern Eurasian margin associated with collision-induced continental extrusion tectonics (Şengör and Yılmaz 1981; Kissel et al. 1993; Tatar et al. 2002; Robertson et al. 2003; Çiner et al. 2008; Elitok and Dolmaz 2010; Piper et al. 2010).

2 Stratigraphy

2.1 North of the Isparta Angle

In the apex of the IA and the south and southeast of Isparta city, the stratigraphic sequences of Triassic to Cenozoic sedimentary basins, from base to top, comprise (Figs. 2 and 3):

- (1) Late Triassic-Early Cretaceous Ispartaçay Formation (Yalçinkaya et al. 1986; Robertson 1993; Vrielynck et al. 2003) composed, from bottom to top, of the following sedimentary units: (a) marls and turbiditic sandstones; (b) limestones interbedded with marls and (c) cherty limestones. These three lower units involve Triassic fossils (i.e., *Halobia* and *Ammonites*); (d) calcareous breccia including Middle Jurassic foraminifera; (e) Jurassic to Early Cretaceous green and red radiolarites; (f) red silts and clays with interbedded calcareous breccia. The observed whole sequence points to a deep marine setting and reaches a thickness of > 600 m (Yalçinkaya et al. 1986; Poisson 1977; Robertson 1993; Tekin 1999; Tekin and Yurtsever, 2003; Vrielynck et al. 2003).

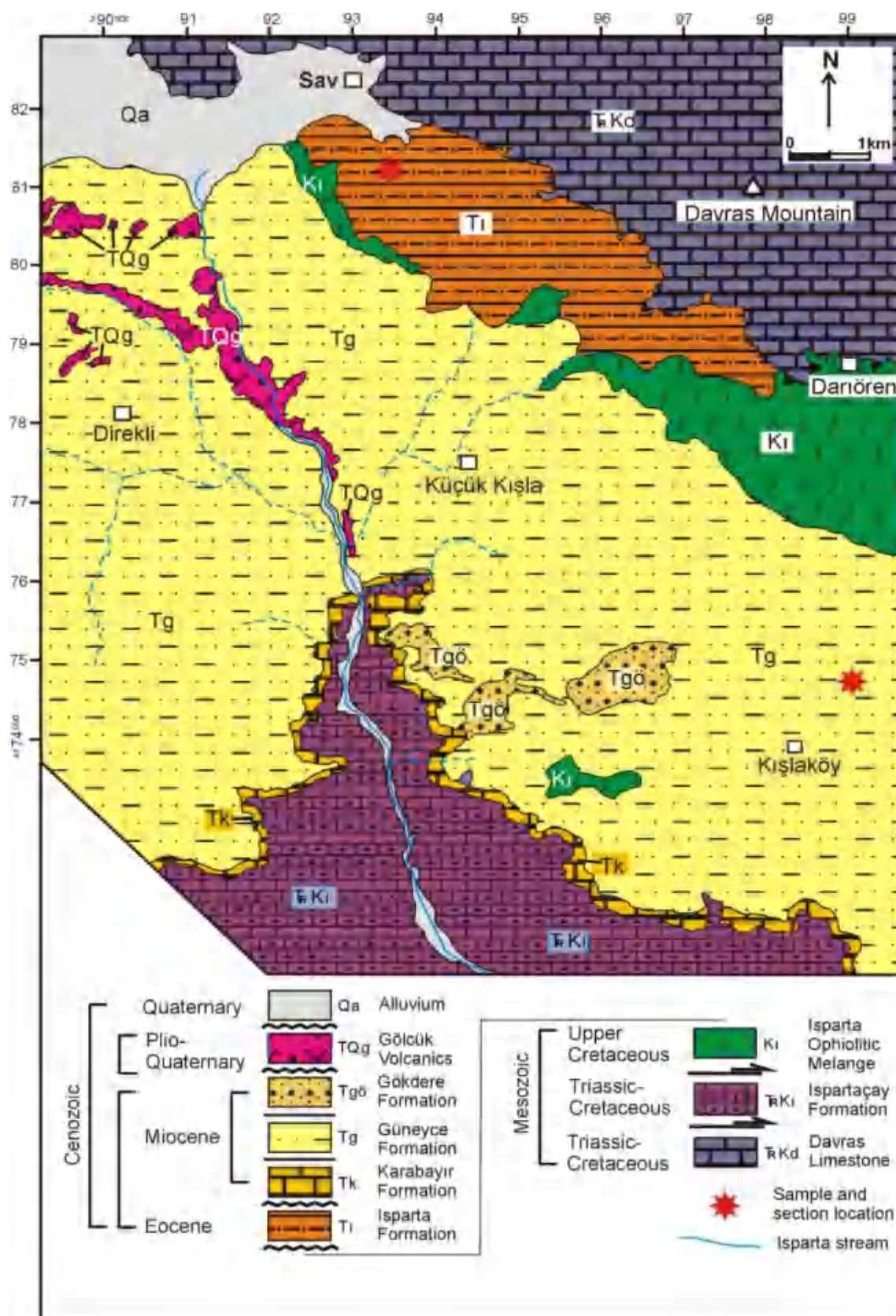


Fig. 2 Geological map of the study area (SE of Isparta region) (remodified from Gutnic et al. 1979; Şenel 1997; Uysal 2004) and locations of sampled section areas.

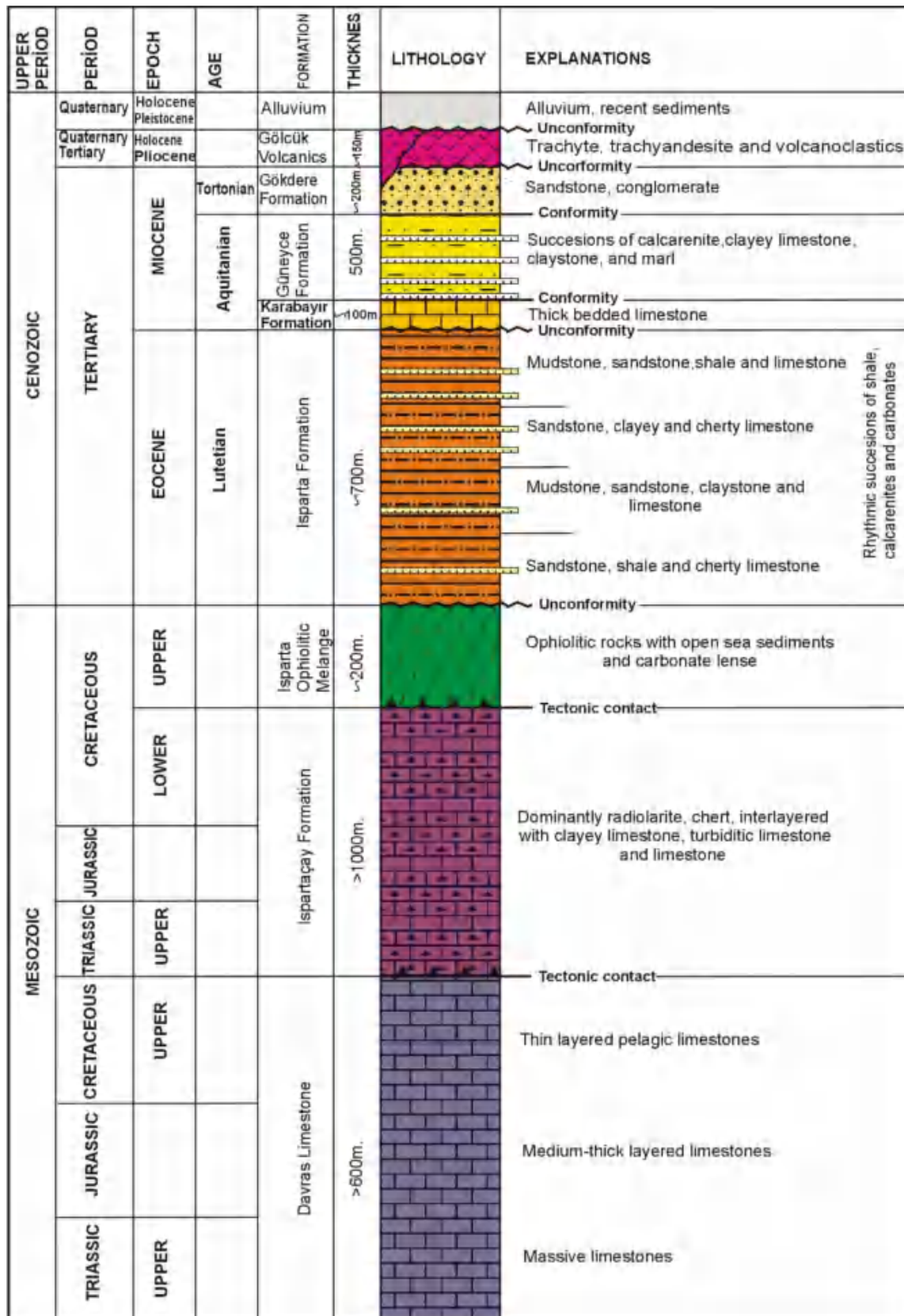


Fig. 3 Tectonostratigraphical column of the study area in the Isparta Angle (remodified from Nielsen et al. 2010).

- (2) The para-autochthonous Davras Limestone (Late Triassic–Upper Cretaceous, Robertson 1993), also known as Mesozoic-aged platform-type carbonate, starts with grey, beige-coloured, medium- to thick-layered intrabiomicritic and intrabiosparitic neritic limestones interlayered with dolomite and dolomitic limestones and passes thin-layered clayey biomicritic pelagic limestones upward.
- (3) Upper Cretaceous Isparta Ophiolitic Melange (Robertson 1993) comprises serpentinite, gabbros, basalt and chert. This unit is approximately 200 m thick and has tectonic contact with other units (Yalçinkaya et al. 1986; Uysal 2004). The Ispartaçay Formation and Isparta Ophiolitic Melange represent part of the Antalya Complex on the northern side of the IA. They are thrust over the Davras limestone of the Beydağları autochthon. Cenozoic marine sediments unconformably overlie the Upper Cretaceous ophiolitic melange (Sagular and Görmüş 2006; Yavuzlar and Sagular 2018).
- (4) The Eocene Isparta Formation consists of transitional facies ranging from marine clastics to carbonates. It consists of clayey limestone, claystone, siltstone, sandstone and limestone succession units, and it also contains clayey and cherty interbeds dominantly of turbiditic origin. The formation, which contains benthic foraminifera (*Alveolina sp.*; *Nummulites sp.*, etc.) and fossils such as *Globigerina sp. ve Morozovella sp.*, was deposited in a shallow environment with a thickness of approximately 700 m (Nielsen et al. 2010).
- (5) The abundant coral fossiliferous Karabayır Formation, known for its thick-bedded limestones (20–100 m thick), is Lower Miocene (Aquitanian) and deposited in a reef environment (Gutnic and Poisson 1970; Yalçinkaya 1989; Görmüş and Hançer 1997; Uysal 2004).
- (6) Planktic (*Globigerinoides sp.* *Rotalia sp.*, etc.) and benthic (*Lepidocyclina sp.*, *Discocyclina sp.* and *Amphistegina sp.*, etc.) foraminifera and calcareous nannofossils are found in the 700-m-thick Güneyce Formation (Aquitanian), which is formed by stratified sandstones, claystone-marl succession and limestone lenses in a shallow marine deposition environment (Uysal 2004; Nielsen et al. 2012; Yavuzlar and Sagular 2018).
- (7) The Upper Miocene Gökdere Formation comprises fossil-free, medium- to coarse-grained clastic conglomerates and sandstones and reaches a thickness of about 200 m (Görmüş and Hançer 1997). Plio-Quaternary Gölcük trachytic and trachyandesitic volcanic and volcanoclastic rocks (e.g., Kumral et al.

2006; Elitok et al. 2010; Caran 2016; Atıcı et al. 2025) and Quaternary recent alluvium are the youngest units in this area (Fig. 3).

Studied Cenozoic sandstone samples were collected representatively from different levels of the (1) Savköy section (≈500 m thick) of the Eocene Isparta Formation (≈700 m thick) and (2) Kışlaköy section (≈280 m thick) of the Lower Miocene Güneyce Formation (≈500 m thick) (Figs. 2 and 4). The Isparta Formation is composed mainly of sandstone and mudstone alternations and also contains clayey and cherty limestone and marl interbeds (Fig. 4). Eocene calcarenitic sandstones (Fig. 5A–B) are medium- to coarse-grained, moderately to poorly sorted, and appear as calciclast-rich litharenite having 40%–50% carbonate cement, 20%–30% mainly feldspar and lithic fragments such as radiolarite, chert, quartz, volcanic rock fragments, bioclasts, ophiolite-fragments, pyroxene and rare mica minerals (Fig. 5C–D). Reworked carbonate bioclasts, pelagic forms, transported fossils and clayey limestones are also recorded in the samples. In the coarse-grained sandstones, pyroxene and mica minerals are transported materials of ophiolites. The fossil fauna (*Alveolina sp.*, *Nummulites*, *Discocyclina*, *Sphaerogypsina*, *Rotalid*, *Globigerina sp.*, *Morozovella sp.*) data indicate a shallow marine environment for Eocene samples. The studied Kışlaköy succession of the Güneyce Formation comprises mainly sandstone, calcarenite, clayey limestone, claystone and marl alternations (Fig. 3). Sandstones (Fig. 6A–B) are mainly yellowish brown to green coloured and moderately to thinly bedded. Lithic clasts from the older sediments such as Davras limestone, ophiolitic melange, quartz and feldspar minerals are the main components (Fig. 6C–F). Cement is neomorphic calcite. The pebbles are made up of limestone (40%), quartz-feldspar (50%) and ophiolitic (10%) fragments. Plant traces, angular-shaped glauconite, pyroxene, mica and goethite are also recorded in thin sections. The determined fossil fauna indicates an open marine environment for Miocene samples (Nielsen et al. 2010, 2012; Yavuzlar and Sagular 2018).

2.2 East-northeast of the Isparta Angle

The AC is thrust over the Anamas-Akseki and Mesozoic Karacahisar carbonate platforms in the east and northeast of the IA (Fig. 1B). In the northeast of the IA, the AC is dominated by deep-water and slope basin deposits (Pazarköy Group of Waldron 1982) and detailed geological information has been given in previous works (Waldron 1984; Robertson 1993; Elitok 2012). In the east of the IA, the Triassic Kasımlar Formation within the Karacahisar autochthon (Dumont 1976, 1978) consists mainly of shale

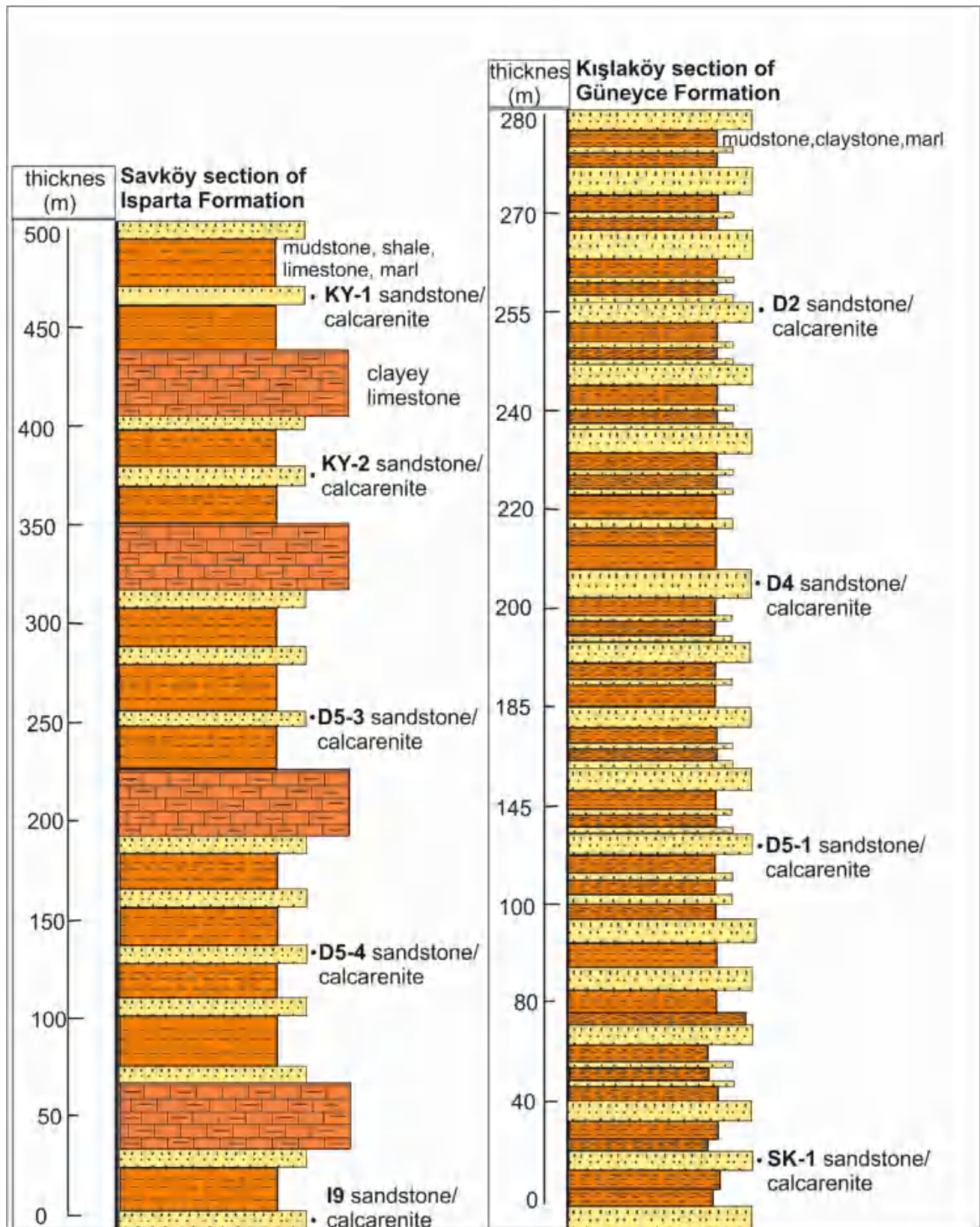


Fig. 4 Measured and simplified **A** Savköy section of the Eocene Isparta Formation and **B** Kışlaköy section of the Lower Miocene Güneyce Formation.

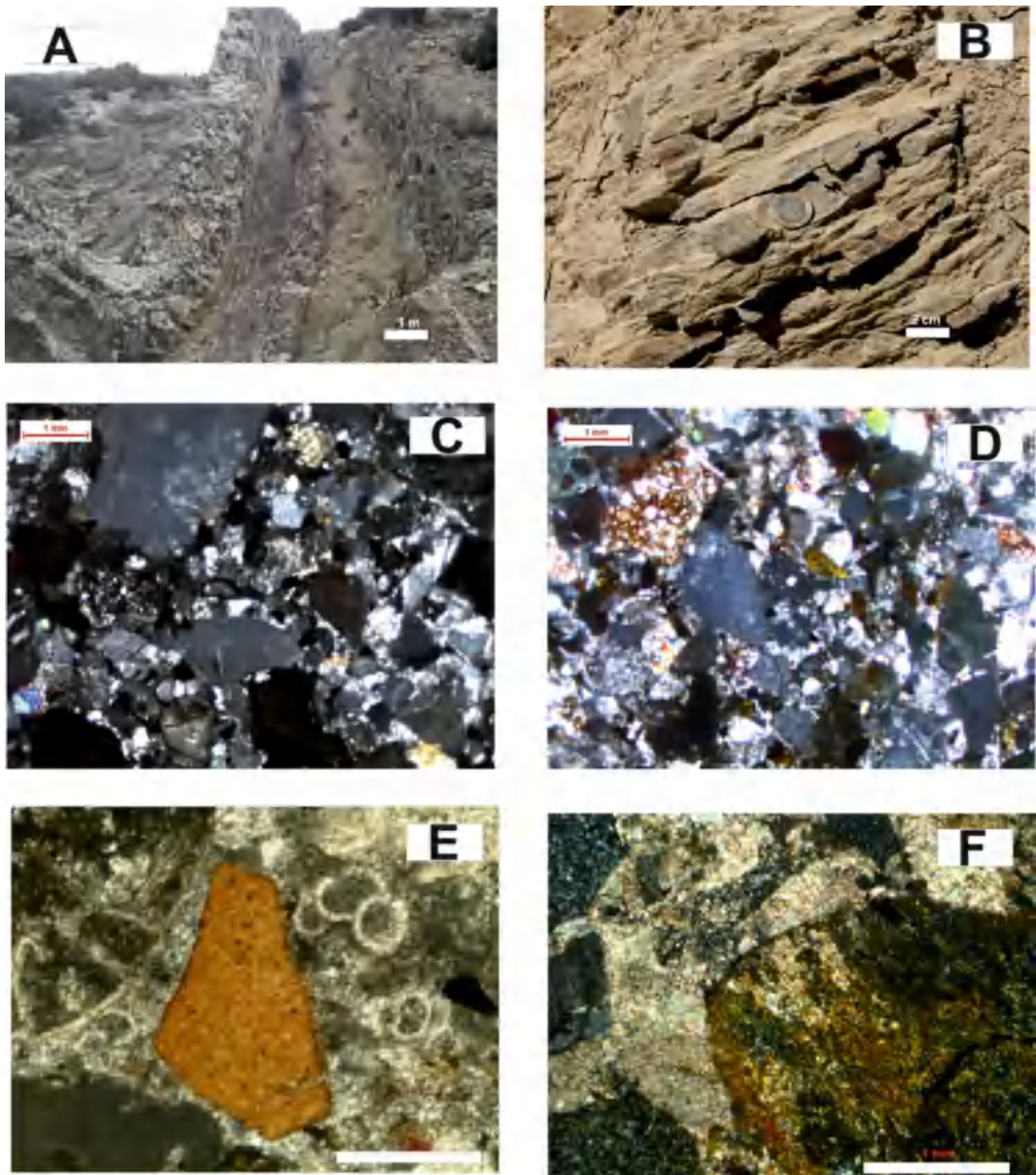


Fig. 5 A–B Field views of the clastic rocks from the folded Eocene-aged Isparta Formation. Bedded sandstones intercalated with slate and clayey and cherty limestone. In thin section photomicrographs of the sandstone (C–D), as lithic clasts, coarse-grained quartz (E) and ophiolitic clast (F).

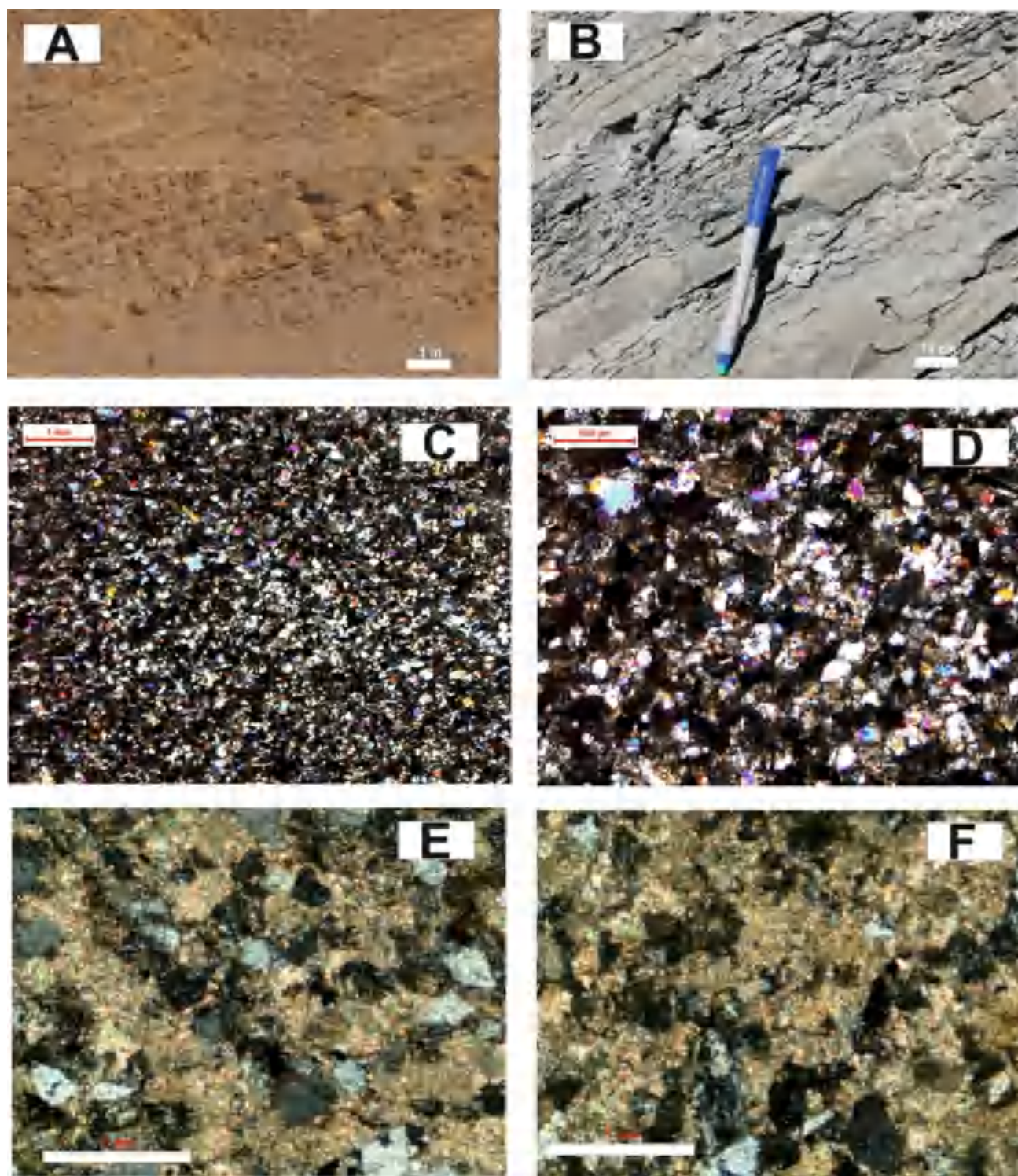


Fig. 6 A–B Field views of the clastic rocks from the Miocene-aged Güneyce Formation. Bedded sandstones intercalated with clayey limestone and marl. In thin section photomicrographs of the fine-grained sandstone (C–D), in a carbonate cement, fine-grained feldspars and quartz (E–F).

and sandstone units, and its thickness reaches 1300 m. This formation starts with mostly conglomerate and sandstone in the lower part, with siltstone and blackish and dark greyish shales. It passes to blackish shales and brecciated limestone lenses with abundant algal fossils in the middle to upper levels. The uppermost levels consist of platy limestone with greenish claystone intercalations (Kuşçu et al. 2016). Considering the fossil content and lithological features, the depositional environment of the Kasımlar Formation is considered a shallow marine environment. Trace and rare element geochemistry of the black shales within the Kasımlar Formation were studied by Kuşçu et al. (2016). These geochemical data were used in this study.

2.3 Southwest of the Isparta Angle (western side of Antalya Bay)

In the southwest of the IA, the Antalya Complex is composed of a complex assemblage of sedimentary, mafic and ultramafic igneous rocks, mostly Mesozoic age, which characterise the platform, slope-basin deposits and ophiolites. The rock assemblage in this area is subdivided into N-S trending tectonic units and named, from bottom to top, (1) the Kumluca Zone, the Gödene Zone, the Kemer Zone and the Tekirova Zone (Woodcock and Robertson 1982), (2) Lower, Median and Upper Antalya Nappes (Marcoux et al. 1987 and references therein), (3) the Çataltepe, Alakırçay, Tahtalıdağ and Tekirova ophiolitic nappes (Şenel et al. 1992, 1996) and (4) continental margin unit, rift assemblages, marginal plateaus or platform rafts, ophiolites and mantle slabs (Dilek and Rowland 1993). Geology, geochemistry and depositional-tectonic settings of the clastic sedimentary rocks over the allochthonous sheets were studied in detail by Robertson et al. (2020a, b), and the geochemical data were used for comparison in the current study.

3 Sampling and analytical method

Representative Eocene to Miocene sandstones were sampled from different levels of successions of the Isparta and Güneyce Formations for their petrographic descriptions and geochemical analyses (Fig. 4, Table 1). Five fresh samples (KY-1, KY-2, D5-3, D5-4, I9) from the Eocene Isparta Formation and four fresh samples (D2, D4, D5-1 and SK-1) from the Lower Miocene Güneyce Formation were collected (Fig. 4). Geochemical analyses of the major, trace and rare earth element contents in whole-rock samples were conducted at the commercial ACME Laboratories, Ltd., in Vancouver, Canada. The compositions of the samples were measured by ICP-AES for major oxides (0.2 g pulp sample by LiBO_2 fusion). To prepare whole-rock powders, 1–2 kg of the fresh samples was crushed in a steel crusher; then, the samples were ground in an agate mill to a grain size of < 200 mesh. Major

element detection limits are about 0.001%–0.04%. For the trace elements, 0.2 g of sample powder and 1.5 g of LiBO_2 flux were mixed in a graphite crucible and heated to 1050 °C for 15 min in a muffle furnace. The molten sample was then dissolved in 100 ml of 5% HNO_3 (American Chemical Society-grade nitric acid in demineralised water). Sample solutions were shaken for 2 h, and then an aliquot was poured into a polypropylene test tube and aspirated into a PerkinElmer Elan 600 ICP mass spectrometer. Calibration and verification standards and reagent blanks were added to the sample sequence. The detection limits of trace elements range from 0.01 to 0.5 ppm. The analytical precision is generally better than 1% for elements with concentrations of > 200 ppm and 1%–3% for elements with < 200 ppm concentrations.

4 Results and discussion

4.1 Major and trace element signatures of siliciclastic sediments

Representative major, trace and rare earth element (REE) analyses of nine samples collected from the Eocene and Miocene sandstones within the IA are given in Table 1. For comparison, the available 9 geochemical data of Upper Cretaceous ophiolite-related clastic sediments (mainly sandstones) from the southwest of the AC (data from Robertson et al. 2020b) and 21 geochemical data of Late Triassic shales from the Kasımlar Formation in the east of the IA (data from Kuşçu et al. 2016) (Table 1) were evaluated in the geochemical diagrams along with the analytical results of Cenozoic clastic sediments (mainly sandstones) from this work. In general, all analysed Cenozoic sandstone samples collected from the Eocene Isparta Formation and the Miocene Güneyce Formation are calcium-rich (CaO , 30.54–43.87 wt%), being akin to those of Upper Cretaceous ophiolite-related clastic sediments (CaO , 7.94–46.55 wt%), and three samples of Late Triassic shales (CaO , range between 11.27 and 26.71 wt%) possess an excellent carbonate dilution effect on the majority of major oxides with a low but a wide range, for example, in SiO_2 (12.86–33.19 wt%), Al_2O_3 (1.4–4.39 wt%), $\text{Fe}_2\text{O}_3+\text{MgO}$ (2.89–6.78 wt%; total Fe as Fe_2O_3), K_2O (0.13–0.58 wt%) and Na_2O (0.14–0.84 wt%) and TiO_2 (0.08–0.26 wt%) (Table 1). As shown in Fig. 7A, relative to some Late Triassic sediments illustrating the effects of both silica and carbonate dilutions on major element compositions, the carbonate dilution appears much more dominant in Upper Cretaceous ophiolite-related clastic sediments and Cenozoic sandstones. There is also a remarkable positive correlation between CaO and LOI , which suggests that LOI and CaO are incorporated into calcite rather than other elements (Tobia and Shangola 2016). Compared to the major element geochemistry of Late Triassic and Upper Cretaceous

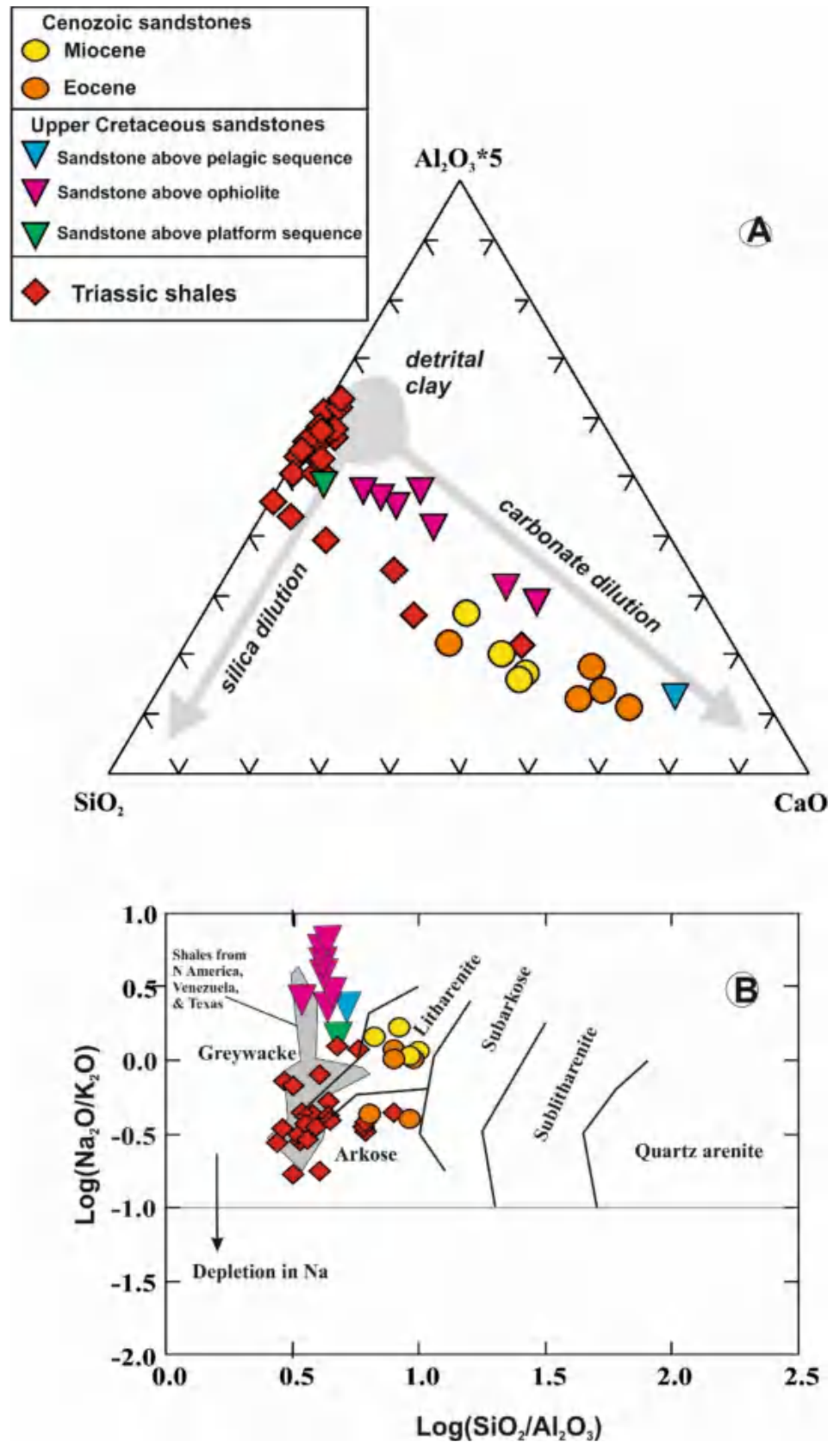
Table 1 Major (wt%), trace and REE (ppm) concentrations of Cenozoic sandstones from Isparta Angle and average geochemical compositions of Upper Cretaceous sandstones (Robertson et al. 2020a, b) and Late Triassic shales (Kuşçu et al. 2016). Average geochemical compositions are shown as bold

Sample	Cenozoic sandstones										Upper Cretaceous sandstones			Late Triassic shales		
	Eocene-aged sandstones					Miocene-aged sandstones					Number of samples: 9			Number of samples: 21		
	D5-3	D5-4	I9	KY-1	KY-2	D2	D4	D5-1	SK-1	Average	Min	Max	Average	Min	Max	Average
SiO ₂	15.49	18.31	33.19	15.11	12.86	22.72	28.75	25.44	23.46	21.70	7.91	57.42	32.91	38.40	71.88	57.60
TiO ₂	0.16	0.09	0.13	0.09	0.08	0.19	0.26	0.21	0.15	0.15	0.09	0.79	0.51	0.31	1.08	0.79
Al ₂ O ₃	2.54	1.74	3.61	1.86	1.40	2.37	4.39	3.07	2.19	2.57	1.60	12.43	8.01	4.77	19.90	14.25
Fe ₂ O ₃ TOT	1.49	1.14	1.58	1.66	1.38	1.54	3.13	2.23	1.40	1.72	1.64	6.68	4.90	3.38	18.88	7.25
MnO	0.13	0.15	0.05	0.15	0.12	0.07	0.11	0.12	0.04	0.10	0.10	0.32	0.13	0.02	0.28	0.08
MgO	1.40	2.26	1.81	2.26	2.36	3.33	3.65	2.57	3.97	2.62	2.41	10.11	4.53	1.04	3.29	2.02
CaO	42.47	42.00	30.82	42.39	43.87	36.29	30.54	34.88	35.49	37.63	7.94	46.55	24.91	0.40	26.71	5.32
Na ₂ O	0.20	0.16	0.52	0.18	0.14	0.34	0.84	0.53	0.32	0.35	0.23	2.09	1.51	0.10	2.54	1.13
K ₂ O	0.47	0.15	1.28	0.15	0.13	0.33	0.58	0.31	0.28	0.40	0.10	0.95	0.46	0.56	4.43	2.42
P ₂ O ₅	0.06	0.09	0.02	0.03	0.01	0.03	0.04	0.03	0.03	0.04	0.05	0.17	0.08	0.05	0.31	0.16
Cr ₂ O ₃	0.01	0.02	0.04	0.11	0.20	0.04	0.07	0.08	0.02	0.07	0.01	0.12	0.04	0.00	0.02	0.01
LOI	35.60	33.80	26.90	35.90	37.30	32.60	27.40	30.30	32.50	32.47	12.20	38.80	21.81	2.60	23.40	8.77
Total	100.02	99.91	99.95	99.89	99.85	99.85	99.76	99.77	99.85	99.81			99.80			99.83
Fe+Mg	2.89	3.40	3.39	3.92	3.74	4.87	6.78	4.80	5.37	4.35	4.05	17.61	9.43	5.13	22.14	9.27
K ₂ O/Na ₂ O	2.35	0.93	2.46	0.83	0.92	0.97	0.69	0.58	0.87	1.17	0.15	0.69	0.34	0.79	5.60	2.59
Trace (ppm)																
Sc	4.00	3.00	5.00	5.00	5.00	3.00	8.00	6.00	4.00	4.77	4.00	28.00	2.53	6.00	42.00	16.20
V	27.00	18.00	28.00	31.00	33.00	27.00	50.00	40.00	30.00	31.55	26.00	207.00	26.50	31.00	148.00	105.50
Cr	75.26	143.68	280.50	766.31	1395.78	280.50	485.78	574.73	150.52	461.45	61.60	786.00	437.00	20.50	116.00	80.50
Co	12.70	12.70	10.80	12.30	11.80	8.40	20.10	14.10	9.20	12.45	6.80	30.00	9.60	6.10	19.60	13.90
Ni	58.90	67.90	98.40	97.70	141.40	57.00	123.40	132.90	28.90	89.61	31.00	249.00	75.50	17.30	82.20	42.30
Ga	3.60	2.10	3.90	2.40	1.70	3.20	4.90	3.50	2.50	3.08	0.50	11.90	2.45	5.60	22.30	16.40
Rb	18.90	5.70	31.90	5.50	4.60	10.80	24.60	12.80	8.40	13.68	4.70	29.80	10.95	25.10	152.20	93.20
Sr	904.30	442.10	228.40	364.80	312.10	573.90	559.30	823.20	412.90	513.44	205.60	672.00	363.80	32.60	500.40	164.40
Y	11.70	13.60	10.30	8.60	5.70	8.50	17.20	13.60	7.30	10.72	7.70	25.10	7.91	14.20	44.90	26.80
Zr	31.50	21.20	47.40	14.10	11.90	66.90	51.00	47.70	35.80	36.38	7.60	90.00	30.53	61.00	256.00	187.30
Nb	3.60	1.60	3.50	1.10	1.10	3.40	3.90	3.60	2.20	2.66	0.70	4.50	2.08	6.70	23.70	15.70
Ba	247.00	42.00	134.00	240.00	33.00	50.00	49.00	70.00	31.00	99.55				67.00	612.00	281.30
Hf	0.80	0.30	1.40	0.60	0.30	1.80	1.40	1.30	0.80	0.96	0.20	2.50	0.84	2.00	6.80	5.31
Ta	0.10	0.10	0.10	0.10	0.10	0.10	0.40	0.20	0.10	0.14	0.10	0.30	0.12	0.40	1.50	1.00
Pb	3.90	3.60	3.00	2.90	1.30	2.80	7.80	5.20	2.60	3.67				3.70	21.40	16.22
Th	1.70	0.90	2.90	0.50	0.30	2.30	2.30	1.70	1.40	1.55	0.20	3.50	1.26	3.70	14.70	9.90
U	0.50	0.50	1.40	0.90	0.90	0.80	1.70	1.00	1.10	0.97	0.20	1.00	0.86	0.90	3.10	2.34

Table 1 (continued)

Cenozoic sandstones		Miocene-aged sandstones					Upper Cretaceous sandstones			Late Triassic shales		
Eocene-aged sandstones		Miocene-aged sandstones					Upper Cretaceous sandstones			Late Triassic shales		
Sample	Eocene-aged sandstones					Average	Number of samples: 9			Number of samples: 21		
	D5-3	D5-4	I9	KY-1	KY-2		Min	Max	Average	Min	Max	Average
Eu/Eu*	0.71	0.79	0.93	0.86	0.74	0.82	0.74	1.08	0.98	0.48	1.34	0.68
Th/U	3.40	1.80	2.07	0.55	0.33	1.70	0.40	3.50	1.78	3.07	5.22	4.20
Yb/Sm	0.49	0.56	0.53	0.60	0.56	0.45	0.68	1.25	1.05	0.30	0.57	0.42
La/Sm	5.76	5.79	5.63	6.09	6.07	4.83	2.05	7.28	3.43	3.60	6.02	5.15
La/Yb	11.70	10.30	10.44	10.00	10.66	10.62	2.00	10.62	3.62	8.33	15.35	12.33
Th/Sc	0.42	0.30	0.58	0.10	0.06	0.28	0.05	0.23	0.04	0.33	0.88	0.65
Cr/V	2.78	7.98	10.01	24.71	42.29	14.36	0.91	10.80	2.89	0.53	0.88	0.76
Cr/Ni	1.27	2.11	2.85	7.84	9.87	4.32	1.71	4.72	2.66	0.72	3.43	2.05
REE (ppm)												
La	11.00	10.20	9.30	6.40	4.80	10.20	2.60	14.70	4.65	11.50	43.40	31.40
Ce	14.30	10.40	15.10	8.30	6.10	18.20	3.30	24.40	6.97	24.20	93.20	64.00
Pr	2.43	2.20	2.04	1.29	0.95	2.52	0.57	3.18	1.04	2.93	10.18	7.64
Nd	10.00	9.10	7.30	5.00	3.80	10.30	2.90	13.00	4.84	1.80	42.90	27.90
Sm	1.91	1.76	1.65	1.05	0.79	2.11	0.70	3.17	1.38	3.19	9.41	6.00
Eu	0.46	0.46	0.37	0.29	0.24	0.55	0.19	0.82	0.50	0.82	2.27	1.30
Gd	1.87	1.97	1.51	1.20	0.78	2.10	0.80	3.60	1.88	3.17	9.30	5.78
Tb	0.29	0.31	0.26	0.19	0.14	0.35	0.14	0.61	0.34	0.44	1.31	0.83
Dy	1.61	1.97	1.67	1.20	0.82	2.08	0.85	4.10	2.21	2.96	7.91	5.11
Ho	0.32	0.39	0.30	0.23	0.16	0.43	0.20	0.90	0.50	0.42	1.55	0.94
Er	0.97	1.05	0.94	0.60	0.48	1.18	0.59	2.68	1.50	1.52	4.08	2.73
Tm	0.14	0.16	0.15	0.09	0.07	0.14	0.08	0.39	0.20	0.19	0.51	0.39
Yb	0.94	0.99	0.89	0.64	0.45	0.96	0.48	2.82	1.43	1.38	3.68	2.52
Lu	0.14	0.16	0.14	0.10	0.07	0.14	0.08	0.43	0.22	0.22	0.53	0.38

Fig. 7 **A** Triangular diagram (modified from Klaver et al., 2015) showing the relative concentrations of Al_2O_3 , SiO_2 and CaO in the Cenozoic, Upper Cretaceous and Late Triassic sediment samples from the Isparta Angle. Most Cenozoic and Upper Cretaceous samples plot on a carbonate dilution trend because of the biogenic carbonate. In contrast, some of the Late Triassic samples show significant dilutions with both carbonate and silica. **B** Plots of whole-rock chemical data from Cenozoic, Upper Cretaceous and Late Triassic samples in the $\log(\text{Na}_2\text{O}/\text{K}_2\text{O})$ versus $\log(\text{SiO}_2/\text{Al}_2\text{O}_3)$ diagram (after Pettijohn et al. 1987).



ophiolite-related clastic sediments samples, Cenozoic samples are compositionally distinct from them. Late Triassic and Upper Cretaceous ophiolite-related clastic sediments samples are characterised by relatively higher SiO_2 (average 57.6 wt% for Late Triassic, average 32.9 wt% for Upper Cretaceous), TiO_2 (average 0.79 wt% for Late Triassic, average 0.51 wt% for Upper Cretaceous), Al_2O_3 (average 14.25 wt% for Late Triassic, average 8.00 wt% for Late Cretaceous) and $\text{Fe}_2\text{O}_{3\text{Total}}+\text{MgO}$ (average 9.27 wt% for Late Triassic, average 9.43 wt% for Upper Cretaceous) contents than those of Cenozoic samples with lower SiO_2 (average 21.70 wt%), TiO_2 (average 0.15 wt%), Al_2O_3 (average 2.57 wt%) and $\text{Fe}_2\text{O}_{3\text{Total}}+\text{MgO}$ (average 4.35 wt%). Cenozoic samples with $\text{SiO}_2/\text{Al}_2\text{O}_3$ ratios between 6.0 and 10.5 mostly fall into the field of litharenite, and three samples in the arkose field in the chemical classification diagram of ($\text{Log}[\text{SiO}_2/\text{Al}_2\text{O}_3]$

versus $\text{Log}[\text{Na}_2\text{O}/\text{K}_2\text{O}]$) proposed by Pettijohn et al. (1987) and Herron (1988) (Fig. 7B), whereas Upper Cretaceous ophiolite-related clastic sediments fall into the greywacke field and Late Triassic samples (previously identified as shale, Kuşçu et al. 2016) within the field of shales from North America, Venezuela and Texas (Herron 1988, references therein). Considering trace element variations of the studied clastic rocks, on the PAAS (Post-Archean Average Archean Shale, McLennan 2001, references therein)-normalised (for selected elements) spider variation diagram (Fig. 8A), similar to Upper Cretaceous ophiolite-related clastic sediments, Cenozoic samples display strike depletions in Nb–Ta, Th, Rb, Zr and Hf and enrichments in Ni and Cr relative to Late Triassic shales with close affinity to PAAS. The Ni, Co and Cr enrichments of Cenozoic and Upper Cretaceous ophiolite-related clastic sediments are

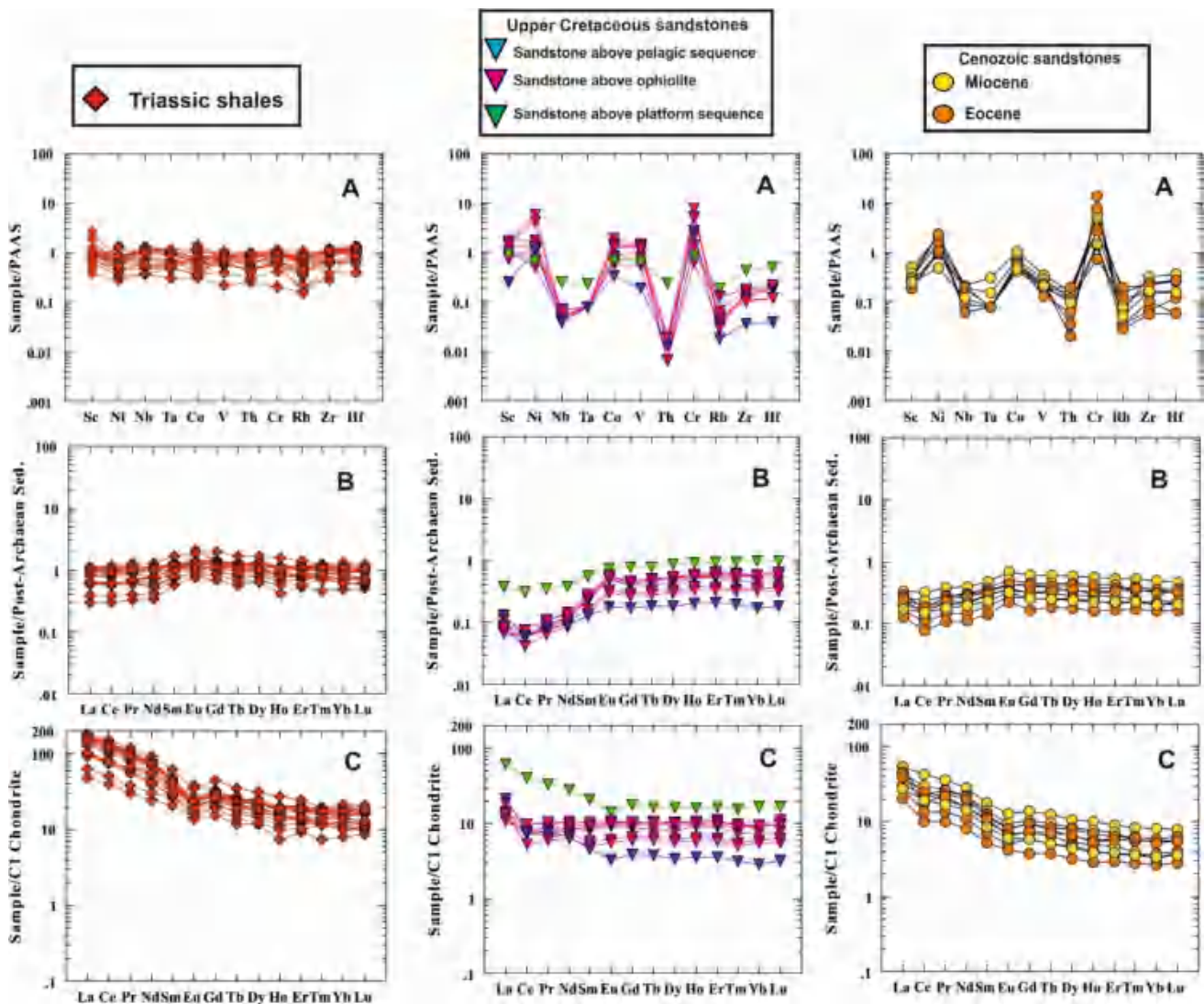


Fig. 8 PAAS-normalised multi-element variations (A) and REE patterns (Taylor and McLennan 1985) (B), and chondrite-normalised REE patterns (McDonough and Sun 1995) (C), for the Cenozoic and Upper Cretaceous sandstones and Triassic shales from the Isparta Angle.

consistent with the presence of ophiolitic clasts in these sandstones. Similarly, the presence of significant Cr (average 80.5 ppm), V (average 105.5 ppm), Ni (average 42.3 ppm) and Co (average 13.9 ppm) enrichments in Late Triassic shales (Table 1) can be attributed to the influences of Triassic OIB-like basaltic rift magmatism in their source regions. Accordingly, the Rb contents of all the samples are positively correlated with K_2O (not shown; see Table 1), also implying that the K and Rb abundances are mainly controlled by K-feldspar and K-bearing clay minerals (e.g., illite, kaolinite) (McLennan et al. 1983).

Considering PAAS-normalised REE patterns (Fig. 8B), relative to REE-enriched Late Triassic shales showing compatibility with the pattern of PAAS, Cenozoic and Upper Cretaceous ophiolite-related clastic sediments display remarkable depletion in LREE. However, compared to Cenozoic sandstones in the chondrite-normalised REE diagram (Fig. 8C), Upper Cretaceous ophiolite-related clastic sediments demonstrate more enrichments in HREE, but both have obvious negative Eu and Ce anomalies in all samples. In contrast, Late Triassic shales with negative Eu anomaly show elevated enrichments in LREE relative to others. Notably, since Upper Cretaceous and Cenozoic sediments have high CaO contents, negative Ce anomalies can be attributed to carbonates because oxic seawater is characterised by strong ‘negative’ dissolved Ce anomalies (Bolhar et al. 2004), and negative Ce anomalies can be transferred to carbonates (Planavsky et al. 2020). Similarly, high contents of Sr in CaO-rich Upper Cretaceous (average 462 ppm) and Cenozoic clastic sediments (average 513 ppm) also coincide with this conclusion in terms of the lower Sr concentrations (average 164 ppm) of Triassic shales.

4.2 Inferences for the nature of parent rocks of the sediments

As shown in Fig. 9A, plots of La/Th against Hf (Floyd and Leveridge 1987) favour arc-derived mafic or mixed igneous source rocks for Cenozoic and Upper Cretaceous ophiolite-related clastic sediments. In contrast, the Late Triassic shales show distribution towards a passive margin felsic source. Elevated concentrations of Cr, Co, Ni and V in sediments particularly record the geochemical signal of mafic and ultramafic rocks in a source region (Taylor and McLennan 1985; Wronkiewicz and Condie 1987; Garver et al. 1996; Botzolis et al. 2023). Accordingly, higher Cr/Ni ratios (≈ 2 and greater, Garver and Scott 1995) typify mafic-volcanic detritus, which is typical of Late Triassic shales (average 2.05), Cenozoic sandstones (average 4.7) and Upper Cretaceous ophiolite-related clastic sediments (average 2.66). The presence of ophiolitic ultramafic detritus in Upper Cretaceous and Cenozoic samples from IA units can be inferred from the relative depletion in V and Th concerning enrichment in

Ni in the V-Ni-Th*10 ternary diagram (Fig. 9B) in which fields representative of ultramafic, mafic and felsic rocks plot separately (Bracciali et al. 2007). As seen in the diagram, Late Triassic samples are close to felsic and tend to mafic rocks, and Cenozoic samples shift from felsic to ultramafic compositions. However, Upper Cretaceous samples overlap between the mafic and the ultramafic compositions except for the sandstone above the platform sequence. Similar trends can also be seen in Co/Th versus La/Sc (McLennan et al. 1993; Gu et al. 2002; Wang et al. 2012) and Cr/V versus Y/Ni (McLennan et al. 1993) diagrams (remodified from Totten et al. 2000) (Fig. 9C–D). Enrichments in Co/Th and Cr/V ratios in Upper Cretaceous and Cenozoic samples appear to be related to a larger contribution of ultramafic material from the ophiolite complexes or intraoceanic arc rocks. In contrast, the mafic detritus of Triassic shales can be attributed to Triassic intra-plate basaltic rift magmatism in their source region. Notably, Triassic shales possess higher Nb/La ratios (average 0.49) relative to Upper Cretaceous (average 0.29) and Cenozoic (average 0.28) sediments. This can be attributed to influences of OIB-like mafic intra-plate basaltic detritus (average Nb/La ratio in Triassic OIB-like intra-plate basalts in Antalya Nappes, 1.29, Maury et al. 2008) in Triassic shales.

4.3 Tectonic settings in IA based on sediment geochemistry

The K_2O/Na_2O versus SiO_2 (wt%) discrimination diagram of Roser and Korsch (1986), separating the ARC, ACM (active continental margin) and PM (passive margin) settings, is one of the most widely used ones. However, these authors indicated that the chemistry of sediments with high CaO contents cannot be plotted directly in the diagram (Roser and Korsch 1986) because of the effect of carbonate dilution on major element compositions. Hence, all data for CaO-rich IA sediments studied have been recalculated to 100% CaO- and LOI-free and are projected into the related fields (Fig. 10A). This diagram shows the enrichment in K_2O/Na_2O ratio from the volcanic arc to active continental margin to passive margin setting. Late Triassic samples plot mainly along the field of the passive margin (PM), and four samples plot in the active continental margin (ACM) field (Fig. 10A). In contrast, Upper Cretaceous ophiolite-related clastic sediments are projected into the ARC field, except one sample (sandstone above platform sequence) plotted in the ACM field.

Notably, Upper Cretaceous ophiolite-related sandstones, plotted in ARC field, either correspond to the intra-oceanic arc setting or are directly derived from an ophiolitic source of the SSZ setting. Since there was no evidence of arc setting, we interpreted the arc field as direct ophiolites of the SSZ setting. Cenozoic sediments are mainly clustered

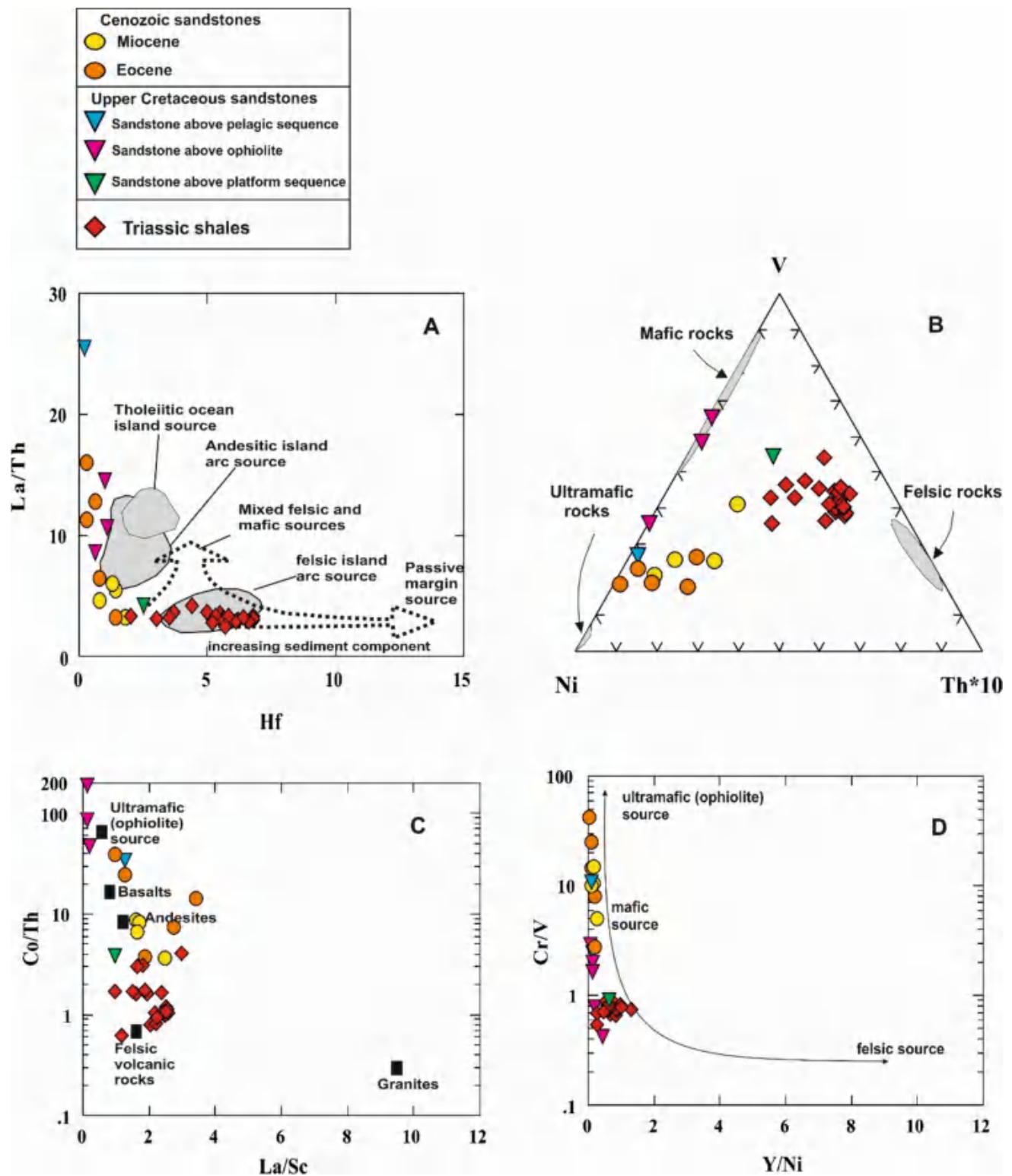


Fig. 9 Source rock discriminations for the Cenozoic, Upper Cretaceous and Triassic sediments in Isparta Angle on: **A** La/Th versus Hf (Floyd and Leveridge 1987) variation diagram, **B** Ni-V-Th*10 ternary diagram (Bracciali et al. 2007), **C** La/Sc versus Co/Th bivariate diagram (McLennan et al. 1993) and **D** Y/Ni versus Cr/V bivariate diagram (McLennan et al. 1993) (ultramafic source in the diagram taken from data of Isparta Angle ophiolites, Caran et al. 2010).

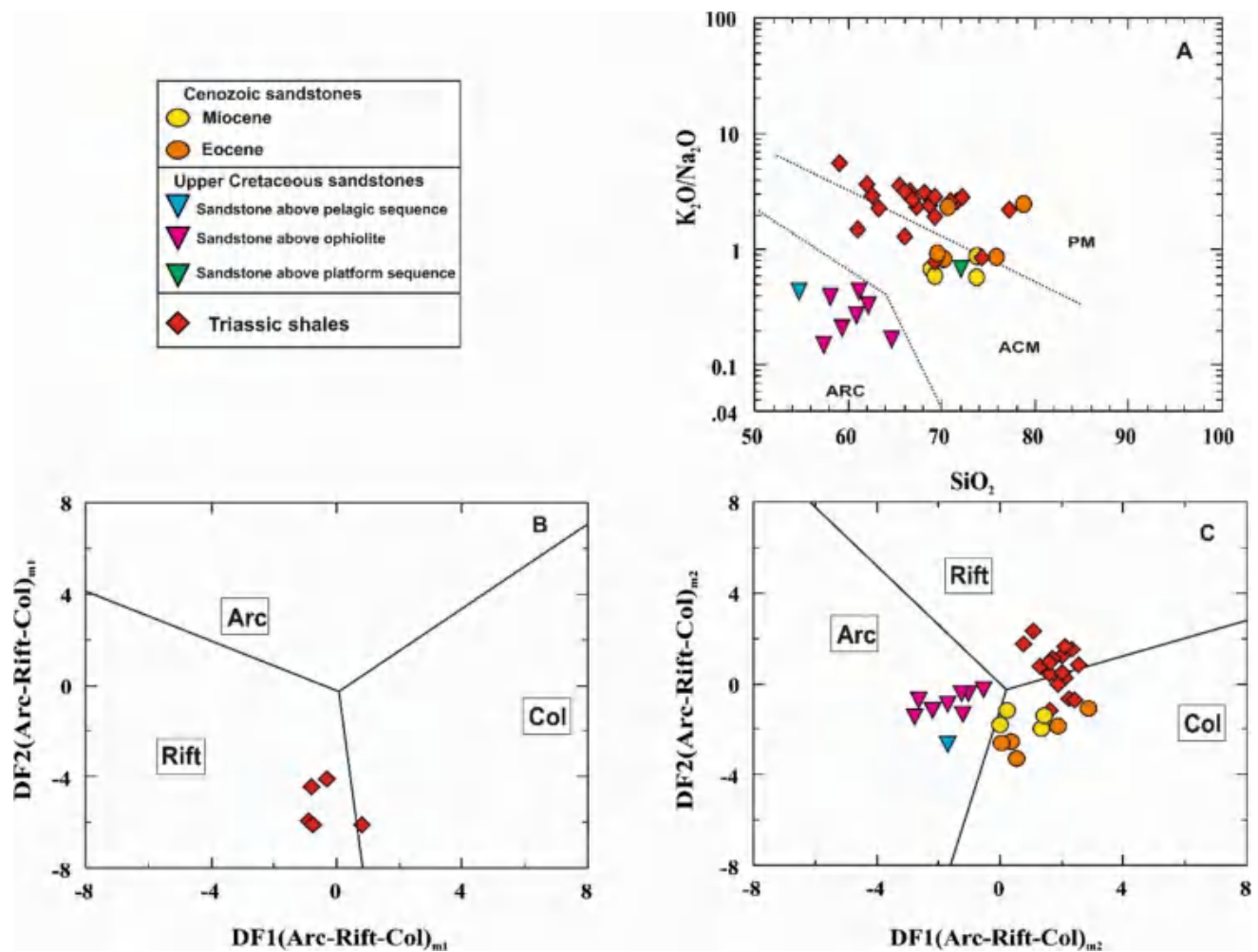


Fig. 10 **A** K_2O/Na_2O – SiO_2 (after Roser and Korsch 1986) plot for the Cenozoic, Upper Cretaceous and Triassic sediments in Isparta Angle. **B** Discriminant function multidimensional diagram (Verma and Armstrong-Altrin 2013) for Triassic high-silica shales from the Isparta Angle. The subscript m1 in DF1 and DF2 represents the high-silica diagram based on loge-ratios of major elements. The discriminant function equations are: $DF1(Arc-Rift-Col)_{m1} = (-0.263 \times \ln(TiO_2/SiO_2)_{adj}) + (0.604 \times \ln(Al_2O_3/SiO_2)_{adj}) + (-1.725 \times \ln(Fe_2O_3t/SiO_2)_{adj}) + (0.660 \times \ln(MnO/SiO_2)_{adj}) + (2.191 \times \ln(MgO/SiO_2)_{adj}) + (0.144 \times \ln(CaO/SiO_2)_{adj}) + (-1.304 \times \ln(Na_2O/SiO_2)_{adj}) + (0.054 \times \ln(K_2O/SiO_2)_{adj}) + (-0.330 \times \ln(P_2O_5/SiO_2)_{adj}) + 1.588$. $DF2(Arc-Rift-Col)_{m1} = (-1.196 \times \ln(TiO_2/SiO_2)_{adj}) + (1.604 \times \ln(Al_2O_3/SiO_2)_{adj}) + (0.303 \times \ln(Fe_2O_3t/SiO_2)_{adj}) + (0.436 \times \ln(MnO/SiO_2)_{adj}) + (0.838 \times \ln(MgO/SiO_2)_{adj}) + (-0.407 \times \ln(CaO/SiO_2)_{adj}) + (1.021 \times \ln(Na_2O/SiO_2)_{adj}) + (-1.706 \times \ln(K_2O/SiO_2)_{adj}) + (-0.126 \times \ln(P_2O_5/SiO_2)_{adj}) - 1.068$. **(c)** Discriminant function multidimensional diagram (Verma and Armstrong-Altrin 2013) for Cenozoic, Upper Cretaceous and Triassic low-silica clastic sediments from the Isparta Angle. The subscript m2 in DF1 and DF2 represents the low-silica diagram based on loge-ratio of major elements. Discriminant function equations are: $DF1(Arc-Rift-Col)_{m2} = (0.608 \times \ln(TiO_2/SiO_2)_{adj}) + (-1.854 \times \ln(Al_2O_3/SiO_2)_{adj}) + (0.299 \times \ln(Fe_2O_3t/SiO_2)_{adj}) + (-0.550 \times \ln(MnO/SiO_2)_{adj}) + (0.120 \times \ln(MgO/SiO_2)_{adj}) + (0.194 \times \ln(CaO/SiO_2)_{adj}) + (-1.510 \times \ln(Na_2O/SiO_2)_{adj}) + (1.941 \times \ln(K_2O/SiO_2)_{adj}) + (0.003 \times \ln(P_2O_5/SiO_2)_{adj}) - 0.294$. $DF2(Arc-Rift-Col)_{m2} = (-0.554 \times \ln(TiO_2/SiO_2)_{adj}) + (-0.995 \times \ln(Al_2O_3/SiO_2)_{adj}) + (1.765 \times \ln(Fe_2O_3t/SiO_2)_{adj}) + (-1.391 \times \ln(MnO/SiO_2)_{adj}) + (-1.034 \times \ln(MgO/SiO_2)_{adj}) + (0.225 \times \ln(CaO/SiO_2)_{adj}) + (0.713 \times \ln(Na_2O/SiO_2)_{adj}) + (0.330 \times \ln(K_2O/SiO_2)_{adj}) + (0.637 \times \ln(P_2O_5/SiO_2)_{adj}) - 3.631$.

in the ACM field, except for one Eocene sample plotted within the PM field. However, more recently, Verma and Armstrong-Altrin (2013) proposed two discriminant

function-based major element diagrams for the tectonic discrimination of siliciclastic sediments from three main tectonic settings: island or continental arc, continental rift and collision, created for the tectonic discrimination of high-silica $[(\text{SiO}_2)_{\text{adj}} = 63\%–95\%]$ and low-silica rocks $[(\text{SiO}_2)_{\text{adj}} = 35\%–63\%]$. These two diagrams were constructed based on worldwide examples of Neogene-Quaternary siliciclastic sediments from known tectonic settings. The authors suggested that these diagrams can be useful for successfully discriminating the tectonic setting of older sedimentary basins, which may consist of one or more tectonic assemblages. These discriminant function-based major element diagrams (Fig. 10 B–C) were also used in this study to identify the older tectonic environments of the Triassic, Upper Cretaceous and Cenozoic sediments from the IA. The application shows that, in the low-silica diagram (Fig. 10C), most Triassic samples were plotted in the rift field, except five samples in the collision field but close to the rift field. Upper Cretaceous samples (having average $\text{SiO}_2_{\text{adj}}=40.75\%$) fit well within the arc field, interpreted as a SSZ setting, and all Cenozoic samples mainly clustered in the collision field. Accordingly, in the high-silica diagram (Fig. 10B), five Triassic high-silica samples fell into the area assigned for the rift field.

5 Comparison with the earlier proposed tectonic models

The AC, bounded by the NE-SW trending Beydağları carbonate axis in the west and the NW-SE trending Anamas-Akseki carbonate axis in the east, shows different characteristics and geological histories from north to south and east to west. Due to this complexity, different tectonic models were proposed for the evolution of the AC, especially using geochemical and petrological data mainly from igneous and metamorphic formations. The results of detailed geological, petrographical and geochemical studies of siliciclastic rocks from the northern part of the AC obtained in this study were evaluated together with the Triassic clastic sedimentary rocks of the autochthonous Kasımlar Formation in the east of the IA and ophiolite-related clastic rocks in the southwest of the AC. The paleotectonic reconstructions suggested for this area were discussed using a holistic approach below.

5.1 Triassic siliciclastic rocks

Triassic formations can be found as autochthon and allochthon units in the IA area. The Triassic Kasımlar Formation occurs within the Mesozoic Karacahisar autochthon to the western side of the Anamas-Akseki carbonate axis. While

most black shales in the Kasımlar formation are litharenite, fewer have arkose and greywacke compositions (Fig. 7B). Most have a clay composition and include some carbonate material (Fig. 7A). Major and trace element ratios of the Triassic samples clustered mainly in the passive continental margin to active continental margin in the tectonic setting discrimination diagrams, but most fall into the area of passive continental margin (Fig. 10A) and in the rift field (Fig. 10 B–C), suggesting the Triassic rifting phase in the area (Fig. 11A).

Allochthonous Triassic formations can be traced from north (the apex of the IA) to south within the AC. To the westward at the northern part of the Anamas-Akseki carbonate axis, the Late Triassic to Early Jurassic is dominated by an allochthonous slope basin and deep-water deposits within the Pazarköy Group. Further west from this region, the allochthonous Triassic Isparta Çay Formation is characterised by deep-sea deposits such as thin-bedded radiolarite-chert, platy limestones and turbiditic limestones between Davras Mountain and Beydağları carbonate axis. Generally, to the west from the Anamas carbonate axis, the Triassic to Early Jurassic depositional setting changes from a neritic to a slope basin and deep-sea environment (Fig. 11B).

In the southwestern side of the IA, between the Gulf of Antalya and Beydağları, the AC is characterised by thrust-imbricated nappe stacks, including ophiolite-related lithologies and sedimentary deposits from shallow to deep-sea environments. Different classifications have been used for these imbricated tectonic units (see the Stratigraphy section in this study). In the classification of Woodcock and Robertson (1982), the Kumluca Zone is composed of thrust-imbricated deep-water passive margin sediments of Late Triassic to Upper Cretaceous age; (2) the Gödene zone, characterised by thrust-imbricated succession of Late Triassic-Cretaceous volcanics and deep-sea sediments, is overlain by Upper Cretaceous ophiolite-related clastic sediments, and the Upper Cretaceous sediments are largely intersliced in the Gödene Zone, but they are transgressive on the Kemer Zone carbonate platforms; (3) the Kemer Zone is dominated by steeply dipping slices of Paleozoic (Ordovician-Permian) sedimentary rocks overlain by Mesozoic shallow water carbonates; (4) the coastal Tekirova ophiolite is overlain by supra-ophiolite sediments, which include fossil contents from the Santonian-Maastrichtian. Also, radiolarians within the red chert of the breccia-conglomerates over the ophiolitic units were Early Cretaceous (Robertson 1998; Robertson et al. 2020b). While the Lower-Middle Triassic represents a shallow environment, the Upper Triassic has a more pelagic character around the Kemer region in the southwest of the AC (Önder et al. 1998). Tekin and Sönmez (2010) interpreted that a major deepening event, characterised by radiolarian cherts, occurred between

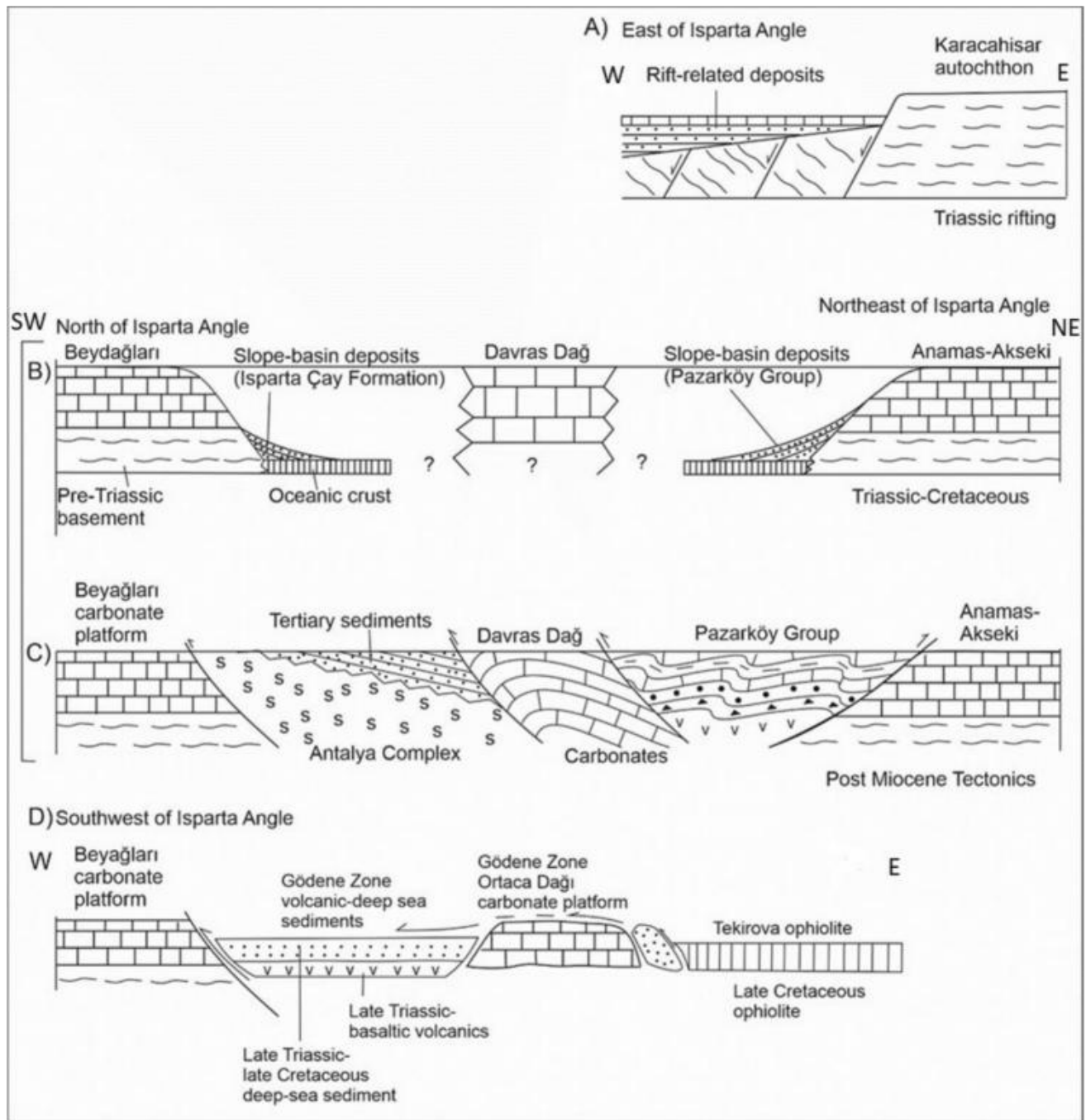


Fig. 11 **A** Triassic rifting in the east of the Isparta Angle, **B** Triassic-Cretaceous and **C** post-Miocene tectonic setting north and northeast of the Isparta Angle, **D** emplacement-related deformation of the Antalya Complex-triggered Upper Cretaceous clastic deposition in the southwest of the Isparta Angle (modified from Robertson et al. 2020a, b).

the Middle and Late Ladinian (Middle Triassic) within the succession of the Tahtalıdağ. However, Late Triassic thick-bedded neritic limestones show shallow conditions, resulting from the horst-like rising of the Tahtalıdağ Nappe during the Late Triassic block faulting.

5.2 Upper Cretaceous siliciclastic rocks

Four different geological settings are suggested for the Upper Cretaceous ophiolite-related clastic sediments: (1) overlying the Tekirova ophiolite in the Kemer zone (Kemer area); (2) overlying pelagic sedimentary rocks in the east of Ortaca Dağı (Gödene Zone eastern outcrop);

(3) overlying the Mesozoic carbonate platform within the Gödene zone (Ortaca Dağı, Teke Dağı); (4) intercalated with ophiolite and rift-passive margin lithologies to the west of the above carbonate platform (west and northwest of Gödene; Gödene zone western outcrop) (Robertson et al. 2020b, Fig. 11D). Based on the whole-rock chemical analyses, they suggested a mafic igneous rock source for the sandstones above the ophiolites and the pelagic succession and terrigenous origin for the sandstone above the Mesozoic carbonate platform. Generally, it can be concluded that the depositional setting of the Upper Cretaceous siliciclastic rocks over the platform carbonates is a continental margin setting, but over the ophiolites and pelagic sequence is an ophiolite-derived setting.

5.3 Eocene-Miocene siliciclastic rocks

On the tectonic setting discrimination diagrams, Eocene sediments display distribution from the active continental margin to the passive margin setting, and Miocene sediments fall into the active continental margin (Fig. 10A), consistent with the forearc suggestion of Sagular and Görmüş (2006). All the Cenozoic samples fall into the collisional setting (Fig. 10C), consistent with the paleotectonic evolution of the region (Fig. 11B–C).

Considering all the data of siliciclastic rocks from the AC, it can be concluded that the depositional setting of autochthonous Triassic sedimentary rocks is a continental margin setting (mostly passive continental margin) in the north of IA, and allochthonous ones, from Triassic to Early Jurassic, change from a neritic to slope basin and deep-sea environment throughout the AC, suggesting rift-related block faulting in the continental crust and then slope basin and deep-sea basin evolution. Geological and geochemical characteristics of the Cenozoic siliciclastic rocks from the north of the AC display a collisional setting consistent with the paleotectonic evolution of the region.

6 Conclusions

A specific geological and petrological study has been conducted on Cenozoic (Eocene to Miocene) sandstones north of Isparta Angle (IA) (SW Anatolia, Eastern Mediterranean domain). The obtained data and the available whole-rock chemical data of Triassic and Upper Cretaceous siliciclastic sediments from the Antalya Complex have been evaluated to clarify their source rocks and tectonic settings. The results led to the following conclusions.

(1) Relative to unweathered Cenozoic and Upper Cretaceous sandstones, Late Triassic shales of the

IA were influenced by low to moderate degrees of weathering processes.

- (2) Based on the inferred source materials for the studied IA Mesozoic to Cenozoic sediments, it is concluded that: (a) Mesozoic carbonates of IA are major sedimentary source rocks of Upper Cretaceous and Cenozoic ophiolite-related calciturbiditic sandstones; (b) Triassic shales were derived from mixing of both mainly felsic-intermediate detritus of a continental source and lesser mafic materials of the intra-plate (perhaps basaltic) rift magmatism in a passive margin setting; (c) Upper Cretaceous ophiolite-related sandstones were derived from mixing of both mainly mafic-ultramafic detritus of ophiolite-related igneous rocks and lesser felsic detritus of a continental igneous source; (d) Cenozoic sandstones were derived from mainly felsic materials of a continental source and lesser mafic-ultramafic detritus of an ophiolite-related igneous rocks in their parent regions.
- (3) The relevant discrimination diagrams of the tectonic setting of IA clastic rocks confirmed that the Cenozoic sandstones of the IA (SW Türkiye) occurred in a collision-related setting associated with ophiolite emplacement, whereas the Upper Cretaceous ophiolite-related sandstones were derived in an ophiolite-derived tectonic setting. In contrast, Late Triassic shales in IA formed in a passive margin setting.
- (4) The depositional setting of autochthonous Triassic sedimentary rocks is continental margin (mostly passive continental margin) in the north of IA, and allochthonous ones, from Triassic to Early Jurassic, change from neritic to a slope basin and deep-sea environment throughout the AC, suggesting rift-related block faulting in the continental crust and then slope basin and deep-sea basin evolution.
- (5) The results are also comparable to those in the Mamonia Complex and Kyrenia Range of Cyprus (McCay and Robertson 2012; Chen et al. 2019; Chen and Robertson 2020), where the Mesozoic and Cenozoic sandstone compositions represent the combined influences of the (Late Triassic–Early Cretaceous) passive margin setting, (Upper Cretaceous–Oligocene) ophiolite-related magmatism and (Late Miocene–Recent) diachronous continental collisional setting.

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Writing—review and editing; Şemsettin CARAN: Investigation, interpretation, writing and editing.

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Declarations

Conflict of interest The authors declare that they have no known competing financial interests or personal relationships that could have influenced the work reported in this paper.

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