TECTONIC DEVELOPMENT OF THE ANKARA-ERZINCAN SUTURE AND THE EASTERN PONTIDE MOUNTAINS, NORTHEAST ANATOLIA, TURKEY

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Key Points:

- Pontides, Tectonics, Evolution
- The Pontide Mountains of eastern Turkey is the northmost tectonic belt of the Anatolian tectonic collage. It was developed synchronously with the Ankara-Erzincan suture formed due to the gradual demise and total elimination of the NeoTethyan Ocean. This paper describes consecutive stages of this orogenic development which witnessed four collisional events.

Abstract

The Eastern Pontides are the northmost component of the Anatolian orogen. Its geological development closely associated with the evolution of the Ankara-Erzincan Suture. It exhibits records of the events from the opening to the eliminations of the surrounding oceans. During the Late Paleozoic, the Pontides were located in the north of Gondwana, facing the Paleo-Tethys Ocean. The southward subduction of the Paleo Tethyan oceanic lithosphere generated an active continental margin and opening of the Neo-Tethys Ocean as a back-arc basin during the Early Mesozoic. Throughout the Jurassic-Early Cretaceous, the Pontides remained a passive continental margin facing the Neo-Tethys in the south. Arc reversal occurred as the Neo-Tethys began subducting under the Pontides during the late Early Cretaceous (?)-Late Cretaceous.

The Pontides experienced four collisional events throughout the development of the Ankara-Erzincan Suture; (1)- a forearc-arc collision occurred when the accretionary complex, which formed along the southern edge of the Pontides was backthrust over its leading edge during the Late Campanian. (2)- This was followed by a continent-arc collision when the Kırşehir Massif and the underlying NeoTethyan ophiolite nappe collided with and thrust over the Pontides at the end of the Early Eocene, (3)- Following the oceanic lithosphere's total demise, the remnant basin located between the Pontides, and the Taurus was closed under the northerly advancing Taurus nappes during the Late Eocene.

The latest collision is related with the collision of the Arabian Plate with the Anatolian plates. The Arabian Plate's continuing northward advance after the demise of the NeoTethyan Ocean squeezed and shortened the Eastern Anatolia. From this time onward, the Eastern Pontides were thrust to the north and the south over the surrounding tectonic belts and started to rise as a coherent block.

1 Introduction

Geological research for the last four decades has determined that an integral part of the complex Alpine-Himalayan orogenic belt was developed in Anatolia, Turkey, as a result of the closure of the PaleoTethyan and NeoTethyan Oceans (e.g., Şengör and Yılmaz 1981; Okay et al., 1994; Yilmaz et al 1997A; Nikishin et al., 2002; Dilek, 2008; Hinsbergen et al., 2016., Yılmaz 2017A). The present tectonic framework of Eastern Turkey, formed mainly due to the closure of the multi-branched NeoTethyan Ocean during the Late Mesozoic-Cenozoic. Gradual elimination and final closure of the NeoTethys Ocean initially generated the Pontide Range (Fig. 1a). The closure affected the southern regions and formed the Bitlis-Zagros suture mountains in southeastern Anatolia, during the Cretaceous-Eocene period and the East Anatolian High Plateau of Turkey during the Neogene. This paper describing the development of the Pontide Range is the first of the three chapters in this book that describe the evolution of the east of Anatolian orogenic collage.

The Pontides, an east-west-trending orogen, is a composite tectonic entity consisting of two genetically different belts, the Western and the Eastern Pontides (Fig. 1a). They are tectonically juxtaposed along a broad zone in the Central Pontides (Fig. 1a). The Eastern Pontides contains records of the events from the opening to the closing of the surrounding oceans that finally generated the Ankara-Erzincan suture.

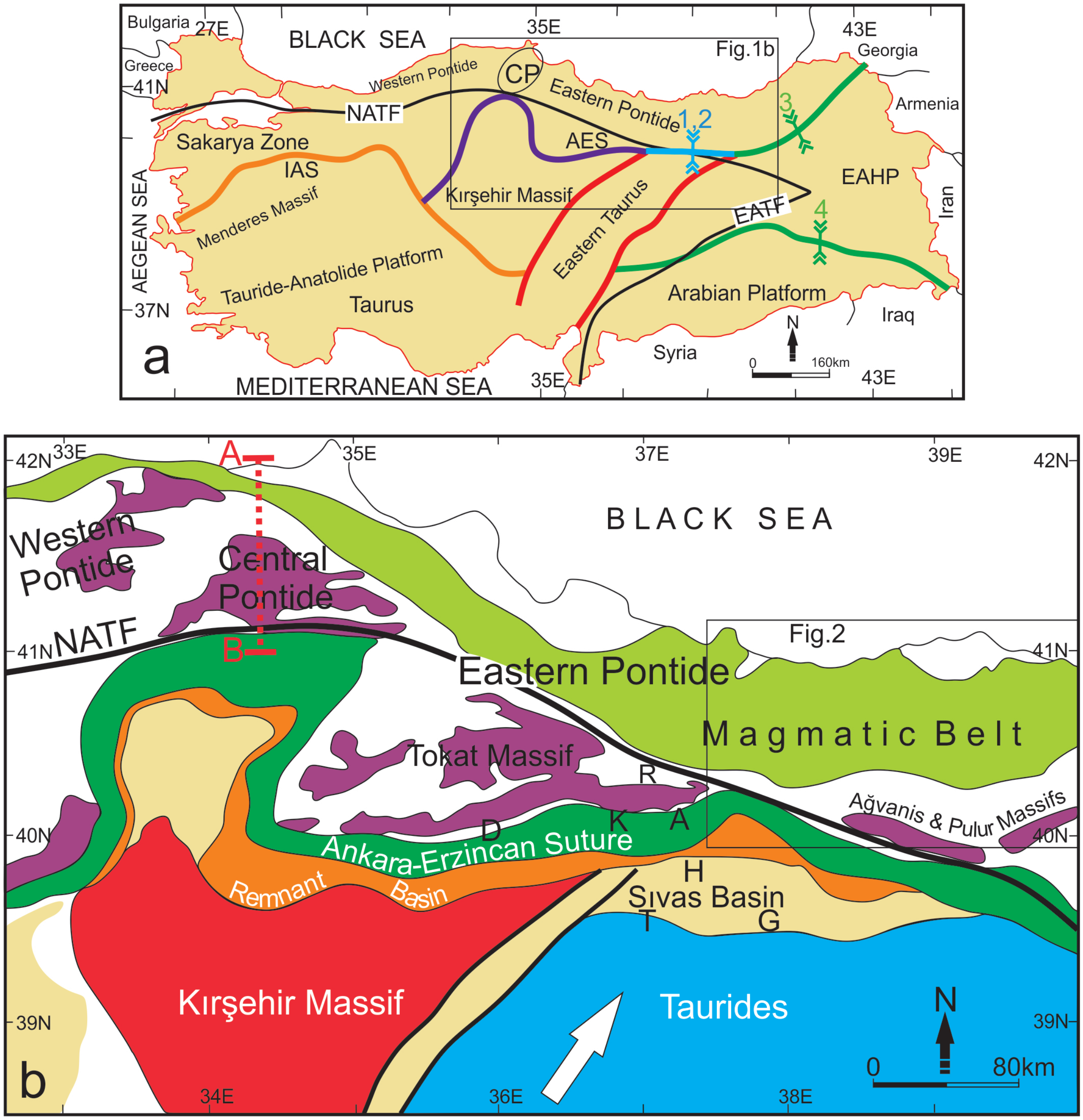

Figure1A. Major tectonic units of Anatolia (Modified after Şengör and Yılmaz 1981).

1 and 2; the belt along which the first and second collisions occurred, 3; the zone of the third collision, 4; the region of the latest collision.

The ellipse surrounding CP marks the approximate location of the Central Pontides.
Abbreviations. EAHP; Eastern Anatolian High Plateau, CP; Central Pontide. IAS and AES represent the western (the İzmir-Ankara Suture) and eastern (Ankara-Erzincan Suture) parts of the Izmir-Ankara-Erzincan Suture. NATF; The North Anatolian Transform Fault. EATF; the East Anatolian Transform Fault.
Figure 1B. Major tectonic components of the Eastern Pontides and the surrounding regions. The red line bounded by A and B shows the cross-section direction displayed in fig 4. The white arrow indicates motion direction of the Taurus Mountains. Abbreviations; NATF; The North Anatolian Transform Fault. A; Asmalıdağ, D; Devecidağ Mountains, G; Gürlevik Mountains, T; Tecer Mountains, K; Kızıldağ Mountains, H and R; Hafik and Reşadiye towns.

Over the last century, numerous studies have focused on general or specific geological aspects and the tectonic development of the Pontides (Şengör et al. 1980; Şengör 1990; Robinson et al. 1995., Yılmaz et al 1997A., Okay and Şahintürk 1997; Bektaş et al. 1999., Çiftçi and Harlavan 2005; Okay et al 2006., Rice et al. 2006: 2009., Boztuğ and Harlavan 2008., Kaygusuz et al. 2010; Çinku et al. 2010; Nikishin et al. 2013. 2015 a; b., Eyüboğlu 2015; Aydın et al., 2016; Robertson et al 2016., Roland et al., 2016; Sosson et al., 2017; Meijers et al., 2017. Barrier et al. 2018; Alkan et al. 2019., and the references therein). Recently, Kandemir et al. (2019) studied the tectonic development of the Eastern Pontides in an area of limited size near the Artvin region. Based on detailed paleontological data, they verified the major geological events demonstrated by Yılmaz et al. (1997A).

In this paper, we discuss the tectonic development of the Eastern Pontides in light of new field and analytical data gathered since the publication of Yılmaz et al. (1997A). The field-based geological observations include (a)-transport direction of the ophiolite nappes, (b)-vergence direction of structures based on evidence measured from the autochthonous sequence, (c)-locations and migrations of the magmatic belts and the sedimentary basins, and also discuss the relative location(s) of the NeoTethyan ocean and its subduction polarity with respect to the Eastern Pontides, which have long been controversial.

The Eastern Pontides displays the following approximately east-west-trending belts (Fig. 2);1- the basement, 2-the magmatic rocks, 3-the nappes, and 4-the basins.

2 The Basement

There are two subparallel belts within the basement: a-the metamorphic massifs (Mt in Fig. 2a) and b-the Mesozoic limestones (Lst. in Fig. 2a).

2.1 The Metamorphic Massifs

The metamorphic massifs (Mt) are Paleozoic in age and crop out in two isolated inliers sharing similar geological and petrological properties: the Tokat

Massif in the West and the Agvanis-Pulur Massif in the East (Fig. 1a). They were formed before the development of the NeoTethyan Ocean. A brief information about their geological/structural characteristics is given for correlative and comparative purposes.

The Tokat Massif is a tectonic mosaic consisting of three different tectonic components (Yılmaz et al. 1993;1997B). There is a thick metaophiolite tectonic wedge (the Turhal metaophiolite) located between two entirely different metamorphic assemblages; a low-grade metasedimentary sequence of Late Paleozoic age (the Amasya metamorphic Massif) above, and a high-grade metapelite-metalavas sequence of possibly Permo-Triassic age (the Yeşilırmak Metamorphic Massif) below (Yılmaz et al. 1993 and 1997B). The Liassic basal sandstones are the oldest sedimentary cover that rests on this amalgamated tectonic pile (Yılmaz et al. 1997A and 1997B). Major tectonostratigraphic components and their age ranges of the Tokat Massif compared closely and thus genetically linked with the Paleo-Tethyan Oceanic units (Yılmaz et al. 1997B). The Paleo-Tethys Ocean was located along the northern margin of the Gondwana that existed during Paleozoic and Early Mesozoic times and closed during the Late Triassic-Jurassic (Şengör et al 1980; 1981; Yılmaz and Şengör 1985; Şengör 1990).

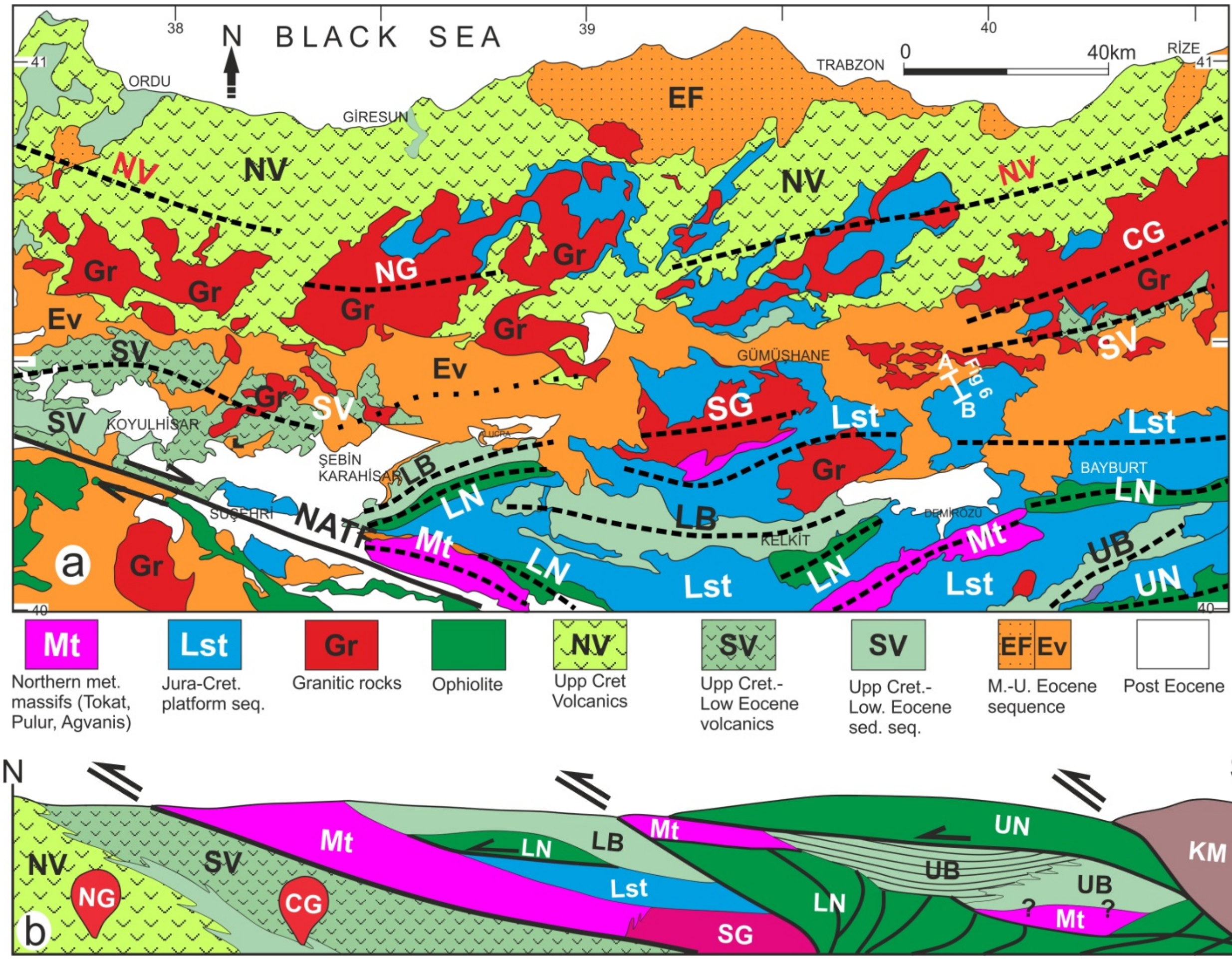

Figure 2A. Geological map of the Eastern Pontides. Dotted black lines correspond to the axes of the major belts. Abbreviations; EF; Eocene Flysch, EV; Eocene volcanics., CV; young (Post Oligocene) volcanic centers. NV; northern volcanic belt, SV; Southern volcanic belt, EV; Eocene volcanic rocks, NG; Northern plutonic belt, CG; Central plutonic belt, SG; Southern plutonic belt, LB; Lower basins. UP; Upper basins, LN; Lower ophiolite nappe, UN; Upper ophiolite nappe, Mt; Belts of metamorphic massifs, KM; Kırşehir Massif, Lts; Limestone belt, Gr; Granitic rocks, NATF; The North Anatolian Transform Fault. The line bounded by A and B shows the direction of the cross-section in Fig 6.

Figure 2B. Schematic structural relations diagram for the Eastern Pontides The cross-section across the Eastern Pontides shows the structural arrangement of the major tectonic components. Abbreviations, the same as in Fig 2A.

2.2 The Mesozoic Limestones

Mesozoic limestones (Lst) crop out subparallel to the basement rocks. They cover the metamorphic basement (Figs. 2a and 2b) (MTA 2016) and are made up of a thick neritic limestone sequence ranging in age from Early Jurassic to Early Cretaceous (Yılmaz et al. 1997A) (1; Fig. 3). In the northern regions, the Mesozoic limestones are observed as para-autochthonous successions thrust over the younger rocks (Fig. 2) (Yılmaz et al., 1997A).

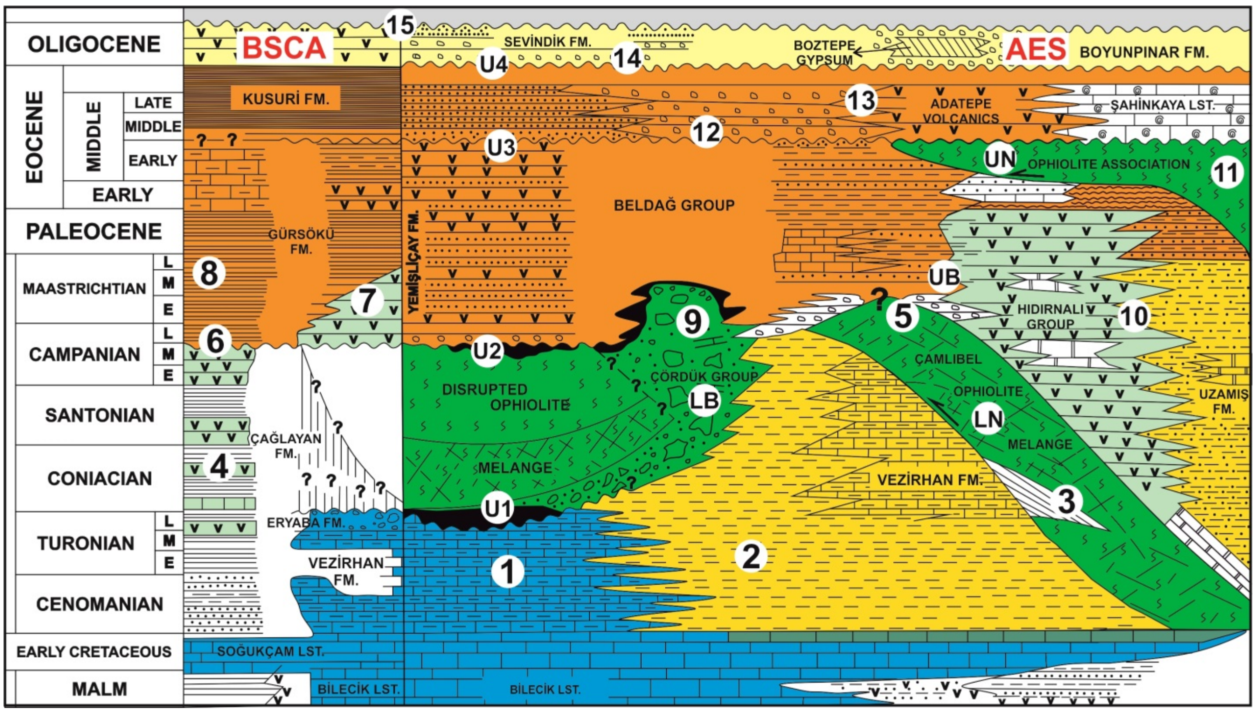

Figure 3. Generalized stratigraphic sections across the Eastern Pontides (modified after Yılmaz et al. 1997). The black vertical line corresponds approximate division between the main belt (AES) and the coastal zone (BSCA). Numbers mark the major geological events described in the text. Abbreviations, U1 to U 4 represent the four regional unconformities; LB and UB represents the lower and upper basins respectively. LN and UN are the lower and upper nappes.

3 The Magmatic Rocks

The magmatic rocks of the Eastern Pontides crop out as a wide belt between the Black Sea coastal zone in the north and the Ankara-Erzincan suture in the south (Fig. 2a). It consists of intricately intermingled volcano-plutonic associations (Figs. 2a and 2b). The volcanic rocks form the bulk of the magmatic assemblage. Previous studies demonstrated their Andean-type magmatic arc character (Kaygusuz et al., 2010; Nikishin et al., 2013. 2015 a; b., Eyüboğlu 2015.; Aydın et al., 2016; Roland et al., 2016; Sosson et al., 2017; Alkan et al. 2019). The geological data indicates that magmatic activity in the Pontides interruptedly continued from Cretaceous to Miocene (Yılmaz et al. 1997A; Schleiffarth et al. 2018; Hassig et al. 2000). The available isotope data are yet insufficient to state firmly if the magmatic activity was episodic.

The eastern Pontides' volcanic rocks form two partly overlapping subparallel belts; the northern volcanic belt (NV) and the southern volcanic belt (SV) (Figs. 2a and 2b).

3.1 The Northern Volcanic Belt

The northern volcanic belt (NV) ranges between 50 and100 km wide belt, extending along the Black Sea coastal region (Kandemir et al 2019). It consists mainly of basalts and basaltic andesites alternating with pyroclastic and volcano-sedimentary rocks of Turonian-Lower Campanian age (4; Fig. 3).

The NV lavas display narrow modal and chemical compositions varying from tholeiitic basalts (Moore et al., 1980; Akıncı et al., 1991) to calc-alkaline (Manetti et al., 1983; Ercan and Gedik,1983), high-K, calc-alkaline andesitic lavas (Manetti et al. 1983), which display affinities of volcanic arc lavas (e.g., Yılmaz et al. 1997A, Çiftçi and Harlan 2005; Eyüboğlu 2010., Nikishin et al. 2013., Aydın et al., 2016).

3.2 The Southern Volcanic Belt

The southern volcanic belt (SV) extends E-W in the southern regions of the Pontides (SV in Fig. 2a). It was developed during the Late Campanian-Maastrichtian times (7 in Fig. 3) Eyüboğlu et al., 2011).

The SV comprises a wide variety of volcanic rocks, from basaltic lavas to andesite-dacite lavas and associated pyroclastic rocks. They were also formed in an Andean-type magmatic arc setting. In this suite, andesitic lavas are more prominent. They are commonly calc-alkaline. Adakitic lavas were also locally identified in this belt (Eyüboğlu et al. 2011; Karslı et al. 2011).

3.3 The Belts of Plutonic Rocks

Three approximately E-W trending subparallel belts of granitic plutons (Figs. 2a and 2b) may be distinguished in Eastern Pontides as the southern (SG), central (CG), and northern (NG) belts. Each belt was developed by multiple magma injections (Yılmaz and Boztuğ 1996; Kaygusuz and Aydınçakır 2011 and the references therein). In the areas where the plutonic belts are close to one another, separating members of the different plutonic belts can only be achieved by detailed field, petrological and isotopic age data.

Most of the previous work focused mainly on geochemical characteristics and isotopic ages of the plutonic rocks (Boztuğ et al. 2004.,2006; 2007; Aslan and Aslan 2006; Kaygusuz et al. 2008; 2014; Kaygusuz and Şen 2011; Karslı et al. 2010; 2011., Topuz et al 2011., Kaygusuz and Aydınçakır 2011; 2014., Kaygusuz et al. 2013, 2014; Vural and Kaygusuz 2019). However, available isotope ages are still insufficient to effectively cover aerial extents, limits, and volumes of the plutons (Fig. 2). Despite this, three plutonic belts may be separated based on the stratigraphic data derived from the associated volcano-sedimentary units supported by the available isotope ages.

The southern belt comprises the plutons of commonly Paleozoic and Early Mesozoic ages (500-250 my) (Çoğulu 1967., Dokuz et al. 2010., Kaygusuz et al. 2016; Vural and Kaygusuz 2019 and the references therein). Therefore, they are not directly related to the NeoTethyan evolution. The northern belt comprises the plutons of commonly Upper Cretaceous age range (80-70 My; Boztuğ et al. 2004., Temizel and Kurt 2019 and the references therein). They display close spatial and temporal connections with the members of the NV (Fig. 1a). The Central plutonic belt commonly consists of Tertiary plutons. Their ages cluster around ~ 41- 44 My (Topuz et al 2005., Aslan and Aslan 2006., 2013., Karslı et al. 2011 and the references therein).

Modal compositions of the plutonic rocks commonly vary between diorite and granite and according to petrologic studies, they are mostly calk-alkalic and metaluminous (Kaygusuz et al 2008, Karslı et al 2010; 2012; Kaygusuz and Çakır 2011). The granitic magmas underwent fractional crystallization, mixing, and crustal contamination (Eyüboğlu et al., 2011., Kandemir et al., 2019 and the references therein). In the discrimination diagrams, most of them fell into the I-type granite field (Kaygusuz et al., 2008; Kaygusuz and Şen 2011; Kaygusuz et al. 2014; 2020) displaying a close genetic link with the coeval arc lavas. The geological field data reveal further that the granitic magmas commonly reached shallow levels in the crust and evolved in collapsed caldera environments (Yılmaz and Boztuğ 1996; Yılmaz et al. 1997A; Karslı et al. 2012; Kaygusuz et al., 2013).

4 The Nappes

The nappes of the Eastern Pontides consist mainly of ophiolitic rocks. They are divided into two nappe packages: the lower ophiolite nappe (LN; Figs. 2a and 2b) and the upper ophiolite nappe (UN; Figs. 2a and 2b). The nappes were thrust onto the leading edge of the southern Pontides over three stages: 1-Late Cretaceous, 2-late Early Eocene, and 3-Late Eocene. The metamorphic massifs located to the south were thrust onto the ophiolite nappes during the second nappe emplacement stage.

4.1 The Lower Ophiolite Nappe

The lower ophiolite nappe (LN) consists mainly of ophiolitic mélanges, which rest tectonically above the Mesozoic limestone succession (Figs. 2b and 3). Blocks of the Mesozoic limestones tectonically incorporated into the mélange may locally be observed within a serpentinite and spilitic basalt matrix. Tatar (1977) and Çelik et al (2016) reported glaucophane-lawsonite facies high-P metamorphic ophiolite fragments within the mélange. The oldest sedimentary rocks resting above the lower nappes are Upper Campanian-Lower Maastrichtian basal clastics of a transgressive sequence, which tightly constrain the thrusting time to the Late Campanian (Yılmaz et al. 1997A) (6; Fig. 3). The Pontides underwent a north-vergent shortening deformation during the lower nappe transport phase (Yılmaz et al 1997A) (Fig. 2b). Ophiolitic mélange slivers imbricated with the basement metamorphic rocks during this period may be observed in the Çorum, Amasya, Turhal, Çamlıbel, and Erzincan areas.

4.2 The Upper Ophiolite Nappe

Our field observations showed that the Upper ophiolite (UN) is a more than 1 km thick nappe (Figs. 2a and 2b and 11 in Fig. 3), cropping out extensively in the southern and central regions of the Eastern Pontides. They consist essentially of discontinuous layers of serpentinized peridotite. The upper layers of an ophiolite suite in the UN are commonly missing. The peridotite belongs to ultramafic cumulates, in which intrusive dunites may locally be identified. Thin ophiolitic mélange slivers, observed locally between the ophiolite thrust sheets, represent tectonic slices dragged during the nappe transport.

The Pontides sequence suffered from a severe shortening deformation during the thrusting of the UN (Figs. 4 and 6) as revealed by regional north-vergent tight folding and thick-skinned deformation that involved the metamorphic basement**.** The thrusting period is tightly constrained to the late Early Eocene because the underlying sedimentary sequence comprises an uninterrupted succession (Kurtman 1970; Gökten 1983.Yılmaz et al 1997A; Legeay et al 2018) from Upper Cretaceous to the Ypresian (the Hıdırnalı Group and the Uzamış Fm,10; Fig. 3 and Fig. 6), and the oldest sedimentary rocks resting on the nappes are Lutetian basal clastic units (13; Fig. 3) (Yılmaz et al 1997A). The Kırşehir Massif tectonically overlies the upper ophiolite nappe (KM; Fig. 2b).

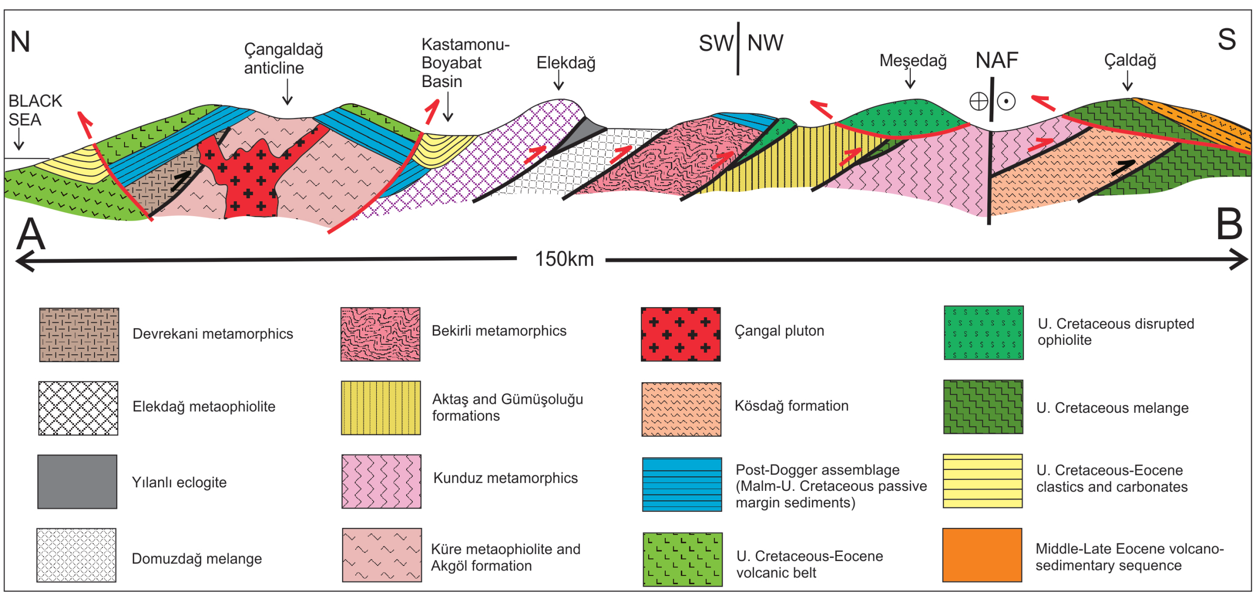

Figure 4. Schematic geological cross-section across the Central Pontide along the direction indicated in Fig 1B showing the structural arrangement of the

major tectonic units. The precise age range of tectonostratigraphic units formed in association with the PaleoTethys has not yet been fully resolved.
The cross-section displays the opposite nappe transport directions during the demise of the Paleo and Neo Tethys oceans. The older nappes were obducted from North to South whiles the younger nappes from south to north. (Modified after Yılmaz et al. 1997).

5 The Basins

Based on the tectonostratigraphic, structural, and geographic positions, two belts of basins may be identified in the Eastern Pontides, as the inner (lower) and the outer (upper) basins (LB and UB; Figs. 2a, 2b, and Fig. 3). The nappe emplacements played an active role in the development and destruction of the basins.

5.1 The Lower Basins

The lower basins (LB) were bounded by the Mesozoic limestone high in the south and the SV in the north (Fig. 2a and Fig. 3). The two elevations were also the source areas for the sediments in the basin (Fig. 3). The locations with respect to the SV reveal their forearc tectonic setting. The basin fill comprises a thick sequence beginning with chaotic coarse clastic sediments deposited over various older units, including the ophiolitic mélange (Fig. 3) (Yılmaz et al 1997A). The coarse clastic rocks transit laterally and vertically to a flysch-like succession of Upper Cretaceous age (the Çördük Group of Yılmaz et al. 1997A; 9 in Fig. 3) (Kurtman 1970; Meşhur and Dellaloğlu 1980; Gökten 1983). The LN was thrust over the LB during Campanian (Figs. 2b and 3).

5.2 The Upper Basins

A chain of connected basins located between the Pontides to the north and the Kırşehir Massif and the Taurus Plate to the South. They define stratigraphically and structurally higher (outer) basins (UB) (10; Fig. 1b, Fig. 3 and Figs. 5b, b1, c, and c1) (MTA 2016), formed within and along the suture zone. The elevated accretionary complex delimits the basin's northern boundary (Fig. 3) (Yılmaz et al 1997A).

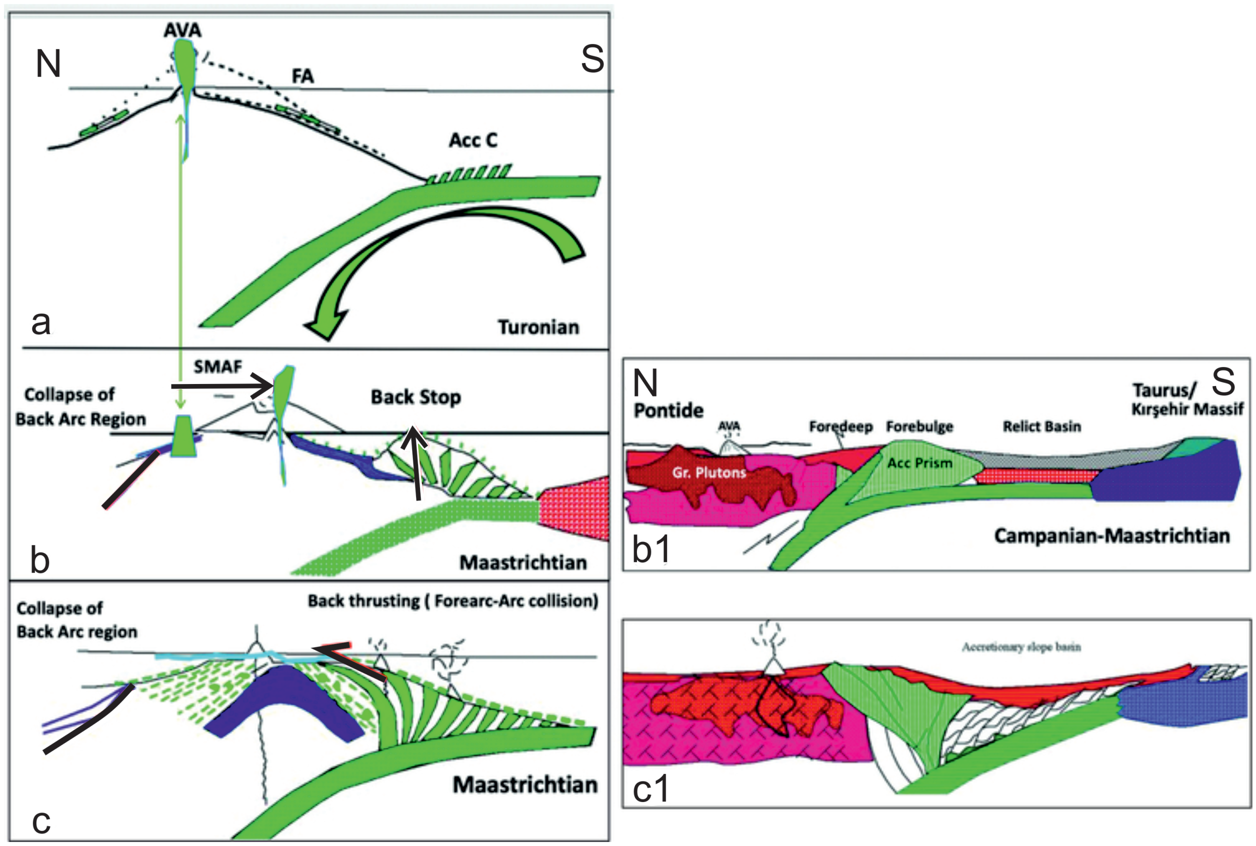

Figure 5. Cartoons showing consecutive stages of the tectonic development of the Eastern Pontides during the Late Cretaceous time. Figures A and B display southward migration of the magmatic arc, B and C progressive growth and back-thrust of the mélange-accretionary complex. B1 and C1; detailed cross-sections of the figures B and C. Arrows show motion directions. Thick lines delimiting the Black Sea Basin display the normal fault along which the Black Sea basin collapsed. Abbreviations. AccC; Mélange-Accretionary complex, AVA; Andean-Type volcanic Arc, FA; Forearc region, SMAF; Southerly migrating arc front.

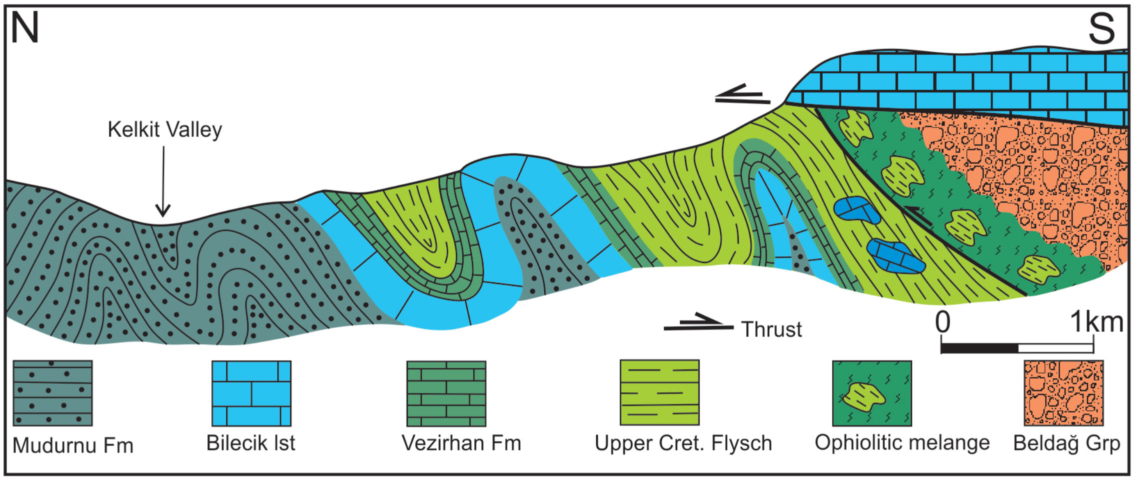

Figure 6. Geological cross-section from the Eastern Pontides along the direction indicated in Fig 2B showing the Mesozoic limestones detached from the base and moved northward as décollement style above severely deformed younger units (Modified after Yılmaz et al. 1997).

The upper basin-fill begins with deep-sea sediments resting on the accretionary complex (10; Fig. 3 and Figs. 5b and 5c) (Yılmaz et al 1997A; Bilgiç and Terlemez 2007; MTA 2016; Legeay et al. 2019A). At the beginning stages, the tectonic position of the UB may be correlated with a remnant basin that remained after the elimination of the oceanic lithosphere (Figs. 5b and 5c). The pelagic sedimentary rocks consisting of thinly bedded white-red limestones alternate with andesite lavas of the Upper Cretaceous age (10; Fig. 3) (Yılmaz et al 1997A). The deep-sea sediments grade into a turbiditic flysch, which in turn, transits to shallow marine sediments (the Hıdırnalı Group and Ilıca Fm; Fig. 3) (Kurtman 1970; Yılmaz 1980; Meşhur and Dellaloğlu 1980; Tekeli et al. 1991., Yılmaz et al. 1997A; Darin and Umhoefer 2019).

The destruction of the UB occurred in two stages during collisions of the Pontides with the southerly located plates (Fig. 1a). The western part of the UB was destroyed earlier when the upper ophiolite nappe overlain by the Kırşehir Massif (block) was thrust over the basin fill during the late Early Eocene (Fig. 1a, Figs. 2b and 13; Fig. 3). The eastern parts of the UB remained until the Taurus collided with the Pontides in the Late Eocene-Oligocene times (Yılmaz et al 1997A).

5.2.1 The Sivas Basin

The Sivas Basin, located between the Kırşehir Massif and Taurus in the south-southwest and the Pontides in the north (Fig. 1b) (Yılmaz et al. 1997A; MTA 2016), represents eastern part of the upper basins (SB; Fig. 1b) (MTA 2016).

The Sivas Basin was developed above an ophiolitic mélange foundation (3; Fig. 3), which is exposed along the northern boundary from Devecidağ-Asmalıdağ to Kızıldağ (Fig. 1b) (MTA 2016). The basin fill is thick (Kurtman 1973., Gökten,1983; Meşhur and Dellaloğlu 1980; Önal et al. 2008). Legeay et al. (2019A; 2019B) and Darin and Umhoefer (2020) estimated more than 12 km sedimentary sequence from Upper Cretaceous to Quaternary. Within the basin fill marine sedimentary sequence ranging in age from the Upper Cretaceous to the Lower Eocene are commonly uninterrupted (Kurtman 1973; Meşhur and Dellaloğlu 1980; Gökten 1983; Yılmaz et al 1997A; Legeay et al.2018). The lower part of the succession is represented by Upper Cretaceous pelagic sedimentary rocks, which grade into progressively shallowing clastic sediments of the Paleocene-Lower Eocene age. Middle Eocene basal sandstones of a transgressive sequence lie above them over a marked angular unconformity of late Early Eocene age (12 and 13; Fig. 3) (Meşhur and Dellaloğlu, 1980; A. Yılmaz; 1980; Tekeli et al 1991., Yılmaz et al 1997A; Legeay et al 2018).

Upper Eocene-Oligocene red beds unconformably rest above the Eocene marine sequence. They grade vertically to lacustrine, siltstone-marl alternations containing thick evaporites. Lower Miocene shallow marine sediments interfingering with the terrestrial sedimentary rocks were also locally identified, i.e., the Hafik region. From Middle Miocene onward, terrestrial clastic sediments and evaporites were extensively deposited in the Sivas Basin (Meşhur and Dellaloğlu, 1980; Tekeli et al. 1991; Ribes et al. 2018).

The previous studies differentiated many partly coeval tectonostratigraphic rock groups within the Sivas Basin fill. Lateral and vertical transitions and distributions of coarse to fine grained sedimentary materials that form these units reveal that boundaries and character of the Sivas Basin changed continuously due to substantial tectonic control of the bordering plates (Meşhur and Dellaloğlu 1980; Aktimur et al. 1990; Tekeli et al. 1991; Cater et al. 1991; Poisson et al. 1995., Guezou al 1996; Bilgiç et al. 2007; Callot et al. 2014; Kergaravat et 2016; Darin et al. 2018; Ribes et al 2018; Legeay et al. 2019A and 2019B., Darin and Umhoefer 2019; 2020). The upward lithofacial transitions reveal progressive shoaling of the southern shelf edge of the basin. Coarse clastic materials were continuously supplied into the basin from the northerly advancing high (Fig. 7) (Cater et al., 1991; Tekeli et al., 1991; Yılmaz et al.,1997; Darin and Umhoefer 2019; 2020).

Oblique faults with reverse slip components bound the Sivas Basin along its northern and southern borders (Fig. 7) indicating that the basin has undergone a N-S shortening deformation between the surrounding mountain ranges in periods since the Late Eocene (Fig. 8) (Yılmaz 2017A; Darin and Umhoefer 2019). Darin and Umhoefer, 2019, presented evidence indicating the escape in southern Sivas fold-thrust belt (SSFTB) was initiated in latest Miocene to Pliocene time. The western boundary along with the Kırşehir Massif is a major left-lateral strike-slip fault (Fig. 1) (MTA 2016).

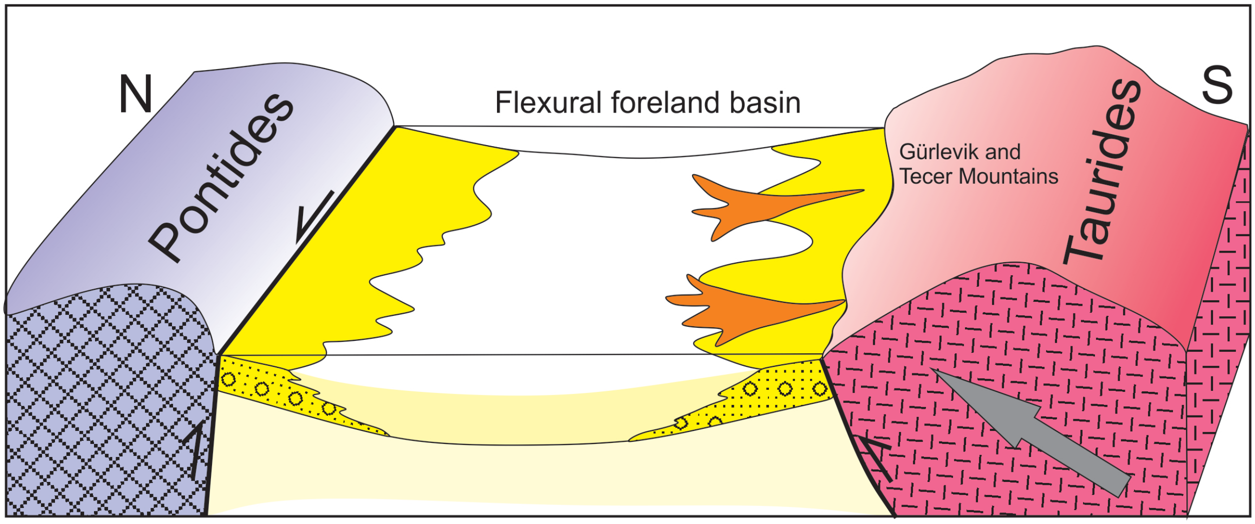

Figure 7. Schematic block diagram showing tectonic development of the Sivas basin in front of the northerly advancing Taurus nappes (Taurides) during the Cenozoic. The overlapping lateral fans were formed in front of the nappes. The thin black arrows indicate revers-slip and left-lateral strike-slip displacements of the faults formed along the southern boundary of the Pontides. The thick gray arrow shows the motion direction of the Taurus Mountains.

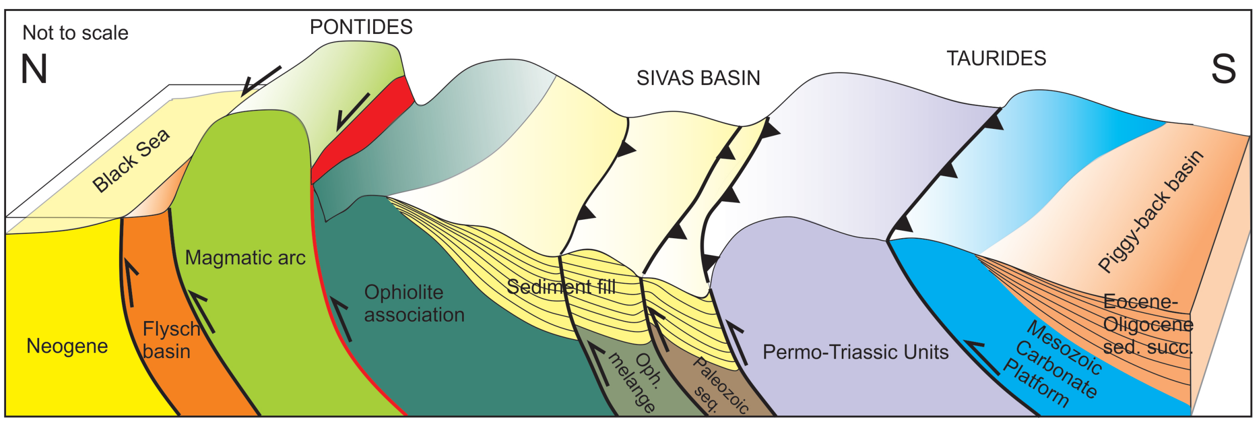

Figure 8. Schematic block diagram displaying destruction of the Sivas Basin in front of the northerly advancing Taurus nappes (Taurides) during the Late Eocene-Early Oligocene times. Thrust-bounded blocks display relative locations of the major tectonic components of the Eastern Pontides. The lateral distance from the Black Sea coastal region to the Sivas Basin is approximately 150 km. The arrows indicate the sense of displacements of the thrust-bounded blocks. Hachures display relative motions of the thrusts.

6 Stratigraphy

In this section, only the post-Paleozoic stratigraphy of the Pontides is described because the older rocks are out of the scope of this paper.

Ophiolitic rocks and epi-ophiolitic deep-sea sediments of Liass-Dogger age are observed along the Çoruh River valley in northeastern part of the eastern Pontides (Şengör et al 1985; MTA, 2016). In contrast, the coeval shallow marine sedimentary successions resting on the old metamorphic basement crop out extensively in the central and the southern regions (MTA 2016). They begin with Lower Jurassic sandstones transiting to neritic limestones of the Lower to Upper Jurassic age (the Bilecik Limestone; Fig. 3) (Yılmaz et al.1981; 1995; 1997). The neritic limestones grade into the Lower Cretaceous pelagic limestones, which in turn grades into the turbiditic flysch of Turonian-Campanian age (1 and 2; Fig. 3) (Yılmaz et al. 1995;1997A). The Mesozoic units with similar litho-facial characteristics extend along the Pontides from the Sakarya Zone in the west to the Caucus in the east (Fig. 1) (Yılmaz 1981; Yılmaz et al. 1995; 1997A; Okay et al 2006).

Four regional unconformities are identified in the Upper Mesozoic-Cenozoic successions of the Eastern Pontides (6,12,14 and 15; Fig. 3) corresponding to 1-Turonian, 2- Late Campanian-Early Maastrichtian, 3-late Early Eocene, and 4-Late Eocene. They were developed in association with the nappe emplacements, which caused tight folding and thrusting.

The Turonian unconformity (Kurtman 1970; Meşhur and Dellaloğlu 1980; Gökten 1983 Tekeli et al 1991) separates the Jurassic-Cretaceous successions, including the flysch of the Cenomanian-Lower Turonian age (6; Fig. 3), from the Turonian-Lower Campanian volcano-sedimentary units (Fig. 3).

The Late Campanian-Early Maastrichtian unconformity (Yılmaz et al. 1997A., Kandemir et al 2019) (6; Fig. 3) separates the volcanic edifices of the NV from the sedimentary rocks and intercalated volcanic layers of the SV.

The axis of NV migrated to the south during the Late Campanian period (Figs. 2 and 6 and 7; Fig. 3) (Yılmaz et al 1997A). Consequently, the areas previously occupied by the volcanic edifices were later blanketed by volcanogenic flysch (7 and 8; Fig. 3) (Yılmaz et al. 1997A; Sarı et al. 2014; MTA 2016). During the same period, the SV was built on the flysch-limestone belt of the Turonian-Campanian age (Fig. 3) (Yılmaz et al 1997A; Kandemir et al 2019). Away from the SV to the south, a coeval flysch and volcanogenic flysch accumulated (the lower or Inner basin). They begin with olistostrome deposits (Fig. 3) derived from southerly located structural highs represented by the Mesozoic limestones and the ophiolitic mélange nappes (5; Fig. 3). The basal clastics grade vertically into turbiditic flysch, followed by a sandstone-conglomerate alternation of the Paleocene-Lower Eocene age (the Beldağ group; Fig. 3) (Tekeli et al 1991., Yılmaz et al 1997A). In the southern regions the Beldağ Group developed above deep-sea sediment-lava alternation of Upper Cretaceous-Lower Eocene aged (the Hıdırnalı Group in the upper or outer basin).

The late Early Eocene unconformity (12; Fig. 3) separates the Upper

Cretaceous-Lower Eocene sediments from the SV and the coeval sedimentary successions (13; Fig. 3). The basal units resting above the erosional surface display lateral facies changes from thick clastic sediments to thin reefal limestones (13; Fig. 3) (Yılmaz et al. 1997A), indicating that they were formed above an irregular topography resulting possibly from the presence of fault-controlled local basins (Yılmaz et al. 1997A). The limestones grade into deep-sea sediments (Meşhur and Dellaloğlu 1980; Tekeli et al.1991; Yılmaz et al. 1997A) as the fault-controlled basins get deepened.

The Lutetian basal sandstones and limestones alternate commonly with intermediate (andesite-trachyandesite) and felsic (dacite and rhyolite) lavas (Adatepe volcanics; Fig. 3) (Yılmaz et al. 1997A; Aslan et al. 2013; 2014; Aydınçakır and Şen 2013; Yücel et al 2017) of mildly alkaline to calc-alkaline composition yielding 49.4, 45.31, 44.6, 44.87 Ma ages (Temizel et al. 2012; Yücel et al 2017). The Eocene volcanic edifices blanketed the entire Pontide and the surrounding regions (Fig. 2a) (MTA 2016). The Middle Eocene volcanic rocks displaying volcanic arc affinities possibly owed their origin to the regional N-S extension that affected the entire eastern Anatolia

Lower Miocene transgressive units overlie the Late Eocene-Oligocene erosional surface (14; Fig. 3). They show rapid lateral and vertical facies changes (15; Fig. 3). The lower Miocene shallow sea sediments represent the youngest marine units of the Pontides Range. They are commonly observed along the mountain range's southern periphery. Volcanic rocks of intermediate composition alternate with the Miocene sediments (Temizel et al., 2012; Yücel 2017). After Middle Miocene, the terrestrial sediments were deposited in the Pontides. The latest regional angular unconformity corresponds to the Late Miocene. Elevation of the Pontides to the present height began during this period.

7 Discussion

Geochemical and geological data indicate that the eastern Pontides were an Andean-type subduction related magmatic arc during the Late Mesozoic-Early Tertiary period (Yılmaz et al. 1997A; Bektaş et al. 1998; Yılmaz-Şahin 2005; Çinku et al. 2010; Kaygusuz et al. 2008; Eyüboğlu 2010; Karslı et al. 2010; Delibaş et al. 2016; Aydın et al. 2016; Sosson et al. 2017; Barrier et al. 2018; Hassig et al. 2010). However, the locations of the ocean and the subduction polarity with respect to the Eastern Pontides have long been controversial. The following three models have been repeatedly proposed.

1-The Tethyan Ocean was located to the north of the Pontides throughout Paleozoic, Mesozoic, and Tertiary. Its closure by southward subduction under the Pontides generated an Andean-type magmatic arc on the Pontides. The Neo-Tethys Ocean was formed in a back-arc setting in the south during this period (Eyüboğlu, 2010, 2015; Eyüboğlu et al., 2011; Liu et al., 2018).

2- The Neo-Tethys was located to the south of the Pontides. Its northward subduction under the Eurasian margin during the Mesozoic and Tertiary built the Pontide magmatic arc and caused the opening of the Black Sea as a back-arc basin (Dercourt et al., 1986; Stampfli et al., 2001; Adamia et al., 2011; Sosson et al., 2017; Barrier et al., 2018; Hassig et al., 2020 and the references therein).

3- The Paleo-Tethys Ocean was located to the north of the Cimmerian Continent during the Paleozoic and the Early Mesozoic times (Şengör et al 1980; 1981; Yılmaz and Şengör 1985; Şengör,1990). The Pontides and Sakarya Zone were in the western part of the Cimmerian Continent. Southward subducting of the Paleo-Tethyan oceanic lithosphere generated a back-arc rifting during the Triassic-Liassic, which opened the Neo-Tethys Ocean in the south. The southward subduction polarity was reversed during the Mesozoic (Şengör et al. 1980; 1985; Şengör 1990; Forte et al. 2014). The northward subduction of the NeoTethyan oceanic lithosphere under the Pontides generated an Andean-type magmatic arc during the Cretaceous, which also caused the development of the Black Sea Basin in the north.

The models proposed for the subduction polarity of the Paleo-Tethys and the Neo-Tethys oceans were based mainly on isotope and geochemical data derived from the randomly collected volcanic rock samples. In the following paragraphs, we discuss the polarity problem in light of the geological-structural data together with isotope and geochemical evidence. We start with the Permo-Triassic events from the Sakarya Zone (Fig. 1a) and the Central Pontides, where outcrops of the arc-related magmatic rocks and the associated sedimentary units are better exposed.

7.1 The Sakarya Zone and the Central Pontides

The stratigraphical and petrological data summarized in the proceeding paragraphs show close similarities between the Eastern Pontides and the Sakarya Zone of the northwestern Anatolia (Fig. 1a), which lead to assume a genetic connection between the two regions (e.g., Okay and Şahintürk 1997; Yılmaz et al. 1997A; Dokuz et al 2013., Hassig et al. 2014.,2017., 2020). Yılmaz et al. (1997A) stated that the Sakarya Zone was a part of the eastern Pontides and slid away westward along a strike-slip fault during the Late Cretaceous. In both regions, granitic plutons were injected into the old metamorphic basement in an Andean-type magmatic arc setting during the Permo-Carboniferous period (Yılmaz 1974; 1977; 1981; Yılmaz et al. 1997A., Dokuz et al 2013., Xypolias et al 2016). In the basement associations of the two regions, two interconnected Permo-Triassic successions are distinguished (Genç and Yılmaz 1995); 1- a non-metamorphic sequence; representing a north-facing passive continental margin. Its lower section reveals a rifting that occurred during the Early Triassic (Genç and Yılmaz 1995). The upper section defines the drifting stage; facies distribution of the rock units reveals that the basin progressively deepens to the north where deep-sea sediments alternate with basalt lavas (Genç and Yılmaz 1995; Pickett and Robertson 2004). At the top of the sequence are Upper Triassic olistostrome deposits containing ophiolite fragments corresponding possibly to the onset of the ophiolite nappe's obduction onto the Sakarya Continent (Genç and Yılmaz 1995).

2- A thick metamorphic and non-metamorphic ophiolite nappe (i.e., the Geyve Ophiolite) and tectonic slivers of a disrupted ophiolite (the Almacık ophiolite) (MTA 2016), commonly together with the epi-ophiolitic pelagic sediments of the Triassic age (Genç and Yılmaz 1995; Pickett and Robertson 2004; Federici et al 2010; Xypolias et al 2016). The ophiolitic rocks are exposed as a discontinuous belt to the north of the coeval continental margin sequence (Genç and Yılmaz 1995; Sarıfakıoğlu 2017). The HP metamorphism was determined from fragments of the ophiolitic mélange in the Boğazköy area, north of Bursa (Genç and Yılmaz 1995) where Sarıfakıoğlu (2017) obtained 227-268 Ma isotope ages from the plagiogranite. The ophiolite nappes were obducted on the Sakarya Continent during the Late Triassic (Genç and Yılmaz 1995). The nappe emplacement caused severe south-verging deformation and imbrication of the ophiolitic slivers with the continental basement (Genç and Yılmaz 1995). Following the ophiolite emplacement, Liassic basal sandstones of a transgressive sequence were deposited and sealed the tectonically amalgamated units similar to eastern Pontide region (Yılmaz et al. 1995).

The north-facing Permo-Triassic passive continental margin sequence and the south-vergent deformations associated with the late Triassic ophiolite nappe obduction reveal that the Paleo-Tethys Ocean was located to the north of the Sakarya continent during the Paleozoic and early Mesozoic periods (Genç and Yılmaz; 1995; Sarıfakıoğlu et al.2017). This data is critical supporting evidence to the polarity problem of eastern Pontide Range.

Despite the Late Triassic ophiolite emplacement, the Paleo-Tethys Ocean remained open in the north, possibly until Dogger, as evidenced by the isotope ages from the Küre Ophiolite (170 and 181 Ma, Sarıfakıoğlu, et al. 2017) (Fig. 4) and the paleontologically determined Triassic-Liassic ages from the epi-ophiolitic pelagic sediments transported with the ophiolite nappes to the central and eastern Pontides during Dogger (Aydın et al. 1986.,1995; Önder et al. 1987; Okay et al. 2015; Marroni et al. 2019). The Dogger ophiolite nappes are exposed as a northern belt extending from the Küre region in the Central Pontides to the Yusufeli-Artvin region in the Eastern Pontides.

A north-south cross-section across the central Pontides displays major components and structural order of the ophiolitic and metamorphic nappes (Fig. 4) (Yılmaz and Tüysüz 1991; Yılmaz et al. 1997B) and associated south-vergent thrusting. The ophiolitic tectonic slivers were imbricated with the metamorphic basement (Fig. 4).

Within the south-vergent nappe pile consisting of ophiolite nappes and tectonic slivers of continental crust origin, the Küre ophiolite and its epi-ophiolitic layer (the Akgöl Fm) are at the top (Fig. 4). They were thrust above a medium P/T metamorphic ophiolite association, the Elekdağ meta ophiolite (Yılmaz et al. 1997B), which was interpreted as a part of the Küre Ophiolite (Yılmaz et al. 1997B). A thin slice of metaophiolite dragged under the Elekdağ Ophiolite is known as the Yılanlı ophiolite (Yılmaz et al. 1997B. It displays eclogite facies metamorphism (Milch 1907; Eren 1977; Okay et al. 2006). The Elekdağ Ophiolite rests tectonically above the metamorphic subduction-accretion complex (the Domuzdağ mélange), containing blueschist fragments. The latter unit lies on the Kunduz metalavas of the Permo-Triassic age representing the coeval passive continental margin sequence. The underlying metamorphosed unit, the meta flysch of the Triassic age, defines the continental slope of the passive margin (Yılmaz et al., 1997B).

The structural section outlined from the Central Pontides may be correlated closely with the Sakarya Zone. The structural arrangement of the ophiolitic tectonic slices and their vergent directions indicate that the Paleo-Tethys Ocean was located to the north and the nappe transport occurred from north to south (Şengör et al. 1980; 1981; Yılmaz and Şengör,1985; Şengör 1990). Therefore, the geological data disapprove the claims viewing location of the Paleo-Tethys Ocean in the south of the Pontides during the Late Paleozoic-Early Mesozoic period.

The first common cover sedimentary sequence resting on the amalgamated tectonic pile is Upper Jurassic marine conglomerates (Yılmaz et al., 1997A; 1997B). They display rapid lateral facies changes (Tüysüz et al. 1990). The thickened and elevated continental crust under the ophiolitic nappe load collapsed rapidly during Malm, as indicated by a new sea invasion (Tüysüz et al. 1990). Consequently, the region underwent N-S extension, which caused the development of a horst-graben morphology over which basal sediments of the Malm transgression were deposited (Tüysüz et al., 1990).

8 Development of the Ankara-Erzincan Suture and the Geological Evolution of the Eastern Pontides

The Izmir-Ankara-Erzincan Ocean was formed when the Menderes-Taurus block (Menderes Massif and the Tauride-Anatolide Platform; Fig. 1a) was separated from the Sakarya Continent during the Early Mesozoic (Fig. 1a) (Şengör and Yılmaz 1981., Yılmaz 1995., Yılmaz et al. 1997A; Candan et. al, 2011; Van Hinsbergen et al. 2010). The Taurus and the metamorphic massifs of southeastern Anatolia (e.g., the Bitlis Massif) were also drifted away from the African Arabian Plate of the Gondwanaland during this period (Şengör and Yılmaz 1981; Candan 2011; Yılmaz 2017A). The data documented in the preceding paragraphs disapprove of an oceanic realm between the Pontides and the African-Arabian plate during the Permo-Triassic period (Şengör and Yılmaz 1981; Şengör et al. 1985; Okay et al. 2006; Robertson et al. 2012).

The regionwide neritic limestone succession (1; Fig. 3) developed during the Jurassic defines a carbonate platform. The overlying Lower Cretaceous pelagic limestones represent an open-marine shelf environment. The Upper Cretaceous turbiditic flysch (2; Fig. 3) indicates a continental slope (Yılmaz et al.1995;1997A). The Mesozoic succession reveals collectively the development of a passive continental margin facing the southerly located NeoTethyan Ocean (Fig. 5a) (Yılmaz et al. 1997A). The evidence for the ocean floor is obtained from the ophiolitic mélange elevated above the sea and formed an E-W trending belt during Turonian (Figs. 2 and 5b). This belt consists of ophiolitic mélange and fragments of the continental crust incorporated into the mélange (Yılmaz et al. 1997A). Together, they formed an accretionary complex lying along the edge of the Pontides due to the northward subduction of the oceanic lithosphere (LN; Figs. 2 and 3). The data indicate further that two partly coeval events occurred during this period; 1-the northward subduction of the NeoTethyan oceanic lithosphere generated an E-W trending magmatic arc (NV) on the Pontides during Turonian (Fig. 5a) (Yilmaz et al. 1997A., Eyüboğlu 2010., 2015., Delibaş et al. 2016). 2-subparallel volcanogenic flysch belts were developed around the volcanic belt (Fig. 3 and Fig. 5) (Yılmaz et al. 1997A; Okay et al. 2006). Progressive growth and the consequent thickening and elevation of the mélange prism formed a barrier (a backstop or a forebulge; Figs. 5b and 5b1), which delimited a forearc depression located between the volcanic arc and the mélange (the inner-Lower basin; LB in Fig. 2) (Fig. 5b). The inner basin was filled with the coarse conglomerates and olistostromes shed from the surrounding highs (9; Fig. 3). The coarse clastic sediments were followed upward by a flysch during Turonian- Campanian period (9; Fig. 3. The flysch grades upward into sandstone-conglomerate alternation during the Santonian-Early Campanian (Yılmaz et al.,1997A).

8.1 First Collision

The thickened and elevated accretionary complex was later backthrust over the continental margin sequence during Late Campanian (5; Fig. 3 and Figs. 5c and 5c1). This event may be evaluated as a forearc-arc collision, which caused a north-vergent shortening deformation and wholesale elevation of the Eastern Pontides (Figs. 5c and 5c1). As a result, a regional unconformity (6; Fig. 3) was formed above the tightly folded and faulted pre-Upper Campanian successions.

An extensional regime followed the shortening deformation during the Late Campanian-Early Maastrichtian. This is evidenced by the basal units of a younger transgression deposited above the Lower Maastrichtian succession. This event corresponds to the volcanic front's southward migration (SV; Fig. 2, and 7; Fig. 3; Figs. 5a and 5 b), which is thought to be due to the southerly retreat, roll-back of the subducting lithosphere of the northern branch of the Neo-Tethys Ocean (Fig. 5a) (Şengör and Yılmaz 1981; Çinku et al. 2010; Kandemir et al. 2019) with a similar mechanism to that of progressive growth of the Eastern Mediterranean accretionary prism and the associated southward migration of the Aegean volcanic arc (Jolivet et al 2013;Yılmaz 2017 A; 2017B and the references therein). Consequently, the regions previously occupied by the magmatic arc were left in the back-arc region (Yılmaz et al.1997A) and thus rapidly collapsed (Fig. 5b). The deepening of the Black Sea Basin appears to be related to this event (Yılmaz et al., 1997A; Stephenson and Schellart 2010; Nikishin et al. 2015). Compositions of the magmas that passed through the thickened continental crust changed from the low-K Calc-alkaline to High-K Calk-Alkaline during this period (84 Ma. Liu et al. 2018; Karslı et al. 2010; 2012; 2013; Eyüboğlu 2010; Eyüboğlu et al. 2011; Hassig et al. 2020).

8.2 Second Collision

Despite the mélange nappe's (LN) emplacement onto the Pontides, a deep basin remained above the mélange foundation between the Pontides and the Kırşehir Massif (Figs. 2 and 5b) during the Campanian-Maastrichtian (Meşhur and Dellaloğlu 1980; Tekeli et al. 1991; Yılmaz et al. 1997A; Hassig et al. 2020). Considering its tectonic setting, this basin can be interpreted as an accretionary slope basin (Fig. 5c1), in which pelagic, thinly bedded limestones and cherts of Maastrichtian-Paleocene age were deposited (Fig. 3) (Yılmaz et al. 1997A; Tekeli et al. 1991). Along the slope of the basin, sediment deposition continued uninterruptedly until the Early Eocene (Meşhur and Dellaloğlu 1980; Yılmaz et al. 1997A; Darin et al., 2018., Legeay et al. 2019 and Darin and Umhoefer,2020). The pelagic sediments transited vertically to shallow sea sediments during the Paleocene-Early Eocene (Ilıca Fm; Fig. 3). The basin fill was later deformed under the northerly advancing nappe pile consisting of the ophiolite nappe (UN) and the overlying Kırşehir Massif (UN; Figs. 2b, and 11; Fig. 3). The nappe emplacement is tightly constrained to the end of the Early Eocene because the youngest underlying sediments are Lower Eocene in age, and the oldest common cover rocks are the Lutetian shallow marine conglomerates and sandstones (12 and 13; Fig. 3) (Yılmaz et al. 1997A). The stratigraphic ages are confirmed by the Apatite fission-track data (Boztuğ et al., 2004; Cavazza et al., 2012; Espurt et al., 2014).

The Kırşehir Massif's thrusting over the southern edge of the Pontides magmatic arc is a continent-arc collision (1; Fig. 1a and Fig. 2b). During this collisional event, the entire Pontides underwent a new north-vergent shortening deformation, which caused tight folding, thrusting (Fig. 6), and thick-skin deformation. In front of the nappes, thin-skin deformations were developed when the Mesozoic limestones were forced to detach from the base and thrust over the younger units in a decollement style (Fig. 6) (Yılmaz et al. 1997A).

8.3 Third Collision

The Sivas basin sediment fill has stratigraphic, sedimentologic and structural records revealing the development of the third collision and post-Mesozoic tectonic evolution of the Ankara- Erzincan suture (Figs. 1b and 3).

The lower part of the succession is identical to that of the other outer basins represented by the Upper Cretaceous-Early Tertiary deep-sea sediments resting above a mélange foundation. They were formed in the remnant oceanic basin delimited by the Taurus Range in the south. The deep-sea basin began to be filled with a fast accumulation of clastic materials (Fig. 3) (Yılmaz et al. 1997A). The transport indicators such as southward coarsening data from the clastic sedimentary sequence reveal that they were sourced from the northerly advancing Gürlevik and Tecer Mountains of the Taurus nappes (Fig. 7) (Cater et al. 1991; Darin and Umhoefer 2020). In front of the nappes, the depression turned into a foreland basin (Fig. 7) (Cater et al. 1991; Kergaravat et al. 2016). The size and depth of the basin were progressively reduced (Meşhur and Dellaloğlu,1980; Aktimur et al., 1990; Cater et al., 1991; Tekeli et al., 1991; Temiz et al. 1992; Poisson et al. 1995; Legeay et al. 2018; 2019; Darin and Umhoefer 2020). The Taurus nappes were then thrust over the basin's southern edge during the Late Eocene-Early Oligocene (Meşhur and Dellaloğlu 1980; Cater et al. 1991; Darin and Umhoefer 2020) (Figs. 1 and 8). The basin fill was internally imbricated to accommodate the N-S shortening deformation, which did not allow a head-on collision of the bordering plates (Meşhur and Dellaloğlu 1980; Yılmaz et al. 1993; Kergaravat et al. 2016, Darin and Umhoefer 2020) (3; Fig. 1a and Fig. 8). The N-S orthogonal compressional stress generated north-vergent thrust tectonics and overturned folds in the southern Pontide regions. Progression of the shortening deformation caused escape tectonics and associated lateral extrusions in the later stages. The oblique faults with significant left-lateral strike-slip coupled with reverse slip displacements were formed during this phase when the stress permutation occurred (Fig. 7). The faults divided the southern Pontide mountains into fault-bound blocks (Fig. 8). Age of the development of thrusts and oblique faults may be tightly constrained to the Late Eocene, because the Upper Eocene sediments stacked within the fault-bounded blocks were unconformably overlain by the Oligocene red beds (Yılmaz et al., 1993). The Oligocene terrestrial sediments were commonly deposited in the E-W trending fault-bounded, long, and narrow depressions (Fig. 8) (Yılmaz et al. 1993). While the Sivas Basin fills underwent a severe deformation in front of the Taurus nappes, a coeval piggy-back basin was developed on the nappes where the low energy sedimentary rocks were deposited. Therefore, they remained relatively undeformed (Fig. 8) (Cater et al. 1991; Darin and Umhoefer 2019; 2020).

The Upper Eocene regressive units and the subsequent Oligocene red beds (Fig. 3) may be interpreted as a response to the rise and the consequent erosion of the Pontides mountains during the Late Eocene-Early Oligocene collisional event (14 in Fig. 3).

8.4 Fourth (the latest) Collision

The latest collision occurred when the Eastern Pontides were actively involved in the development of eastern Anatolian orogen**.** The NeoTethyan oceanic lithosphere was eliminated from the entire eastern Turkey by the late Eocene (Yılmaz 2017A and Yılmaz et al this volume). Its demise generated a wide and thick mélange-accretionary complex that underlies the entire eastern Anatolia (Şengör and Yılmaz 1981; Şengör et al 2008). The mélange-accretionary prism behaved like a wide and thick cushion, which did not allow a head-on collision of the bordering continents (the Turkic-type orogen; Şengör et al, 2008); the Eastern Pontides in the North and the Southeast Anatolian orogenic belt in the South (Fig. 9). The N-S compressional stress that was generated from the continuing northward advance of the Arabian Plate has deformed the entire eastern Turkey. As a result of this, the Pontides were thrust to the north and the south over the surrounding tectonic belts and began to rise as a coherent block from Late Miocene onward (see Yılmaz et al. this volume).

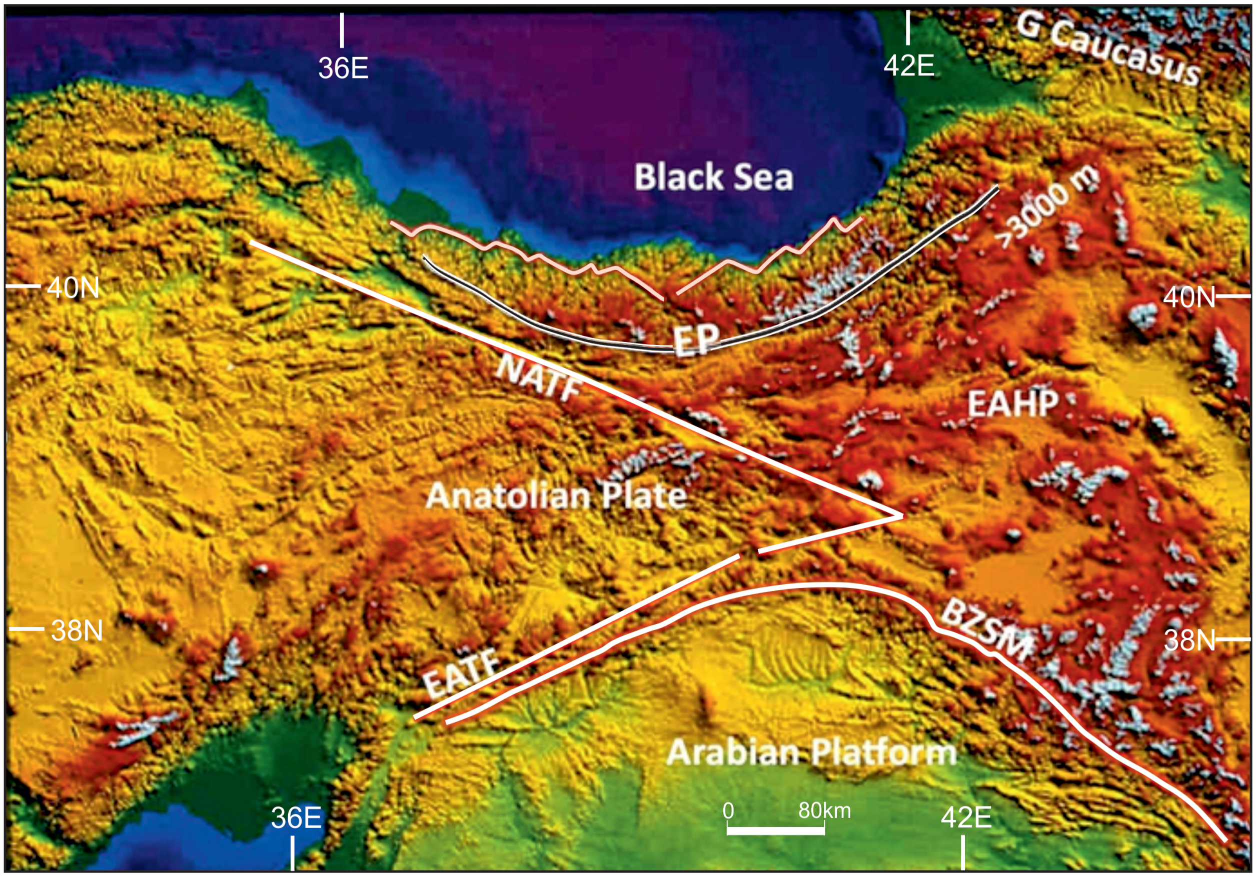

Figure 9. Morphotectonic map of eastern Turkey. The curvilinear blackline with white glows, and the yellow line with red glows display trendlines of the eastern Pontides (EP) and the Southeastern Anatolian Orogenic Belt (BZSM; the Bitlis-Zagros suture mountains). The red zigzagging lines extending along the Black Sea shore correspond to the strikes of the faults measured from the bedrocks.
Abbreviations: EATF; The East Anatolian Transform Fault, NATF; The North Anatolian Transform Fault.

8.5 Late-post collisional phase

The latest transgression occurred briefly along a narrow strip of lowland in the southern areas of the Eastern Pontides during the Early Miocene (Yılmaz 2017A). This was possibly related to the Miocene global sea-level rise.

The Eastern Pontides started to gain their present elevation during the Late Miocene (Yılmaz 2017A and Yılmaz et al. this volume). This statement is based on the following evidence: 1- Between the Bayburt and Uzundere towns, along the southern Pontide Range (Fig. 1a), the Lower-Middle Miocene marine sediments were elevated on the shoulder of the mountains (MTA, 2002). 2- poorly sorted, non-lithified fanglomerates, derived from the rising mountains continually deposited in the adjacent lowlands during Pleistocene-Quaternary (Yılmaz et al. 1997A., Softa et al. 2018; Yılmaz 2017A).

An escape tectonic regime replaced the N-S orthogonal shortening during the Late Pliocene-Pleistocene (Yılmaz 2017A and the references therein). Therefore, numerous strike-slip faults with reverse-slip components were formed (Fig. 10). Several normal faults (Figs. 10 and 11) were developed on both flanks of the mountains as a gravitational response to the rapid uplift. Due to the intersections of the oblique and the normal faults zigzagging coastlines were generated (Figs. 10, and 11). The fault couples caused the development of several small bays and promontories (Figs. 9, 10, and 11). Along the sea terraces, the fluvial deposits are Quaternary in age (Keskin et al., 2011; Yılmaz 2017A., Softa et al. 2018).

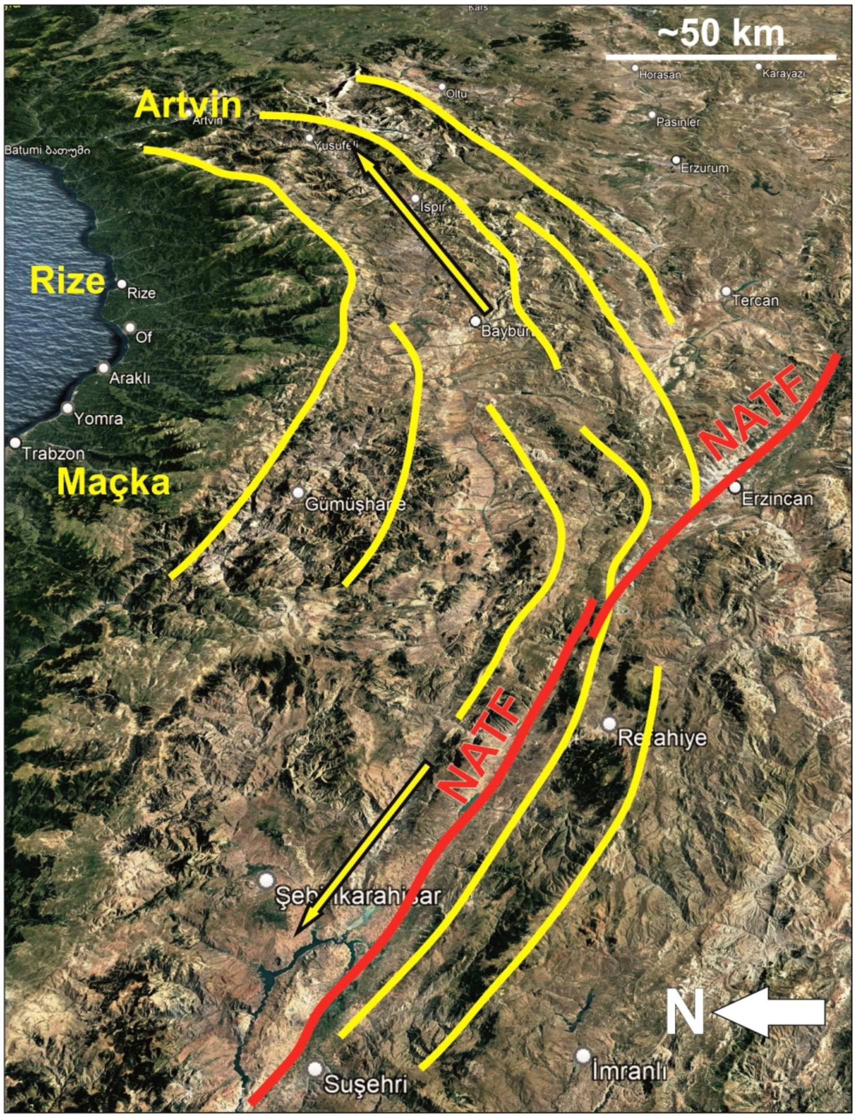

Figure 10. Google satellite image showing eastward view of the Eastern Pontides. The yellow curvilinear lines correspond to the trendline of the Black Sea Mountains. The yellow arrows display the divergence of the two limbs from the region of maximum indentation. The red lines show the North Anatolian Transform Fault zone.

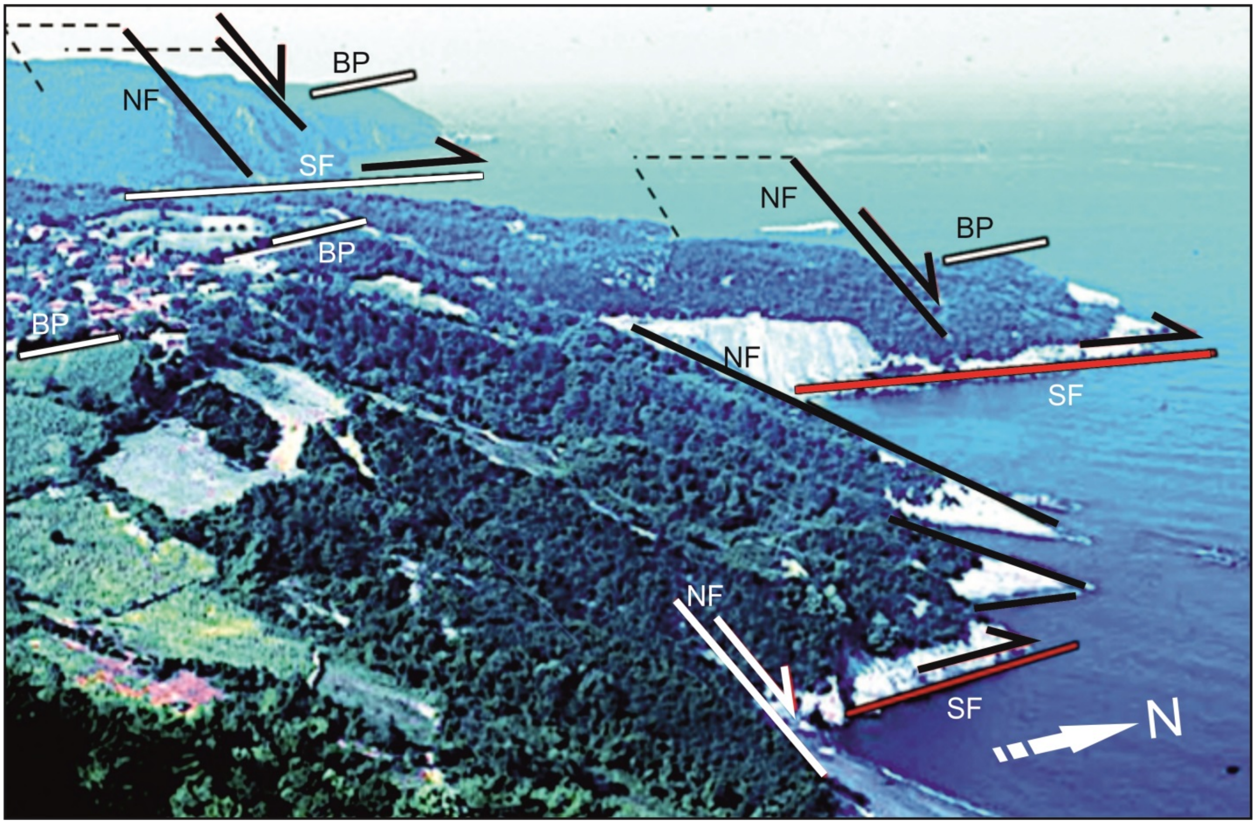

Figure 11. A photo from the Black Sea coastal region. The zigzagging shoreline resulted from intersections of the two sets of faults; the strike-slip faults wedging the lands into the sea and the gravity faults causing the sub-horizontal steps and steep slopes. Abbreviations; NF; Normal fault, SF; Strike-slip fault, BP; bedding plane. Attitudes of the bedding planes indicate back-tilting in the hanging wall due to the rotational movement along the fault planes of the normal faults (Modified after Yılmaz 2017 Fig 2.26 b).

9 Concluding Summary

The Anatolian Orogen is a tectonic mosaic formed from the collisions of the continental slivers rifted off from the African-Arabian Plate. They were later accreted to Eurasia following the closure of the Tethyan Oceans.

The consecutive stages of the collisional history may be summarized as follows.

The thick Mesozoic neritic limestone succession above the Pontides indicates an extensive carbonate platform development during the Late Jurassic.

-The carbonate platform subsided steadily during the Early Cretaceous, as revealed by the gradual replacement of the neritic limestones with pelagic limestones.

-The platform extended to a slope environment as indicated by the transition from the pelagic limestones to the Upper Cretaceous turbiditic flysch.

- The Upper Cretaceous ophiolitic mélange belt lying along with the southern border reveals that the oceanic lithosphere began subducting under the Pontides along with a north-dipping subduction zone during the late Early Cretaceous(?)-Late Cretaceous. Fragments and tectonic slivers were incorporated into the mélange from the continental crust's leading-edge, which formed a mélange-accretionary complex.

- The northward subduction of the oceanic lithosphere under the Pontides generated an Andean-type magmatic arc in the northern region (the NV). From the volcanic arc, voluminous basalt lavas were extruded during the Turonian-Campanian period. The volcanoes also produced volcanoclastic material to the surrounding lowlands during the Santonian-Early Maastrichtian period.

-The mélange-accretionary complex was elevated and then was backthrust onto the continent's southern edge during Campanian. Tectonic forces associated with the ophiolite obduction caused a north-vergent shortening deformation. Consequently, the region was deformed, shortened, and rose above the sea, as revealed by a regional unconformity (the lower unconformity). The emplacement of the ophiolitic mélange nappe (LN) may be interpreted as a result of the collision of the forearc with the magmatic arc (the first collision).

-An extensional regime followed the shortening deformation during the Late Campanian-Early Maastrichtian. This is evidenced by the basal units of a younger transgression deposited above the Lower Maastrichtian succession. This event corresponds to the volcanic front's southward migration (SV; Figs. 2, and 7; Fig. 3; Figs. 5a and 5b), which is thought to be due to the southerly retreat, roll-back of the subducting lithosphere of the northern branch of the Neo-Tethys Ocean (Fig. 5a) (Şengör and Yılmaz 1981; Çinku et al. 2010; Kandemir et al. 2019). Consequently, the regions previously occupied by the magmatic arc were left in the back-arc region (Yılmaz et al.1997A) and thus rapidly collapsed (Fig. 5b). The deepening of the Black Sea Basin appears to be related to this event (Yılmaz et al.,1997A; Stephenson and Schellart 2010; Nikishin et al. 2015). The magmas passing through thick continental crust produced more evolved volcanic edifices, the andesitic to rhyolitic lavas, and the pyroclastic rocks. The previous arc region was occupied by volcanogenic flysch during this period.

- A remnant basin survived on the mélange foundation. The pelagic sediments were continually deposited during the Late Cretaceous-Paleocene period. Shallow marine sediments gradually deposited atop the deep-sea sediments during the Late Paleocene-Early Eocene. The elevated accretionary complex formed a backstop, which defined the basin's northern boundary.

- A second nappe package (UN) was thrust over the remnant basin fill from the south to the end of Early Eocene. This nappe package consisted of thrust sheets detached from the ophiolite and the associated deep-sea sedimentary rocks dragged in front of the colliding Kırşehir Massif. The nappe transport caused a new phase of a north-vergent shortening deformation in the entire Eastern Pontides. The Kırşehir Massif's thrust over the Pontides may be evaluated as an arc-continent collision (the second collision).

-The shortened, thickened, and elevated Pontides collapsed rapidly towards the end of the Early Eocene. The sea invaded the Pontides regions once again during Lutetian. Rapid lateral transitions of the marine sediments from limestones to the thick, coarse clastic sedimentary rocks indicate that the new extensional regime generated regionwide horst-graben structures. The calc-alkaline andesitic-rhyolitic lavas accompanied the sediment deposition. The Middle Eocene volcanic edifices covered the entire Pontides region. The CA volcanic suite of intermediate composition owed its origin possibly to the N-S extension that affected entire Anatolia during this period.

-The shallow marine sediment deposition continued until Late Eocene when the Gürlevik and Tecer Mountains of the northern Taurus were thrust over the Sivas Basin's sediment fill. The thrusting signifies a continent-continent collision (the third collision) when both the Sivas Basin and the Pontides underwent a new north-vergent deformation. Several left-lateral strike-slip faults with reverse slip components divided the Eastern Pontides into fault-bounded blocks.

-Except for a narrow E-W corridor in the south that the sea invaded during the early Miocene, the Pontides have remained above the sea since the late Eocene-Oligocene.

- The final stage of the uplift of the Eastern Pontides occurred during the late Miocene when the compression due to the Southeast Anatolian-Arabian collision began to deform eastern Turkey (the fourth collision), which also caused westward escape of the Anatolian plate and associated lateral extrusion (Şengör and Yilmaz, 1981; Çemen et al., 1993; Yılmaz 2017A).

In summary the Pontide mountains suffered four collisional events in association with the demise of the NeoTethyan Oceanic realm. During each phase the Pontides were shortened, thickened, elevated, and then collapsed as manifested by the regionally developed folds, thrust and the subsequent regional unconformities.

Based on the field-oriented geological work and detailed geochemical studies, the general tectonic framework of the Ankara-Erzincan suture and the eastern pontide mountains in northeast Anatolia, Turkey, are well-understood. However, there are still many questions that need to be answered. One of the major questions is crustal thickness variations from the suture zone to the Black Sea basin. This question can be answered by a well-designed, well-thought-out high-resolution seismic tomography study.

Another critical question involves the stratigraphic, sedimentological, and structural relationship between the offshore and onshore tectonostratigraphic units in the southern Black Sea region. How did the Pontide structures control sedimentation in the Black Sea basin? This question may be answered with problemoriented 3D seismic data, which will enhance future hydrocarbon studies in the eastern Black Sea Basin.

The problem that still remains to be resolved is the differential uplift rate of tectonically controlled blocks along and across the Pontide Mountains (Yılmaz 2017A). This problem may be answered by detailed satellite geodesy studies supported with rich isotope age dating on the surface rocks.

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